GEOLOGINEN TUTKIMUSLAITOS

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N:o 166

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

HELSINKI
MAI 1954

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POSSIBLE NATURAL FRACTIONATION OF THE MAGNESIUM ISOTOPES [1]

BY

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ABSTRACT

The natural fractionation of the stable magnesium isotopes is found probable. Several possibilities of fractionation are discussed on the basis of the geochemistry of magnesium.

The most recent measurements showing the conventional relative abundance of the three stable magnesium isotopes, ^{24}Mg, ^{25}Mg, and ^{26}Mg, are those made by Hibbs and Redmond (1948) and by White and Cameron (1948). The results of the two series of determinations and the ^{24}Mg/^{26}Mg ratios calculated from the data are given in Table 1.

Table 1. Conventional relative abundance of the magnesium isotopes

Author	Abundance, % ^{24}Mg	^{25}Mg	^{26}Mg	^{24}Mg/^{26}Mg
Hibbs and Redmond (1948)	78.98	10.05	10.97	7.20
White and Cameron (1948)	78.60±0.13	10.11±0.05	11.29±0.08	6.96

The sample analyzed by White and Cameron consisted of magnesium metal. Unfortunately, I have been unable to get information about the ultimate source of the metal. Moreover, attempts to obtain information about the material investigated by Hibbs and Redmond proved futile. Therefore, nothing can be said of the true abundance of the magnesium isotopes on the basis of the analyses quoted in Table 1. It would appear,

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however, that the difference between the ^{24}Mg/^{26}Mg ratios is significant, because it probably is too great to be caused by instrumental effects. Similar instruments, viz. 60-degree magnetic spectrometers, were used for both determinations of the isotopic constitution of magnesium. Unless fractionation of the magnesium isotopes has taken place in some phase during the metallurgical and other processes used for the making of the samples analyzed, it appears probable that some variation occurs in the isotopic constitution of magnesium in Nature.

The geochemical behavior of magnesium suggests several possibilities for its isotopic fractionation. Thus, fractionation might be established between the early crystallized magnesium minerals of igneous rocks and the hydrous magnesium minerals formed by autometamorphism. Additional changes in the ^{24}Mg/^{26}Mg ratio might occur during the hydrothermal crystallization of magnesite and brucite, and the formation of magnesian clay minerals might incorporate fractionation by cation exchange and sorption. In the minor cycle, possibilities for fractionation might be present in the deposition of magnesium carbonate, dolomitization, and crystallization of the salt minerals of the evaporites. In the biosphere, marine organisms might be able to incorporate the different magnesium isotopes at different rates that might depend on the prevailing water temperatures, because it is known that the intake of magnesium carbonate by organisms is a function of temperature. Finally, because magnesium is a constituent of chlorophyll, one might search for a fractionation caused by green plants taking up magnesium from their substratum. This fractionation, if verified, might be useful for studying the occurrence of chlorophyll and the history of photosynthesis in past geological times.

Ingerson (1953, pp. 355, 357, 358) suggested that during the very slow crystallization of a great dunite reservoir there might occur a detectable concentration of ^{26}Mg in the early olivine crystals that accumulate on the floor of the chamber, because of the tendency of ^{26}Mg to become incorporated into crystals at a higher temperature than does ^{24}Mg and because of the higher specific gravity and higher sinking rate of crystals containing the heavier isotope. He also thought that diffusion caused by gravity could produce fractionation of the magnesium isotopes at a great depth in quiet lake and ocean waters. Moreover, he suggested the investigation, by means of the ^{24}Mg/^{26}Mg ratio, of the magnesium metasomatism, i. e., the metasomatic metamorphism incorporating the introduction of magnesium.

Furthermore, because there is definite proof of the existence of isotopic fractionation above magnesium in the mass range from 28 to 34 (Silicon: Allenby, 1953; Marsden, 1951; Reynolds and Verhoogen, 1953. Sulfur: Thode, Macnamara, and Collins, 1949), the chances are that isotopic fractionation occurs in the case of magnesium as well. The mass ratio $26/24 = 1.08$ for magnesium is similar to the ratio $30/28 = 1.07$ for silicon and

34/32 = 1.06 for sulfur, and therefore at least some fractionation may be expected for magnesium, even though the geochemical cycle of magnesium is rather simple, as is also the cycle of silicon (Reynolds and Verhoogen, 1953, p. 233). With modern mass instruments, however, small variations in the isotopic constitution of the elements also are detectable.

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ORIGIN OF CARBON IN SOME EARLY PRE-CAMBRIAN CARBONACEOUS SLATES FROM SOUTHEASTERN MANITOBA, CANADA [1]

BY

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ABSTRACT

The isotopic constitution of carbon was investigated in five specimens of early pre-Cambrian carbonaceous slates from the Rice Lake group in southeastern Manitoba. A discussion is presented of the general geology and lithology of the said rocks, and the specimens investigated are described. The age of the slates is estimated to be approximately 2.55×10^9 years. The $^{12}C/^{13}C$ ratios range from 90.06 to 90.86 for the oldest slates. It is concluded that at least the highest ratios lie in the biogenic range. Consequently, the carbon is biogenic in origin. Geologic evidence supports this conclusion. The manner of occurrence of carbon in the original argillaceous sediments is discussed with particular reference to the hypothesis of a reducing early pre-Cambrian atmosphere.

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INTRODUCTION

Following a suggestion of Wickman (1941, p. 419) and looking for other than geologic evidence of the origin of carbon in pre-Cambrian, particularly Archean, metamorphic rocks, I investigated the isotopic constitution of carbon in graphite, coal, carbon-bearing schists, and the suggested fossil, *Corycium enigmaticum* Sederholm from eastern Fennoscandia (Rankama, 1948). The results indicated that the isotopic constitution of carbon in the said materials fell into the biogenic range, thereby producing evidence of the biogenic origin of carbon in the rocks investigated.

This conclusion caused a number of critical remarks on geologic (v. Straaten, 1949), statistical (Hutchinson, 1949), and isotope chemical (Craig, 1953) grounds. I answered these arguments in two papers (Rankama, 1950, 1954) and am still satisfied that no new evidence, geologic or other, has been presented that contradicts the validity, in principle, of using the isotopic constitution of carbon as a criterion of its biogenic origin in metamorphic rocks of argillaceous derivation and of concluding that the *Corycium* and the other materials that I investigated are of biogenic origin. Shaw (1952, p. 142) gave independent proof of the biogenic origin of carbon in Archean carbonaceous schists. He investigated the occurrence of thallium in minerals and rocks and concluded that the *Corycium*-bearing phyllite owed its increased thallium content to biologic action. According to Shaw, carbonaceous shales and schists on an average contain nearly twice as much thallium as argillaceous sediments. He also attributed a biogenic origin to the graphite from Mäntyharju in Finland, thereby confirming the conclusion that I reached on the basis of its isotopic constitution.

The age of the *Corycium*-bearing phyllite formation in the Tampere area in Finland may be estimated as approximately 1.5×10^9 years. Of a somewhat younger age, approximately 1.2×10^9 years, are the sinuous marks on the bedding surfaces in the ripplemarked Middle Huronian Ajibik quartzite in Michigan that Faul (1949, 1950) interpreted as fossil burrows. It appears, however, that the fossil character of these marks is not unambiguously verified inasmuch as they are, to judge from Faul's description, very similar to the *Manchuriophycus* and other marks of the same kind that Häntzschel (1949) interpreted as desiccation cracks.

Anxious to continue my search for carbon of biogenic origin in rocks still more ancient than the phyllites of the Tampere area, I took notice of the occurrence of carbonaceous slates of extremely ancient age in the sedimentary series of southeastern Manitoba in Canada. In 1949 and 1950, I obtained several specimens of these rocks through the good offices of Dr. J. D. Allan of the Mines Branch at Winnipeg, Manitoba. The abundance of the stable carbon isotopes in the specimens was determined, in 1950 and

1952, in Stockholm through the courtesy of Professor Frans E. Wickman of the Riksmuseet. An account of the material and of the results obtained will be presented in the following sections.

GENERAL GEOLOGY AND LITHOLOGY OF THE CARBONACEOUS SLATES IN SOUTHEASTERN MANITOBA

A series of ancient rocks covers an area east and southeast of the southern end of Lake Winnipeg in the southeastern part of the Province of Manitoba in Canada. According to a general account of the geology of this area given by Wright (1932, pp. 10, 12-17, 26, 30, 34), the bedrock there is of pre-Cambrian age and may be divided into two major groups. These are a group of generally steeply dipping sedimentary and volcanic strata and another group consisting of igneous rocks ranging from peridotite to granite in composition that cuts through the first group. The sediments comprise slate, chert, iron formation, graywacke, quartzite, arkose, and conglomerate, and schists and gneisses also occur as their metamorphic derivatives. The interbedded volcanic rocks range from basalt to rhyolite in composition and often are altered to chlorite, sericite, and carbonate schists. For the sedimentary-volcanic rock complex, Wright proposed the name, Rice Lake series. The Rice Lake formations are severely deformed and steeply folded. Wright subdivided this series on the basis of lithologic features into three groups, or phases, viz., in the order of decreasing age, the Manigotagan phase of sediments and their metamorphic derivatives, the Beresford Lake phase, predominantly composed of volcanic rocks, and the Wanipigow phase of sediments. He, however, stated that the lithologic subdivision is not altogether satisfactory.

According to Wright's description, the Manigotagan phase comprises bedded quartzose and argillaceous sediments ranging from fine to medium in grain size and from light to black in color. The abundant members of this phase are quartzite, graywacke, and black bedded or dense rocks with poorly developed slaty cleavage called metargillites and considered to represent clayey sands. The fine-grained quartzose and argillaceous sediments of the Manigotagan phase were deposited, perhaps, in part under considerable depths of water.

Stockwell (1938, p. 3), after remapping much of the area surveyed by Wright, presented new conclusions relative to the Rice Lake series. He divided the rocks into two groups, retaining the name, Rice Lake series, for the sediments and interbedded lavas and using a new name, San Antonio formation, to designate a younger series of sediments consisting of feldspathic quartzite and overlying the Rice Lake series. The San Antonio formation comprises a part of Wright's Wanipigow phase. Stockwell's division of the Rice Lake series, again, is lithologic.

According to Allan (1949), Stockwell in 1939 changed the name, Rice Lake series, to the name, Rice Lake group. This practice was followed by the Manitoba Mines Branch in their mapping in southeastern Manitoba. Stockwell's division was used because the San Antonio formation obviously is a younger formation that has not undergone the folding and metamorphism of the Rice Lake group. Because in many localities the sedimentary rocks can be mapped separately from the volcanic rocks, the subdivision of the group into a volcanic, a sedimentary, and a metamorphic division was introduced on lithologic grounds, without implying definite age relationships (see Davies, 1949, pp. 3,5). The metamorphic division comprises altered equivalents of the other two divisions, plus injection gneiss.

In the reports covering the geology of parts of the Rice Lake mining division Russell (1948, p. 10; 1949, p. 5) subdivided the Rice Lake group into an upper sedimentary division and a lower volcanic division, the sedimentary member being the younger of the two.

Allan (1949), however, stated that he was not entirely satisfied that the Rice Lake group is one continuous series, and that it may consist of two or more series which have been interfolded. Therefore, the Rice Lake group actually may cover a wide range of time, and rocks closely associated in space may be widely apart in age. This possibility, however, is still tentative and inadequately supported by facts. At any rate, all the Rice Lake group is older than the San Antonio formation, and older than most of the exposed granitic rocks and the dikes containing the lepidolites and uraninites dated by radioactivity methods.

AGE OF THE SLATES

Davies (1950, p. 3) presumed the Rice Lake group to be early pre-Cambrian.

Wilson (1949, 1950), on the basis of available age determinations, divided the pre-Cambrian Canadian Shield into several geologic provinces and subprovinces. Among the provinces, believed to be primary orogenic belts, the Superior and the Yellowknife are considered the oldest known. In the Superior province, Wilson (1949, p. 236) indicated an overall age of 2.2×10^9 years for the intrusive bodies. This province includes the pre-Cambrian formations of southeastern Manitoba.

The said age is based on results obtained by radioactivity methods, viz., the lead-isotope and strontium methods. The age measurements refer to uraninite, monazite, and lepidolite from albite pegmatites associated with granites that are intrusive into rocks of the Rice Lake group (Wright, 1932, pp. 35, 36, 47). The individual ages determined are listed in Table 1.

Table 1. Age of pegmatite minerals from southeastern Manitoba

Mineral	Locality	Method of dating	Age in 10^9 years	Reference
Uraninite	Huron Claim nr. Winnipeg River	Lead isotope	1.985	Holmes, 1948, p. 190
Monazite	»	»	1.827 — 2.590	Holmes, 1948, p. 192
Lepidolite	Falcon Island, Lake of the Woods	Strontium	2.2	Ahrens, 1947
Lepidolite	Winnipeg River	»	2.3	Ahrens, 1947
Lepidolite	Silver Leaf Mine Winnipeg River	»	2.1	Ahrens, 1947
Lepidolite	»	»	2.35	Ahrens, 1947
Lepidolite	Winnipeg River	»	2.38	Ahrens and Gorfinkle, 1950

On the basis of several determinations, Ahrens (1947) considered an age of 2.1×10^9 years reasonable for the lepidolites; whereas Ahrens and Gorfinkle (1950) used the new, higher, age of 2.38×10^9 years to replace the earlier, more approximate, strontium ages. Later, Ahrens (1952, p. 194) computed an approximate mean age of 2.2×10^9 years for the pegmatite minerals of southeastern Manitoba. Although the age of approximately 2.2×10^9 years evidently cannot be taken too rigorously, it agrees better with the ^{207}Pb/^{206}Pb ages of 2.2×10^9 years for the Huron Claim uraninite and 2.57×10^9 years for the Huron Claim monazite computed by Nier (1939, p. 159) and by Nier, Thompson, and Murphey (1941, p. 113), respectively, than does the age of approximately 2×10^9 years recommended by Holmes (1948, p. 190) for the said uraninite; the higher lead-isotope ages are further supported by the new precise determination of the half-life of ^{235}U (Ahrens, 1952, pp. 193, 195). Finally, the corrected helium age of pillow lavas from Beresford Lake in southeastern Manitoba is approximately 2×10^9 years (Horwood and Keevil, 1943, p. 30; see also Holmes, 1948, p. 192), and the helium ages of three magnetite specimens from an east-west trending orogenic belt of infolded rocks in Ontario are from 2×10^9 years to 2.4×10^9 years (Hurley, 1949). It appears that the Rice Lake group lies on the line of this orogenic belt if extended westward. The said two independent age determinations support the ages assigned to the pegmatite minerals of southeastern Manitoba.

The ages determined for the pegmatite minerals may be used to estimate the age of the Rice Lake group. It is obvious that all recent determinations of age of minerals and rocks have produced values considerably greater than those previously believed to be representative. Consequently, applying this result to the pegmatite minerals of southeastern Manitoba, one would not consider an average approximate age of 2.4 × 10^9 years unreasonable for them. Furthermore, the pegmatites are the youngest members of an igneous sequence that comprises a series of rocks ranging from peridotite to granite, and consequently one may conclude, with a reasonable margin of safety, that the average age of this igneous complex is no less than 2.4 × 10^9 years. The Rice Lake group that represents a previous geologic cycle in Holmes' (1948, p. 193) opinion comprises a span of time of at least 0.15 × 10^9 years. Therefore, the conclusion follows that the age of the Rice Lake group is approximately 2.55 × 10^9 years.

The occurrence of granite and granodiorite pebbles in conglomerate layers in the Rice Lake group (Wright, 1932, pp. 13, 46; Russell, 1948, p. 5; Davies, 1951, p. 7) indicates the presence of exposed silicic rocks in southwestern Manitoba at the time of deposition of the Rice Lake sediments, even though no outcrops of such rocks have been recognized and their origin remains unknown. The conglomerates seem to afford proof of the existence of at least one older geologic cycle preceding the Rice Lake cycle (Holmes, 1948, p. 193). Once more adding at least 0.15 × 10^9 years to the age of the Rice Lake group, one obtains the approximate age of 2.7 × 10^9 years for the pre-Rice Lake cycle or for a previous phase of intrusion of silicic rocks into the early Rice Lake metasediments within the same orogenic cycle (Ahrens, 1952, p. 195). A ready check of this age could be obtained by dating by the argon method the potash feldspar in the granite pebbles, some of which are coarse-grained.

DESCRIPTION OF SPECIMENS INVESTIGATED

Dr. J. D. Allan of the Mines Branch at Winnipeg very kindly placed six specimens of carbonaceous slates from the sedimentary division of the Rice Lake group at my disposal. He remarked that the slates were all fairly well metamorphosed. One of the specimens, No. 4-A-48, a light-colored greenish gray schistose rock from Silver Falls on Wanipigow River turned out to be a chlorite schist that did not contain native carbon enough to make a mass-spectrometric determination of its isotopic constitution possible. Short descriptions of the other specimens will be given in the following paragraphs.

Specimen No. 1-A-48. Slate. Shore of Conley Bay, Wallace Lake.

This is a black dense rock with a well-developed fissility. Narrow pyritiferous quartz stringers less than 1 mm in thickness generally run parallel to the bedding. Under the polarizing microscope, the slate is found to be composed of carbonaceous matter, quartz, chlorite, and muscovite. Quartz stringers containing pyrite in idiomorphic crystals, muscovite flakes, and a little limonite run parallel to the schistosity. The rock is very rich in carbon.

According to Russell (1948, pp. 3, 4), the rocks of the Wallace Lake area are complexly folded. Graphitic slate occurs in a transitional, or intermediate, member between a predominantly volcanic and a predominantly sedimentary member.

Specimen No. 2-A-48. Slate. Same locality.

This is a black dense slate very similar to No. 1-A-48, but its bedding planes are undulate. The content of carbon is still higher than in No. 1-A-48. The quartz stringers penetrating the rock are less numerous and almost pyrite-free.

Specimen No. 3-A-48. Slate. First tram cut east of Clangula Lake.

This specimen consists of a dull black dense rock with a rather poor fissility and contains groups of idiomorphic magnetite octahedra not exceeding 2 mm in diameter imbedded in calcite on one of its weathered surfaces. Microscopic investigation reveals the presence of chlorite and carbonaceous matter as chief constituents and of quartz and limonite in subordinate amounts. The carbonaceous matter is largely concentrated to make separate layers of varying thickness running parallel to the bedding and accumulations of irregular shape. The rock is cut by a network of calcite as small irregular spots and stringers of later origin that carry magnetite with scarce hematite inclusions. It is obvious that this rock is much altered hydrothermally and is more strongly metamorphosed than the previously described slates.

Specimen No. 5-A-48. Slate from a drill core. Locality unknown, but the rock comes from the Rice Lake district. According to Dr. Allan (personal information), this rock may not be from the Rice Lake group but is, rather, from younger rocks.

Megascopically, this dense black slate bears a considerable resemblance to No. 1-A-48. Carbon, quartz, and chlorite appear under the microscope as its main constituents. There is a little calcite, sometimes associated with pyrite. The slate is varved. Its degree of metamorphism is rather low.

Specimen No. C-1-R-50. Slate. From a band 300 ft. - 400 ft. east of the eastern shore of the northwestern arm of Beresford Lake; from the old Orogrande Dock. The rock occurs as pyritiferous carbonaceous bands interbanded with conglomerate and striped chlorite-bearing quartzite. According to Allan (1949), this rock contains considerable pyrite and, with a fair degree of certainty, is older than the dikes from which the dated uraninites and lepidolites were obtained.

Megascopically, the rock is a somewhat rusty dull black dense fissile slate with numerous miniature folds. Under the microscope, it is found to be composed of quartz, carbon, muscovite, and chlorite, and a little limonite. There are ribbonlike layers of carbon alternating with quartz-muscovite-chlorite layers that contain finely disseminated carbon. Small lenticles of quartz of later origin are present, obviously introduced from aqueous solutions during the folding of the rock. As in No. 3-A-48, also in this rock a part of the carbonaceous matter is concentrated in the form of separate layers of varying thickness running parallel to the bedding.

ISOTOPIC ANALYSES

The isotopic constitution of carbon in the slate specimens was determined in Stockholm by the courtesy of Professor Frans E. Wickman. The finely ground rock powder was boiled twice in 6N hydrochloric acid to remove carbonate, and the carbon was oxidized to carbon dioxide in a current of oxygen in a combustion furnace. The carbon dioxide was precipitated as barium carbonate by passing it in a barium hydroxide solution, and carbon dioxide released from the barium carbonate was analyzed on a mass spectrometer. The details of the analytical procedure are those previously published by Wickman, Blix, and von Ubisch (1951, pp. 142, 143) and Wickman (1952, p. 243).

The results, in the form of $^{12}C/^{13}C$ ratios, are listed in Table 2. The reproducibility of the measurements made in Stockholm is within ± 0.10 units of the ratio, usually within about ± 0.05 units (Wickman and v. Ubisch, 1951, p. 120; Wickman, 1952, p. 243). The results refer to the so-called Stockholm carbon standard, viz., barium carbonate with a $^{12}C/^{13}C$ ratio of 89.5 (Wickman, Blix, and von Ubisch, 1951, p. 143). Referred to Nier's (1950, p. 793) primary standard with a $^{12}C/^{13}C$ ratio of 88.97, the Stockholm standard has a ratio of 89.76 (Dansgaard, 1953, p. 256).

Table 2. Isotopic constitution of carbon in carbonaceous slates from southeastern Manitoba

Specimen No.	Description	$^{12}C/^{13}C$
1-A-48	Black slate, shore of Conley Bay, Wallace Lake	90.66
2-A-48	Black slate, same locality	90.86
3-A-48	Black slate, Clangula Lake	90.06
5-A-48	Black slate from a drill core, unknown locality in the Rice Lake district	91.20
C-1-R-50	Black slate, Orogrande Dock, E. of Beresford Lake	90.32

DISCUSSION AND CONCLUSIONS

In another paper, I reviewed the geologic and isotopic evidence serving to indicate the biogenic or nonbiogenic origin of carbon in pre-Cambrian rocks of argillaceous origin (Rankama, 1954). The discussion indicated that, in the absence of contradicting geologic evidence, one is entitled to conclude that the finely disseminated carbon even in early pre-Cambrian slates and schists, and the carbonaceous accumulations in such rocks, are of biogenic origin if their isotopic constitution falls within the biogenic range, that is, if the $^{12}C/^{13}C$ ratio of such carbon is greater than 90.5.

The geologic and petrographic evidence relative to the specimens of carbonaceous slates from Manitoba described on pp. 10-12 indicates that these rocks originally were pelitic sediments, viz., silts and muds containing carbonaceous matter. I summarized the geologic theory of the origin of carbon in such rocks in another paper (Rankama, 1954). It may be added here that the richness in carbon of many pre-Cambrian schists is commonly interpreted as a proof of their biogenic origin (see, e. g., Backlund, 1941, p. 35; Pantin, 1941; Chamberlin, 1952, p. 256), and that in many schists the carbon is very finely disseminated, forms the cementing material between the clastic grains, and occurs in close association with the original constituents of the rocks. Carbon of this kind may have originally been biogenic detritus, perhaps partly of colloidal dimensions, that made an interstitial paste in the original sediment (Sederholm, 1912, p. 541; Wilson, 1931, p. 121; Twenhofel, 1939, p. 1 178). The carbon content of the carbonaceous slates and graywackes of the Canadian Shield is comparable to the carbon content of younger sediments. Therefore, if the amount of carbonaceous matter is a rough quantitative measure of the abundance of life, the early pre-Cambrian seas supported as much life as did the later seas (Pettijohn, 1943, p. 948). One may note here that one of the outstanding discoveries made during the past decade by geophysical exploration in the Canadian Shield is the common occurrence — much more common than previously believed — of graphite and graphitic schists (Russell, 1952).

It appears that there are no geologic reasons contradicting the assumption of biogenic origin of the carbon in the Manitoban slates, in spite of their very ancient age.

Consideration of the isotopic constitution of carbon in the slates indicates that three of the specimens, viz. 1-A-48, 2-A-48, and 5-A-48 fall within the biogenic range. The last-mentioned slate, however, probably belongs to a complex younger than the Rice Lake group, and consequently, although satisfied that its carbon is of biogenic origin, I will omit it from the following discussion. Specimens 3-A-48 and C-1-R-50 which are rather strongly affected by crustal movements fall below the biogenic range.

The specimens 1-A-48 and 2-A-48 which are lithologically and petrographically similar are from the same locality. The isotopic constitution of carbon in them, however, is significantly different. The slate 2-A-48 with the higher carbon content which also shows fewer signs of alteration by mineralizing solutions of later origin has the higher $^{12}C/^{13}C$ ratio of the two. This pair of specimens indicates that later mineralization has decreased the $^{12}C/^{13}C$ ratio in carbon, but the mechanism of this change, whether diffusion or some other process, cannot be ascertained from the present material. In any case, the isotopic constitution of carbon in the said two specimens lies in the biogenic range, and because the slates, geologically and lithologically, are not different from their younger counterparts in which the biogenic origin of carbon has been established beyond doubt, I feel authorized to conclude that their carbon is of biogenic origin. If the other two specimens are considered, the same conclusion may be extended to their carbon also, even though the evidence appears less convincing. But these four slates belong to the same formation, and there is nothing to indicate a different manner of origin for them. Consequently, if a biogenic origin of carbon is assumed for the specimens with the highest $^{12}C/^{13}C$ ratios, it may just as well be assumed for the specimens with the lower ratios.

Like the Tampere carbonaceous schists, also the S. E. Manitoban carbonaceous slates appear to be, in general, surprisingly weakly metamorphosed, even though they now lie in an almost vertical position. This suggests that the original isotopic constitution of carbon in these rocks has remained almost or totally unchanged.

Shaw (1952, p. 142), by means of thallium analyses, found independent evidence indicating that the carbonaceous schist from Beresford Lake, like the *Corycium* phyllite, contained carbon of biogenic origin.

To recapitulate, geologic and isotopic evidence indicates that the carbon, in some of the carbonaceous slates of southeastern Manitoba at least, is of biogenic origin. In other words, living organisms were in existence already in the early pre-Cambrian times, some 2.55×10^9 years ago. There is, of course, no possibility of telling what kind of organisms these carbonaceous remains from the rocks of the Basement Complex represent. But with an age of 3.5×10^9 years (Collins, Russell, and Farquhar, 1953) for the upper lithosphere it appears possible that conditions were favorable for the creation of life soon after the making of the lithosphere.

I feel tempted to trespass, with due caution, on the domain of the origin of life. It is reasonably postulated that life did not appear on the Earth contemporaneously with the formation of the three uppermost geospheres, viz., the lithosphere, the hydrosphere, and the atmosphere, but was preceded by a nonbiologic period in the geologic evolution. It appears possible that the atmosphere during this prebiologic time was reducing rather than oxidizing and that reducing conditions still existed after the first living matter had evolved (see discussion and references in Urey, 1952 b).

Experiments have been made to investigate whether highly complicated chemical reactions could take place in reducing surroundings before the existence of biologic processes. The results so far obtained indicate that complex chemical compounds could be produced in thermochemical, photochemical, and electrochemical processes (MacNevin, 1953). Furthermore, it has been proved that several amino acids actually are produced by electrochemical means in an atmosphere of methane, ammonia, water, and hydrogen (Miller, 1953). These results support the idea suggested by Oparin (1938) and seconded by Bernal (1949, pp. 545, 546; 1951) and Urey (1952 a, pp. 149-157; 1952 b, pp. 356, 357) that numerous prebiologic organic compounds formed in a primitive reducing atmosphere, or before the advent of living organisms in general (Blum, 1951, pp. 161, 163, 164).

It is universally agreed that the first life was aquatic and that the first great expansion of life took place in the sea or in adjoining shallow waters (Gulick, 1948, p. 267). The primordial ocean obviously provided very favorable surroundings for the development of primitive organisms because it must have contained in solution numerous inorganically made organic compounds as Urey (1952 a, pp. 152, 153; 1952 b, p. 357) suggested.

Thode, Macnamara, and Fleming (1953) investigated the isotopic constitution of sulfur in sulfides and sulfates of marine origin and concluded that isotopic fractionation of sulfur between sulfide and sulfate did not begin until approximately 0.8×10^9 years ago. Their interpretation of this result was that living organisms which obtain their free energy from oxidation of hydrogen sulfide and sulfur did not appear much before the said time, even though biogenic matter and living organisms already certainly existed. They also felt that their date probably preceded the start of large-scale photosynthesis. Urey (1952 a, pp. 153, 154; 1952 b, pp. 359, 360) said that this date might mark the transition from reducing to oxidizing conditions in the atmosphere, because extensive oxidation of sulfur and sulfides to sulfates would take place only after the appearance of free oxygen or the development of advanced photosynthesis. He concluded that it seemed just barely possible that reducing conditions were maintained until some 0.8×10^9 years ago.

Urey *(loc. cit.)* also claimed that great quantities of limestone were deposited early in the Earth's history while graphite was not deposited, because consideration of thermodynamic equilibria indicates that limestone and graphite would not be stable at the same time in a hydrogen-methane atmosphere. I fail to comprehend Urey's logic, inasmuch as it is very hard to visualize the formation of graphite deposits on the Earth's surface by direct atmospheric action. Furthermore, there is plenty of evidence of the existence of finely disseminated carbon in pre-Cambrian argillaceous sediments, even among the most ancient formations known, such as the carbonaceous slates of southeastern Manitoba. Geologic evidence therefore contradicts Urey's hypothesis. He admitted, to be sure, that thermodynamic equilibria need not have been attained after all and that the presence of living organisms almost certainly would cause important deviations from such equilibria.

If one assumes that the time, 0.8×10^9 years, actually marks the transition from a reducing to an oxidizing atmosphere, it is possible to explain, as follows, the manner of formation of the finely disseminated carbon with an isotopic constitution in the biogenic range found in pre-Cambrian slates and schists older than 0.8×10^9 years. This carbon is assumed to represent the remains of primitive organisms of an unknown kind capable of living under reducing conditions. Because atmospheric oxygen was absent, the carbon in their remains could not be oxidized into carbon dioxide but rather remained in the form of reduced native carbon. Because of the occurrence of this carbon as a finely disseminated pigment in the metamorphosed sediments, it is possible that the organisms were very small. If that is the case, the carbonaceous matter in the slates and schists may be considered to be constituted of microfossils.

There is, however, another possible explanation. One may justly ask whether the isotopic constitution of carbon could not become affected and change into the biogenic range in the photochemical and electrochemical processes in which organic molecules were produced in the early atmosphere and the early hydrosphere. This problem has not been investigated, and it is uncertain whether experiments like those made by Miller (1953) can illustrate this point, because they refer to a completely closed system. Unfortunately, there is no safe way of deducing the isotopic constitution of carbon in the methane and carbon dioxide of an early pre-Cambrian atmosphere. Furthermore, if such a change is postulated, or if it is proposed that the isotopic constitution of carbon in methane and carbon dioxide inherently lay in the biogenic range and was not changed in the subsequent photochemical and electrochemical reactions, then geologic evidence becomes void and the uniformitarian principle ceases to be valid. It appears that, because of the lack of evidence, this problem is academic, and its further discussion must be relegated to limbo. The existence of a reducing early pre-Cambrian atmosphere, on the other hand, appears thoroughly possible geologically, and I hope to return to some problems connected therewith in a later paper.

Since Hutton's times, it has been the belief among geologists that no trace is visible of the oldest geologic happenings on the Earth, and that everything has been erased by later evolution. But, as Eskola (1951, p. 35) said, this thesis is no longer infallible, and every effort should be welcomed to trace the vestiges of a beginning. One of the avenues that leads to the earliest geologic frontier is the investigation of carbon in ancient rocks of argillaceous origin. One may assume, in addition, that other ways and means will become available with the increased knowledge of the geology of early pre-Cambrian formations. With particular reference to the most ancient pre-Cambrian terranes it may be remarked that southeastern Manitoba makes a part of the early pre-Cambrian Superior province which, along with the Yellowknife province in western Canada, is the oldest known part of the Canadian Shield (Wilson, 1949, p. 235). They differ from all the other provinces of the Shield in their high ratio of lavas to sediments. Wilson suggested that when the Superior province was formed the conditions were rather different from what they are today. A detailed investigation of such terranes is certain to shed new light on the earliest geologic happenings.

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3

MINERALOGY OF MAFURITE [1]

BY

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ABSTRACT

This paper deals with the mineralogy of mafurite from Kyambogo crater (westernmost or dry Mafuru crater), Bunyaruguru field, S. W. Uganda. The rock, a volcanic lava consisting mainly of olivine, clinopyroxene, and kalsilite, was first described from that locality by A. Holmes. Another specimen of lava from the original locality was investigated in detail. Chemical analyses of the bulk rock and of kalsilite, olivine, and clinopyroxene extracted from the specimen are presented. Microscopic description of the rock and optical and other properties of its main constituents are given.

The rocks of the Bunyaruguru volcanic field south of Lake George in southwestern Uganda, that were geologically mapped by the late Dr. A. D. Combe, of the Geological Survey of Uganda, have been subjected to extensive petrographic studies in a series of papers by Professor Arthur Holmes, of Edinburgh. In addition to very important and interesting results dealing with these highly potassian ultrabasic alkaline lavas, his investigations led to the discovery of a new natural polymorph of $KAlSiO_4$, called kalsilite. The mineral, found in the westernmost of the two Mafuru craters (Holmes, 1942a), occurs in a lava that consists mainly of olivine, clinopyroxene, and kalsilite. For this rock, a new name, mafurite, was given by Holmes. The chemical composition, crystal structure, and physical properties of kalsilite were determined by Bannister and Hey (1942). Their work was carried out on material supplied by Holmes from Combe's original collection, that had been placed at his disposal by the Uganda Geological Survey.

According to the description of mafurite from the Mafuru crater given by Holmes (1942a), kalsilite occurs in the interstitial background of the rock, poikilitically enclosing the melanocratic constituents. Together with

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[1]) Received November 9, 1953.

pyroxene and other constituents, it also forms segregations (pegmatoids) in which the grain size is larger than in the surrounding lava. The specimens available to Holmes were ordinary hand specimens collected for microscopic and other work. Therefore, the amount of material from which kalsilite was extracted for closer study was very limited. In fact, the kalsilite material used by Hey (Bannister and Hey, 1942) for microchemical analysis was extracted from one single segregation, about 1 cm in diameter. The small amount of material available for mineralogical study offered a considerable difficulty in purifying the kalsilite sample for analysis. The impurities present in the segregation could not be removed completely from the kalsilite to be analyzed without too greatly reducing the size of the sample.

After kalsilite had been discovered in the lavas of the Mafuru crater, its presence was established in a number of other localities of the Bunyaruguru volcanic field. Because of the exceptionally pronounced potassian character of the lavas of that field, kalsilite, and not nepheline, together with leucite, seem to form the dominating leucocratic constituent of the rock suite. In addition to the Bunyaruguru field proper, kalsilite was found in the lava of the isolated Katunga volcano (Holmes, 1950) that lies approximately ten miles southeast from the Bunyaruguru field. No chemical analysis or physical data have been published so far of the mineral from that volcano. A paper dealing with the mineralogy of the Katunga lava, called katungite by Holmes, is under preparation by Dr. Neuvonen. Recently, kalsilite has also been found at two localities of the North Kivu area, viz. at Mt. Goma and Mt. Nyiragongo, in Belgian Congo (Sahama, 1953b). There the mineral is perthitic and contains interlaminated strips of nepheline and occurs in the lava together with nepheline as separate grains and as margins around the microperthitic kalsilite crystals. Outside East Africa, kalsilite has been found only in San Venanzo, Italy (Holmes, 1942b; Bannister, Sahama, and Wiik, 1953).

In connection with the author's work dealing with the general mineralogy of the natural nepheline-kalsilite series, now in progress, it seemed very desirable to obtain some additional material of kalsilite from the original locality in the Bunyaruguru field. This was one of the reasons why the author undertook an expedition to East Africa in 1952 (Sahama, 1953a). During the visit in Uganda, an excursion was made to the locality, together with Dr. K. J. Neuvonen, of the Geological Survey of Finland, and Mr. Kai Hytönen, of the University of Helsinki.

The central part of the Bunyaruguru field is shown in Fig. 1, reproduced from Combe's original map (unpublished). The Mafuru crater proper, the bottom of which is covered by a small lake, is situated immediately on the road that runs from the Kazinga Channel and Fort Portal straight through the volcanic field down to the towns of Mbarara and Kabale. On the western side of the Mafuru crater, there is another crater the bottom of which is dry. Its truncated western side slopes gently down into a swampy terrain. This crater, called the westernmost or dry Mafuru crater by Holmes, bears the name of Kyambogo crater on Combe's map. Accordingly, in the following, only the easternmost of Holmes' two Mafuru craters will be called the Mafuru crater and the vesternmost of his craters will be called the Kyambogo crater.

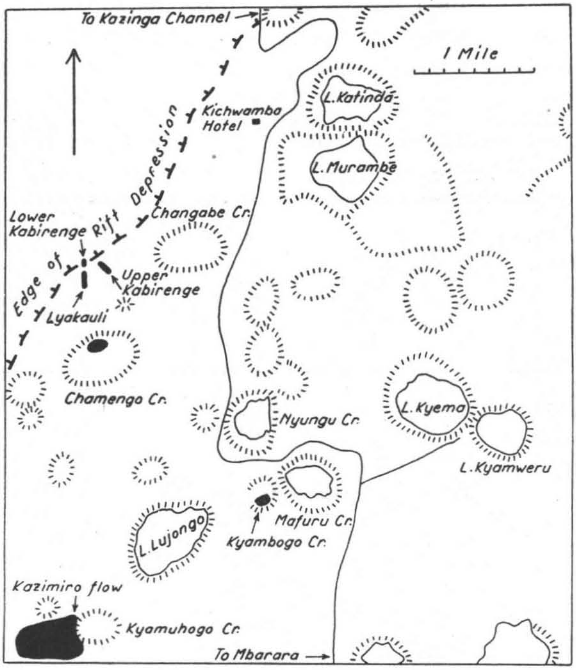

Fig. 1. Central part of the Bunyaruguru volcanic field, southwestern Uganda. According to the unpublished geological map by A. D. Combe, of the Geological Survey of Uganda.

Combe found no lava exposures on the very steep walls of the Mafuru crater. On the truncated western edge of the south side of Kyambogo crater, he found several exposures of mafurite lava that were overlain by tuffs with mafurite lapilli. The contact between lava and tuff is not exposed. From these lava exposures lying under the tuff, several specimens of considerable size were collected by the author and his coworkers. One of these specimens, FEAE No. 41, was selected for closer study.

Thanks to the courtesy of Professor Holmes, a thin section of his original specimen, C 6073, was available. Of that specimen, micrometric and chemical analysis as well as complete microscopic description has been given by Holmes (1942a). The material for the chemical analysis of kalsilite published by Bannister and Hey (1942) was also taken from the same specimen. Microscopic comparison of the two mafurite specimens, C 6073 and FEAE No. 41, indicates that the two rocks are virtually identical in mineral composition and texture. Therefore, a detailed microscopic description of the specimen FEAE No. 41 is not necessary. The similarity of the two specimens mentioned is also seen from the bulk composition of the rocks that is given in Table 1.

Table 1. Chemical composition of mafurite from Kyambogo crater. Bunyaruguru field, southwestern Uganda

Constituent	Specimen FEAE No. 41 From the lava sheet		Specimen C 6 073 Ejected block from the tuff above the lava sheet
	%	Molecular ratio multiplied by 10 000	Analysis published by Holmes (1942 a) %
SiO_2	40.65	6 765	39.06
TiO_2	2.40	300	4.36
Al_2O_3	8.44	828	8.18
Fe_2O_3	4.71	295	4.61
FeO	5.49	764	4.98
MnO	0.16	23	0.26
MgO	17.06	4 231	17.66
CaO	11.03	1 967	10.40
Na_2O	0.89	144	0.18
K_2O	6.79	721	6.98
P_2O_5	0.57	40	0.61
H_2O+	1.12	622	1.42
H_2O-	0.50	—	0.50
CO_2	0.10	23	trace
Total	99.91		99.94 [1]
Sp. gr.	3.02		

In the mafurite specimen FEAE No. 41, the following constituents were detected under the microscope: clinopyroxene, olivine, kalsilite, black ore, perovskite, a pale mica, apatite, and calcite.

The kalsilite grains that enclose poikilitically the dark constituents are often prismatic with well-developed crystal form. The prisms measure up to 0.3 mm in length. To prepare a sample of pure kalsilite for calorimetric work, about 15 kg of the rock was crushed and ground. The material was separated by means of the Clerici solution in a centrifuge. Two final concentrates were obtained the purities of which were tested in powder preparations mounted in Canada balsam. The first concentrate, approximately 40 g, has an average grain size less than 0.01 mm. It is hard to give any accurate figure to indicate its purity. Separate grains other than kalsilite are virtually absent. Many of the grains contain some tiny pigment

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[1] The total includes: F 0.13 S 0.13 Cr_2O_3 0.08 BaO 0.32 SrO 0.18 less O 0.10.

spots or a few small inclusions that apparently are clinopyroxene. The total amount of impurities might be as high as 1 % or 2 %, but is probably less. The second concentrate, approximately 160 g, has the same grain size and is slightly more impure. Powder patterns of the two concentrates show only kalsilite lines.

Chemical analysis of kalsilite from specimen FEAE No. 41 was made of the first (more pure) concentrate. The result and the physical properties of the mineral are given in Table 2. For comparison, similar data for kalsilite from the specimen C 6 073 are added to the table.

It is evident that a part of the iron, magnesium, and calcium found in the analysis is contained in the inclusions and pigment that are present in the kalsilite. On the other hand, the total amount of impurities in the analyzed kalsilite material is certainly not large enough to explain the bulk of these elements shown by the analysis. The figures for iron in the analyses of kalsilite from the specimens C 6073 and FEAE No. 41 are about the same. The kalsilite material extracted from the specimen C 6073 apparently was, however, less pure than that from specimen FEAE No. 41. This is shown also by the considerably higher figures for MgO and CaO in the analysis of kalsilite from specimen C 6073. Accordingly, the kalsilite from the specimen FEAE No. 41 seems to be richer in ferric iron and may even be called a ferrian variety of that mineral. The supposedly higher

Table 2. Chemical composition and physical properties of kalsilite from Kyambogo crater, Bunyaruguru field, southwestern Uganda.

Constituent	Specimen FEAE No. 41					Specimen C 6 073	
	%	Molecular ratio multiplied by 10 000	Unit cell content based on 0 = 8.00			Published by Bannister and Hey (1942) %	
SiO_2	38.50	6 407	Si	2.11		SiO_2	39.6
TiO_2	0.09	11	Al	1.70	2.00	Al_2O_3	21.3
Al_2O_3	26.27	2 577	Fe^{III}	0.21		Fe_2O_3	5.9
Fe_2O_3	5.07	317	Fe^{II}	0.02		MgO	3.7
FeO	0.53	74	Mg	0.07		CaO	5.0
MnO	0.02	3	Ca	0.03	1.99	Na_2O	1.6
MgO	0.87	216	Na	0.22		K_2O	20.1
CaO	0.44	78	K	1.74			
Na_2O	2.07	334	OH	0.38			
K_2O	24.85	2 638	0	8.00			
H_2O +	1.04	577					
H_2O —	0.05	—					
CO_2	0.00	—					
Total	99.80					97.2	
Sp. gr.	2.663					2.59 ± 0.03	
ω	1.554					1.542	
ε	1.550					1.537	

content of ferric iron in the kalsilite of the specimen FEAE No. 41 is in agreement with the considerably higher refractive indices and specific gravity of this kalsilite. The difference in composition of the two kalsilites might be explained by the fact that the analyzed kalsilite material from the specimen C 6073 was extracted from a pegmatoid segregation whereas the material of kalsilite from the specimen FEAE No. 41 is mainly from the poikilitical kalsilite grains of the groundmass of the lava.

The olivine, mostly occurring as phenocrysts, was extracted from the rock for chemical study. The result of the analysis and the physical properties of the mineral are given in Table 3. A microscopic grain count, made of the analyzed material, indicates 1 %—2 % clinopyroxene as impurity. A correction for this admixture was made, and the calculated composition of pure olivine is given in the third column of Table 3. According to the analysis, the mineral contains approximately 9 mol. % Fe_2SiO_4. The physical properties agree well with the composition of the mineral.

Table 3. Chemical composition and physical properties of olivine from mafurite (specimen FEAE No. 41). Kyambogo crater, Bunyaruguru field, southwest Uganda

Constituent	Original analysis %	Corrected for admixed clinopyroxene %	Molecular ratio multiplied by 10 000	Physical properties	
SiO_2	40.86	40.57	6 752	2Vγ	87°
TiO_2	0.12	0.10	13	α	1.654
Al_2O_3	0.25	0.24	24	β	1.669
Fe_2O_3	1.74	1.73	108	γ	1.686
FeO	7.15	7.19	1 001	Sp. gr.	3.338
MnO	0.08	0.08	11		
MgO	49.66	49.97	12 393		
CaO	0.33	0.00	—		
H_2O +	0.12	0.11	61		
H_2O —	0.00	0.00	—		
Total	100.31	99.99	—		

Clinopyroxene mostly occurs as tiny colorless prisms that form the main part of the groundmass. These prisms are, on an average, approximately 0.04 mm long and approximately 0.01 mm thick. Because of the small size of the prisms, the material had to be ground down to an average grain size of a few thousanths of a millimeter. This fine-grained material was floated several times in the Clerici solution in a centrifuge. It soon became evident that considerable difficulties were connected with a purification of the clinopyroxene concentrate. The small grain size of the material and the relatively small difference in specific gravity between clinopyroxene and olivine made a quantitative removal of olivine from the pyroxene concentrate impossible. To obtain an approximate view of the composition of this pyroxene, the final concentrate that contained up to 10 % foreign material (mostly olivine) was considered worthy of analysis. The result of the analysis is the following: SiO_2 50.96; TiO_2 1.21; Al_2O_3 0.80; Fe_2O_3 2.18; FeO 3.21; MnO 0.11; MgO 18.84; CaO 21.49; Na_2O 0.58; K_2O 0.29; H_2O+ 0.38; H_2O- 0.06; CO_2 0.00; Total 100.11. The specific gravity of the concentrate is 3.313. The following optical properties were measured on the pyroxene: $\alpha = 1.682$, $\beta = 1.694$, $\gamma = 1.712$, $2V\gamma = 62°$, $c \wedge \gamma = 44°$. The result of the analysis and the optical properties indicate that the mineral belongs to the diopside-hedenbergite series and is a salite.

In addition to the groundmass pyroxene, some few clinopyroxene phenocrysts occur in the rock. They are colorless and usually strongly zoned.

In the groundmass, small amounts of a pale brownish mica occur, poikilitically enclosing the pyroxene prisms. Its pleochroism is extremely weak, almost lacking. $\gamma \backsim \beta = 1.604$, $2V\alpha = 26°$. After several experiments, it was found impossible to purify the mineral for analysis. The black ore is very magnetic (magnetite). Perovskite (sp. gr. 4.04) is abundant in the groundmass. A chemical analysis of a similar perovskite from katungite lava from Katunga will be published by Dr. Neuvonen. A powder pattern of the carbonate mineral that is seen in the thin section agrees with that of calcite. The apatite needles are very scarce.

The chemical analyses of the bulk rock and of kalsilite were made by Dr. H. B. Wiik, and the analysis of the clinopyroxene was made by Mr. Pentti Ojanperä, M. A., both of the Geological Survey of Finland. The analysis of the olivine was made by Mr. Aarno Juurinen, M. A., and the optical determinations by Mr. Kai Hytönen, both of the Institute of Geology, University of Helsinki. To the gentlemen mentioned the author expresses his sincere gratitude.

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STRUCTURE AND STRATIGRAPHY OF THE MUSTIO AREA, SOUTHERN FINLAND [1]

BY

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ABSTRACT

A description is given of the structure of the Mustio cupola produced by the uplift of microcline granite. The stratigraphical succession of Archean supracrustal rocks at Mustio is determined. The lowermost known member in the succession is leptite, originally arkose, which contains interstratified layers of limestone. The leptite is overlain by a subsilicic volcanic bed. Mica schist, originally an argillaceous sediment, is the topmost member. The sedimentary rocks are described, and the evolution of the sedimentation is discussed.

CONTENTS

INTRODUCTION

The Mustio (Svartå) area is situated about 60 km W. of Helsinki, in southwestern Finland. Geologically it belongs to the leptite zone of the Svecofennian mountain chain which is of Archean age. The first special study of the Mustio area was published by Metzger (1928). His investigation

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[1] Received November 24, 1953.

dealt with the tectonics of Mustio, and he observed that the strikes run around the centre of the area and that the dips are outwards (»Diese Gesteine streichen ringförmig um Bruksträsk herum und fallen kuppelförmig nach aussen». Metzger, 1928, p. 5). He supposed also that a *massif* of microcline granite underlays the rocks in the central part of the Mustio area. A block stereogram completed Metzger's paper. Later Metzger dealt with the granite of Mustio (1947).

In 1937, the Mustio area was mapped by Dr. Erkki Mikkola. His notes were, however, incomplete and his death interrupted the investigations. His geological map of the Mustio area was published by Eskola (1941) in the necrology of Mikkola. In this paper Eskola considered the Mustio area structurally a »tumorartige Kuppel», and concerning its origin he wrote. »Die Entstehungsweise solcher Tumoren ist unklar, am wahrscheinlichsten erscheint mir, dass sie Magmadiapire ohne entblösste Plutonmassen darstellen» (Eskola, 1941, p. 460).

The ring-shaped structure appearing at Mustio is not rare in the Svecofennian territory of Finland. Several similar structural forms occur on a geological map published by Parras (1941).

The geological remapping of southwestern Finland has been continued since 1946. It has been necessary to revise certain areas. For the stratigraphical classification of some gneisses of supracrustal origin it was necessary to determine the stratigraphical succession in some structurally simple instances. One of these is the Mustio area. The present author re-examined the Mustio area during the summers 1950—1951 and in more detail in the summers 1952—1953. The new map (Figs. 9 a and 9 b) does not essentially differ from the map of Mikkola, but the more detailed observations make it possible to find the structure of Mustio cupola as well as the sequence of the supracrustal rocks there. The stratigraphical conclusions are based in part on observations made outside the Mustio area, and hence this paper is also a preliminary note of the field investigations and revisions made for the general geological map of southwestern Finland.

PETROGRAPHY

In the centre of the Mustio area (see Fig. 9 b), there occur stratified quartz-feldspar gneisses which have also been called leptites. These gneisses contain carbonate-bearing layers and interstratified layers of limestone. Intrusive lenses and dikes of microcline granite occur among the leptite gneisses.

The leptitic central part is surrounded by a ring of subsilicic rocks, called by E. Mikkola »banded diorite». These rocks consist of both homogeneous and banded as well as layered rock types. On the whole, the author considers this zone of subsilicic rocks to represent a bed of volcanic material which contains both ash layers and lava flows.

Outside the zone of the subsilicic volcanic rocks there is a belt of mica schist. This belt is surrounded on its northern and southern sides by zones of subsilicic rocks which are volcanic types rather similar to the rocks of the zone inside the ring of mica schist.

On the southern side of the area with the ring structure, there is a belt of leptite gneisses rather similar (also carbonate-bearing) to those in the central part. A zone of gneissose granites and granitized leptitic gneisses occurs on the northern side.

LEPTITES

The leptites in the Mustio area are pale-coloured, grey or reddish-grey quartz-feldspar gneisses which are commonly poor in mica. In spite of metamorphism the stratification is visible, but indistinct current bedding is rare, and graded bedding has not been observed.

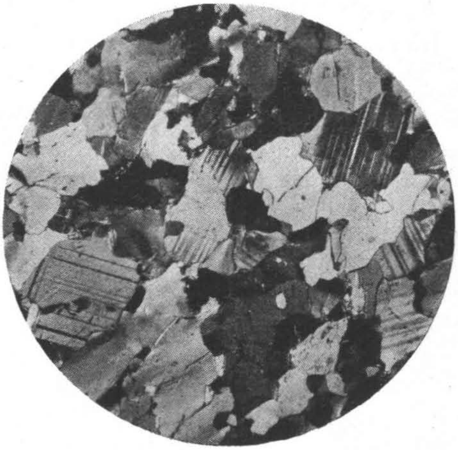

Fig. 1. Plagioclase leptite. E. of Brukssträsk. Nicols +. 20 ×.

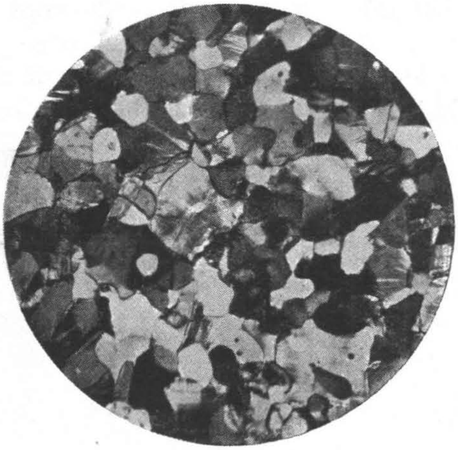

Fig. 2. Microcline leptite. Knapsby. Nicols +. 20 ×.

The leptites are commonly rich in feldspar. Both plagioclase and microcline occur (Figs. 1 and 2), although in several layers either feldspar may dominate (Table 1). In some quartzitic interstratified layers the feldspars are totally absent. Biotite occurs only sparsely and sometimes is absent altogether. Muscovite is rare. Sphene is commonly present as rounded grains, and usually is evenly distributed. The texture of the rock is granoblastic, but also blastoclastic features occur.

The leptites are moderately even-grained, and their grain size is usually

Table 1. Mineral composition of leptitic gneisses at Mustio in per cent, measured on an integration stage

Constituent	1	2	3	4	5	6
Quartz	34.75	46.22	39.92	37.81	40.71	78.06
Microcline	—	—	37.59	49.60	53.46	—
Plagioclase	57.30 [1]	43.68 [2]	9.28 [3]	—	0.36	—
Biotite	7.78	10.10	13.21	0.48	5.40	—
Diopside	—	—	—	0.61	—	19.75
Amphibole	—	—	—	10.03	—	—
Chlorite	—	—	—	1.11	—	—
Sphene	0.17	—	—	0.32	0.07	2.19
Apatite	—	—	—	0.04	—	—
Total	100.00	100.00	100.00	100.00	100.00	100.00

1. Plagioclase leptite, Lövkulla (N. of the mapped area).
2. Plagioclase leptite, Knapsby.
3. Leptite, Knapsby.
4. Amphibole-bearing microcline leptite, Knapsby.
5. Microcline leptite, Knapsby.
6. Diopside-bearing quartzite, Knabsby.

less than 1 mm. Some quartz grains may be greater, up to 2 mm in diameter; such grains are sometimes observed on the weathered surfaces.

Table 2 shows the chemical composition of a limestone, interstratified in leptite, found at Mustio (see p. 30). The limestone contains quartz and feldspar grains (Fig. 3), both scattered and as bands; these grains are dis-

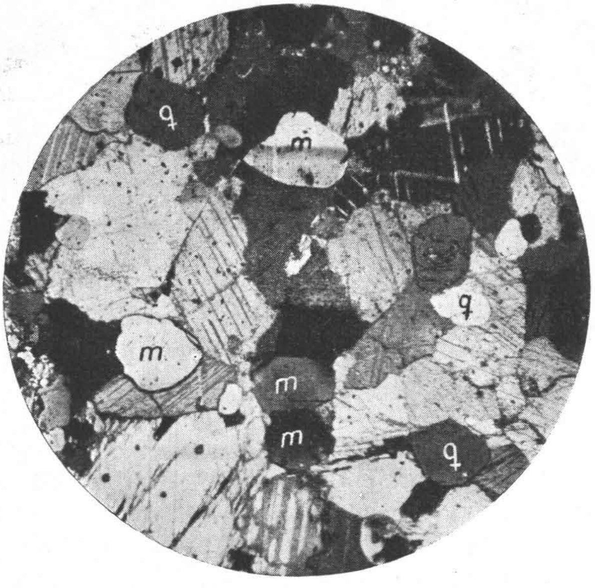

Fig. 3. Embedded grains of quartz (*q*) and microcline (*m*) in limestone. N. W. shore of Bruksträsk. Nicols +. 20 ×.

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[1] An_{10-20}. [2] An_{28-33}. [3] An_{30}.

tinctly rounded and obviously have retained their original shape. Some rounded grains of sphene occur in the limestone also.

Table 2. Chemical composition of limestone from Mustio.
Analyst, P. Ojanperä

Constituent	%
Insoluble	23.92
Al_2O_3 Fe_2O_3	3.88
CaO	40.51
MgO	0.29
CO_2	31.14
Total	99.74

Diopside- and/or amphibole-bearing, and almandite- and even cordierite-bearing varieties occur among the leptitic gneisses. The diopside-bearing leptites are, however, obviously produced by metamorphism from an originally carbonate-bearing material. The rocks 4 and 6 in Table 1 are examples of such rocks. These specimens were taken from different layers of the same exposure. Rock 6 obviously originally was a carbonate-bearing quartzite; in rock 4 the pyroxene is largely altered into amphibole, and in this rock potassium dominates because plagioclase is wholly absent.

It is characteristic of garnet-bearing leptites that they are normally granitized by microcline granite. Because the leptite is sedimentary in origin, it probably contained an excess of Al_2O_3 which allowed the formation of almandite. But it should be noted that the intrusive microcline granite also is regularly garnet-bearing. Hence it is difficult to tell in these migmatitized garnet-bearing gneisses when the almandite originates from the primary material of the leptite and when its material has been imported by granitizing microcline granite. Both origins seem to be probable.

Although the diopside-bearing gneiss has been produced from carbonate-bearing leptite, the skarnmaking reaction between siliceous matter and carbonate in the limestone layers has not been very pronounced. The rounded quartz and feldspar grains in the limestone are commonly undisturbed. From this blastoclastic texture (Fig. 3) it is also possible to conclude that the leptite was a coarse-grained sediment, obviously arkose. On the whole, it represents disintegration material which has undergone only a partial decomposition that has not greatly affected the feldspar. Sorting during transportation separated the more decomposed clayey material from the arenaceous debris. The mineralogical composition, the structure and texture, along with the presence of calcareous and limestone layers in the leptite, show that sedimentation took place in shallow water, where carbonate is known to be deposited. The original sediments were distal sediments and represented the foreland sedimentation facies.

Similar leptites nearly always occur together with limestones in southern Finland. The bulk of the pale-coloured diopside gneisses and diopside-scapolite gneisses of the so-called pyroxene gneiss area (Parras, 1941; Härme, 1954) north of Mustio are comparable to the diopside-bearing gneisses at Mustio. Also in the pyroxene gneiss area, the diopside gneisses were originally carbonate-bearing arenites, and they occur in the company of limestones.

In the pyroxene gneiss area north of Mustio and in the adjacent leptite areas in southern Finland, the limestones are mostly connected with stratified microcline-rich leptites. In the centre of the Mustio area, the limestones occur along with both microcline and plagioclase leptites. These two leptite types occur at Mustio irregularly interstratified, although the microcline leptites are more commonly met in connection with the limestone.

Metzger (1928, p. 4) mentioned a conglomerate-like rock near the railway station of Mustio. In this rock both the pebbles and the matrix are of leptitic composition. The pebbles are subangular and have been elongated by strong movement. Brecciation has taken place also. Because of deformation it is not possible to conclude whether this rock is a real intraformational conglomerate or whether it is a tectonic breccia only.

What, then, is the source of the material of the leptites? The presence of much feldspar and quartz shows that the original weathered rock was silicic. The grain size of the leptites indicates the possibility that it was a plutonic rock or at least a comparatively coarse-grained rock. The amounts of plagioclase and microcline vary greatly in the different layers, and there is no reason to suppose that any sorting of the feldspars took place during sedimentation. Thus it seems rather possible that at least two silicic (plutonic ?) rock types were weathered, one of them being rich in sodium and the other rich in potassium.

SUBSILICIC ROCKS

Subsilicic banded rocks, or »banded diorites» (see p. 30), are a distinct type remarkably common in southern Finland, and they make, both genetically and petrologically, a special problem.

In the Mustio area, field observations indicate that the zones of subsilicic rocks on the two sides of the ring of mica schist are lithologically comparable. This does not mean that the zones are of one coherent rock type, because there are both stratified and banded, and homogeneous rocks. The zones are bordered on both sides by stratified amphibolite (Fig. 4), which, towards the central part of the zones, gradually passes over into a banded amphibolite

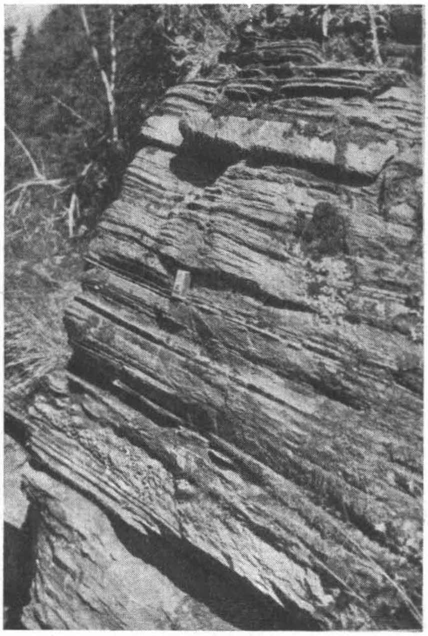

Fig. 4. Stratified amphibolite of volcanic origin at the N. N. E. border of the leptite of the centre. The rock contains interstratified layers of leptite. The bedding dips 30° N. 1/15 natural size.

which further grades into a homogeneous fine-grained gabbroic rock. The central parts of the zones are of medium-grained rocks of gabbroic and dioritic composition. This is the principal order of the rocks in these zones. But sometimes the homogeneous rocks cut the stratified and banded rock types, although most of the homogeneous rocks are clearly schistose. The average composition of the rocks varies from gabbro to diorite. Nearly all the subsilicic rock types are more or less deformed, and the degree of metamorphism is remarkably high, but the stratified structure still remains in the amphibolite.

The stratified amphibolites are related to banded types which have no clear stratification. Uralite porphyrites and some plagioclase porphyrites occur among them also.

The stratified rocks contain pale- and dark-coloured layers. The variation in colour is chiefly a result of a change in the amount of femic minerals. The pale-coloured layers contain plagioclase (An_{25-35}), quartz, biotite, and amphibole in varying amount. Sphene is a common accessory constituent. The dark layers consist mainly of plagioclase (An_{30-35}), amphibole, and a little quartz and biotite. Pyroxene is sometimes present. The texture of the rocks is granoblastic.

In the homogeneous rocks, the chemical composition varies from gabbroic to dioritic. The more fine-grained varieties in contact with stratified and banded amphibolites are mostly of gabbroic composition. The medium-grained rock types in the central part of the zones also contain dioritic portions. A small outcrop of peridotite was observed to adjoin the gabbro at the W. N. W. border of the leptite in the centre.

A homogeneous gabbroic rock in the northern part of the ring area contains plagioclase ($\alpha = 1.566$, $\beta = 1.570$, $\gamma = 1.574$; extinction $\perp$ PM $=$ 42°), amphibole ($c \wedge \gamma = 14°—15°$, $2V\alpha = 80°—82°$; $\alpha = 1.636$, $\gamma = 1.659$), pyroxene ($c \wedge \gamma = 43°$, $2V\gamma = 58°$; $\alpha = 1.675$, $\gamma = 1.702$), and a little pyrite. A chemical analysis (Table 3) was made of this rock. The pyroxene is partly uralitized and varies in amount. Sometimes it is absent. The plagioclase is weakly zoned. Sometimes a little sphene and apatite also

Table 3. Chemical composition of gabbro [1] from Bällby, Mustio. Norm, Niggli numbers and mineral composition determined on an integration stage. Analyst, H. B. Wiik

Constituent	%	Mol. prop.	Norm		Niggli numbers		Mineral composition	
SiO_2	48.75	.813	or	2.22	*si*	103.1	Plagioclase (An_{70-75})	46.13
TiO_2	0.61	.008	ab	11.53	*ti*	1.0	Amphibole	39.97
Al_2O_3	15.23	.149	an	34.19	*al*	18.9	Pyroxene	13.63
Fe_2O_3	1.22	.008	di	27.60	*fm*	47.1	Pyrite	0.27
FeO	7.77	.108	en	7.33	*c*	30.7		
MnO	0.17	.002	hy	5.13	*alk*	3.3		
MgO	9.83	.246	fo	4.20	*qz*	—10.1		
CaO	13.57	.242	fa	3.67	*k*	0.15		
Na_2O	1.37	.022	mt	1.86	*mg*	0.66		
K_2O	0.41	.004	il	1.22	*c/fm*	0.65		
P_2O_5	0.00		H_2O	1.45				
H_2O +	1.45							
H_2O —	0.00							
Total	100.38			100.40				100.00

occur. The texture (Fig. 5) is granoblastic and only partly blastohypidiomorphic. The grain size in the type analyzed is less than 1 mm but some grains may reach a diameter of 3 mm.

Fig. 5. Gneissose gabbro. Bällby. Nicols +. 40 ×.

The peridotite contains a pale-coloured amphibole ($c \wedge \gamma = 16° — 20°$, $2V\gamma = 64° — 78°$; $\alpha = 1.630$, $\beta = 1.643$, $\gamma = 1.654$, pleochroism weak)

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[1] Sp. gr. 3.01.

as its main constituent. It is obviously a product of uralitization. The rock also contains olivine ($\alpha = 1.690$, $\beta = 1.710$, $\gamma = 1.729$; $2V\alpha = 74°—78°$) and green spinel ($n = 1.77$). The olivine is only partly serpentinized. A little sulphide ore occurs as an accessory mineral.

A petrologically interesting variety of gabbro occurs in the N. E. part of the outer zone of subsilicic rocks. It is very rich in plagioclase ($\alpha = 1.575$, $\beta = 1.580$, $\gamma = 1.584$; extinction $\perp$ PM $= 38°$, maximum extinction 49°) and contains amphibole ($2V\alpha = 82°$, $c \wedge \gamma = 14°$; $\alpha = 1.631$, $\gamma = 1.653$), a little epidote, and carbonate. The rock is even-grained (Fig. 6), and its structure is weakly gneissose. The plagioclase is weakly zoned and wholly clear. The grain size is less than 1.7 mm.

Fig. 6. Anorthosite. Kleven. Nicols +. 20 ×.

The rock is homogeneous and igneous-like. An analysis, Table 4, was made of it. Its chemical composition indicates that it belongs to the anorthosite group of the system of Niggli (1923). The manner of occurrence of this rock type will be discussed on p. 42.

The stratified amphibolites surround the zones of subsilicic rocks on both sides. According to the field observations, the author considers these zones to be a bed of subsilicic volcanic rocks. The stratified amphibolites in the lower part of the bed represent the volcanic ash sediments of an early stage of eruption. The material was deposited in water. It also contains weathering products deposited contemporaneously with the ash. The homogeneous part of these zones is made of a comparatively coherent

Table 4. Chemical composition of anorthosite[1] from Kleven, Mustio. Norm, Niggli numbers and mineral composition determined on an integration stage. Analyst, H. B. Wiik

Constituent	%	Mol. prop.	Norm		Niggli numbers		Mineral composition	
SiO_2	45.32	.755	or	1.67	*si*	96.6	Plagioclase (An_{90-95})	68.63
TiO_2	0.14	.002	ab	5.76	*ti*	0.3	Amphibole	27.54
Al_2O_3	25.20	.247	an	64.77	*al*	31.6	Epidote	3.60
Fe_2O_3	1.13	.007	di	8.66	*fm*	23.6	Carbonate	0.23
FeO	2.30	.032	fo	7.00	*c*	43.0		
MnO	0.06	.001	fa	2.04	*alk*	1.8		
MgO	5.52	.138	wo	7.62	*qz*	—10.6		
CaO	18.82	.336	mt	1.62	*k*	0.21		
Na_2O	0.67	.011	il	0.30	*mg*	0.75		
K_2O	0.25	.003	H_2O	1.07	*c/fm*	1.82		
P_2O_5	0.00							
H_2O +	1.07							
H_2O —	0.00							
Total	100.48			100.51				100.00

lava flow. The coarser grain size in the central part of the zone is a result of the considerable thickness of the bed, because in the central parts of a thick lava bed the slow crystallization enhances the grain size, as the crystallization of the young lavas indicates. Furthermore, recrystallization commonly has caused an increase in grain size.

The stratified amphibolites of the upper part of the bed consist of tuffitic material deposited in water. These layers also contain debris from the partial decomposition of other rocks.

MICA SCHIST

The mica schist of Mustio is clearly stratified. Graded bedding is rare, and current bedding was not observed. The rock is rich in quartz, the amount of biotite is considerable, and the quantity of feldspar varies. Table 5 presents mineralogical data of two types of mica schist at Mustio. The excess of Al_2O_3 must have been considerable in the original sediment because in places the rock contains abundant almandite, cordierite, and sillimanite porphyroblasts. In the mica schist, the degree of metamorphism is remarkably high. The texture of the rock is granoblastic. In some places microcline granite has granitized the mica schist.

Rock 1 in Table 5 contains microcline, cordierite, and sillimanite as porphyroblasts. In this rock, however, the microcline may have been produced by granitization. The partly pinitized porphyroblasts of cordierite contain quartz, biotite, and plagioclase as inclusions. The rather large

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[1] Sp. gr. 3.01.

Table 5. **Mineral composition, in per cent, of two mica schists from Mustio, as determined on an integration stage**

Constituent	1	2
Quartz	26.12	43.82
Biotite	24.87	24.42
Plagioclase (oligoclase)	7.28	31.17
Microcline	23.75	0.42
Cordierite	17.42	—
Sillimanite	0.56	—
Apatite	—	0.14
Epidote	—	0.03
Total	100.00	100.00

1. Mica schist. Bällarby.
2. Mica schist. Mjölnarby.

porphyroblasts of microcline enclose grains of quartz, biotite, plagioclase, and cordierite. The quartz included in the microcline often has round and even hexagonal cross-sections. The sillimanite often occurs in connection with biotite.

Commonly these mica schists are plainly foliated. The grain size (except in the porphyroblasts) is less than 0.7 mm. The occurrence of aluminium-rich porphyroblasts and the high content of biotite indicate that the original sediment was argillaceous. The material was laid down in water, and most closely represents an axial geosynclinal sediment.

MICROCLINE GRANITE

Microcline granite occurs in the Mustio area as veins or intrusive lenses transecting all rocks in varying amounts. In the central part, the massive lenses follow the strike of the leptite gneisses, and the gneisses have also been granitized by microcline granite.

The massive microcline granite is commonly medium or coarse-grained and reddish in colour. Its main minerals are quartz and microcline. A little plagioclase and biotite occur in addition. Almandite is a common accessory, others being apatite and, sometimes, sericite and epitote.

In the centre, the leptitic gneisses are often injected by microcline granite. Granitic material has intruded as veins and thin veinlets along the planes of schistosity, so that the rock may sometimes contain more than 50 % of microcline granite. Such arteritic migmatites have, however, the general strike of the leptites. The thinnest individual granitic veinlets may be only 2 mm thick, and the gneissic bands may be equally thin. In some places the granitic veins cross the schistosity and even brecciate the gneiss. Such granitized gneisses always contain almandite both in the granitic and/or in the gneissic part. As a sedimentogeneous rock, the leptitic gneiss has probably contained an excess of Al_2O_3 which has been sufficient for the production of almandite, but the occurrence of almandite seems to be connected with the granitization by microcline granite. In other words, the microcline granite has also contained an excess of Al_2O_3 sufficient for the production of almandite. This is shown also by the fact that the massive microcline granites, as a rule, are almandite-bearing both at Mustio and in the neighbourhood, even when they occur with sharp contacts in environments where they have been unable to assimilate their excess of Al_2O_3.

STRUCTURE OF THE MUSTIO AREA

The map of the Mustio area shows that many rocks occur as ring-shaped zones. The strikes of bedding and of foliation run around the centre of leptitic rocks. The dips extend outwards from the centre; in the central part the dips are flat (20°—30°), and father away from the centre they steepen.

In the leptitic gneiss of the centre and in the inner ring of subsilicic rocks, the majority of lineations are directed outwards from the centre if not quite radially. In the outer ring zones, however, the lineations have a tendency to follow the strike of bedding and of schistosity. On the other hand, the majority of fold axes are directed according to the strike of bedding, although there also occur other fold axes which are directed outwards from the centre.

At the northern border of the leptitic centre, at its contact with the inner ring of subsilicic rocks, there occurs on the slope of a hill an outcrop of stratified amphibolite (Fig. 4), in which the dip is 30° N. Here the amphibolite contains more silicic, leptitic, interstratified layers in a number which increases downwards. The amphibolite grades over into leptite downwards at a distance of some ten metres. The dip indicates that the amphibolite vaults over the leptite. Similar phenomena have also been met in some other places.

Taking into consideration the ring-shaped structure and the attitude of the beds, one may state that the central part of the Mustio area has the form of a cupola as Metzger (1928) suggested.

The rocks, their manner of occurrence, and the similarity of structural features show that the zones of subsilicic rocks on both sides of the belt of mica schist are wholly comparable to each other. The zones are bordered by a stratified amphibolite which, towards the central parts of the zones, gradually passes over into homogeneous igneous-rock types. At the contacts between the amphibolite and leptite (see Fig. 9 b), the rocks grade by interstratification into each other. On the other hand at the contacts between the amphibolite and the mica schist, these rocks pass, also by interstratification, into each other. Graded bedding is rare in the mica schist but was once observed at the N. W. border of the northern mica schist belt. This observation shows the bottom of mica schist to be towards N. W. or, in other words, towards the stratified amphibolite. Consequently, the bed of subsilicic rocks is overlain by the mica schist. Therefore, it is reasonable to suppose that both the outer and the inner zone of subsilicic rocks belong to a single horizon, to an originally volcanic bed. This bed forms a ring-shaped syncline on which the mica schist lies (Fig. 7).

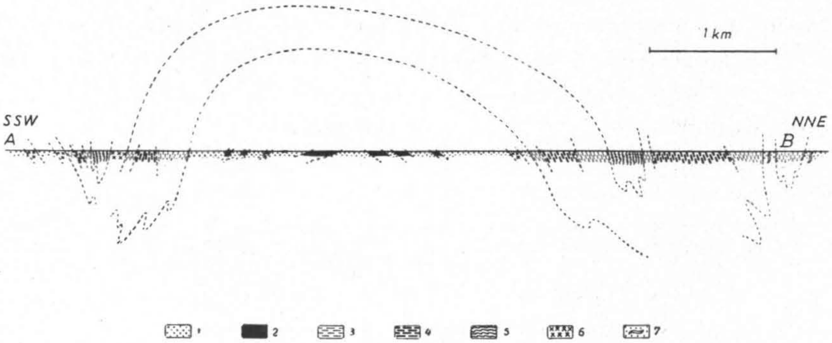

Fig. 7. Vertical section of the Mustio cupola. 1, leptite; 2, limestone; 3, stratified and banded subsilicic rocks; 4, homogeneous subsilicic rocks; 5, mica schist; 6, microcline granite; 7, mixture of granitized leptite and gneissose granite.

This opinion is also supported by the fact that southwards of the ring zones the leptite is of the same type as in the centre, and that it also contains carbonate-bearing layers. On the northern side the leptite and the granitized gneisses also are of the same type, although no carbonate is present. Therefore the horizon of the leptite also forms a syncline below the bed of subsilicic rocks.

In the leptitic central part, microcline granite occurs abundantly as intrusive lenses. The contacts of these lenses follow the common strike running around the centre, but the dips of the contacts are sometimes steeper than the dips of bedding and of schistosity of the leptite. Mostly the leptites of the middle part are strongly granitized migmatitic gneisses. These facts give clues to the process which has brought about the cupola structure. A pluton of microcline granite must exist below the centre of the area. Metzger (1928, 1947) and Eskola (1941) previously arrived at the same conclusion.

The fold axis and other tectonic features give no reason to suppose that the area of the ring-shaped structure represents any culmination of an anticline. On the contrary, the belt of mica schist continues as a syncline towards both the west and east, and this observation makes it possible to assume that the cupola was produced in the trough of the syncline. The updoming was caused by the upheaval of the microcline granite in a liquid state, pushing up the overlying strata. This movement stretched the strata and caused a laminar gliding in the layers. This gliding produced the secondary lineations now radially directed in the central part of the cupola. The cupola, however, has not a regular round shape, and therefore the laminar gliding outwards from the centre was not exactly radially directed and equally strong in different directions. Hence the lineations in the centre are not quite radial. The radial arrangement of lineations occurs also in the intrusive pluton itself, as is the case in the Värmälä *massif* described by Härme and Seitsaari (1950). Furthermore, the uplift of an intrusive pluton causes frictional drag in the immediate vicinity of the body, and the drag produces secondary lineations in the direction of the movement which is commonly vertical.

With regard to the uplift of the central part, a relative subsidence of the marginal zone of the cupola has taken place. In addition, by the uplift of the cupola the layers of the overlying strata relatively flowed down along the flanks of the cupola or at least caused a pressure in this direction. On the other hand, a pressure directed horizontally outwards is generated in an intrusion. These factors produced a secondary folding with nearly horizontal axes running around the raised part of the cupola. This folding is now observable in the flat axes of small folds which run around the cupola.

On the southern border of the cupola, the rock zones are narrower than on the northern border. On the southern border the zones were pressed strongly together. This is revealed by the dips which are steeper on the southern border than on the northern border of the cupola. Thus the cupola is to some degree tilted towards the south. This tilt can scarcely have been caused by any general thrust directed towards the south. The tectonic observations outside the Mustio area do not show signs of any common thrust southwards, and so it may be that here only a local force from the north is involved. On the other hand, in the N. E. part of the outer subsilicic rock zone there occurs a local overthrust towards the north, as Figs. 8 a and 8 b indicate. Fig. 8 a is a photograph of an outcrop, and Fig. 8 b is a tracing of this photograph on the same scale. In the central part of the rock wall there occur folds overturned towards the north. The gabbro was thrust over these folds, and the overthrust has cut away parts of the folds. The thrust plane occurs at the contact between the gabbro and amphibolite. This gabbro is of the even-grained foliated anorthositic type described on p. 37. The strike of the anorthosite is nearly at right angles to the plane of the picture, and the dip is downwards to the left and thus nearly perpendicular to the limbs of the folds of the amphibolite layers.

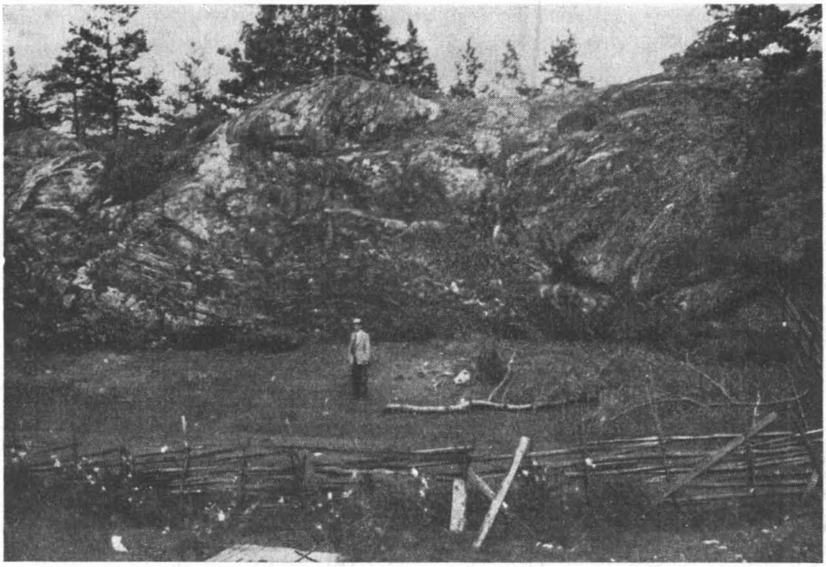

Fig. 8 a. Photograph of an outcrop at the N. E. part of the outer subsilicic rock zone, Hållsnäs, S. E. of Kleven.

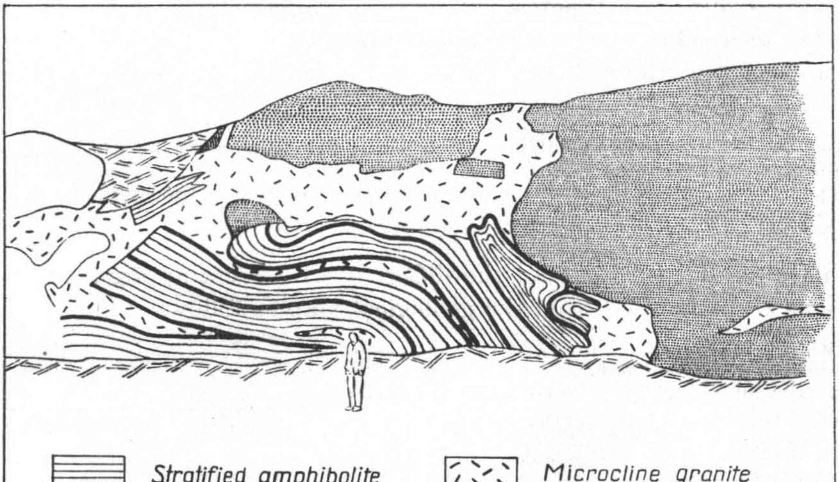

Fig. 8 b. Profile drawn from Fig. 8 a. The photograph was taken in the direction N. 30°E.

As the youngest rock, the microcline granite transects both the amphibolite and the anorthositic gabbro.

It seems evident that in this instance the overthrust of anorthosite took place in the solid state and obviously contemporaneously with the intrusion of the microcline granite. The chemical composition suggests that this anorthositic rock is not connected with the subsilicic volcanic rocks but represents an intrusive synkinematic rock.

This overthrust seems, however, to be a local phenomenon, and no general conclusions can be drawn from it.

Some faults transect the ring zones. On the northern side they are directed to the N. W., but on the southwestern side a fault series intruded by microcline granite transects the zones in a W. S. W. direction. These faults were produced in connection with the updoming.

STRATIGRAPHY AND AGE RELATIONSHIPS

The bed of subsilicic rocks vaults over the leptite gneiss, and the bottom part of the subsilicic bed grades by interstratification into leptite. This shows that stratigraphically the bed of subsilicic rocks lies over the leptite. On the other hand, both the outer and inner zones of the subsilicic rocks at Mustio belong to the same horizon. The mica schist belt lies in the syncline formed by this horizon (see p. 41). The amphibolite in the upper part of the bed of the subsilicic rocks passes by interstratification into mica schist, and the few observations on the graded bedding show that the mica schist overlies the subsilicic rocks.

Consequently, the sequence, from top down, of the supracrustal rocks at Mustio is as follows:

Mica schist, originally argillaceous sediment, with an excess of Al_2O_3.
Subsilicic volcanic bed of tuffs and flows.
Leptite, originally arenaceous sediment, with interbedded limestone.
Base unknown.

The minimum thickness of the different strata was approximated from the vertical section and from the geological map as follows:

Mica schist	100 m
Subsilicic volcanic bed	500 m
Leptite, including the interbedded limestone	1 100 m
Total thickness	1 700 m

These numbers reveal the thicknesses only in Mustio. In Kasaby, 10 km W. of Mustio, the subsilicic bed is only approximately 60 m thick. The mica schist, the topmost member in the sequence at Mustio, is not present everywhere in the neighbourhood and is not directly comparable to the other mica schists and slates in southern Finland because they represent a lower horizon below the volcanics (Simonen, 1953).

The anorthosite and the peridotite occur in connection with the zones of subsilicic rocks, but it is likely that they do not belong to the volcanic rock series. They may be synkinematic intrusions.

Gneissose granite (granodiorite—oligoclase granite) occurs on the northern and northeastern borders of the Mustio area and transects as veins the leptite, subsilicic volcanic rocks, and mica schist. It is therefore younger than the supracrustal rocks. On the other hand, the microcline granite transects and granitizes both the supracrustal rocks mentioned and the gneissose granite, and hence it is the youngest plutonic rock in the Mustio area.

Recently Saksela (1953, p. 31) discussed the structure and the granites of Mustio. The present author will not here deal with the general basis of Saksela's classification of granites, but will only state that Saksela's opinion of the age of the microcline granite of Mustio is erroneous. The age of an intrusive body cannot be determined from its shape.

The uplift of microcline granite in a liquid state causes breaks in the overlying strata. The cracks traverse the layers, but they also follow the planes of schistosity and bedding and so make possible the intrusion of microcline granite. This is the reason why the microcline granite may occur as lenses following the strike of the gneisses in the roof of the cupola. Many lenses of microcline granite at Mustio follow the strike and dip of the leptite gneisses in the centre, but there also exist lenses which follow the strike but not the dip of the gneisses. According to the field observations made in connection with the geological mapping of southwestern Finland, the microcline granite is here late-kinematic in age. Consequently, at its periphery the late-kinematic intrusion may follow the »old designs» of the wall rock (Wegmann and Kranck, 1931).

Southwest of the centre of Mustio, the microcline granite cuts the ring zones of subsilicic volcanic rocks and mica schist as lenses following the fault series (see Fig. 8 b).

The leptitic gneisses are granitized by the microcline granite, and they consequently represent the migmatitic granitization zone in front of the upheaval of the microcline granite dome (*Migmatitfront*, Wegmann, 1935).

SUMMARY

The leptite gneiss of sedimentary origin is the lowermost member in the stratigraphic sequence. Its base is unknown. Both the carbonate-bearing and the limestone layers occur in the centre, i. e., in the lower part of the leptite bed which is exposed. The arenitic material of the leptite was laid down in water, most probably in shallow water in the distal area where carbonate is known to be deposited (the foreland facies, Pettijohn, 1949, p. 451).

Outwards from the centre, i. e., towards the top in the sequence, the leptite passes over to a type which is more fine-grained and contains more biotite. This may indicate the subsidence of the sedimentation base.

Upwards the leptite passes into a stratified amphibolite which is the lowermost part of the volcanic bed. A lava flow overlies the stratified amphibolite, and the uppermost part of the subsilicic volcanic bed again consists of stratified amphibolite. The material of both these amphibolites was deposited in water.

The upper amphibolite passes into a mica schist which was originally an argillaceous sediment. Its material was deposited in deep water in the axial area (the geosynclinal facies, Pettijohn, 1949, p. 443).

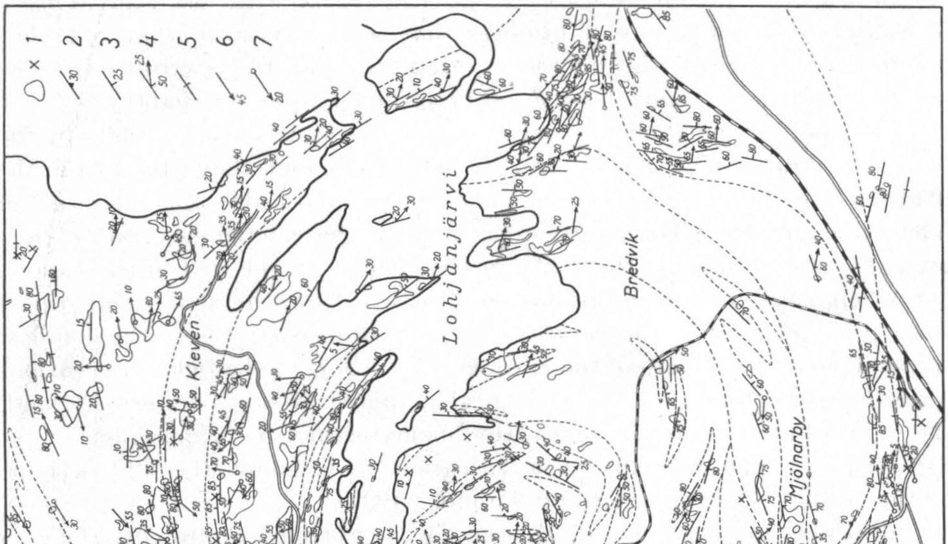

Fig. 9 a. Tectonic map of the Mustio area. 1, outcrops; 2, bedding; 3, foliation; 4, foliation and lineation; 5, vertical foliation; 6, lineation; 7, fold axis.

Consequently, it seems that a subsidence of the sedimentation level took place. The arenaceous distal sediments pass into argillaceous axial sediments, and the eruption of subsilicic volcanic rocks represents an interruption in the evolution from the platform facies to the geosynclinal facies.

These opinions relative to the character of the sedimentation at Mustio are supported by the observations made by the present author in the geological remapping and revision of the area of the general map sheet Turku.

Fig. 9 b. Lithologic map of the Mustio area. 1, leptite; 2, limestone; 3, stratified and banded subsilicic rocks; 4, homogeneous subsilicic rocks; 5, mica schist; 6, microcline granite; 7, gneissose granite and granitized gneiss; 8, gneissose granite; 9, conglomerate-breccia.

The observations also show that in the Svecofennian mountain chain in southwestern Finland there exists a comparatively gently folded zone and that the cupola-like intrusions are not rare in this zone.

In the Mustio area also, the supracrustal strata seem to have been only gently folded during the Svecofennian folding. The diapire-like intrusion of microcline granite then took place during the late-kinematic phase; and this intrusion caused the vaulting of the overlying strata. There is nothing to suggest that there is any true anticline, but it seems most likely that the cupola structure is caused simply by the vertical uplift of the microcline granite.

ACKNOWLEDGMENTS

Mr. Jyry Saastamoinen assisted the author in the field work. Miss Thyra Åberg drew the maps and figures, and Mrs. Toini Mikkola, M. A., made a part of the optical determinations. The chemical analyses were made by Dr. H. B. Wiik and Mr. P. Ojanperä, M. A. To all the persons mentioned the author is greatly indebted.

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5

MEASUREMENTS OF POLLEN OF *ALNUS GLUTINOSA* AND *A. INCANA*[1]

BY

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ABSTRACT

Measurements of recent pollen of *Alnus glutinosa* and *A. incana* showed that there were significant differences between them in the a/b ratio, where a is the height of the pore and b its greatest diameter. An example of the use of this difference in identifying fossil *Alnus* pollen is given.

Measurements of recent pollen of *Alnus glutinosa* and *A. incana* were made in order to see if it is possible to distinguish between the pollen of these two species, and to obtain a basis for measurements of fossil pollen. Although the material used was not sufficient for a more detailed investigation, it gave some idea of the possibility of this method. Only one example of measurements of fossil pollen is given here.

Erdtman (1936, 1943) and Bertsch (1942) pointed out that the pollen of *A. incana* has broader pores than the pollen of *A. glutinosa*, and that it is thus possible to distinguish between the pollen of these two species, but they did not give any results based on a large number of observations. The measurements here discussed showed that the width of the pores compared with the distance between them did not show consistent differences between the two species, and it is therefore not suitable to use these values; on the other hand the height of the pore compared with the diameter of the pollen showed a marked difference, and this ratio, which was used here, is comparatively easy to measure under a microscope.

In the measurements which were made during 1952 and 1953 with a magnification of approx. 1 350 × and an oil-immersion objective, one pore and the diameter (the greatest) were measured on each pollen grain with an accuracy of $^1/_2$ and 1 μ, respectively. For each pollen the a/b

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[1] Received November 3. 1953.

ratio (Fig. 1) was determined and given as a percentage, and these pore-values were used in the diagrams to the nearest whole number. The averages for the pore values are, however, determined from the original values and given to one decimal place.

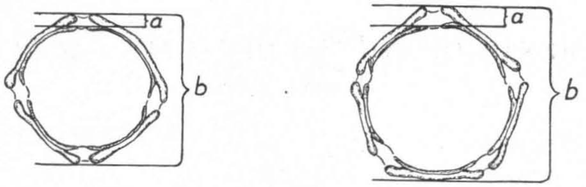


Fig. 1. The way in which pore value a/b has been measured.

The acetolysis method, as carried out by Faegri and Iversen (1950), was used for the recent and the fossil material, and the preparations were mounted in glycerine jelly. The recent material was from the herbariums in the Finnish Forest Research Institute and the Department of Botany, University of Helsinki. The following tabulation shows the material used; the number of reparations; where, when, and by whom the material was collected.

No. 1.	*Alnus glutinosa,*	Turku (Åbo), 1907, V. A. Pesola.
» 2.	» »	Botanic Gardens, University of Helsinki, 1949, L. Korhonen.
» 3.	» »	Sipoo (Sibbo), Hitå, 1952, G. Marklund.
» 4.	*Alnus incana,*	Sippola, Ruotila, 1941, V. Kujala.
» 5.	» »	Sippola, Ruotila, 1941, V. Kujala.
» 6.	» »	Pasila (Kottby), Helsinki, 1946, S. Saarnijoki.

In each of the 6 preparations, 50 pollen grains with 4 pores, and 50 with 5 pores were measured. The results are given in the curves in Figs. 2 and 3, where the x-axis shows the different pore values (the a/b ratio as a percentage), and the y-axis the number of pollen grains in each group. Each curve has the number of the preparation. The peaks for the curves of *Alnus glutinosa* with 4 pores are separated, but from Table 1, in which the results are given, one sees that the average does not show great differences, being 8.5 %, 7.7 %, and 8.5 %. The curves coincide very closely for *A. incana* with 4 pores, as the curves for pollen grains with 5 pores do for both species. In Fig. 4 the results are put together in the same diagram, and here the y-axis shows how many percent of the pollen grains there are in each group. The diagram shows clearly that the curves for pollen grains with 4 and 5 pores coincide, and the average (Table 1) for *A. glutinosa* is 8.2 % and 8.3 %, and for *A. incana* 9.9 % and 10.1 %. Because the averages for

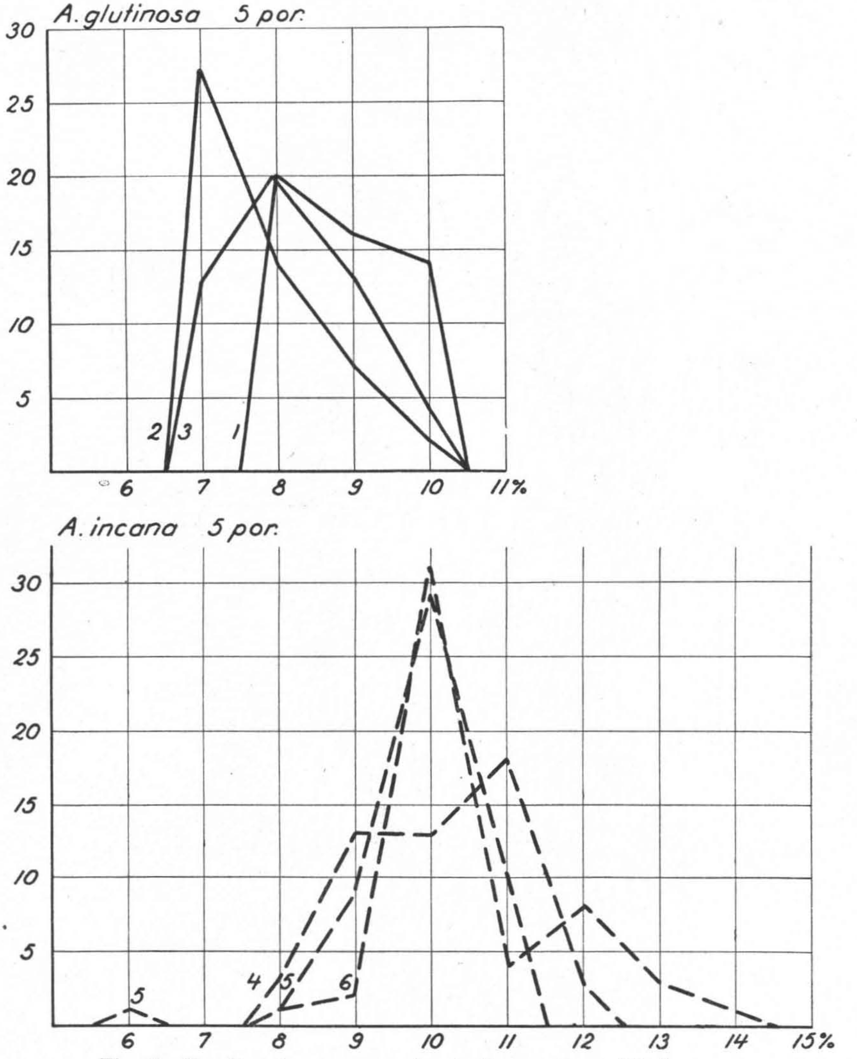


Fig. 2. Results of measurements of *Alnus* pollen with 5 pores.

pollen grains with 4 and 5 pores are nearly the same, the measurements can be made without separating these two. The difference between the two species is about 2 %, or great enough to make it possible to distinguish between them, but the variations in *a/b* are so great — for *A. glutinosa* 6 %—12 % and for *A. incana* 6 % — 14 % — that it is necessary to have a great number of measurements to get significantly different averages. The values are so near each other that the two species have several identical types. The diameter showed the same variations for both, being 23 μ — 35 μ.

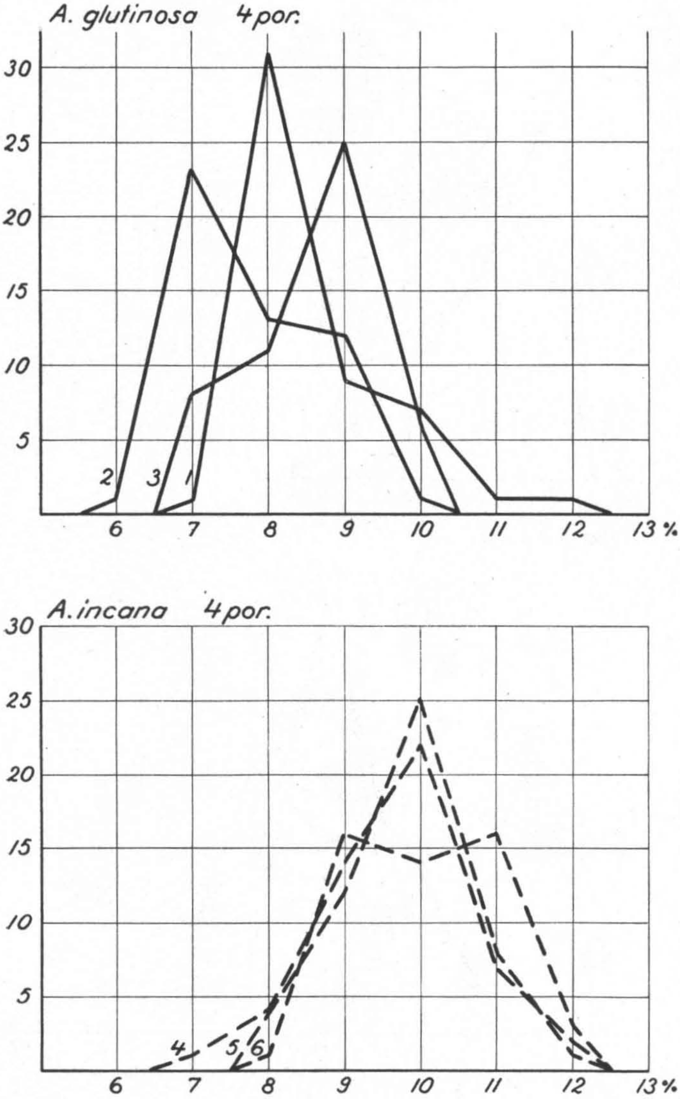


Fig. 3. Results of measurements of *Alnus* pollen with 4 pores.

As an example on applying the method to fossil material, some *Alnus* pollen were measured in a pollen diagram (Fig. 5) from Hanaböleträsk, a small lake about 20 km N. E. of Helsinki. Before the rise of the *Alnus* curve at the transition from zone V to zone VI, there is more than 10 % of *Alnus* in the upper part of the pre-Boreal zone IV. In a preparation from 4 m depth 20 pollen grains were measured, and the results are given in Fig. 6, where the diagram is drawn in the same way as in Figs. 2 and 3. Even if the number of measured pollen grains is only 20, the curve shows a marked maximum at 8 %, and the average is 8.1 %, which is nearly the same as the average for *A. glutinosa* (8.2 %—8.3 %). With this regular curve and the small variations of the values, one may assume that these pollen represent *A. glutinosa.*

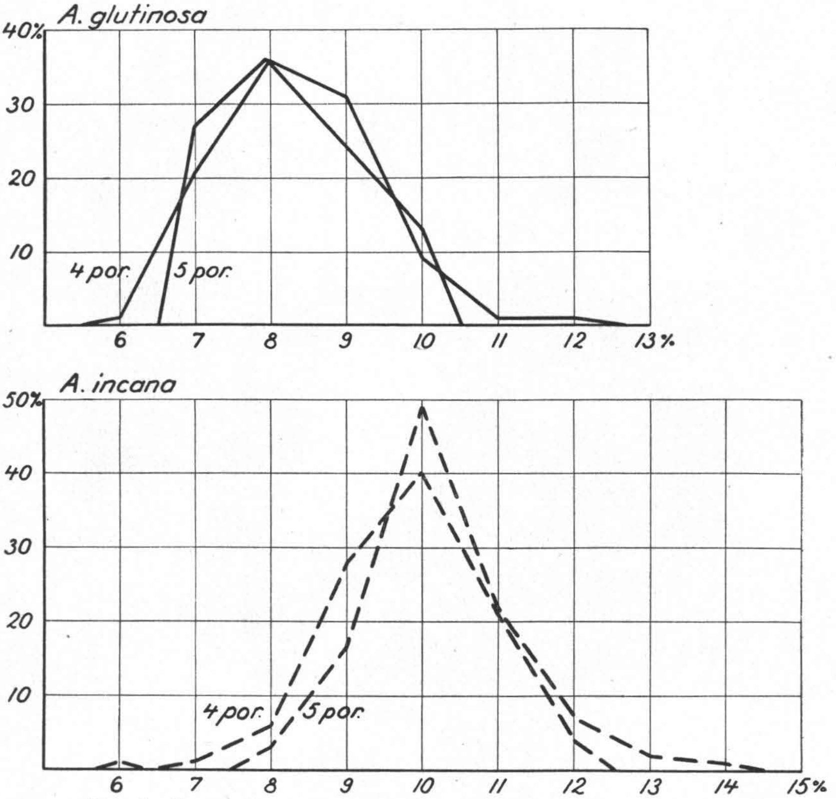
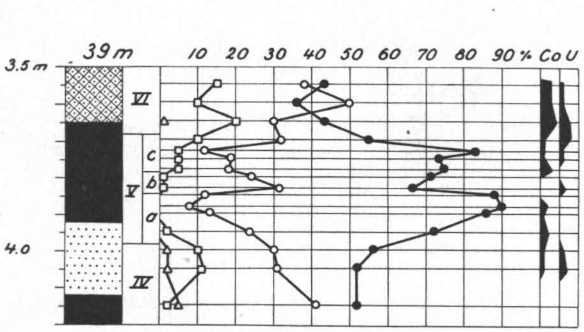


Fig. 4. Curves for measurements of *A. glutinosa* and *A. incana.*



Fig, 5. Pollen diagram from Hanaböleträsk, 20 km. NE of Helsinki. Stratigraphy from the bottom upwards, clay, sand, clay and claymud; circles, *Betula;* dots, *Pinus;* rectangles, *Alnus;* triangles, *Picea;* Co, *Corylus;* U, *Ulmus.*

The high percentage values for *Alnus* in zone IV do not constitute a local feature in this diagram. It is common in most of the diagrams in the Helsinki area, as shown by Sauramo (1954), who analysed a diagram from another part of Hanaböleträsk. The high *Alnus* percentage is con-

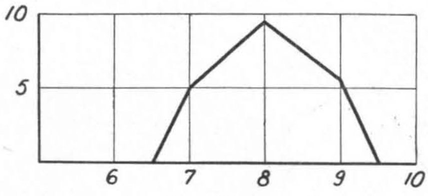


Fig. 6. Curve for measurements of pre-Boreal *Alnus* pollen from Hanaböleträsk.

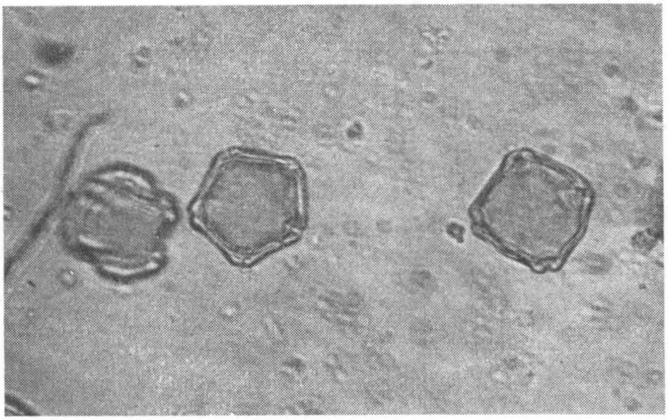


Fig. 7. Photograph of *A. glutinosa*, prep. no. 1. (Photo E. Viluksela, 1953).

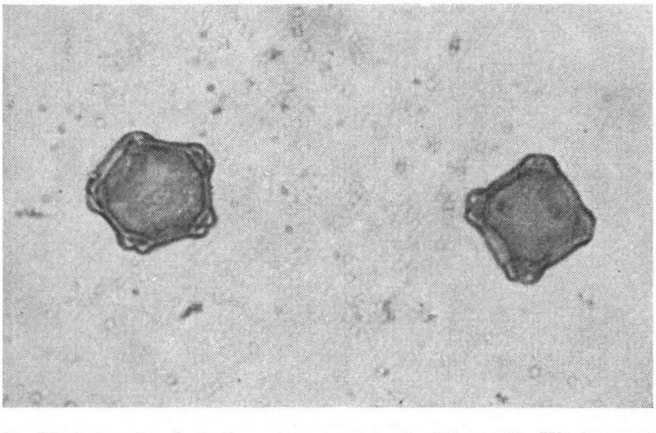


Fig. 8. Photograph of *A. incana*, prep. no. 4. (Photo E. Viluksela, 1953).

temporary with the regression of the Yoldia Sea in the latter part of the pre-Boreal period, and this regression was followed by a transgression of the Echineis Sea (Sauramo, 1954), which in the diagram (Fig. 5) is represented by the clay overlaying the sand. In Ruotsinkylä, about 5 km N. W. of Hanaböleträsk, roots of *Alnus* were found in a sand layer, which is covered by another sand layer formed during the transgression of the Echineis Sea (Sauramo, 1954). Determinations by prof. V. Kujala (Sauramo, 1936, 1954) showed that these roots are from *A. glutinosa*, and thus the macroscopic remains confirm the assumption that *A. glutinosa* was common in the Helsinki area already during the regression of the Yoldia Sea in the end of the pre-Boreal period.

Table 1. Results of measurements of recent pollen of *Alnus glutinosa* and *A. incana*.

		Alnus glutinosa										*Alnus incana*									
		4 pores					5 pores					4 pores					5 pores				
		No. of preparation			Total	%	No. of preparation			Total	%	No. of preparation			Total	%	No. of preparation			Total	%
		1	2	3			1	2	3			4	5	6			4	5	6		
ratio *a/b* in %	6	—	1	—	1	1	—	—	—	—	—	—	—	—	—	—	—	1	—	1	1
	7	1	23	8	32	21	—	27	13	40	27	1	—	—	1	1	—	—	—	—	—
	8	31	13	11	55	36	20	14	20	54	36	4	4	1	9	6	3	1	1	5	3
	9	9	12	25	46	31	16	7	13	36	24	14	12	16	42	28	13	9	2	24	16
	10	7	1	6	14	9	14	2	4	20	13	22	25	14	61	40	13	29	31	73	49
	11	1	—	—	1	1	—	—	—	—	—	7	8	16	31	21	18	10	4	32	21
	12	1	—	—	1	1	—	—	—	—	—	2	1	3	6	4	3	—	8	11	7
	13	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	3	3	2
	14	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	1	1	1
Number of pollen		50	50	50	150		50	50	50	150		50	50	50	150		50	50	50	150	
Average *a/b* ratio in %		8.5	7.7	8.5	8.2		8.9	7.7	8.2	8.3		9.9	9.8	10.1	9.9		10.0	9.9	10.5	10.1	

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6

FORMATION PAR MÉTASOMATOSE DE QUELQUES STRUCTURES ZONÉES DANS LES PLAGIOCLASES VOLCANIQUES [1]

PAR

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RÉSUMÉ

On décrit les phénocristaux de plagioclase avec structure zonée, et les associations de ce minéral avec de la sanidine et de l'opale, dans les dacites tertiaires du Campo de Vera (Almeria, Espagne). Les caractères structurels de ces phénocristaux ne peuvent pas s'expliquer d'une manière satisfaisante par les mécanismes de croissance qui jusqu'à présent avaient été admis pour leur formation dans les roches magmatiques. L'on déduit que dans ces feldspaths il n'y a pas eu le même nombre de périodes de croissance que de zones, sinon seulement deux: le premier de simple cristallisation du plagioclase basique (83 % An en moyenne), l'autre postérieur de métasomatose sélective avec la cristallisation simple de plagioclase acide (48 % An en moyenne) ou de sanidine; cette dernière période peut se continuer avec la cristallisation simple de plagioclase acide de la même composition. On admet que le procès de métasomatose s'effectue sans la disparition du cristal solide déjà formé, au moyen d'un interchange ionique »liquidus-solidus» des éléments Na-Si et Ca-Al pour la formation du plagioclase, et, K-Si et Ca-Al-Na pour la formation de la sanidine. L'origine de cet interchange est un brusque changement physico-chimique dans le magma comme conséquence de l'assimilation de roches acides et alcalines. Finalement après la consolidation totale de la roche il s'est produit une autre métasomatose dans les plagioclases qui a donné l'origine à des associations de plagioclase acide et d'opale avec des caractères morphologiques analogues à ceux des associations précédentes.

TABLE DES MATIÈRES

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[1] Reçu le 4 Novembre, 1953.

INTRODUCTION

Au cours d'une étude pétrologique de la région volcanique tertiaire du SE de l'Espagne, dans les provinces de Murcia et Almeria, nous avons trouvé très fréquemment des phénocristaux de plagioclase avec des structures zonées complexes, dont les caractères ne peuvent pas être expliqués par les théories sur l'accroissement des feldspaths que nous connaissons.

Nous avons choisi comme exemple plus démonstratif les plagioclases des dacites de Vera, dans la province d'Almeria, dans lesquelles ces structures complexes sont plus facilement perceptibles. Dans la première partie du travail nous ferons la description des principaux types de plagioclases des roches de cette région; dans la deuxième nous tâcherons d'expliquer leur formation avec des critères un peu différents de ceux qui ont été utilisés jusqu'à présent. Cependant nous devons faire remarquer que nous n'avons pas la prétention d'élaborer une théorie générale qui puisse s'appliquer directement à tous les cas de formation de plagioclases. Nous pensons que les roches de cette région constituent un cas en quelque sorte spécial, quoique nullement unique, car nous avons souvent trouvé dans la littérature pétrographique des descriptions et des photographies de types de plagioclases très semblables à ceux que nous avons étudié.

Toute la partie experimentale de ce travail a été exécutée dans les laboratoires de l'Institut Polytechnique Fédéral de Zürich, pendant l'année 1952 et sous la direction des Professeurs Dr. P. Niggli et Dr. C. Burri; nous voulons ici leur témoigner notre sincère reconnaissance pour les grandes facilités et les précieuses indications que nous avons reçu d'eux. Les travaux de laboratoire ont été réalisés avec la collaboration de Mme Elisa I. de Fúster, à laquelle une grande partie des conclusions obtenues sont dûes.

DESCRIPTION DES PRINCIPAUX TYPES DE PLAGIOCLASES ZONÉS

Dans les roches dacitiques de Campo de Vera (Almeria) il existe de nombreux types de plagioclases zonés, qui se présentent non seulement dans les roches des divers gisements mais aussi dans un même exemplaire et dans une même coupe mince.

Les deux seules caractéristiques communes à tous les types sont les suivantes: 1) la grande différence de composition chimique entre les diverses zones principales et 2) l'existence d'une limite claire et coupée entre elles. A l'interieur de chaque zone il existe à son tour des bandes étroites, en géneral oscillatoires qui se distinguent les unes des autres par un très petit changement dans le pourcentage d'anorthite (en général 5 % An).

La composition quantitative des diverses zones a été determinée en tous les cas au moyen de très nombreuses mesures avec la platine universelle de Fédoroff. Cette composition est très semblable pour chaque groupe de zones, quelle que soit la roche étudiée; en général la composition pour les bandes plus basiques varie entre 80 % et 85 % d'anorthite, avec une moyenne de 83 %, quoiqu'il existe aussi quelques unes, dans des cristaux à zones alternantes, dont la basicité est plus réduite (60 % à 70 % An); les zones les plus acides, qui présentent une composition encore plus constante, contiennent des pourcentages de An compris entre le 44 % et le 53 % An, avec une moyenne générale de 48.5 % An. On trouve par conséquent entre les diverses zones d'un même cristal de très grandes différences de composition, qui atteignent sans difficulté 40 % de An.

Les types de zonage sont très nombreux: les cristaux avec structure normale non-oscillatoire sont très fréquents, des fois avec le noyau idiomorphe, d'autres avec le noyau allotriomorphe. Très souvent aussi on trouve des cristaux avec structure oscillatoire constitués par plusieurs zones de composition alternativement acide et basique, généralement disposées normalement, c'est-à-dire avec le noyau riche en An, d'autres fois avec la disposition inverse. Dans la plupart des cas le noyau de plagioclase basique est pénétré par de nombreuses inclusions de plagioclase acide de la même composition que celle de la zone externe.

Outre ces cristaux de plagioclase on observe, dans les roches des gisements du Campo de Vera, des associations cristallines de plagioclase et de sanidine que nous décrirons aussi à cause de la similitude morphologique qui existe entre elles et les cristaux de plagioclase à structure oscillatoire. Dans ces associations la sanidine prend (dans l'espace) la même position que les zones riches en An dans les cristaux oscillatoires; le reste du plagioclase a toujours une basicité assez réduite (à peu près de 50 % An), c'est-à-dire une composition analogue à celle des zones acides des cristaux oscillatoires.

À la fin de ce chapitre descriptif nous étudierons quelques associations de plagioclase et d'opale, dans lesquelles ce dernier occupe une position analogue à celle de la sanidine dans les associations mentionnées ci-dessus et à celle de plagioclase basique dans les cristaux oscillatoires.

CRISTAUX AVEC STRUCTURE NON-OSCILLATOIRE

Tous les cristaux appartenant à ce groupe ont une disposition normale; la zone plus basique (∽ 80 % An) occupe le centre du cristal et la zone de moindre basicité (∽ 50 % An) le pourtour.

Certains de ces cristaux possèdent un noyau idiomorphe dont les contours sont souvent plus parfaits que leur contour externe, mais plus généralement le noyau est allotriomorphe; dans ce dernier cas il arrive fréquemment que les plans de macle du noyau ne se continuent pas dans la partie externe et vice versa.

On trouve très souvent des cristaux de plagioclase dont le noyau basique idiomorphe ou allotriomorphe présente des nombreux »orifices» qui peuvent être ou non accompagnés de digitations périphériques, remplis par du plagioclase de la même composition chimique que la zone externe du cristal; la caractéristique plus frappante de ces inclusions est que, quoiqu'en apparence elles soient isolées entre elles et de la zone périphérique, elles ont exactement la même orientation cristallographique qui correspond à celle du reste du cristal; du point de vue optique elles se comportent de la même façon qu'une association perthitique ou myrmekitique. La fig. 1 représente un cristal de ce type; il est constitué dans sa plus grande partie par un noyau de 84 % An, recouvert par une bande périphérique de 46 % An; à l'intérieur de ce noyau idiomorphe, on trouve de très nombreuses inclusions irrégulières et digitées de plagioclase avec un 46 % An, ayant la même orientation et par conséquent maclées suivant les mêmes lois que les différents individus cristallins de l'association à laquelle elles appartiennent.

L'interpénétration entre les parties les plus acides et les plus basiques de ces cristaux est très intime; ainsi, dans la figure citée ci-dessus il existe une inclusion de plagioclase acide, située dans la partie gauche supérieure du schéma, qui sépare à son tour une petite partie du noyau basique, et celle-ci, englobe de nouveau une autre petite partie du plagioclase acide.

La fig. 2 est un exemple très intéressant; on peut voir simultanément le développement de ces digitations profondes (dans la partie gauche du cristal) et de pénétrations apparemment isolées à l'intérieur du noyau basique (dans la partie droite de l'association); dans ce cas le noyau du

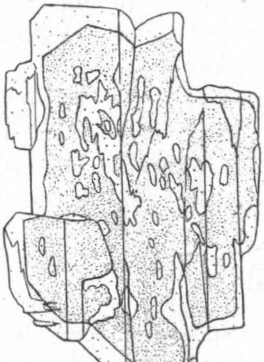

Fig. 1.

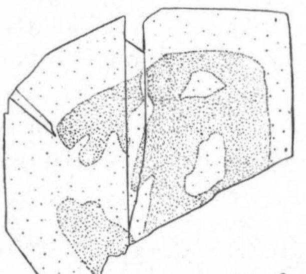

Fig. 2.

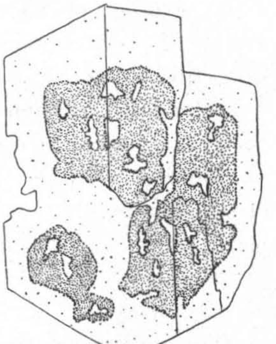

Fig. 3.

Dans toutes les dessins, le pointillé plus dense représente le plagioclase avec 80 % An; le pointillé plus clair le plagioclase avec 47 % An; les points plus gros représentent la sanidine; les parties entièrement noires sont de l'opale.

cristal est complètement idiomorphe sauf dans sa partie gauche où il semble avoir été corrodé.

Dans les fig. 3 et 4, nous voyons aussi d'autres exemples démonstratifs de ce genre de plagioclases. Dans ces deux cas extrêmes, le noyau est divisé en plusieurs morceaux apparemment isolés dans le plan d'observation; chacun de ces morceaux est perforé à son tour, mais ils conservent toujours l'orientation cristallographique et les lois de macle qui correspondent à l'ensemble du cristal.

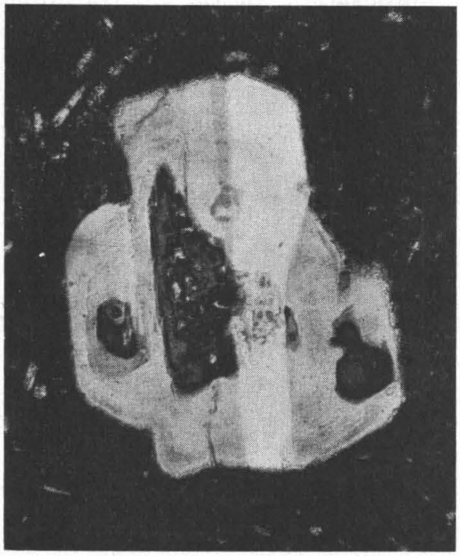

Fig. 4. Les trois parties plus foncées sont de plagioclase basique; le reste de plagioclase acide.

CRISTAUX AVEC STRUCTURE OSCILLATOIRE

Dans les dacites de Vera on rencontre aussi très fréquemment des phénocristaux de plagioclase dans lesquels les bandes riches et les bandes pauvres en An se répètent alternativement deux ou trois fois; comme dans le cas précédent, non-oscillatoire, il existe toujours une coupure très nette au début et à la fin de chaque zone. Ces structures oscillatoires sont uniformes (Homma) puisque à l'interieur de chaque zone quoiqu'il puisse exister un subzonage oscillatoire, le pourcentage général de An demeure constant; cette uniformité se maintient également dans l'ensemble des zones d'un même groupe, c'est-à-dire les zones basiques d'une part et les zones acides d'autre part gardent la même composition quelle que soit leur distance au centre du cristal.

A. AVEC LE NOYAU RICHE EN ANORTHITE

La fig. 5 constitue un exemple typique de ce groupe; le noyau très réduit et complètement allotriomorphe est entouré d'une bande de plagioclase avec 51 % An; cette bande est separée de la zone externe qui a exactement la même composition, par un anneau de plagioclase basique, semblable à celle du noyau; la limite interne de cet anneau présente une surface plus irrégulière que la limite externe.

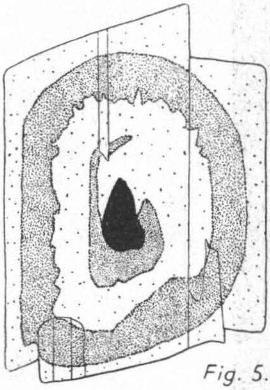

Fig. 5.

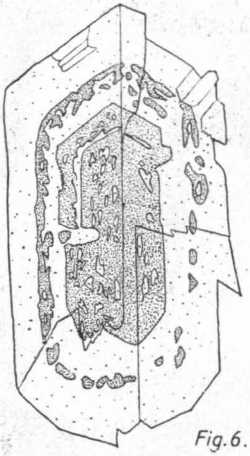

Fig. 6.

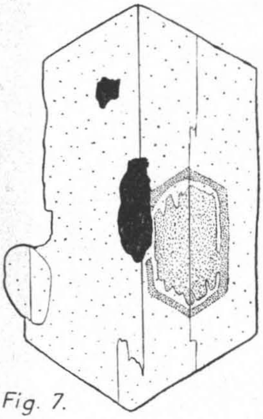

Fig. 7.

Dans les cristaux de ce type il existe aussi des digitations et des pénétrations du plagioclase acide à l'intérieur des zones basiques (fig. 6); ces digitations se développent parfois suivant des directions privilégiées, parallèles à certaines faces du cristal, et, lorsqu'elles sont très développées, comme il arrive dans l'exemple cité ci-dessus, elles arrivent presqu'à limiter une nouvelle zone qui partage en deux le noyau de plagioclase basique. Il arrive parfois que ces zones basiques sont tellement pénétrées par le plagioclase acide, qu'elles sont réduites finalement à une serie de petites plaques isolées qui laissent encore déviner la forme de la zone primitive (renvoi à la même figure).

La distinction entre un cristal à structure oscillatoire et un cristal à structure non-oscillatoire dans lequel se sont développées les digitations parallèles à ses faces cristallines, est souvent difficile. La fig. 7 est un exemple démonstratif de ce que nous venons de dire; à première vue il semble que l'on puisse distinguer les zones suivantes: 1. Un noyau basique en partie allotriomorphe, avec des digitations peu développées; 2. Une zone étroite acide avec sa limite externe idiomorphe; 3. Une zone basique idiomorphe dans ses deux limites; 4. La zone acide qui constitue la plus grande partie du cristal. Cependant à un examen attentif on peut voir que la zone acide intercalée entre les deux basiques est reliée à la zone principale du côté gauche, tandis que le noyau et la zone basique sont reliées sur le côté opposé.

B. AVEC LE NOYAU PAUVRE EN ANORTHITE

Les cristaux avec les zones oscillatoires rangées suivant la disposition normale sont les plus fréquents, mais il existe aussi des plagioclases qui ont une disposition oscillatoire inverse, c'est-à-dire, dont le noyau a un pourcentage de An plus petit que celui de la zone immédiate. Dans ces cristaux les zones de plus grande basicité n'atteignent jamais celle des cristaux étudiés auparavant, leur composition arrivant au maximum à 67 % An.

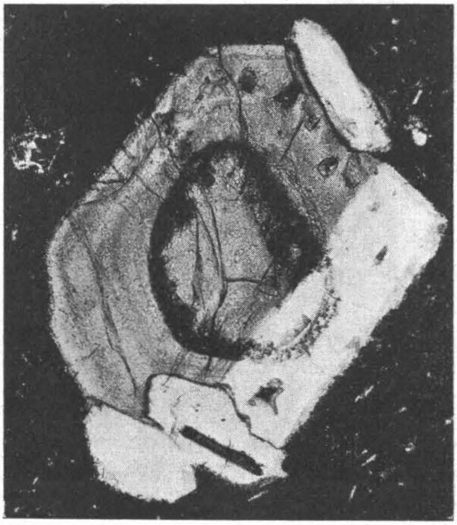

Fig. 8. L'anneau interne est du plagioclase basique; le reste est du plagioclase acide.

La fig. 8 constitue un exemple assez simple de ce genre de plagioclases; en effet il s'agit d'un cristal de plagioclase acide qui renferme dans son interieur un anneau complet de plagioclase riche en An; la forme de cet anneau, quoique irrégulière, suit à grands traits la forme générale du cristal, sa limite interne étant moins idiomorphe que l'externe.

Il existe aussi dans ce groupe, des plagioclases avec un nombre de zones plus élevé; dans la fig. 9 le cristal est constitué par cinq zones au total, deux basiques et trois acides; les zones basiques comme dans le cas précédent sont beaucoup plus régulières dans leur partie externe et souvent pénétrées par du plagioclase acide.

Ici encore on trouve tous les termes de transition qui conduisent aux cristaux dans lequels les zones riches en An sont tellement pénétrés par les digitations et les inclusions de plagioclase acide, qu'elles sont réduites à quelques plaques isolées (fig. 10); de la même façon que dans le cas des cristaux à disposition oscillatoire normale, certaines zones peuvent ainsi être presque complètement partagées en deux, si les digitations se développent parallèlement aux faces du cristal.

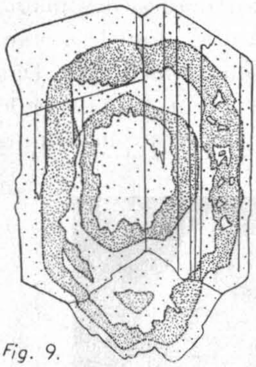

Fig. 9.

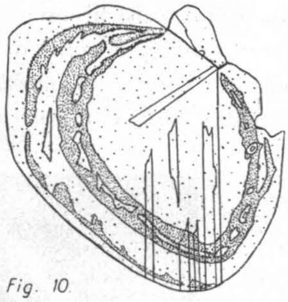

Fig. 10

ASSOCIATIONS DE PLAGIOCLASE ET DE SANIDINE

La sanidine est un mineral qui se trouve en abondance dans toutes les roches dacitiques du Campo de Vera; elle forme souvent de grands phénocristaux, qui atteignent parfois 5 cm. de longueur parsemés irrégulièrement dans la roche. Dans certains gisements, justement dans ceux qui contiennent les plagioclases à disposition oscillatoire inverse que nous venons d'étudier, on rencontre de petits phénocristaux formés par l'association de deux feldspaths, alcalin et calcosodique. Dans toute une série de ceux-ci, le feldspath alcalin, sanidine, occupe le centre du cristal, tandis que la zone externe est formée par du plagioclase; d'autres présentent un anneau subidiomorphe de sanidine qui sépare le noyau de plagioclase de la partie périphérique de composition identique; finalement il y a des cas où le plagioclase constitue le centre de l'association et la sanidine forme le pourtour.

Le feldspath calcosodique se conserve toujours parfaitement; la sanidine, au contraire, est plus ou moins altérée et remplie d'inclusions.

Dans les cristaux du premier groupe (sanidine au centre) les lois de macle du feldspath alcalin se conservent souvent dans le plagioclase (fig. 11), quoique ce dernier puisse présenter en outre d'autres macles particulières qui ne seraient pas possibles dans le feldspath monoclinique. Cependant les cristaux simples de sanidine entourés de plagioclase maclé, sont plus fréquents. Le noyau de ces associations est toujours plus ou moins irrégulier et présente des perforations et des digitations de plagioclase, tout à fait semblables à celles qui ont été décrites au sujet des cristaux de plagioclase formées par des zones acides et basiques.

Dans le deuxième cas (anneau interne de sanidine) l'anneau adopte en général la forme externe du cristal, et peut suivre, ne soit-ce qu'en partie, les directions de la macle du plagioclase; ces anneaux présentent des digitations ou forment des apophyses et des îlots indépendants au milieu des zones immédiates. Leur position et leurs caractères morphologiques sont par conséquent analogues à ceux des anneaux basiques de plagioclase dans les cristaux avec zonage oscillatoire (comparer la fig. 12 avec la fig. 10).

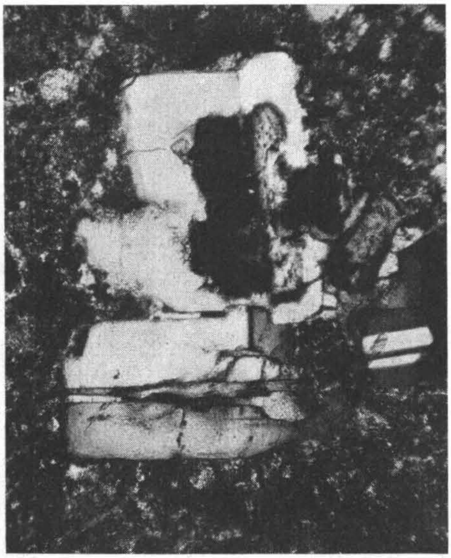

Fig. 11. Le cristal supérieur présente un noyaude sanidine, maclée suivant la loi de Karlsbad, entouré par une zone plus claire de plagioclase acide maclé suivant la même loi.

Les cristaux dans lesquels la sanidine entoure le plagioclase sont assez rares; ce dernier mineral forme un ensemble de cristaux maclés indépendamment de la sanidine, et celle-ci adopte toujours des formes tout à fait irrégulières.

Les problèmes génétiques qui se présentent au sujet de ces associations feldspathiques sont difficiles à résoudre, puisque pour certains de ces cristaux il faudrait admettre l'existence dans le magma fluide de conditions physiques et chimiques tellement différentes, qu'elles puissent permettre en certains moments la déposition d'un des feldspaths et en d'autres moments son associé.

Une circonstance qu'il ne faut pas oublier dans les discussions que ces cristaux pourraient susciter, c'est la ressemblance morphologique qui existe

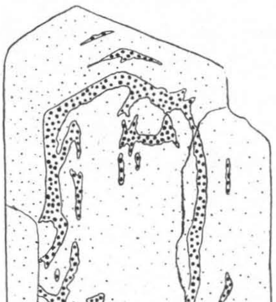

Fig. 12.

entre les alternances de plagioclase et de sanidine et celles de plagioclase acide et basique dans les cristaux avec zones oscillatoires inverses; d'autre part le plagioclase de ces associations mixtes a toujours une basicité assez réduite.

ASSOCIATIONS DE PLAGIOCLASE AVEC DE L'OPALE

Dans les roches volcaniques d'Alifragas beaucoup de phénocristaux de plagioclase sont substitués en partie par une masse transparente d'opale, associé parfois à la limonite. Les parties qu'occupe l'opale, adoptent les mêmes formes et la même disposition que les bandes de plagioclase basique dans les cristaux à structure oscillatoire; pour cette raison elles sont aussi très analogues du point de vue structurel à beaucoup d'associations mixtes de sanidine et de plagioclase décrites dans le paragraphe précédent.

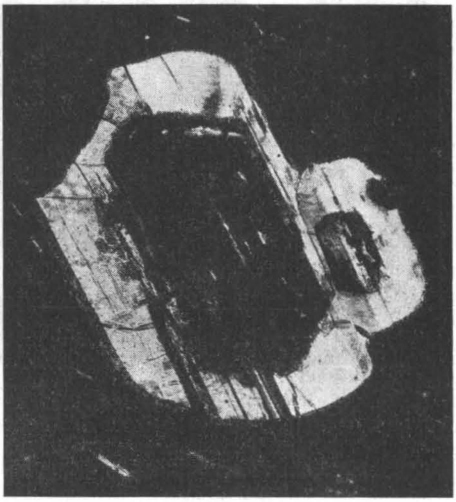

Fig. 13. Toutes les zones internes en noir sont de l'opale.

Dans certains cas l'opale occupe le centre du cristal et constitue sa plus grande partie; parfois, la masse d'opale suit, avec une étrange perfection, la forme externe du cristal (fig. 13) et à l'interieur elle laisse souvent de petits morceaux de plagioclase périphérique, complètement isolés de la masse principale et cependant orientés cristallographiquement de façon identique (fig. 14). Il existe aussi des cas où l'opale est distribué en plusieurs couches que l'on peut homologuer aux zones basiques des cristaux de plagioclase à disposition oscillatoire. Ainsi, le cristal de la fig. 15, est tout à fait semblable à celui de la fig. 5, avec la seule différence que le plagioclase basique a été substitué par de l'opale. De la même façon dans le cristal de la fig. 16 il existe une bande très digitée et des nombreuses inclusions irrégulières dans la partie centrale très semblables par ses caractères morphologiques à celles du cristal de plagioclase avec des inclusions de sanidine de la fig. 12.

Pour expliquer ces associations il faut admettre des processus de substitution dans certaines zones ou parties d'un cristal primitif de plagioclase par les gels de silice. Il est impossible d'admettre des périodes alternantes de déposition de plagioclase et d'opale, car ce serait contraire à toutes les théories et à toutes les données que nous possedons sur la cristallisation du magma. D'autre part les masses d'opale présentent des contours idiomorphes qui obligent à admettre l'existence d'un processus de pseudomorphisme métasomatique, qui a dû avoir lieu après la formation du cristal de plagioclase. La ressemblance entre les formes de l'opale, de la sanidine et des zones de plagioclase basique permet de supposer, tout au moins pour le cas qui nous occupe, que la métasomatose est sélective à travers des zones basiques, à présent disparues.

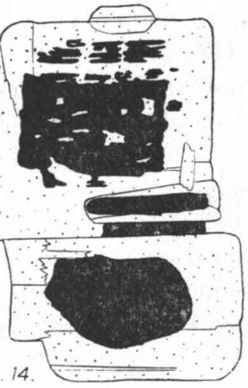
Fig. 14.

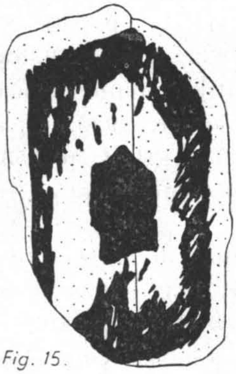
Fig. 15.

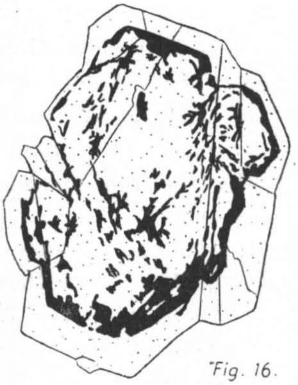
Fig. 16.

SUR LE MÉCANISME DE DÉVELOPPEMENT DES PLAGIOCLASES

Afin d'expliquer les structures zonées des plagioclases dans les roches magmatiques, les auteurs qui les ont étudié, ont élaboré diverses théories ou mécanismes. Toutes ces théories peuvent se classer du point de vue physico-chimique dans les trois groupes suivants:

A. Causes inhérentes au procès physico-chimique de la cristallisation dans les magmas.
   1. Sursaturation (Harker, 1909, p. 233; Iddings, 1909, p. 177):
   2. Déséquilibre entre la vitesse de diffusion des ions Ca et la vitesse de cristallisation. (Harloff, 1927 p. 103—104.)

   3. Variations de la concentration de l'ion Ca dans la phase fluide comme conséquence de la formation d'autres mineraux. (Guimaraes, 1946; p. 18—21; Wahrharftig en partie 1951).
- B. Modification des conditions physiques du système obéissant à des causes externes.
   1. Circulation du cristal, ou du magma, entre les zones plus chaudes et plus froides du réservoir magmatique, par des courants de convection (Bowen, 1913, p. 597; 1915 a, p. 161; 1915 b, p. 38; Fenner, 1926, p. 700—703; Homma 1932, p. 350; 1936 a, p. 141; 1936 b, p. 39—40).
   2. Perte d'éléments volatils comme conséquence de l'éruption des magmas sur-jacents (Fenner, 1926, p. 700—103; Homma, 1936 b, p. 37, 32; Phemister, 1932—1934, p. 353).
- C. Modification de la composition chimique du magma par des causes externes.
   1. Mélange de deux magmas de différente composition (Fenner, 1926, p. 700—703; Phemister, 1932—1934, p. 554; Homma, 1936 a, p. 141; 1936 b, p. 36—37; Kuno, 1936, p. 107; Alling, 1936, p. 146).
   2. Assimilation de roches solides (Kuno, 1936, p. 107; Guimaraes 1946, p. 19).

Tous ces auteurs ont appuyé leurs théories, au moins implicitement, sur les données suivantes qu'ils considèrent définitivement prouvées:

1. Chaque zone représente une période dans laquelle les conditions physico-chimiques du magma sont différentes de celles qui conditionnent la formation des zones immédiates.

2. Les diverses zones ou couches d'un cristal proviennent de l'accroissement simple de celui-ci suivant un procès de cristallisation que nous pourrions appeler normal. On en déduit qu'une zone est d'autant plus ancienne qu'elle se trouve plus proche du centre du cristal et que la succession des zones représente une succession dans le temps.

3. Lorsque deux zones sont séparées par une limite nette régulière, cela signifie que le passage des conditions qui determinèrent la formation de la première, à celles qui ont conditionné la déposition de la deuxième a dû être très brusque. Si la surface qui les sépare est irrégulière, la zone interne a été en partie dissoute ou »digérée» par le magma, suivant le procès que l'on appelle »corrosion magmatique».

En admettant en termes absolus les postulats cités ci-dessus les divers investigateurs ont été obligés, en étudiant ce problème, d'imaginer des mécanismes extrêmement compliqués, et toujours plus ou moins problématiques, pour expliquer les différentes successions de zones que l'on peut trouver dans les roches volcaniques.

D'autre part, un grand nombre des données d'observation dans les plagioclases du Campo de Vera et aussi dans beaucoup d'autres de diverses régions, sont inexplicables si l'on admet auparavant tous et chacun des postulats précédents.

Dans les plagioclases décrits plus haut on trouve souvent, on pourrait dire même presque constamment, des pénétrations de plagioclase acide sous forme d'îlots indépendants à l'intérieur des zones plus internes de plagioclase basique; quoique dans certains cas on puisse trouver une relation directe entre les »îlots» de plagioclase acide et la bande de même composition, il est aussi évident que dans beaucoup d'autres cas on n'observe pas de communication entre eux; il serait possible d'imaginer que cette communication existe dans un plan différent de celui d'observation, mais, même dans ce cas il semble difficile d'admettre le réseau de canalicules qui relieraient les inclusions entre elles et celles-ci avec l'extérieur. En tous cas si nous admettions que les inclusions se sont produites par une pénétration de magma fluide, à l'interieur du plagioclase basique à travers d'un réseau de petits canaux qui ont respecté sa partie externe et transformé l'interne en une éponge, il resterait encore quelques problèmes à résoudre; en effet comment se peut-il, s'il y a eu lieu un processus de corrosion aussi profond par la phase fluide, qu'aucun petit fragment de plagioclase basique ne ce soit separé? Et si un de ces petits fragments était resté isolé et submergé dans une phase fluide il devrait être désorienté par rapport à la zone à laquelle il appartenait; cependant nous n'avons pas trouvé, en aucun cas, des inclusions de plagioclase basique qui soient quelque soit peu désorientées par rapport au reste du cristal.

La disposition morphologique des zones dans les plagioclases de Vera est encore une donnée difficile d'accorder avec les postulats classiques. Il y a des cristaux dans lequels on peut déterminer exactement le nombre des zones visibles dans la section car elles sont bien délimitées; mais, dans de nombreux exemples il est plus difficile de le dire: ainsi, dans le plagioclase de la fig. 2 on pourrait penser à l'existence de trois bandes différentes (acide-basique-acide), quoique, à vrai dire, il n'existe que deux zones, une basique pénétrée jusqu'au centre par une zone acide qui l'entoure en même temps. Le plagioclase de la fig 6 paraîtrait formé en haut et à gauche par six zones, causées suivant les principes cités, au cours d'un nombre équivalent de périodes de croissance, tandis que dans l'angle opposé il y aurait uniquement quatre zones suivant certaines directions et deux dans d'autres.

On pourrait penser que les processus de corrosion magmatique ont été plus poussés dans ces dernières directions; néanmoins les caractères morphologiques nous font penser à une autre solution: de la même façon que dans l'angle supérieur à gauche des digitations de plagioclase acide ont pénétré le noyau en le séparant en deux zones formées en apparence au cours de deux périodes différentes, on pourrait admettre que la zone basique fragmentée qui est complètement séparée du noyau et située plus à l'extérieur a été formée par un développement plus poussé d'un processus de pénétration.

La fig. 7 représenterait la phase intermédiaire de ce processus; le noyau idiomorphe est pénétré à gauche par une invagination du plagioclase acide qui arrive presque à séparer une nouvelle zone du noyau primitif; si le processus avait été complèt il y aurait dans le cristal quatre zones visibles qui se rapporteraient non pas à quatre périodes différentes et consécutives de l'histoire génétique du cristal, mais simplement à deux. Ces raisonnements peuvent s'appliquer à de nombreux cristaux de plagioclase parmi ceux qui ont été étudiés dans les dacites de Vera (voir aussi la fig. 10).

Nous arriverons ainsi à la conclusion que dans certains plagioclases il n'y a pas eu autant de périodes de formation que de zones observables, mais un plus petit nombre, peut-être seulement deux, l'un au commencement de la cristallisation du plagioclase, avec pourcentage de An élevé, et un autre subséquent pendant lequel s'est formé le plagioclase acide qui en partie a fragmenté le noyau en une série de zones indépendantes en apparence.

En partant de cette idée on peut mieux expliquer certaines structures des plagioclases oscillatoires qui jusqu'à présent n'avaient pas reçu une explication satisfaisainte; par exemple, la simple alternance de zones acides et basiques d'un nombre parfois élevé; pour expliquer ces alternances il fallait recourir à des changements répétés dans les conditions du magma, provoqués à leur tour par une série de facteurs qui dévraient se répéter dans le temps; l'existence de ces rythmes, s'il s'agit de zones de composition très différente est très improbable. Un autre caractère de solution encore plus difficile qui se reproduit souvent dans les structures oscillatoires c'est l'identité complète ou presque complète de la composition des zones acides, d'une part, et des zones bàsiques, d'autre part, caractère qui est presque constant dans les plagioclases de Vera. Quelle probabilité mathèmatique existe-t-il pour que des modifications anormales, toujours exactement les mêmes puissent se répéter au cours de l'évolution d'un procès sous l'action de causes multiples et diverses?

Admettons que les plagioclases oscillatoires, comme conséquence de la formation du plagioclase acide suivant des directions parallèles aux faces cristallographiques, peuvent se former en un nombre de périodes plus réduit que le nombre de zones présentes, nous trouverons encore de nouveaux problèmes à résoudre; l'un d'eux qui paraît fondamental c'est le mécanisme de pénétration; un autre qui lui est subordonné c'est l'explication de l'orientation cristallographique parfaite de toutes les parties du noyau primitif qui restent isolées entre elles par la pénétration du plagioclase acide.

Pour expliquer la pénétration du plagioclase acide à l'intérieur des zones basiques on retrouve les mêmes difficultés que pour l'explication de l'existence des »îlots» si fréquents dans ces dernières; bien que dans certains cas on puisse observer une relation directe entre les différentes zones acides d'une part et les différentes zones basiques d'autre part, il existe beaucoup d'autres cas dans lesquels l'indépendance des divers éléments semble complète.

D'autre part si la pénétration était dûe à simple processus de »corrosion» (dissolution, fusion) ou un autre processus quelconque qui fasse disparaître la phase solide pour faire place à la phase fluide, les parties isolées devraient rester dans des positions arbitraires et en aucune façon parfaitement orientées comme il arrive dans tous les cas.

Cette orientation peut uniquement se conserver en admettant la persistance de la phase solide au cours du processus; pour ceci il faut admettre l'existence d'une réaction entre la phase solide du magma et la phase encore fluide sans que la première arrive à disparaître; dans le cas des plagioclases cela suppose une échange ionique des paires Na-Si et Ca-Al; ces derniers ions se détacheraient du réseau structurel du plagioclase riche en anorthite et émigreraient à travers les espaces vides et les imperfections du réseau vers la phase fluide du magma; les ions Na et Si passeraient, à partir de celle-ci et dans le sens contraire, pour rétablir une composition plagioclasique plus riche en albite. L'émigration ionique serait dûe au gradient de potentiel chimique créé en conséquence du changement des conditions du magma, changement qu'il faut admettre forcément quel que soit le processus imaginé pour expliquer les structures zonées.

Ce processus d'échange ionique ou de réaction »solidus-liquidus» dans les roches volcaniques en cours de cristallisation s'appuie sur une série d'arguments:

1. La diffusion dans un solide s'active par la plus grande proportion d'espaces vides, d'ions en position intersticielle ou de quelque autre genre d'imperfection dans le réseau structurel. Toutes ces imperfections augmentent à mesure qu'augmente la température et sont tellement effectives à proximité du point de fusion que dans telles circonstances la pénétration ou l'émigration de certains ions peut s'effectuer beaucoup plus rapidement qu'à basse température, où elle est pratiquement inefficace.

D'après quelques auteurs il y aurait un seuil de température à partir duquel la diffusion serait efficace; ce point pour les silicates serait compris entre 0.8 et 0.9 T, T étant la température de fusion absolue. La transformation d'un plagioclase basique en un plagioclase acide se réaliserait dans presque tous les cas à températures tres proches de leur point de fusion, auxquelles ces processus peuvent avoir lieu.

2. D'après Buerger (1934), à l'intérieur des cristaux il peut exister des zones causées par l'accroissement rapide du cristal, dans lesquelles la facilité d'émigration ionique est plus grande que dans le reste du cristal (»lineage»). D'autre part le cristal d'anorthite qui se substitue en partie, n'est pas complètement homogène, car en général il peut avoir un léger subzonage avec de petites différences de composition.

Ces hétérogénéités seraient en même temps des facteurs qui pourraient expliquer la pénétration en quelque sorte sélective du plagioclase acide à l'intérieur de la partie basique.

3. Dans les roches volcaniques il existe de très nombreux cas dans lesquels les cristaux déjà formés sont en déséquilibre avec la phase fluide à mesure que la cristallisation augmente; dans beaucoup de cas la forme de ces cristaux se conserve et ceux-ci sont remplacés par un agrégat pseudomorphique de minéraux qui sont en équilibre avec les nouvelles conditions. De la même façon se conserve parfois la forme externe des xénolites introduits dans le magma et qui sont complètement remplacés par des minéraux nouveaux. Dans les andesites de Mar Menor, l'auteur a étudié des cristaux de cordierite, quelques fois idiomorphes provenant de roches métamorphiques, qui montrent toutes les phases d'une transformation en un agrégat de plagioclase et de pyroxène orthorhombique, sans que l'ensemble de ces minéraux change la forme primitive de la cordierite (Fuster e Ibarrola, 1952). Aucun de ces phénomènes ne peut être expliqué si l'on suppose que la phase solide qui est en déséquilibre disparaît complètement. Il faut supposer par conséquent que leurs processus de transformation doivent se développer par un mécanisme de diffusion ionique »solidus-liquidus» analogue à celui qui a été décrit pour les plagioclases.

4. Il existe une très grande analogie entre les cristaux de plagioclase zoné, les associations de plagioclase et sanidine et celles de plagioclase et opale, à un tel point que l'on peut établir un parallélisme surprenant entre les cristaux de ces trois catégories. La disposition de la sanidine et de l'opale dans le cristal paraît indiquer qu'ils occupent totalement la place d'une zone primitive de plagioclase basique. Dans le cas de la sanidine il est très difficile de supposer qu'il y ait eu une déposition alternante de plagioclase acide et de sanidine ou inversement, autant pour des raisons génétiques que par la multitude de données morphologiques qui indiquent le contraire. Ainsi dans la fig. 12 on voit une zone de sanidine dans un cristal de plagioclase, à l'extérieur de laquelle s'ébauchent les restes d'une ou deux autres zones réduites à quelques reliques orientées; si l'on supposait qu'elles s'étaient formées successivement par cristallisation à partir du magma il faudrait admettre dans l'histoire génétique du cristal les périodes suivantes: 1. Cristallisation de plagioclase. — 2. Corrosion magmatique. — 3. Cristallisation de sanidine. — 4. Corrosion magmatique. — 5. Cristallisation de plagioclase. — 6. Corrosion magmatique. — 7. Cristallisation de sanidine. — 8. Corrosion magmatique jusqu'à la disparition presque totale de la couche précédente. — 9. Cristallisation de plagioclase. — 10. Corrosion magmatique. — 11. Cristallisation de sanidine. — 12. Corrosion magmatique jusqu'à la disparition presque totale de la couche précédente. — 13. Cristallisation de plagioclase.

Les complications antagoniques qu'il faudrait introduire dans un magma pour expliquer cette succession n'ont besoin d'aucun commentaire. Il est beaucoup plus simple de supposer que dans un cristal de plagioclase préalablement formé il s'est produit une métasomatose sélective dans certaines bandes par l'introduction d'ions Si et K et l'émigration d'ions Al, Ca et Na pour produire le changement du feldspath calcosodique en feldspath potassique.

Les associations de plagioclase et d'opale sont encore plus démonstratives; dans ce cas il est impossible même de supposer qu'il y ait pû avoir des périodes successives de déposition de plagioclase et d'opale, puisque les conditions physico-chimiques nécessaires pour la formation de ces deux minéraux sont totalement différentes. D'autre part, l'opale, minéral sans forme cristalline propre adopte dans beaucoup de cas la forme des cristaux de plagioclase (fig. 13 et 15). Il faut admettre dans ce cas analogue morphologiquement aux précédents, l'existence d'un procès de métasomatose postérieur à la formation du cristal de plagioclase; l'opale, a pénétré à travers tout le cristal et a remplacé certaines bandes de plagioclase en respectant celles qui n'ont pas été remplacées. Naturellement la pénétration de la silice pour former de l'opale a été postérieure aux remplacements de sanidine et de plagioclase acide.

De tout ce qui vient d'être dit, l'on déduit que deux genres de phénomènes ont agi dans la cristallisation des plagioclases des dacites du Campo de Vera et probablement de beaucoup d'autres types de roches andésitiques; le premier de déposition de plagioclase exceptionnellement riche en An; l'autre, postérieur à celui-ci, est une métasomatose, réduite à la zone de cristallisation, par laquelle, et aux dépens du plagioclase primitivement cristallisé, il se forme un autre plagioclase avec une contenance en An beaucoup plus réduite; cette dernière période se continue, après avoir atteint un certain équilibre, par la cristallisation simple du plagioclase acide, qui se poursuit jusqu'au moment de l'éruption. Le procès de métasomatose réduite à la zone de cristallisation, n'a pas affecté de la même manière tout l'ensemble du cristal, ce qui est dû probablement à des differences chimiques ou structurelles préexistentes dans le cristal primitif de plagioclase basique; ce processus n'a pas eu partout les mêmes caractères, puisque dans certains gisements le feldspath produit à été de la sanidine au lieu du plagioclase pauvre en An. Finalement dans presque toutes les roches, mais plus particulièrement dans celles d'Alifragas, se sont produits des phenomènes ultimes de métasomatose par lesquels l'opale s'est déposé suivant un processus de conséquences structurelles identiques aux précédentes.

Il faut admettre pour expliquer ces deux périodes si différents quant à leur mécanisme et à leurs effets, qu'il y a eu, au cours de l'histoire génétique de ces roches, un changement brusque et important dans les conditions de consolidation.

Nous ne croyons pas que ce changement puisse être dû à des modifications dans la composition chimique de la phase encore fluide du magma par la cristallisation des propres minéraux des dacites, car la différence de composition entre les zones des plagioclases (en moyenne 40 % An) est trop élevée. D'autre part nous avons observé dans ces roches une série de caractères structurels et chimiques (présence de xénolites de quartz, transformation de la biotite en flogopite, formation de grands cristaux de sanidine) qui nous permettent d'affirmer l'existence d'une contamination d'un magma primitif rélativement basique par des roches solides plus acides et alcalines qui ont été presque entièrement dissoutes. Comme conséquence de cette contamination la composition chimique du magma a varié suffisamment comme pour expliquer les différences entre les premiers minéraux cristallisés et les minéraux dernièrement formés.

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7

PARAGENESIS OF BYTOWNITE, CHLORITE, AND MANGANOAN GARNET FROM KANGASALA, FINLAND [1]

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ABSTRACT

The chemical and mineral composition of a bytownite-chlorite-garnet schist is described, and the origin of the rock is discussed. The garnet is an almandite-spessartite. The composition of the chlorite approaches the composition of ripidolite, but some difference is apparent in the optical properties. The conditions of formation of the mineral assemblage are discussed and it is suggested that they are those of the amphibolite facies.

CONTENTS

ROCK

In 1946, during field work in the Tampere schist area, the author found some light-coloured intercalations in mica schist S. W. of Lake Rukojärvi, northwestern Kangasala. These layers were 1.5 cm—2 cm wide and not as fine-grained as the associated mica schist. It soon became evident that the layers were rich in CaO and very poor in K_2O. Some relevant data were given in an earlier paper (Seitsaari, 1951, p. 28).

The chemical and mineral composition of the rock is given in Table 1. The mineral composition could be calculated from the composition of the chlorite (p. 79) and the composition of the garnet (p. 77). Absolutely exact

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[1] Received November 30, 1953.

Table 1. Chemical composition of bytownite-chlorite-garnet schist. Southwest of Lake Rukojärvi, Kangasala. Analyst, P. Ojanperä

Constituent	%	Mol. prop.	Niggli numbers		Mineral composition		
					Constituent	Measured with the integration stage	Calculated from the analysis
SiO_2	61.96	10 316	*si*	219	Plagioclase ..	78.3 %	(An_{79}) 40 %
TiO_2	0.83	104	*ti*	2.2	Quartz		37
Al_2O_3	17.46	1 713	*al*	36.4	Chlorite	14.3	11
Fe_2O_3	0.93	58	*fm*	30.8	Garnet	1.7	4
FeO	5.12	713	*c*	28.6	Biotite	2.3	4
MnO	0.31	44	*alk*	4.2	Apatite	—	0.5
MgO	2.34	580	*k*	0.23	Ilmenite	3.4	1.5
CaO	7.57	1 350	*mg*	0.40	Magnetite ...		1
Na_2O	0.94	152	*c/fm*	0.93	Pyrrhotite ..		1
K_2O	0.42	45					
P_2O_5	0.26	18					
CO_2	0.00	—					
H_2O + ...	1.89	—					
H_2O —	0.05	—					
S	0.39	122					
	100.47						
less O for S ..	0.20						
Total	100.27					100.0	100.0

figures for the mineral composition could not be computed, because the composition of the biotite and the K_2O content of the plagioclase were not known and the sulphide mineral was not determined. The Fe content is high enough to form pyrrhotite. An excess of CaO remained after the calculation and will appear as a small difference between the composition of the normative and the modal plagioclase, the former being $Ab_{19.1}$ $An_{80.9}$ and the latter Ab_{21} An_{79}. The refractive indices of the plagioclase were determined by the immersion method with the result: $\alpha = 1.563$; $\gamma = 1.574$. These figures indicate an approximate range of composition from An_{70} to An_{75}. Because the plagioclase and the quartz form an intimate intergrowth of small crystals, the presence of quartz probably caused too low figures for the refractive indices. The composition of the plagioclase could not be determined by any other optical method.

The chlorite flakes are well-defined and 0.1 mm—0.5 mm long. They have a random orientation and are rarely parallel to the schistosity. Some chlorite has crystallized close to the garnet crystals, but the chlorite does not give the impression of an alteration product, because the crystal shapes of both minerals are rather well-developed. The garnet will be discussed in more detail on pp. 77—8.

The associated mica schist consists of biotite, quartz, and chlorite. The chlorite is similar to the chlorite in the light intercalations, but the crystals are smaller and often very well-developed with rectangular shapes.

The chlorite flakes have crystallized at random, whilst the mica crystals have a distinct parallel orientation.

A mica gneiss in the same exposure contains much more biotite, and less plagioclase and garnet, than does the bytownite-chlorite-garnet schist. The plagioclase ($\alpha = 1.550—1.551$; $\gamma = 1.558$) is an andesine (An_{40-42}). Transitional varieties between the bytownite-chlorite-garnet schist and the mica gneiss are also met with. Chlorite is of the same type in all the rocks, but the transitional varieties are mostly poor in chlorite.

Southeastwards, the gneiss gradually becomes more coarse-grained, and about 50 m to the southeast lies a quartzite conglomerate (Seitsaari, 1951, p. 42). Very probably the conglomerate represents the basal part of the afore-described mica schist—mica gneiss sequence.

The sedimentogenic, varved schists in the southern part of the Tampere belt are so well preserved (Seitsaari, 1951; Simonen and Kouvo, 1951) that there is every reason to consider them as having remained approximately unchanged in chemical composition. From this it may be inferred that the original sediment was composed of some argillaceous substance, some form of silica, and carbonate, with small quantities of hydroxides and sulphides of heavy metals. The sediment might be called »siliceous marl» and it is actually not very different from certain shales (see Pettijohn, 1949, p. 274). The poverty in K_2O is conspicuous. The argillaceous substance consisted, perhaps, of kaolinite.

The MnO content is distinctly higher than it generally is in schists. In the process of chemical weathering manganese is enriched in the oxidates, but may enter also into carbonate sediments (Rankama and Sahama, 1950, pp. 647—8; see also Otto, 1936, p. 134). The metamorphic changes of the higher oxides and hydroxides of manganese mostly produce silicate minerals and $MnCO_3$ is also converted into silicates. When the Al_2O_3 content is high enough to form garnet, metamorphic differentiation according to the concretion principle is effective in concentrating within garnet crystals the total MnO present in the rock (*e. g.*, Turner, 1948, p. 138). An attempt was made to separate the garnet from the bytownite-chlorite-garnet schist for a chemical analysis. The separation was done by means of the Clerici solution in a centrifuge and of an isodynamic magnetic separator. Pure material, however, could not be obtained, because all garnet grains contain very numerous inclusions of plagioclase, quartz, and ore, and because the specimen was poor in garnet. The impure garnet fraction was analyzed by Mr. Pentti Ojanperä, M. A., and the following approximate figures may be given for the composition of the garnet: SiO_2, 38.5 %; Al_2O_3, 19.8 %; FeO, 23.8 %; MnO, 5.0 %; MgO, 3.1 %; CaO, 5.5 %; total 95.7 %. Most of the remainder, 4.3 % (TiO_2, Na_2O, S, *etc.*), may not be present in the garnet. Moreover, the figure for CaO is certainly too high, because the impure garnet fraction contained bytownite. The refractive index of the mineral is

$$N_D = 1.809 \pm 0.003 \text{ (immersion method).}$$

It may be inferred with certainty that the garnet is rich in almandite and contains a relatively large quantity of the spessartite component. Almost the entire MnO content of the rock may actually be present in the garnet.

CHLORITE

The chemical composition of the chlorite from the bytownite-chlorite-garnet schist is given in Table 2, Anal. 1 (quoted from Seitsaari, 1951, p. 29). The mineral, however, had remained somewhat impure, because of lack of centrifuge. Therefore, similar chlorite from an associated mica gneiss (see p. 77) was now separated by means of the Clerici solution in a centrifuge and of an isodynamic magnetic separator, and analyzed. The chemical composition of this chlorite is given in Table 2, Anal. 2. The amount of impurities (quartz and a few tiny flakes of biotite) was estimated at 1.5 %.

The two chlorites appear to be almost identical, provided that the impurities in Chlorite 1 were mainly plagioclase, quartz, and ilmenite. These foreign components were actually highly probable. The atomic proportions of the metals were calculated from Analysis 2, and they agree very well with the general chlorite formula

$$(\mathrm{Mg,Fe^{2+},\ Al,Fe^{3+}})_6\ (\mathrm{OH})_8(\mathrm{Si,Al})_4\mathrm{O}_{10}$$

given by Winchell (1951).

The optical properties of the two chlorites (Table 3) also resemble each other closely. A dispersion of birefringence ($v > r$) of the same order of magnitude is observed in each chlorite, as is a multiple twinning on 001. Determinations of the optical properties were made on several different crystals (see Table 3), and the axial angle, in particular, shows a remarkable variation. The extinction angle varies to a smaller extent, whilst the refractive indices may be highly constant, the range of variation being about ± 0.002. It may be inferred that the chlorite is nonhomogeneous.

The composition and most properties of the chlorite are almost identical with those of ripidolite (see Winchell, 1951). Notable differences, however, appear between the figures for 2V and $\gamma \wedge \perp 001$, respectively. Among the chlorites with an Al_2O_3 content comparable with the Al_2O_3 content now indicated, corundophilite would have an axial angle of the same order of magnitude as has, on an average, the chlorite under discussion, *viz.*, about 30° (*op. cit.*). In the last edition of the aforementioned handbook (Winchell, 1951), corundophilite, like all chlorites, is reported to have parallel extinction, whereas in the previous edition (1933) figures up to 10°

Table 2. Chemical composition of chlorite

Constituent	1	2						
	%	%	Mol. prop.	Atomic prop. (O = 18)				
SiO_2	27.95	26.30	4 379	Si	2.73			4.00
TiO_2	0.68	0.33	41					
Al_2O_3	22.58	23.51	2 306	Al	2.87	1.27 1.60	1.89	6.00
Fe_2O_3	2.87	3.35	210	Ti	0.03			
FeO	19.34	20.20	2 812	Fe^{3+}	0.26			
MnO	0.25	0.23	32	Fe^{2+}	1.75		4.11	
MgO	14.17	14.93	3 703	Mg	2.30			
CaO	1.50	0.17	30	Mn	0.02			
Na_2O	0.45	0.00	—	Ca	0.02			
K_2O	0.15	0.12	13	K	0.02			
$H_2O +$	11.02	10.72	5 949	OH	7.40			7.40
$H_2O -$	0.15	0.00						
Total	101.11	99.86						
Sp. gr.	2.968 ± 0.005 (Clerici solution and Westphal balance)	2.948 (pycnometer)						

1. Chlorite from the bytownite-chlorite-garnet schist in Table 1 (Seitsaari, 1951, p. 29).
2. Chlorite from the mica gneiss. Analyst, A. Juurinen.

Table 3. Optical properties of the chlorites (Column numbers 1, 2 refer to Anal. 1, 2, Table 2)

	1		2	
$\alpha \approx \beta$	1.624		1.626	
γ	—		1.630	
$2V\gamma$	51°—52° 49°		47°—58° 32°—33° 30° 22° 20° 18° 14°	(three determinations)
$\gamma \wedge \perp 001$	12.5°—14° 8° 3° —3.5°	(five determinations)	11°—12.5° 9° 9° 5°	(four determinations)

Pleochroism:
$\alpha \approx \beta$ pale greyish green to pale green
γ almost colourless

The refractive indices were determined by the immersion method (those of Chlorite 2 on the U-stage), $2V\gamma$ and $\gamma \wedge \perp 001$ on the U-stage. Measurements were made on several different crystals, and the results of all different measurements are reported in the table.

were given for $\gamma \wedge \_\ 001$. It is very probable that chlorites with an inclined extinction really exist. But corundophilite is definitely richer in MgO than the chlorite now considered. On the other hand, ripidolite and related chlorites, such as prochlorite, are usually reported to occur in hydrothermal and similar veins. Now, the environment and the conditions of crystallization appear to be quite different. A modification of ripidolite, different in certain properties from the common ripidolite, seems to be caused under these conditions[1].

GENESIS OF THE MINERAL ASSEMBLAGE

Discussion of the physical conditions of formation of a mineral paragenesis may be based on any mineral present in the rock, if the conditions of formation of this mineral are known. In this particular case, three minerals might be taken into consideration, *viz.*, the bytownite, the chlorite, and the manganoan almandite. Almandite-spessartites have often been referred to the lower grades of metamorphism (see Harker, 1932, pp. 217—8; Eskola, 1939, p. 399), but this garnet is evidently formed also under the conditions of the amphibolite facies (see Turner, 1948, p. 79). Chlorite is usually considered to be one of the most typical low-grade minerals, but this conception was recently proved to need revision (Yoder, 1952). There remains only the bytownite. As far as is known to the author, plagioclase with a notable anorthite content is never formed at temperatures below those of the amphibolite facies (in the sense of Turner, 1948), provided that a pressure corresponding to this facies also exists (see also Barth, 1952, pp. 286, 338; Rosenqvist, 1952, p. 58). Thus the albite-epidote amphibolite facies and the greenschist facies would be excluded, and the consideration of the mineral assemblage may be based on this assumption. The mineral composition of the associated rocks cannot elucidate the facies problem, because a similar chlorite is present in all the rocks and plagioclase, if present, varies from andesine to bytownite in composition.

The chemical composition of the rock and of the chlorite are plotted on an ACF diagram, and several minerals occurring in the amphibolite

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[1] After this paper was sent to print, Dr. Max H. Hey reported (personal communication) that an inclined extinction and a large axial angle in chlorite would indicate that the chlorite is particularly well crystallized. In Hey's opinion, there is very strong evidence that the chlorites commonly show more or less irregularity in the stacking of the layers of their crystal structure. He concludes: With perfect stacking, the chlorites are monoclinic with a fairly large extinction angle $\gamma \wedge \perp 001$ and often a large axial angle, but the less perfectly crystallized material with considerable stacking irregularities shows a continuous approach to apparent rhombohedral symmetry. Chlorites of some compositions such as corundophilite seem to tend to more perfect stacking than the common run of chlorites and so to distinct inclined extinction. This explanation is supported, *e. g.*, by the complete lack of any correlation between axial angle or extinction and composition. (Hey's investigation on the chlorites should appear in the March number, 1954, of the Mineralogical Magazine.)

facies, particularly those existing in rocks with excess SiO_2 and deficient in K_2O, are added (Fig. 1). The chlorite (Table 2, Anal. 2) is actually not from the bytownite-chlorite-garnet schist, but this analysis, made of pure material, was preferred, because the two analyses are almost identical. This chlorite probably contains a little more Al_2O_3 than does the chlorite from the bytownite-chlorite-garnet schist.

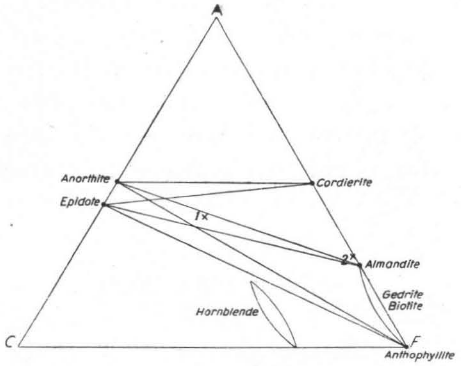

Fig. 1. Composition of the rock in Table 1 (1) and of the chlorite in Table 2, Anal. 2 (2) plotted on an ACF diagram. Minerals occurring in the amphibolite facies, particularly those existing in rocks with excess SiO_2 and deficient in K_2O, are added.

Judging from the manner of crystallization of the chlorite flakes (p. 76), no notable shearing stress can have been acting during the crystallization of this mineral. Cordierite, a typical antistress mineral, would be expected instead of chlorite in a rock deficient in K_2O (Eskola, 1939, p. 352; Turner, 1948, pp. 77—9). The rock is remarkably rich in H_2O, but, according to Yoder (1952, pp. 601, 608), cordierite would be stable in the presence of an excess of water vapour at temperatures down to 400°C, and changes in pressure would not essentially affect the minimum temperature limit. (When Yoder's results are quoted, one must bear in mind that his experiments concerned a system without FeO. In the presence of FeO the equilibria may change somewhat.) The FeO/MgO ratio in the chlorite is about 3/4, corresponding to the upper limit for the FeO/MgO ratio in most cordierite (see Winchell, 1951, p. 471). The excessive FeO has been used to form almandite and the formation of garnet was promoted by the presence of MnO. However, (Mg,Fe)O, as compared with the Al_2O_3 content available, is present in excess over the cordierite ratio, and therefore anthophyllite should have been formed. Now Yoder (1952, pp. 609), from his experiments, came to the conclusion that, if an excess of water is present, anthophyllite would be unstable at all temperatures. The formation of gedrite is, perhaps, controlled in the same way. Insufficient data were available to determine the behaviour of cummingtonite (*l. c.*), but here the FeO/MgO ratio may be of more importance. The Mg-silicate enstatite would become stable in the presence of an excess of water at higher temperatures (Yoder, 1952, pp. 601, 605), but the temperature was evidently not high enough for this mineral. The consideration leads to the surprising conclusion that the pair cordierite-anthophyllite may be totally replaced by a chlorite, provided water is present in fair excess and the bulk composition is otherwise suitable. Although an excess of water is possibly present at almost all temperatures in the amphibolite facies, a mineral rich in water like chlorite has not been regarded as a stable phase in this facies, even though an extreme poverty in K_2O may raise the upper stability limit of chlorite. Yoder (1952, pp. 601, 607), however, found the mineral clinochlore to be stable in the presence of an excess of water up to 680° at about 1 000 atm. More aluminous chlorites, such as the present chlorite mineral, were not produced in Yoder's experiments.

Considering the high content of CaO, Al_2O_3, and H_2O, the complete absence of epidote is somewhat surprising (see Fig. 1). The paragenesis epidote-plagioclase is fairly common (Vogt, 1927; Bugge, 1943; Rosenqvist, 1943; Ramberg, 1944, 1945; Turner, 1948; Seitsaari, 1951), and epidote can be stable together with very basic plagioclase (Ramberg, 1945, p. 51). The following factors may be stated as controlling the formation of epidote: the CaO/Al_2O_3 ratio, intensity of shearing stress, temperature and pressure, and the Fe_2O_3 content.

Under given PT-conditions only the CaO/Al_2O_3 ratio would affect the epidote-plagioclase equilibrium in rocks with excess SiO_2 and H_2O (Ramberg, 1945, p. 52). In systems rich in CaO the equilibrium is displaced towards epidote and acid plagioclase, and in systems rich in Al_2O_3 towards basic plagioclase. Now, the amounts of (Fe,Mg)O and alkalies available to form aluminian minerals are limited, so that most Al_2O_3 must be linked to CaO alone and the CaO/Al_2O_3 ratio will definitely favour the formation of anorthite and not that of epidote. This is, perhaps, the principal reason for the absence of epidote. Some epidote, however, would be expected under favourable physical conditions.

Theoretically, the influence of shearing stress should be insignificant, but it is actually notable, the reason being that the non-elastic behaviour of minerals plays an important rôle in metamorphic deformation (Turner and Verhoogen, 1951, pp. 395—6). Turner (1948) attributes all parageneses with epidote of the amphibolite facies to a considerable shear. Under high shearing stress no basic plagioclase would be formed and epidote would appear in place of anorthite in all rocks whose content of Na_2O is insufficient to permit the treatment of $(Ca,Na_2)O\ Al_2O_3$ as a single component of the system. The absence of epidote in the present instance is conceivable, because the shearing stress was supposed to be low (p. 81).

The range of the stability of epidote increases with increasing confining pressure (Ramberg, 1944, p. 123). Under regional metamorphic conditions, however, the influence of temperature on the epidote-plagioclase equilibrium would exceed the influence of pressure and a basic plagioclase must indicate only a high temperature (Ramberg, 1945, p. 52). Furthermore, Rosenqvist (1952, p. 49) suggests that ferrian epidotes are stable together with a given plagioclase at a higher temperature than is assumed for clinozoisite. In the rock under discussion the Fe_2O_3 content available for the formation of silicates is very low and any notable pistacite content in epidote would be excluded. Again the complete absence of epidote may well be attributed to a relatively high temperature.

Absorption of water to build up minerals is generally assumed to be opposed by high temperature but favoured by increase of pressure (see Turner, 1948, p. 86). According to this conception, a mineral rich in water, such as chlorite, could not be formed at high temperatures, unless the pressure is also very high. Suggestions that water is an essential chemical factor under such conditions are given by Eskola (1939, p. 363) and Turner (1948, p. 102). But the water content of rocks may be of more general significance (Rosenqvist, 1952, p. 32), and the amount of water present in minerals is probably more independent of the physical conditions (Yoder, 1952). Yoder's experiments showed that chlorite is stable up to 680°C at about 1 000 atm., corresponding to a depth of less than 4 km. Thus no high pressure would be necessary to cause minerals rich in water to crystallize at fairly high temperatures. Yoder does not consider the influence of pressure as important as it has been generally believed to be.

The conditions of formation of the paragenesis thus remain somewhat uncertain. Starting from the present facies classification, Turner's (1948, p. 77) cordierite-anthophyllite subfacies of the amphibolite facies may correspond most closely to the data, mainly because the shearing stress is supposed to be low. To follow Yoder's (1952) suggestions, one may actually wonder whether the present classification of the metamorphic facies with their critical mineral assemblages is not unconditionally valid; and whether the bulk composition might be a more important factor in controlling mineral associations of the metamorphic rocks than is assumed.

Acknowledgments — The separation of the chlorite was carried out by Mr. Väinö Makkonen under the author's supervision, and the optical determinations of one of the chlorites were made by Mr. Kai Hytönen. Messrs. Pentti Ojanperä, M. A., and Aarno Juurinen, M. A., made the chemical analyses. The diagram in Fig. 1 was drawn by Miss Thyra Åberg. The author wishes to express his sincere gratitude to these persons.

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8

STRATIGRAPHY OF THE SCHISTS OF THE TAMMELA—KALVOLA AREA, SOUTHWESTERN FINLAND [1]

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ABSTRACT

Structures and stratigraphy of the schists of the Tammela—Kalvola area are described. The sequence and the conditions of the sedimentation are compared with those found elsewhere in southwestern Finland.

CONTENTS

INTRODUCTION

About 100 km south of the well-known Tampere area another region of well-preserved Svecofennian schists occurs in southwestern Finland. This formation, called the Tammela—Kalvola area, was first described by Sederholm (1891). Mainly on the basis of observations in this area, Sederholm (1893) also proposed an interpretation for the stratigraphy of the early Archean supracrustal rocks in Finland.

The schists in the area are indeed suitable for stratigraphical work. In many localities the primary bedding is still observable in the sedimentary and volcanic rocks occurring in gentle folds with more or less horizontal

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[1] Received November 30, 1953.

axes. It is natural, therefore, that the writer while remapping the area for the geological map sheet 2113, Forssa, (Fig. 1) which includes the Tammela—Kalvola schists, collected observations concerning the stratigraphy in the area.

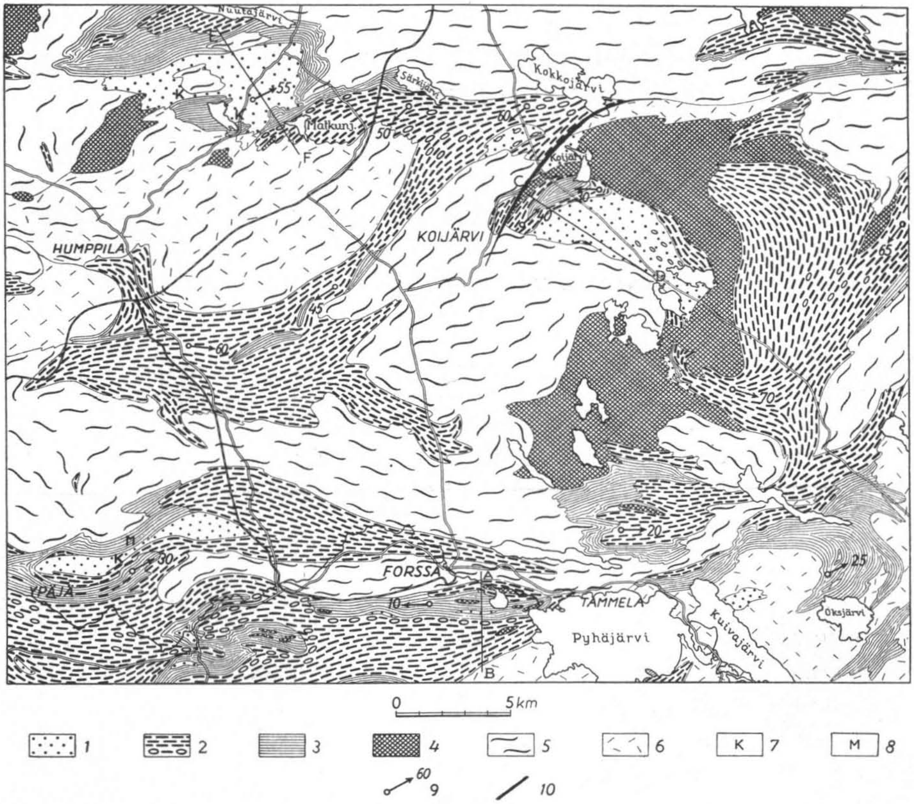

Fig. 1. Geological map of the Tammela—Kalvola schist area (Sheet 2113, Forssa). 1, quartz feldspar schist; 2, subsilicic volcanics and agglomerates; 3, mica schist; 4, diorite and quartz diorite; 5, granodiorite; 6, microcline granite; 7, limestone; 8, anthophyllite-cordierite schist; 9, folding axis; 10, fault line.

The locations of the diagrams presented in Figs. 2, 3, and 4 are indicated by the lines A—B, C—D, and E—F, respectively.

On the basis of observations of Metzger (1945, 1947) and of Simonen and Kouvo (1951) and considering some unpublished investigations recently carried out in southwestern Finland, Simonen (1953) was able to outline the general features of the stratigraphy in the Svecofennian supracrustal formations in this part of the country. For a more complete summary of that kind, however, the strata should be known in detail from a great number of different localities. The purpose of this paper is to fulfill this need for the part of the Tammela—Kalvola schists.

STRUCTURES

The rocks in the area consist of schist belts separated from each other by granodiorites and associated intrusives. Conform intrusive contacts and other structural features indicate that these intrusions penetrated the schists during a period of intensive movements. A long continuous schist belt occurs in the southern part of the mapped area running west from Lake Pyhäjärvi. The schists in this belt are composed of folded sediments and volcanics. Most of the folds are isoclinal and slightly overturned to the south. The strike of the axes follows the elongation of the belt in an east-west direction. The axial plunge is moderate or even horizontal. Along the northern edge of the belt a narrow anticline is situated. The northern limb of this structure is cut off by the intrusives, but towards the south it is followed by a syncline and another anticlinal structure (Fig. 2). Foliation in the schist subparallels the axial plane of the folding, and the lineation plunges down the dip on this plane.

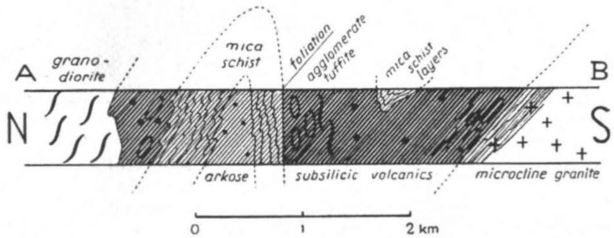

Fig. 2. Vertical section through the southern schist belt 3 km west of Lake Pyhäjärvi, Tammela.

In the northern part of the area another long schist belt is met (Fig. 1). Simonen (1949, p. 32) suggested that its eastern continuation near the town Hämeenlinna has a synclinal structure. The same interpretation is possible also for the structures observed in the belt in the area now considered. These structures are, however, somewhat more complicated. North of the church of Koijärvi, a syncline is found in which the fold axes plunge towards the center of the schist belt from both N. E. and S. W. This double-plunging syncline is, on the other hand, cut by a large vertical displacement in the west (Fig. 3). The observations indicate that the western block has risen in relation to the eastern block. West of the fault line the stratum is nearly horizontal, but towards west and northwest the bedding gets more inclined and forms a wide syncline with a gentle axial pitch to S. W. Along the line Lake Särkijärvi — Lake Matkunjärvi the western limb of this syncline rises up, and a round anticlinorium is observable between those lakes and Lake Nuutajärvi (Fig. 4). The axis of this structure is more or less horizontal but plunges, however, about 50° N. E. in the northeastern part of the anticlinorium. The strike of the folding axes is remarkably constant N. E.—S. W. throughout the whole northern schist belt and independent of the variations in the trend of the belt proper.

Along with the two long schist belts, several smaller areas are occupied by supracrustal rocks. Their investigation, however, yields stratigraphical results of no particular moment. In the western portion of the area (Fig. 1), a fairly large schist body is located. The structures there observed indicate that it makes a syncline in an axial depression. This interpretation is verified by observations about the graded bedding round in the mica schists in the central part of the area.

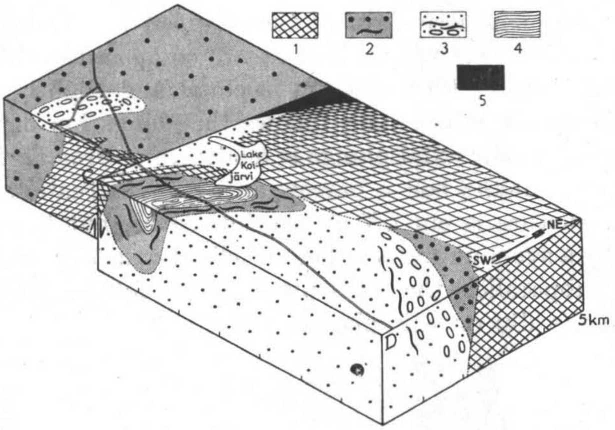

Fig. 3. Synclinal depression and the great displacement near Koijärvi. 1, diorite and granodiorite; 2, uralite porphyrite and subsilicic tuffite; 3, silicic and intermediate lava, agglomerate and tuffite; 4, mica schist; 5, microcline granite.

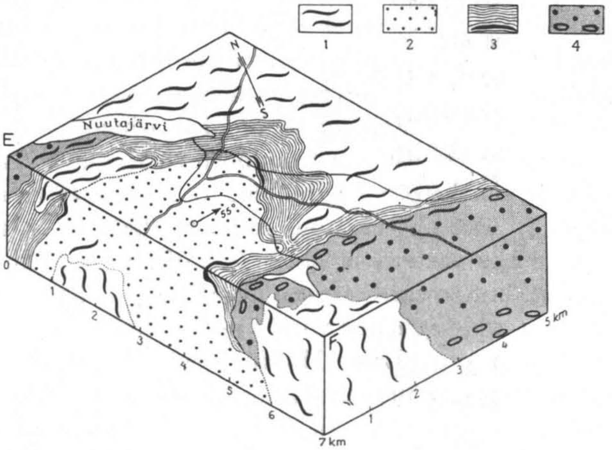

Fig. 4. Nuutajärvi anticlinorium. 1, granodiorite; 2, graywacke (quartz-feldspar schist); 3, mica schist; 4, subsilicic volcanics and agglomerate.

ROCK SEQUENCE

The stratigraphical sequence of the supracrustal rocks in the area can be established mainly in locations where folds are seen. In the narrow anticline occurring in the southern schist belt a quartz-feldspar rock forms the central portion of the fold (Fig. 2). Consequently, this rock represents the lowest stratum in the local sequence. Stratification of this rock and its mineral composition [1] (oligoclase, quartz, and a little biotite) indicate that it originally was an arkose. On the top of this formation there lies a stratum of mica schist about 500 m thick. In less sheared well-preserved places a varved structure with graded bedding is detectable, indicating that the rock was originally an argillaceous sediment. In the eastern end of the schist belt, this stratum is in some places overlain by stratified quartzite (quartz, a little oligoclase, and muscovite). Such layers are not met in the sequence in the western part of the belt. Near the church of Ypäjä, on the other hand, the mica-schist stratum is intensively intercalated by layers of a quartz-feldspar rock and calcium-rich amphibolites. This amphibolite is evidently a metamorphosed calcareous sediment, and it is in places associated with pure limestone.

Above the sedimentary strata described, a thick series of subsilicic volcanics of different types is met. The rock sequence in the volcanics as seen in the syncline of the southern schist belt invariably begins with an agglomerate. The fragments found in a tuffitic cement are mainly basaltic, but pebbles of sedimentary rocks are also met. In few localities where the foliation is not too well developed, the rock appears well bedded and contains narrow strips of argillaceous sediments. Consequently, the rock should in places be called volcanic conglomerate. An estimated thickness for this stratum is about 200 m. Overlying the agglomerate, a complex of alternating layers of tuffites and lavas occurs. These are generally basaltic in composition, but in some places layers of intermediate tuffites are found as intercalations. The thickness of the whole volcanic complex in this part of the area measures about 1 km, but it is not possible to estimate the total thickness of the system, because the top of the formation is evidently not met in the said syncline. The uppermost part of this series is intercalated by some argillaceous layers. It is not possible, however, to determine whether these layers indicate the end of the volcanic period or just casual sedimentation during the volcanic activity.

In the schists enclosed in the synorogenic intrusive granodiorites as veined gneisses, a sequence can be established similar to that described above. Because the folds in the veined gneisses are rather shallow, only a minute portion of the sequence is observable in a single fold. The following

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[1] A complete petrographic description of the rocks considered will be given in the Explanation to the Geological Map Sheet 2113, Forssa.

tabulation summarises the stratigraphic sequence, as determined in the schists of the southern part of the area.

	Thickness
(Lowermost) Sandstone, arkose minimum	500 m
Limestone (found locally near Ypäjä only)	20 m
Argillaceous mica schist (near Ypäjä, intercalated with arkose and calcium-rich pelitic rocks)	500 m—1 000 m
Quartzite (found only in the eastern part of the belt)	0 m— 50 m
Agglomerates, subsilicic	200 m
Basaltic lavas and tuffites with sedimentary layers in the top section	1 000 m
Minimum total of the strata observed	2 200 m

From the northern schist belt supplementary data may be obtained for the areal stratigraphy. Here the lowermost stratum is represented by coarse-grained quartz-feldspar schists occurring in the central portion of the wide anticlinorium in the N. W. corner of the area (Fig. 4). This formation, though more than 2 km thick, is very homogeneous, and shows variation in the grain size and the abundance of quartz only. The amount of feldspar (mainly oligoclase) always exceeds that of quartz. In the best preserved types, the feldspar and quartz grains are fragments in a fine-grained clayey matrix consisting of quartz, feldspar, and mica. Fragments of a quartz-rich fine-grained rock sometimes occur among the quartz and feldspar fragments. It thus seems evident that this rock originates from a sedimentary graywacke. Some argillaceous strips are found as intercalations in this formation. On the top of this stratum, there occurs a series of alternating layers of calcium-rich diopside-bearing amphibolites, limestones, and quartzitic arkoses. In many places, however, the graywackes are in close contact with the mica schists which generally form the next stratum above the limestone series. The mica schists undoubtedly are of sedimentary origin and are often well stratified. Their total thickness hardly exceeds 500 m. The uppermost part of the mica-schist formation is intercalated with layers of quartzitic sandstone.

Overlying the mica schists, a thick complex of subsilicic volcanics occurs. The lowest horizon of this series consists of either stratified tuffite or agglomerates. The thickness of these rocks varies considerably but is usually of the order of a few hundred meters. A complex of lavas occurs above the agglomerates (alternatively tuffites). Rocks of basaltic composition (uralite porphyrites, plagioclase porphyrites, and associated metamorphosed agglomeratic lava rocks) make the main part of this series. Tuffitic layers are met as rare local varieties only. Near the great displacement mentioned earlier a more silicic variety, *viz.*, an agglomerate is found as the uppermost member of the series. Most of the fragments found in this rock and the tuffitic cement are of trachyandesitic composition, but sedimentary as well as basaltic fragments are also present.

The total of the strata worked out from the northern schist belt west of the fault line is as follows:

	Thickness
(Lowermost) Graywacke, unstratified	2 000 m
Limestone and associated calcium-rich pelites and arkose layers	50 m
Argillaceous sediments with quartzite intercalations in the top section	500 m
Agglomerates (or tuffites)	200 m
Bacaltic lava flows	2 000 m
Agglomerates, intermediate	100 m
Total, minimum	4 850 m

According to Simonen (1949, p. 32), the subsilicic volcanics overlie sedimentary schists in the syncline near the N. E. corner of the mapped area. The observations made by the present author are in good agreement with Simonen's interpretation. Because of the inhomogeneity of the volcanic complex in this part of the area, it was not possible, however, to arrange the different members of the complex in detail. The main features of the sequence, however, agree well with that established for the Nuutajärvi—Särkijärvi area. The arrangement met in the central portion of this belt, near Lake Koijärvi, on the other hand, differs considerably from that found at the both ends of the zone.

A synclinal depression is assumed to occur in the schist belt east of the fault line. The central portion of this depression (Fig. 3) is occupied by mica schists with a well-preserved varved structure. Undoubtedly this rock is a metamorphosed argillite. There occur, particularly near the bottom of the formation, intercalations of subsilicic tuffite in the mica schist. The next member downwards in the sequence is a more or less pure tuffite with some narrow basaltic lava beds. The thickness of these volcanics measures about 300 m. Underlying these volcanics, an extremely fine-grained silicic rock exists. The major constituents of this rock are 60 %—70 % quartz, 10 %—30 % albite, $<$ 10 % microcline, and $<$ 10 % biotite [1]. A few larger grains of feldspar and quartz were microscopically observed to occur as phenocrysts in the rock, revealing that it was originally a liparitic lava. Since no primary bedding is detectable in the rock, it is not possible to give any reliable figure for the thickness of this formation.

Towards the east the silicic lavas are followed by amygdaloidal intermediate porphyrites and tuffites. The next member in the sequence is an agglomerate composed of an intermediate tuffitic matrix and of pebbles mostly trachyandesitic in composition. A number of the fragments are sedimentary, and some basaltic pebbles are also found. A well-preserved bedding is locally observable.

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[1] Composition determined by means of X-ray powder diffraction analysis.

The agglomerate is followed by a massive uralite porphyrite that is the last member in the section. The whole sequence met in these schists runs as follows:

(Lowermost) Uralite porphyrite
Agglomerate, intermediate
Tuffites, intermediate
Porphyrites, intermediate
Lavas, silicic
Tuffites, subsilicic
Argillaceous sediments

It is readily noted that the bottom of this rock sequence is similar to the top of the sequence met west of the great fault near Lake Koijärvi. The two agglomerates found on either side of the displacement are also highly similar as far as the structure and composition of their pebbles and matrix is concerned. Because they both are underlain by similar basaltic lavas, it is evident that they represent a single formation. This interpretation is in agreement with the observations concerning the direction of the faulting.

Starting from the bottom, the complete stratigraphic sequence in the northern schist belt runs as follows:

	Thickness
(Lowermost) Graywacke, unstratified	2 000 m
Limestone and associated calcium-rich pelite and arkose layers	50 m
Argillaceous sediments with quartzite intercalations in the top section	500 m
Agglomerates, subsilicic (alternatively tuffites)	200 m
Basaltic lavas	2 000 m
Agglomerates, intermediate	100 m
Tuffites, intermediate	50 m
Porphyrites, intermediate	200 m
Lawas, silicic	(2 500 m)
Tuffites, subsilicic	300 m
Argillaceous sediments greater than	300 m
Minimum, approximately	7 500 m

When this sequence is compared with the sequence found in the southern schist belt (see p. 90) a definite similarity is observed. In fact, these two strata are so much alike that they must be correlated with each other. In the southern belt, however, the upper formations are not represented. In this respect the schist zone in the western part of the area, between the two main belts, resembles the southern belt. Mica schists are found in the synclinal depression occurring in the center of the zone. The argillites, consequently, in this area represent the uppermost formation in the same manner as they do in the southern belt. That the subsilicic volcanics are followed by more silicic ones in the northern schist belt may be the result of local variations in the sequence that may be expected to occur during a period of volcanic activity.

COMPARATIVE REMARKS

The lowest stratum found in the Tammela—Kalvola area is the thick graywacke formation in the Nuutajärvi anticlinorium. Mineral and rock fragments in this rock indicate an incomplete weathering and a relatively short distance of transportation prior to accumulation. Simonen and Kouvo (1951, p. 112) following Pettijohn's (1949) suggestions concluded that the varved graywacke slates in Tampere area represent a geosynclinal accumulation during a period of tectonic activity. Because of the great similarity between the graywackes of Nuutajärvi and those of the Tampere area the writer considers a similar origin very probable for both. Accordingly, the sedimentation took place in a shallow basin which subsided continuously under the sediments accumulated. The main difference between the two graywacke types discussed is that the Nuutajärvi formation lacks the graded bedding so typical of the Tampere schists. This diversity was evidently caused by the more rapid accumulation of the Nuutajärvi sediments.

Although the general features of the rock sequences in Tampere area and in the Tammela—Kalvola area are very much alike — *e. g.*, the subsilicic volcanics are underlain by sedimentary formations in both — certain characteristic differences exist. In the Tampere area, the graywacke slates are overlain by arkose while the graywackes in the Tammela—Kalvola area are followed by a thin layer of limestone, calcareous slate, and intercalated quartzite. The presence of such sediments must indicate a rather radical change in the environments and conditions of sedimentation. Formation of deposits of these kinds requires, at least, a tectonically more stable basin and a lower relief than is typical of the formation of unsorted graywackes. Quartzite intercalations in the limestone and in the argillaceous mica schists indicate the existence of a littoral shallow-water environment of deposition and of fluctuating water level.

Simonen (1953, p. 55) remarked that limestones and other calcium-rich schists occur in abundance in the coastal area of southern Finland but are not found in the schist terranes of the interior, e. g., in the Tampere area. According to Simonen, the calcareous sediments denote a marginal zone of deposition of the geosyncline, *i. e.*, the miogeosynclinal facies. In the Tammela—Kalvola area calcareous sediments occur in the western part only. Near Ypäjä the calcium-rich intercalations and limestones have their largest distribution in the area. Towards the southwest the abundance of the calcareous sediments increases still more. Therefore, it seems probable that the Tammela—Kalvola area is a transitional zone between the eo- and miogeosynclinal belts. This problem and many others arising from the observations made in the area now considered have to wait for a later solution. The geological remapping of southwestern Finland, to be completed in the near future, may be expected to increase the still scanty information about the stratigraphy of the Svecofennian rocks and about the conditions during the ancient times when the said rocks were formed.

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9

THE VÄSTILÄ AREA IN THE TAMPERE SCHIST BELT[1]

BY

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ABSTRACT

Schists and associated plutonic rocks in a part of the Tampere schist belt are described. A potash metasomatism in a section of the schist belt is inferred from a constant occurrence of microcline in the various schist types and from replacement and recrystallization textures. The metamorphism of the schists varies somewhat, but seems as a rule to correspond to the conditions of the amphibolite facies. A small intrusion of non-metamorphic diabase is described.

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[1] Received December 7, 1953.

INTRODUCTION

The area to be described (see map, Fig. 1) comprises a section of the eastern part of the Tampere schist belt and associated plutonic rocks north of it. It is situated in the parish of Längelmäki, about 50 km east-northeast of Tampere and 10 km east of the area previously investigated by the author (Seitsaari, 1951).

The only description of the area so far published was given by Sederholm (1897, p. 120). The description was cited in the explanation to the Mikkeli (S:t Michel) map sheet (Frosterus, 1902). Sederholm was particularly interested in the northernmost part of the schist sequence, where the schists are remarkably rich in feldspar and show gradation into granite. He called them »felsite schist», or »Västilä schist», after a village situated in the area.

GABBRO, DIORITE, AND GRANITE

Large massifs of basic plutonic rocks occur in the northernmost and the eastern parts of the area. An isolated, small massif is encountered in the island of Rekisaari, in Lake Längelmävesi.

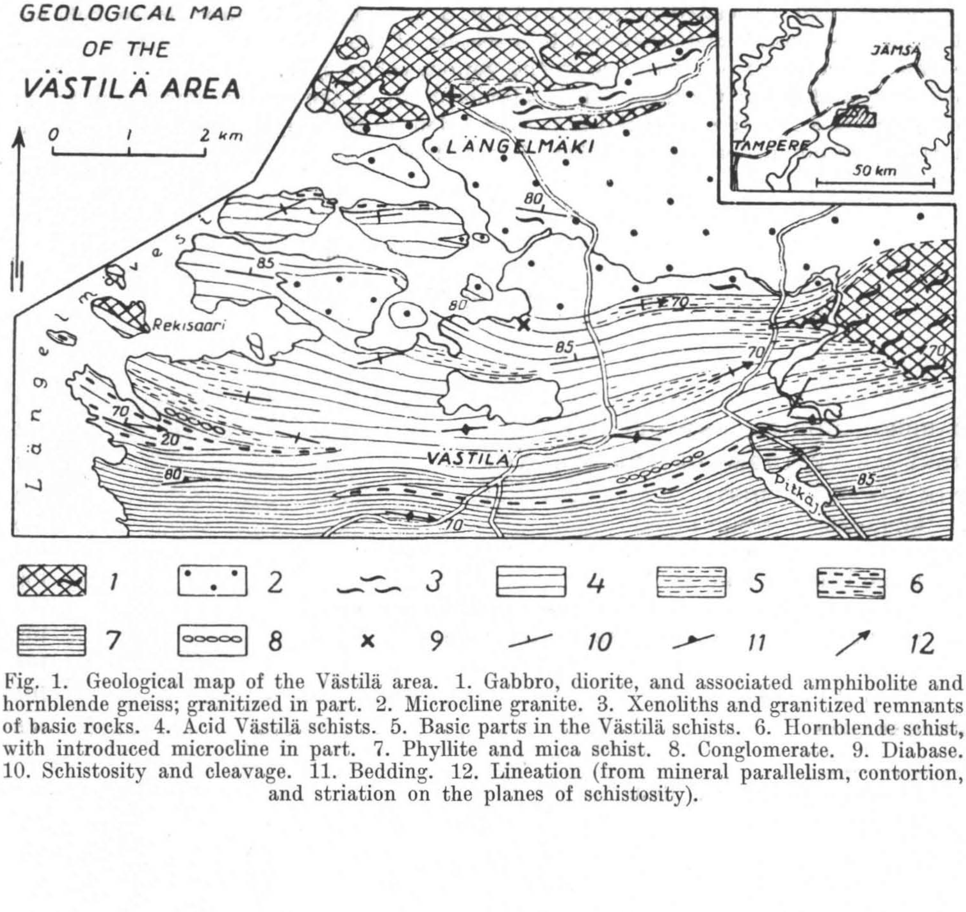

Fig. 1. Geological map of the Västilä area. 1. Gabbro, diorite, and associated amphibolite and hornblende gneiss; granitized in part. 2. Microcline granite. 3. Xenoliths and granitized remnants of basic rocks. 4. Acid Västilä schists. 5. Basic parts in the Västilä schists. 6. Hornblende schist, with introduced microcline in part. 7. Phyllite and mica schist. 8. Conglomerate. 9. Diabase. 10. Schistosity and cleavage. 11. Bedding. 12. Lineation (from mineral parallelism, contortion, and striation on the planes of schistosity).

The northern basic massif consists of hornblende diorite and gabbro. The main minerals are hornblende and plagioclase, with some additional biotite and quartz as a rule. The diorite and gabbro are mostly gneissose, and enclose several amphibolitic portions. The amphibolite forms irregular bodies, or occurs as long, narrow lenses which resemble layers in a supracrustal deposit. The irregular portions, in particular, show gradual transition into diorite and gabbro. Some patches are very rich in hornblende and may contain some augite, approaching hornblendite. A few larger areas of this hornblenditic rock are encountered.

In many places plagioclase porphyroblasts appear in the amphibolite. Almost unaltered crystals of later plagioclase (An_{35-40}) surround sericitized grains of earlier plagioclase (about An_{50}). The earlier grains have sinuous, resorbed boundaries.

The northern basic massif has many features in common with a basic complex in Paattinen in the Turku district, southwestern Finland (Hietanen, 1947, pp. 1 072—3). According to Hietanen, many diorites in Paattinen originate from basic volcanogenic rocks that have recrystallized under a contemporaneous introduction of Na-feldspar. A similar development of diorites and gabbros is in evidence also in the Västilä area.

The massif in the island of Rekisaari consists of fine-grained, gneissose hornblende-biotite quartz diorite.

The basic massif in the eastern part of the area grades into an adjacent hornblende gneiss without distinct boundaries. Each rock, however, is thoroughly granitized, and the primary texture and composition are obliterated. Only the abundance of hornblende in the massif makes it probable that the primary rock was fairly basic. Considerable parts of the northern basic complex are also granitized. Heterogeneous migmatites, and more homogeneous, granodioritic rocks are thus formed. The last-mentioned type of rock is a common transitional variety between the basic rocks and microcline granite.

Gradual granitization of the older rocks is frequently noted at the granite contacts in the whole area. Apart from the contact zones and small sporadic migmatite portions, the granite massif consists of a homogeneous microcline granite showing as a rule a distinct foliation.

GENERAL FEATURES OF THE SCHISTS

In the primary types of schists and in their sequence the area is similar to the other parts of the Tampere belt (see Sederholm, 1897; Seitsaari, 1951; Simonen, 1952). A local, secondary, difference appears in the potash metasomatism that has affected all rocks in the northern part of the belt. These metasomatic rocks, in common called »Västilä schists» by Sederholm (1897, p. 120), will be discussed in greater detail in the following chapter.

At the southern boundary of the above sequence of schists, there are two beds of a particular hornblende schist consisting of hornblende, oligoclase, epidote, quartz, and biotite. The hornblende schist is quite similar to a certain hornblende schist in Orivesi and Teisko, and occupies the same stratigraphical position (Seitsaari, 1953). The western and the west end of the eastern bed has later changed metasomatically and belongs to the »Västilä schists».

The phyllite and mica schist farther south continue far beyond the limit of the area mapped. The margin of the phyllite belt in the western part of the area probably belongs to the metasomatic zone, as does certainly the phyllite wedge north of the eastern bed of hornblende schist. In general, the phyllite and mica schist have not been altered secondarily and mainly consist of biotite, quartz, and muscovite. In places, almandite mica schist, staurolite-almandite mica schist, and andalusite mica schist are found. The almandite, staurolite, and andalusite porphyroblasts are well-preserved. As in the other parts of the Tampere belt, the regional metamorphism took place under the conditions of the amphibolite facies, but the afore-mentioned mineral parageneses show that both shearing stress and temperature have varied. Staurolite points to a notable shearing stress, while this mineral prefers relatively low temperatures of the amphibolite facies. On the other hand, andalusite is a typical antistress mineral that does not avoid high temperatures. Farther west, highly altered andalusite is found in a few places (Simonen and Neuvonen, 1947, p. 254), while sericite pseudomorphs after andalusite are common. Thus some indication, local as it is, of a higher grade of metamorphism exists in the eastern part of the Tampere schist belt, as compared with the central and western parts.

All the Tampere phyllites, also those south of Västilä, contain intercalations of quartz-feldspar schist and basic schists. (The intercalations are not shown on the map.) The texture of the quartz-feldspar schist indicates a sedimentary origin (see Seitsaari, 1951, pp. 38—9, and Simonen, 1952). Of the basic intercalations, a diopside amphibolite may be mentioned. It is situated at the northern end of Lake Pitkäjärvi, near the northern boundary of the phyllite area. The constituents of the rock are diopside, hornblende, fine-grained plagioclase with about 50 % An ($\alpha = 1.555$, $\gamma = 1.563$), some quartz, and a little secondary microcline. In the central and western parts of the Tampere belt, diopside amphibolite is found only in a few places at the contacts of plutonic rocks. Diopside, however, may be formed at all temperatures of the amphibolite facies if the bulk composition is appropriate. The lack of diopside amphibolites in the Tampere area is probably due to the composition of the sediments more than to the physical conditions during the metamorphism. The schists rich enough in CaO usually contain too much Al_2O_3.

THE VÄSTILÄ SCHISTS

GENERAL STATEMENTS

The zone affected by potash metasomatism in the northern part of the schist belt is up to 4 km wide. The microcline-bearing »Västilä schists» thus originated vary greatly in composition. Primary types such as basic volcanogenic rocks, intermediate rocks of various origin, quartz-feldspar schists (»leptites»), and subordinate phyllite and conglomerate can be recognized. To a certain extent, the present variation between the basic, intermediate, and acid types may be caused by variation in the intensity of metasomatism.

Evidence of a secondary introduction of microcline is afforded by the following facts:

1. Abundant microcline is found in many different rocks, also in the most basic ones. Irregular variability in the microcline content, independent of the primary type of rock, is observed. 2. In thin sections microcline is seen to be unevenly distributed and to form lenses and long strings parallel to the schistosity. 3. Plagioclase, mostly sericitized, is often largely replaced by fresh microcline. 4. Microcline forms porphyroblasts, small as they are. A porphyroblast may be formed by coalescence of several different grains into a single crystal, or at the expense of one single, replaced, phenocryst of plagioclase.

Reference is made, *e. g.*, to the author's earlier paper (Seitsaari, 1951, pp. 90, 98).

Crystallization of microcline in rocks that certainly did not contain Al_2O_3 in excess, as in the basic volcanics, shows that Al_2O_3 was also introduced into the rocks.

A secondary mobilization of quartz is in evidence, and quartz has probably increased secondarily. The migration of SiO_2 cannot be referred to any particular stage of metamorphism. Roundish or oval accumulations of quartz are similar to some granulated phenocrysts of quartz in metamorphic porphyries, but every transition from the »phenocrysts» to long strings is represented, and similar accumulations are seen in various rocks.

The metamorphism took place very quietly. The primary fabric is very well-preserved, and the grain size has not greatly increased. In a few places near the microcline granite, however, the »soaked» schist seems to have been rendered mobile and behaves like an intrusive rock. Protrusions from this intrusive part are seen in the associated, better preserved portion of the schist, but under the microscope, the intrusive rock highly resembles the associated schist and is not like a granite. An intrusive behaviour of the soaked rock was suggested by Wegmann (1931, p. 58), and in this particular case there is no reason to believe that any considerable part of the rock had been in a liquid state (see Turner and Verhoogen, 1951, p. 303).

ACID VÄSTILÄ SCHISTS

The definition »acid Västilä schists» on the map comprises all metasomatic schists with abundant quartz, oligoclase, and microcline in variable relative amounts. Biotite in moderate to small amounts is usually the only dark mineral, but some hornblende may be present. Most of these schists contain plagioclase phenocrysts in abundance, but the primary habit of the phenocrysts is seldom in evidence, owing to the metasomatism and recrystallization. Yet the texture is mostly suggestive of a volcanogenic rock. The primary composition of these schists is supposed to have varied from intermediate to acid.

As an interesting and rather typical example of this group of schists, the most acid type with a high K_2O content may be considered in greater detail.

This kind of rock occurs in the southern part of the zone of the Västilä schists. The primary rock was probably also acid, for the primary composition of more basic schists is usually not totally obliterated. Most of these acid schists also seem to have contained, and may still contain plagioclase phenocrysts, and some phenocrysts are angular, like fragments. In some places the schist is stratified, and a few intercalations of conglomerate are found. The schists are supposed to be of sedimentary origin, in part probably volcanic ashes (see Seitsaari, 1951, pp. 20, 38—9; Simonen, 1952). There are sedimentogenic quartz-feldspar schists farther west in the Tampere belt that occupy the same stratigraphical position as do the Västilä schists under consideration (Simonen, 1952).

The main constituents are microcline, quartz, muscovite, plagioclase, and biotite. The rock is mostly heterogeneous and the amounts of feldspars and quartz are fairly variable. The plagioclase is usually an oligoclase (An_{15-25}), in one instance an albite (An_{5-10}). In the albite-bearing rock most of the biotite is chloritized. The usual grain size in the ground mass is 0.04—0.08 mm, and the microcline porphyroblasts are up to 3 mm long. A well-developed parallel texture is indicated by the muscovite flakes and by streaks of muscovite and microcline. The metasomatism and recrystallization probably took place during penetrative movements.

A chemical analysis of a Västilä schist rich in K_2O is quoted from Eskola (1927, p. 76) and given in Table 1, Anal. 1. (The Niggli numbers were calculated by the present author.) In addition, the ionic percentage and the one-cation molecular norm are calculated (Table 1, Anal. 1a), according to a recent proposal by Eskola (1954). The great excess of Al_2O_3 indicates a high content of muscovite, and a considerable part of the excessive Fe_2O_3 is, perhaps, also present in the muscovite. The number of O (see Anal. 1a)

Table 1. Chemical composition of an acid Västilä schist. Västilä, Längelmäki (Eskola, 1927, p. 76)

1					1a					
Constituent	Weight %	Mol. prop.	Niggli numbers		Constituent	Cation number	Cation %	O in oxides	One-cation molecular norm	
SiO_2	70.53	11 743	*si*	363	SiO_2	11 743	66.33	132.66	Q	25.90
TiO_2	0.45	56	*ti*	1.7	TiO_2	56	0.32	0.64	Or	37.80
Al_2O_3	15.67	1 537	*al*	47.6	$AlO_{3/2}$	3 074	17.36	26.04	Ab	26.30
Fe_2O_3	2.11	132	*fm*	13.2	$FeO_{3/2}$	264	1.49	2.24	An	3.75
FeO	0.58	81	*c*	4.1	FeO	81	0.46	0.46	C	3.04
MnO	0.03	4	*alk*	35.1	MnO	4	0.02	0.02		
MgO	0.32	79	*k*	0.59	MgO	79	0.45	0.45	En	0.94
CaO	0.74	132	*mg*	0.18	CaO	132	0.75	0.75	Il	0.64
Na_2O	2.89	466	*c/fm*	0.31	$NaO_{1/2}$	932	5.26	2.63	Mt	0.42
K_2O	6.30	669			$KO_{1/2}$	1 338	7.56	3.78	Hm	1.21
P_2O_5	0.01	1			$PO_{5/2}$	1	0.00	0.00		
H_2O	0.30	166			H_2O	(166)	(0.94)	—		
Total	99.93	—	—		—	17 704	100.00	169.67		100.00
							—O for (OH)	0.94		
								168.73		
							(OH)	1.88		
							Anions pro 100.00 cations	170.61		

greatly exceeds 160. This is a remarkable feature and may be characteristic of this kind of metasomatic rock (see Eskola, 1954, p. 8).

An alkali determination was made of a homogeneous schist very rich in microcline with the following result:

Na_2O 1.84 % K_2O 7.25 %

As far as is known to the author, this K_2O content is the highest ever found in quartz-feldspar schists in Finland, although it is almost matched by a porphyry from Ylöjärvi with 7.20 % K_2O (Simonen and Neuvonen, 1947, p. 252) and another from Vaavujärvi, Orivesi, with 7.10 % K_2O (Sederholm, 1897, p. 68). In the porphyry from Vaavujärvi a secondary increase of K_2O was recently suggested (Seitsaari, 1951, p. 17).

Of course, a maximum content of 7.25 % K_2O does not necessarily indicate a metasomatic origin. In Sweden and Finland there are porphyries with approximately the same K_2O content and without notable traces of secondary changes (see, *e. g.*, Sundius, 1923; Simonen and Neuvonen, 1947, pp. 251—2). For the Archean porphyritic »leptites» in general, both a volcanic origin by some kind of magmatic differentiation (Sundius, 1926; Asklund, 1931) and a metasomatic origin (Backlund, 1938) have been suggested. The interpretation of certain extreme leptites containing up to 13 % K_2O (Asklund, 1929) has been the most difficult. They do not display any relict textures at all, and often contain only little or no quartz and rarely muscovite. In the varieties poor in quartz, the usual dark minerals are hornblende, hedenbergite, and epidote. These rocks thus differ in many respects from both the porphyritic leptites in Sweden and Finland and the Västilä schists.

BASIC VÄSTILÄ SCHISTS

The schists that were primarily basic have not been affected very strongly by the potash metasomatism, and the primary texture is mostly in evidence. Uralite-plagioclase porphyrites, plagioclase porphyrites, and amphibolites can be recognized. In addition to the secondary microcline they contain secondary biotite and quartz. The plagioclase varies from calcic oligoclase to sodic andesine in composition. Some amphibolites among the Västilä schists have escaped the metasomatism and have a more calcic plagioclase than the rest of them.

HORNBLENDE SCHIST WITH INTRODUCED MICROCLINE

The hornblende schists (see p. 98) into which microcline has been introduced differ from the unaltered hornblende schists by having a somewhat coarser texture.

Characteristic of the hornblende schist in general is the presence of abundant epidote together with oligoclase ($An_{20\text{-}30}$) and, in particular, hornblende prisms (up to 7 mm long) with a random orientation. This manner of crystallization and a blue-green colour suggest that the hornblende is similar to another hornblende recently described by the author (Seitsaari, 1953). In the altered hornblende schists, however, the hornblende crystals rarely display any well-defined shape, but have crystallized poikiloblastically by enclosing microcline-bearing ground mass and sometimes biotite crystals (Fig. 2). In a hornblende schist there was observed a diffuse vein consisting of oligoclase, quartz, epidote, and large crystals of blue-green hornblende. The schist is rich in microcline, whilst the vein does not contain microcline at all. The vein seems to be later than the introduction of microcline. It is suggested that all the poikiloblastic hornblende crystals were formed after the potash metasomatism. During this late stage of metamorphism, the penetrative movements were weak, if not terminated.

The Västilä hornblende schist differs from the hornblende schists in Orivesi and Teisko (Seitsaari, 1953) only by its microcline. The composition of the last-mentioned schists was regarded as primary, and it may be supposed that also in the Västilä hornblende schist the hornblende crystallized from the materials primarily present in the rock. In an associated rock, more acid than the usual hornblende schist and devoid of epidote, the porphyroblastic hornblende crystals are surrounded by light aureoles (2.5 mm—10 mm in diameter) consisting of mere feldspar and quartz (Fig. 3). The aureoles indicate that the (Mg,Fe)Ca material has moved by means of diffusion to the hornblende.

Fig. 2. Poikiloblastic crystal of hornblende (grey) in hornblende schist. A biotite crystal (dark grey) is enclosed by the hornblende. Nic. +. Magn. 50 ×.

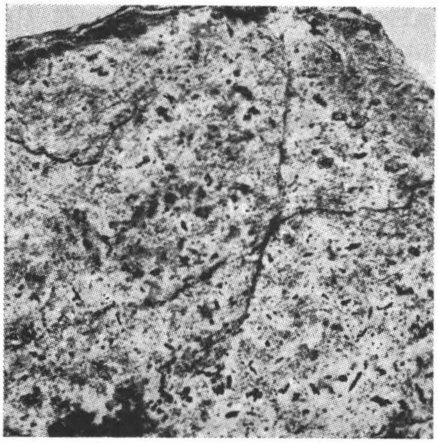

Fig. 3. Hornblende prisms surrounded by light aureoles in a Västilä schist. 4/5 natural size.

In a few instances biotite also forms large crystals like porphyroblasts. Biotite is sometimes chloritized, whereas most of the hornblende is very well-preserved.

METAMORPHIC FACIES OF THE VÄSTILÄ SCHISTS

The mineral assemblage of the acid Västilä schists is oligoclase (at least about 15 % An)—microcline—quartz—biotite, with some muscovite or hornblende in certain instances. The assemblage evidently means that the metamorphism took place under the conditions of the amphibolite facies. The same is the case with the basic Västilä schists with oligoclase-andesine—hornblende—biotite—microcline—quartz.

The hornblende schist shows the mineral assemblage oligoclase (An_{20-30}) —epidote—hornblende—biotite—microcline—quartz. The biotite is sometimes chloritized, whilst the hornblende has been much more resistant. Chloritization of biotite in a rock rich in K_2O should mean a low temperature. If it is true, the hornblende also should have been formed at a low temperature. This interpretation, however, does not seem correct, because plagioclase containing less than 20 % An is never found in the hornblende schist nor in any other rock in the area with the exception of the most acid ones. In a silicified schist containing very basic portions of an older rock (Seitsaari, 1953, pp. 95—6) the author found a fresh labradorite (about An_{60}) associated with large flakes of chlorite, a fresh blue-green hornblende showing a random orientation, and some epidote. The assemblage was believed to mean a disequilibrium, but this is not necessarily so. According to Yoder (1952), chlorite is stable at fairly high temperatures in the presence of an excess of H_2O, and the present author found a chlorite mineral to be stable together with bytownite (Seitsaari, 1954). The mineral paragenesis of the hornblende schist is to be referred to the amphibolite facies (in the sense of Turner, 1948). Biotite is apparently more liable to chloritization than are certain types of hornblende.

This particular, blue-green type of hornblende seems to prove a very low shearing stress, judging from its manner of crystallization in all the respective schists over a wide area. As the temperature probably corresponds to the amphibolite facies, the epidote would be metastable (Turner, 1948), unless a high confining pressure has widened the field of stability of epidote (see Ramberg, 1944, p. 123).

CONCLUDING REMARKS

The essential difference in texture between the Västilä schists and the granitized rocks proper is the fine texture and the lack of myrmekite in the former. Myrmekitic plagioclase is as a rule present in the granitized rocks in the Västilä area, as it is elsewhere in the Tampere belt (see Seitsaari, 1951, pp. 91, 93). In the afore-mentioned points, the Västilä schists are similar to certain potash-metasomatic schists in Orivesi and Teisko (Seitsaari, 1951, pp. 97—8). However, the plagioclase in the last-mentioned rocks is almost an albite (An_{5-15}) and they never contain hornblende, although the primary rock was not usually acid. Thus the metamorphism of the Västilä schists must have taken place at a higher temperature. Their formation seems to be in a very close connection with the granitization, and they grade into more coarse-grained, myrmekite-bearing rocks in many places at the granite contact. The difference between the two processes is possibly explained by a difference in the concentration of the introduced fluids and in the mechanism of migration.

PYROXENE DIABASE

3.2 km south-southeast of the church of Längelmäki, a pyroxene diabase is met with as a separate exposure, about 10 m × 20 m in size. The rock is very well-preserved and distinctly ophitic in texture, and mainly consists of plagioclase and clinopyroxene.

The lath-shaped crystals of plagioclase are mostly 1 mm—1.5 mm long and usually zoned. The composition varies from An_{70} to An_{50} (U-stage). The plagioclase crystals are fairly idiomorphic, but the marginal parts of the crystals often seem to have crystallized later than the pyroxene.

The pyroxene is almost colourless. The optical properties are as follows:

$$\begin{aligned} &(+)2V = 22^\circ\text{—}32^\circ \text{ (U-stage; the variation within five determinations)}\\ &\gamma /\!\backslash c = 45^\circ \text{ (U-stage)}\\ &\left.\begin{aligned}\alpha &= 1.710 \pm 0.002\\ \gamma &= 1.736 \pm 0.002\end{aligned}\right\} \text{(immersion method)}\end{aligned}$$

The axial plane is parallel to 010; twinning on 100 is often seen. The pyroxene is obviously an isomorphous mixture of monoclinic (Mg,Fe) pyroxene and Ca(Mg,Fe) pyroxene (pigeonite). The diopside-clinoenstatite ratio can be only roughly estimated from the optical properties. According to the diagram given by Winchell (1951), the mineral contains a little more of the (Mg,Fe) compound than of the Ca(Mg,Fe) compound. The MgO/FeO ratio may be estimated at about 3/5.

A little hornblende (greenish brown) and biotite are associated with the pyroxene crystals. The other accessories are an oxidic ore mineral, probably ilmenite-magnetite, and apatite.

Pyroxene diabase has not been previously found in this part of Finland. The rock is quite similar to certain diabases in central Finland (Wilkman, 1925), and resembles the quartz-bearing pyroxene diabases in southwestern Finland and Ahvenanmaa (Åland) (*e. g.*, Wahl, 1906). Many of these diabases, if not all, are sub-Jotnian (Asklund, 1931). Another, post-Jotnian, quartz diabase occurs in southern East Karelia (Wahl, 1908).

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