

GEOLOGINEN TUTKIMUSLAITOS

BULLETIN
DE LA
COMMISSION GÉOLOGIQUE
DE FINLANDE

N:o 218

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

XXXVII

OTANIEMI 1965

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XXXVII

HELSINKI 1965

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PENTTI ESKOLA

1883—1964

BY

VLADI MARMO

WITH A SELECTED BIBLIOGRAPHY

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Pentti Eskola

More than three centuries ago there was a farm at Lellainen in the parish of Honkilahti, southwestern Finland, that was owned by a man who had been christened Esko. A tax reform was enacted around that time which forced everybody to adopt a family name. That is how Esko's farm became known as *Eskola*, a name that became attached to its proprietors and their families, too. The official records reveal that in 1640 the owner of the farm was Pentti Eskola and that ever since it has remained in the possession of the same family. Eskola grew into an estate of considerable size, one big enough to be obliged by law to furnish and equip a cavalryman for the king's army.

Throughout the centuries the Eskolas were farmers. But there was one known exception, a character named Lars Aeschonius (because Finnish names were changed at school by the Swedish-speaking authorities) who enrolled as student No. 42 in the recently established Academy of Åbo (Turku to the Finns). He applied himself to learning there in the years 1647 and 1648 and earned himself a clergyman's frock.

It was on the family's hereditary farm that on the 8th of January, 1883, Berndt Eelis Eskola was born, the second child of parents whose ancestors had all been tillers of the soil. Later on, the lad changed his first name from the Swedish Berndt to the Finnish Pentti.

His father was Kustaa Eskola, and his mother, Eeva Kristiina Ellaa, descended from old farming stock, too.

When Pentti was six years old, his mother died giving birth to her sixth child. Only two of her children survived to adulthood.

Father Kustaa was forty when he married, which meant that young Pentti could expect little help from him at that time. That is how it turned out that it was Farmer K. Eino, his aunt's husband, who became mainly responsible for bringing up young Pentti Eskola.

At the age of eight, he was sent to the primary school at the farmhouse of Ellaa, where he lived for two years with his maternal uncle. There he met his cousin Frans Lampola, at that time a student, later to become a clergyman, and some of his friends. Listening to the wise discussions of the students, Pentti Eskola, who was ten at that time, made a firm decision to learn similar things so as to be able to converse in the same way.

He was admitted to the secondary school in Pori. As Pentti Eskola himself remarked later on, he was never again so wise as after the entrance examination to the secondary school. That was in 1893. In this school, botany, Greek and Latin were the subjects that most attracted his young mind, but he was also interested in sports and all kinds of other things. Therefore, he was not a very good pupil at first — as far as marks are concerned. In the higher grades the situation changed, and he matriculated in 1901 with the highest marks in most subjects and with honor. As mentioned, he was in love with Greek, but not very fond of German. So, for instance, he never studied any grammar in this language — and still did well! In the last class at school, in view of approaching matriculation, he decided to buy a German grammar, too. But his classmates decided: Eskola never needed a grammar before, why now? And they hid it. Pentti Eskola never bought a new one, yet he passed his final examination in German — without the benefit of a grammar!

Pentti Eskola was at first inclined to major at the University in Greek. He changed his mind, however, and began by majoring in zoology. He failed in one of the preliminary examinations, and therefore went over to chemistry, in which he graduated in 1906. But, as he wrote in 1951: »Before that, however, I had decided to go over to geology, not because I was unfaithful to chemistry, but in order to apply chemistry to the mineral kingdom.»

Eskola's graduate study in chemistry concerned reactions effected by pressure in the solid state. At that time very little was known about such reactions, and the problems connected with them remained in his mind for most of his life.

We meet the young geologist Pentti Eskola, however, as early as 1904, working as a field assistant, and every year after that not only in Finland, but also abroad. Wilhelm Ramsay was Eskola's teacher in geology. He wanted, according to Eskola, to teach his students independent thinking rather than technical details. When asked for advice, he used to say: »Think it over and then do it so that it is well done.»

In 1914 Eskola won his doctorate with a thesis on the Orijärvi region — a classical work, which will be discussed more on the forthcoming pages.

In 1910 Eskola began his activity as a university teacher, thus inaugurating a distinguished career in this field. His first appointment was as teacher of geology at the department of agricultural economy of Helsinki University. The following year he wrote a basic textbook in geology, which was then used in several revised and expanded editions as the basic textbook in geology courses at Finnish universities continuously up to recently.

October 1, 1913, was a red letter day because on that day he was appointed assistant at the Institute of Geology and Mineralogy, Helsinki University — the institute at which he worked until his retirement in 1953, that is, for forty years. In 1924 he was awarded the chair of extraordinary professor of

geology and mineralogy of the same institute, and in 1928 he succeeded his teacher Ramsay as ordinary professor. This period produced several textbooks in geology and mineralogy. Some of them, being written in German, were adopted as textbooks in many Universities abroad. The most remarkable result of this period of Eskola's activity is the fact that all the contemporary geologists with important geological posts in Finland are his pupils.

Eskola was an excellent teacher, who always took over his students from the very beginning and personally delivered the preliminary lectures and gave out advice. And he closely followed the progress of each of his students. He wrote about them to his friends, among them prominent geologists around the world. As K. J. Neuvonen, one of his students, once observed, no student ever told Eskola thanks and good-bye upon graduating, but then proceeded to cultivate the friendship of their erstwhile professor. In his memorial speech at the funeral of Eskola on December 14, 1964, Väinö Auer said: »Eskola was one of the predestined university men, who could give their time endlessly both to students and to friends, and his receptions usually lasted practically all day.»

It was not only his doctor's thesis that made the year 1914 a memorable one in Eskola's life: That year he also married Mandi Wiiro. He got a wife who brought the life of Pentti Eskola to full flower, and she gave him a son, Matti, in 1916 (killed in the war in 1941) and in 1920 a daughter, Päivätär (now a secondary school teacher of chemistry). She was also of an old farming stock, and like her husband had graduated in the natural sciences, having once, in secondary school, been the present writer's geography teacher. Eskola cherished his wife very deeply. In his response to the presentation of the Penrose Medal in 1951, he wrote: »In the first place I must mention my wife, who has taken part in all kinds of my work and has accompanied me to many lands. Without her help the best part of my work would have remained undone. In especial, she has done much to help and encourage my pupils. In Sweden, she was known by the pet-name Geologmamma, first given her by H. von Eckermann.» Indeed: all the students of Eskola knew Mrs. Eskola well — and she them. Eskola often held final examinations at his home. Such examinations could last long — for hours. A break usually came after a couple of hours of strong thinking in the form of Mrs. Mandi Eskola carrying a coffeepot: »Halloo, boys — isn't it time to relax?» When in 1947 Mrs. Eskola became partially paralyzed and bedridden for a decade (she died in 1959), Pentti Eskola took care of her with such love and devotion, at last being obliged to cope simultaneously with his own health as well, that everybody could not help but admire him also in this respect.

Meanwhile, Eskola also worked with the Geological Survey. From Aug. 1, 1918, to Feb. 1, 1924, he was on its staff as a State geologist his speciality being ore geology and ore prospecting. For this work Eskola also had the

necessary training: in 1907 he had been a student in the Bergakademie, Freiberg, Germany. He spent the time from March 1921 to November 1922 in America, however, mainly working in Washington with the Geophysical Laboratory. His aim was to establish an experimental basis for his mineral facies hypothesis, and he worked on the silicates of barium and strontium. As far as the initial aim of his visit was concerned, he felt disappointed. But at the same time, his belief in the necessity of experimental work was strengthened, exerting a decisive influence upon all his later work. There he met Norman Bowen, and they became lifelong friends, and we, Eskola's students, would often hear from him what progress Bowen had again made in his studies. When the experimental research on hydrated systems under high pressures started, there Eskola's expectations were raised to a high pitch. He hoped that these experiments would finally settle the dispute between the »pontiffs» and the »soaks».

During his stay in America, Eskola also made another contribution, which at the time was just a routine mapping project, but became rather important and topical again some 30 years later. In 1922 he worked for four months with the Geological Survey of Canada in the region north of Lake Huron. From this time dates the geological map of the Blind River area, which Joubin could later find only after some difficulty, and it was this map that he used during his prospecting work there. That led to the discovery of tremendous uranium deposits. »I never had any uranium in my mind there,» said Eskola laughing when the subject of the Blind River was broached to him in Helsinki once in the late 1950's.

The scientific papers of Eskola, before his thesis in 1914, deal with quite a variety of problems. After determining the rock types of neolithic weapons for the National Museum of Finland, he described vesuvianite. Then there followed a concise geological study of the sandstone of Satakunta, SW-Finland. He went on to deal with Quaternary problems connected with the Pöyrisjoki (river), and with the volcanic activity of Martinique and St. Vincent. During his period as State geologist and in succeeding years he published several, mostly popular, papers on ores and prospecting. But at this time he already had begun to cope with several solid problems, some of which gave him work for the rest of his life. These problems will be discussed later on in this paper and in much more detail. Before going into them, however, it may be of special interest to learn something about certain of the non-geological features of Eskola's make-up, which nevertheless were quite characteristic of him.

Going through Eskola's old field notebooks, one is surprised to find a wealth of non-geologic observations in them. His love of Greek and Latin may also explain his keen interest in many linguistic questions. He knew most of the Finnish dialects, and the ones spoken in the parts of the country often

visited by him he could speak well enough to fool the local people into thinking he was one of them. His notebooks had many entries concerning dialects, and his students were often privileged to hear anecdotes told by Eskola in the original dialect.

Another trait typical of Eskola is his extraordinary exactness everywhere and in everything. In the field notes from the summer of 1914, Transbaikalia, there is a note: »I saw a bear, it was at a distance of approx. 12. 5 meters from me.» Contrary to what we might imagine, writing especially as he grew older, did not come very easy to him. He wrote again and again, and often after the corrections and revisions nothing was left of the original manuscript. His daughter, Päivätär Eskola, stressed this matter, and probably this difficulty arose also from his ceaseless searching for exactitude. In 1964, when he was confined to his sickbed and had only a few days more to live, and he could hardly speak, his doctors prescribed a regular daily fluid intake. Eskola then became quite worried about the inexactness of the glasses used for this purpose. And when he tried to hold a glass in a shaking hand himself, he became still more worried: »One should first determine the spilling frequency of the milk and then the constant which could be used as a correlation of the amount given,» he said.

During the last months of his life his love for the Greek and Latin languages burned brighter than ever before. Lying half unconscious in his hospital bed, he spoke at length in these languages, and then translated what he had said to his daughter and to Mrs. Helena Syrjälä-Eskola, his second wife (whom he had married in 1962). She is only about a year his junior, and they were already good friends fifty years ago, back in the golden days of their youth. Helena Eskola had likewise been educated in the natural sciences and she had very much to give Pentti Eskola mentally. She gave him happiness again — but unfortunately only for a short duration.

Eskola's achievements as a scientist have been of such a magnitude that in a small country like Finland few have attained to the same level. As a geologist he must be ranked, also internationally, among the most distinguished representatives of his science. He received honorary doctorates from the following foreign universities: Oslo (1938), Padova (1942), Bonn (1943), Prague (1948). He was awarded the following medals, each of which must be rated a high honor for a geologist: the Gustav-Steinmann-Medaille of the Geologische Vereinigung (10 I 43, Bonn, Germany), the Penrose Medal of the Geological Society of America (8 XI 51, Detroit, USA), the gold Wollaston Medal of the Geological Society of London (30 IV 58, London), the Friedrich Becke-Medaille of Österreichische Mineralogische Gesellschaft (1960, Helsinki), Leopold von Buch-Plakette of Deutsche Geologische Gesellschaft (1962), and the Vetlesen Prize (together with Holmes, in 1964). In addition: in 1963 the Geological Society of Finland established a gold Eskola Medal to be

awarded to geologists who have made notable contributions to the geology of Finland and the Precambrian. The first Eskola Medal was awarded to Pentti Eskola himself, being presented to him in Helsinki on the occasion of his 80th birthday, on January 8, 1963.

He has also been awarded high decorations by the Finnish and Swedish governments.

He has been a member or fellow in numerous geological, mineralogical and other scientific societies and academies round the globe and he was honorary president of the Geological Society of Finland.

On the occasion of his 75 th birthday in 1958, O. Kouvo and Y. Vuorelainen named a new mineral, Cr_2O_3 , from the Outokumpu mine, Eskolaite.

Pentti Eskola is a name that will always have its place in any book dealing with the history of the geological sciences. His scientific contribution to geology embraced numerous problems, mostly of basic importance and of large scope, and his field covered most branches of geology. No exhaustive review of his investigations can be ventured upon here; rather is it the aim of the following pages to consider the most important problems he coped with and what he contributed toward their solution.

MAGNESIA METASOMATISM

Eskola's first major scientific project was his Ph. D. thesis, published in 1914: »On the Petrology of the Orijärvi Region in Southwestern Finland.» This impressive paper has become a classical work on Finnish geology, but for some decades it has been out of print, and nowadays is almost a »geobibliographic» rarity, as far as younger geologists are concerned.

In this paper, Eskola is mainly concerned with the complexity of the geology of the region investigated, including the sulphide mine of Orijärvi. The results of the investigation have been summarized by Eskola as follows: »The rocks of the Orijärvi aureole owe their peculiar character to pneumatolytic agencies which have caused considerable changes in their composition. These changes have, for the greatest part, consisted in a metasomatic replacement of lime, soda and potash by iron oxides and magnesia. These main features are the same in the alteration of the limestone into skarn, of the leptytes or other siliceous rocks into cordierite-anthophyllite rocks, and of the amphibolites into cummingtonite-amphibolites. In some localities the alteration consisted in a considerable enrichment in silica.» And, further »The sulphide as well as the oxide ores originated in the same processes by which the rocks assumed their present habit and composition.» These statements include the basic points of Eskola's hypothesis on magnesia metasomatism. This hypothesis immediately gained wide approval everywhere, but, of

course, especially in Fennoscandia. In Sweden, a couple of years later, Per Geijer was led to the same conclusion in the Falun region, around a very old and famous mining area. Quoting Eskola's own words: »... and some geologists gave it a far wider application than I had ever dreamed of.» Granite was the supposed source of magnesia necessary to produce the respective metasomatism. Barring a few objections, like those submitted by Tilley and Flett in 1930 and by Brögger in 1934 — men who did not believe in the possibility that the residual liquids from a granite very poor in magnesia could bring forth such amounts of it in, for instance, a leptite as to convert it into a really magnesian rock — the hypothesis was soon accepted and introduced into several textbooks as an almost proven theory. It was not until 1950, or 36 years after Eskola first expounded his theory, that two pupils of his, Heikki Tuominen and Toivo Mikkola, brought up really serious objections against the magnesia-metasomatism theory, and, remarkably enough, after re-investigation of the same Orijärvi region, where Eskola came to his conclusions. They suggested that the enrichment in magnesia there may be a result of metamorphic differentiation produced by tectonic deformations — an idea close to that proposed in 1939 by Sven Gavelin for the Malonäs ore district in Sweden. The latter, however, placed the responsibility for the metamorphic differentiation on the intruding sulphide solutions. Eskola's reaction to the paper published by Tuominen and Mikkola came immediately in his answer entitled »Orijärvi Re-interpreted»: »If I feel sceptical concerning some of their conclusions and even believe they are decidedly wrong in certain points, this does not mean that I would deny this first publication by the authors the rank of an important contribution to Archean geology. Their new points of view, stressing the kinematic control of metamorphic differentiation, follows a modern trend in petrology and the study of ore deposits and leads to many interesting working hypotheses which certainly will arouse discussion and may lead to real progress.» Here spoke a grand scientist, who encountered new facts and new ideas with an open mind and was fully prepared, if necessary, to change his own opinions for science's sake. For old Eskola, such an attitude was not uncommon at all. But, basically, Eskola's theory concerning the enrichment of magnesia itself, still stands well.

ECLOGITE THEORY

Throughout his life, primarily as a chemist, Eskola was seriously concerned with reactions in a solid state — both wet and dry — and especially with reactions under high pressures. The idea of rocks composed of heavy minerals and formed at high pressures goes back to 1904, to his student days. At that time, the Belgian chemist Walter Spring made his early experiments

on reactions in the solid state effected by pressure. J. H. van't Hoff discussed the Clausius-Clapeyron equation, which expressed the pressure volume control of reactions in a closed system. These investigations made up the basis of Eskola's work on reactions in crystalline conditions (1906) and earned him the credits to graduate as a chemist from the University of Helsinki.

Later on, in 1913, Sir Lewis Ferrier published a paper on garnet as a geological barometer. It was his conclusion that within the earth's crust there exists a continuous infraplutonic zone where the basic rocks are eclogites.

Including the other heavy minerals, such as jadeite, kyanite, rutile, and diamond, Eskola accepted the eclogite theory, adding, however, that there may be still other heavy minerals that have never been observed, because of release of pressure they are immediately converted into lighter minerals. These questions are dealt with in Eskola's paper on mineral facies (1921), and the ensuing discussion was based on his work together with his good friend V. M. Goldschmidt in Oslo, in addition to the studies on the Norwegian eclogites out in 1919 and 1920.

In 1921 Eskola went to the Geophysical Laboratory in Washington to experiment with synthesizing minerals under high pressures. At that time his experiments failed, however, but the eclogite theory aroused lively discussion. Such petrologists as R. A. Daly, Arthur Holmes and Percy Wagner, the student of kimberlite pipes containing diamonds in addition to eclogite nodules, were ready to accept Eskola's theory.

For many decades the heavy minerals referred to defied all attempts at synthesis, and the belief gained foothold that, owing to the small differences in volume, pressure alone was not sufficient to invert the stability relations, its main effect being to lower the crystallization temperatures by preventing the escape of volatile constituents. In the 1950's, however, also the eclogite minerals could be synthesized. Later on, Francis Birch could say that the Mohorovičić discontinuity in the earth's crust probably marks, not a compositional but a phase boundary, or the upper surface of the eclogite zone, thus realizing the early expectations of Eskola.

MINERAL FACIES THEORY

The eclogite theory is, however, only part of the mineral facies theory of Eskola, which is the most important of his contributions to the classification and study of crystalline rocks. As much as possible, Eskola always endeavoured to apply the principles of physical chemistry to the problems of the origin of

crystalline rocks. His mineral facies theory is the best known result of such an application. In the light of the mineralogical phase rule, Eskola showed clearly that, in a rock which has reached phase equilibrium, the mineral assemblage is determined by the bulk composition, the temperature, and the pressure, whether the rock crystallizes from a magma, recrystallizes from solid rocks under metamorphic conditions, or is deposited from solution.

Eskola himself stated in his response at the presentation of the Penrose Medal to him in Detroit in 1951 that the conception of mineral facies arose from the discussion in 1919—20 with V. M. Goldschmidt, who had first applied the concept of chemical equilibrium to the contact-metamorphic rocks of the Oslo region.

Actually, however, the conception of mineral facies must certainly have been evolving in his mind as early as the presentation of his Ph. D. thesis in 1914, as may be deduced from his *Lectio praecursoria* on this occasion. The next year (1915), in his paper »On the Relations between the Chemical and Mineralogical Composition in the Metamorphic Rocks of the Orijärvi Region», Eskola cited the theories of F. Becke, U. Grubenmann, and C. R. van Hise, mentioned V. M. Goldschmidt, who »by applying this (physico-chemical) principle in a more concise form and making use of the phase-rule», explained the correspondence between the chemical and mineralogical composition of the contact-metamorphic rocks in the Oslo region. He also mentioned P. Niggli's venture in the same direction in dealing with certain Alpine schists. At this time Eskola wrote: »Archean rocks have not hitherto been subject to such treatment. — During the investigation of the petrology of an Archaean area in Southern Finland, the writer was struck by the fact that the mineral paragenesis in the metamorphic rocks seemed to be controlled by simple laws and to stand in definite relation to the chemical composition. — It appeared, in fact, that the problem could be solved satisfactorily and that the Archaean formations have actually reached an almost true chemical equilibrium.» On p. 114 of the paper cited, Eskola defined his conception of metamorphic facies. »In any rock of a metamorphic formation which has arrived at a chemical equilibrium through metamorphism at constant temperature and pressure conditions, the mineral composition is controlled only by the chemical composition.» It is remarkable that before reaching this definition, Eskola had based his discussion chiefly on the saturation diagrams of van't Hoff.

Eskola's theory concerning the mineral facies later became the subject of numerous international discussions and elaborations, and it was then greatly enlarged by Francis J. Turner, especially in his paper of 1948. Since Eskola defined his five original facies (sanidine, hornfels, greenschist, amphibolite and eclogite), a division into subfacies has in some cases been found to be necessary, while several additional facies have also been proposed.

The control of the mineralogical composition of rocks by their chemical composition can now clearly be illustrated by means of triangular composition diagrams. The apices of the diagram represent the three components that mainly account for the observed mineralogical variation within the facies. These components, as used by Eskola, are Al_2O_3 , CaO , and $(\text{Mg,Fe})\text{O}$, and he named the diagrams themselves as triangular ACF diagrams.

The mineral facies classification has the great advantage of elasticity, and may be adopted to meet the needs of descriptive petrography as our knowledge of the equilibrium relations of various mineral associations increases.

In 1951 there appeared a paper by Ivan Rosenqvist, in which he sought to fix the temperature intervals of each mineral facies by using as basis the expulsion temperatures of water from hydrated minerals typical of each facies. It thereby markedly strengthened the importance of Eskola's theory. On the other hand, the experimental work, done mainly in the United States, relating to the actual stability of different minerals has shown that many changes in the facies classification are still to be expected. Especially after the paper by Yoder issued in the Bowen Volume of the American Journal of Science in 1952, Eskola became worried about the future of the classification, at that time accepted, according to his mineral facies theory, and he advised also the present writer to keep the doors open and watch the results of future experimental work along these lines.

The greatest value of Eskola's facies theory is that it stands solidly as the basis for the classification of metamorphic rocks, and that the modern progress of sciences can only improve and modify but not change it.

GRANITE PROBLEM

After returning from America in 1922, Eskola became more and more engrossed in the granite problem, which, with advancing age, became more and more important to him. In early 1940's, after he had lost his son Matti in the war, in 1941, he was very much depressed, and his spirits fall still more in 1944 when the future of Finland during the last year of the war — and after — sometimes looked dismally uncertain. But even at that time the granite problem continued to haunt him, and in a letter to V. M. Goldschmidt, who was at that time in England, Eskola wrote (23 IV 45): »In fact I am seriously pondering the alternative of retiring for ever. During the present conditions in many surroundings it would be far more pleasant for me to devote my remaining years to horticulture and to have the writing of my long-prepared work upon the origin of granites only as a hobby of the holidays!» (As an explanation it may be mentioned here that Eskola had become deeply interested in horticulture and he had an orchard with 450 apple trees).

When Eskola began with granites — some time around 1924 — his scientific attitude was, in his own words (acknowledging speech at the presentation of the Penrose Medal in 1951): »In the great controversy around the granite problem I am a magmatist, perhaps largely because I have seen Bowen at work and, so to say, been soaked with the arguments for the reality and importance of crystallization from liquid magma, contained in the work of the Geophysical Laboratory.» Officially, that is, as he stated himself, he remained a magmatist throughout his life. Especially stiff was his attitude in the late 1940's and early 1950's, as his correspondence with Hans Cloos, of Bonn, reveals. At that time, C. E. Wegmann, representing the »soaks», wrote much in the »Geologisches Rundschau», edited by Cloos. Eskola did not like the situation and wrote about this to Cloos, who agreed with Eskola, but, on the other hand, did not like to bound the freedom of science. Still, on Jan. 3, 1950, he wrote to Eskola: »Wir können uns nicht zu einem einseitigen Propagandaorgan für die Bildung von Granit aus frischer Luft machen lassen» and asked Eskola for a magmatically oriented paper. Eskola's answer was that he was too old and too lazy to engage in scientific polemics here, but he predicted that some day a reaction against the soaks would come, and that it was quite the same whether this reaction come during his lifetime or later. Often, in various connections, he used the names »pontiffs» and »soaks» for different »camps» of petrologists, and poked fun at them through the satirical verse on German philosophers by the Swedish poet Gustaf Fröding (in a very freely translated form):

»What's settled truth in Cambridge and in Oxford
is just a silly joke in Edinburgh.»

His papers on granites do not, however, impress the reader as very strongly identifying Eskola as a »pontiff». I should say that he was only moderately a magmatist — especially when viewed in the light of the papers of his last years.

Eskola's first granite papers deal with rapakivi granite. In 1928 he wrote on the rapakivi rocks at the bottom of the Gulf of Bothnia. The following year he became interested in orthoclase and microcline in the Finnish granites and pegmatites. In this connection he reached a conclusion that was actually far ahead of the time of writing and it did not become really topical again until about 25 or 30 years later: »... orthoclase and microcline may also be almost identical in free energy, and the conversion of one form to the other is not likely to take place when the transformation point is surpassed. The change may be achieved on heating, i.e., microcline may be changed to orthoclase, but a change in the opposite direction can not very well take place on cooling, unless mechanical deformation increases the mobility of the

lattices.» Thus Eskola understood that an additional factor was necessary for the transformation of orthoclase into microcline. He entertained the concept of a deformation, which he still clung to in his paper on the granulites of Lapland in 1952, as the required factor. Later on, when the present author proposed time as such a factor — i.e., sluggish crystallization — Eskola did not contradict this possibility.

Eskola's first — and really important — contribution to the granite problem came in 1932. Entitled »On the Origin of Granite Magmas,» this paper proposed to divide the Finnish granites into three different magmatic types. His starting point was magmatic, but (p. 474): »It is of course difficult and in many cases it will probably always be impossible to tell exactly to what extent truly *juvenile* ichors, derived by squeezing out or by some other way of differentiation from the magma, have been added to the palingenic magma in any particular case.» As a matter of fact, he described all granites accompanied by but small amounts of basic rock types, or those entirely sterile in ore minerals, as being largely or entirely palingenic. In this paper of 1932, Eskola still postulated a juvenile magmatic origin for his synkinematic granites. Regarding late kinematic granites he wrote (p. 477): »... In their composition (they are) close to the final stage of crystallization differentiation (and represent) very uniform potash-rich granites, though frequently containing almandite or cordierite, which are perhaps remains of original sedimentogeneous rocks that have never been entirely dissolved. . . . These granites would seem to be largely palingenic.» It is highly interesting to note that despite Eskola's continuously repeated statements about his being himself strictly magmatically minded, he was up to the last years of his life quite prepared to weigh new ideas, and also to reconsider what he had said in his earlier paper. Thus, one reads with admiration what he wrote on the same two groups of granite two decades later. In 1955, in his »About the Granite Problem and Some Masters of the Study of Granite», p. 129: »Today I must . . . admit that there may be no synkinematic ideal granites of original bulk composition in Finland», and that »among the late kinematic granites, where the typical eutectoid ideal granite composition is common, metasomatic types with preserved ghostly relics or old design of stratification are fairly common.» But: »There are probably always portions, mixed with rocks of the metasomatic mode of origin by replacement, that have had a liquid phase from which the crystallization has taken place in a truly magmatic way.» This concession to the metasomatic school was made when Eskola was already 72 years old! Certainly this change of view was not easy for him, but he accepted it because new observations and theoretical considerations forced him to do so; and that is something which enhances his greatness as a scientist! Still later, Eskola (1960) seemed to have abandoned the name late kinematic using instead the term »serokinematic».

Actually in close connection with his work on granite problems, there arose several other problems, among which the development of the theory on »metamorphic differentiation» warrants first mention. The term »metamorphic differentiation» had been used some years earlier by F. L. Stillwell. Eskola introduced the same term independently, however, but gave credit to Stillwell in 1933. In the light of this principle Eskola explained the formation of veined gneisses. The fact of some of them having been previously designated as injected gneisses could, according to Eskola, be likewise explained on the basis of metamorphic differentiation, together with certain bands or schlieren of eclogite rocks, some of which consisted of garnet, the others again entirely of chloromelanite or jadeite. Eskola also considered possible the extraction and redeposition of the most soluble substances.

As a further development of metamorphic differentiation is the differential anatexis, which Eskola embraced after Goranson's work on the remelting of granites. The results are used by Eskola in explaining certain veined gneisses in migmatites as well as of minor aplitic masses.

To certain extent connected with his granite problem, is the discussion on the domes. In 1941, when reviewing late Erkki Mikkola and his contributions to the geology of Finland, Eskola discussed the dome of Mustio, mapped by Erkki Mikkola, as well. He wrote: »Die Entstehungsweise solcher Tumoren ist noch unklar, am wahrscheinlichsten erscheint mir, dass sie Magmadiapire ohne entblösste Plutonmassen darstellen.» Simultaneously Eskola concluded, however, that in the most parts of Finland, the granitization has led to the formation of microcline-granitic rocks of comparatively constant composition. In 1951, Eskola explained the formation of mantled domes of Pitkäranta as follows: »Granitization under addition of much potash and increase of volume has altered especially the marginal parts and made ancient intrusive masses well up into domes.» Next year he adds: »In domes that have not been granitized, the upheaval may be due to tectonic causes alone,» and that the granitization may be independent of the amount of tectonic movement though contemporaneous with it, as Eskola wrote already in 1949.

GEOLOGY OF FENNOSCANDIA

Notwithstanding the fact that Eskola started his university studies by major in chemistry and also graduated as a chemist, he kept before his mind's eye from the very beginning that goal he had set for himself — that he would be a geologist. Therefore he set out early to do geological field work. In 1904 he worked in northern Finland, the next summer in southern Karelia as well as in southwestern Finland, where he focussed his attention on rapakivi. In 1906 he visited Swedish Lapland and northern Norway. In 1907 he went down

to Germany, but in 1908 he started his investigations of the Orijärvi region, continuing them until 1913. In 1914 he was in Transbaikalia, Russian Asia, searching for radium: but as a result of this field work he published at a much later date however, an excellent petrological description and also a report on his observations bearing on Quaternary geology. After working in 1915 on Finnish limestone occurrences, he undertook in 1916 extensive field investigations in East (Russian) Karelia, which continued into the following year. He made his first visit to this area, however, as early as 1909. He was to return to these investigations 25 years later with the assistance of several pupils who were there during the war in 1942 and 1943, as Finnish army officers. I personally especially remember one such trip to study the volcanic rocks of the so-called Suoju-complex, which Eskola had visited in 1917, too. On this trip, a military escort was put at Eskola's disposal because the area visited was not very safe!

All the experiences gained on his various geological expeditions built up for Eskola an excellent background to promote interest in the general geology of Fennoscandia. Furthermore, in 1924 he became professor of mineralogy and geology. In this capacity, he had to carry on the traditions of his teacher Wilhelm Ramsay. In addition, he had great capacity for discussing large questions. For instance, in his first strictly scientific paper »On Phenomena of Solution in Finnish Limestones and on Sandstone Filling Cavities», issued in 1913, he described the limestone of Illo, containing cracks and cavities in which sandstone had been left behind. On this basis, involving at first glance a simple phenomenon, he was still prepared to draw the following conclusions (p. 50): »There has never been any advanced chemical weathering in the pre-Cambrian rocks of this area during post-Cambrian periods. . . . In post-Cambrian times the area had never been raised to any considerable height above the level of the sea. . . . In post-Cambrian times the area was never covered by very thick beds of sedimentary formations.»

In the field of the regional geology of Finland, during the Eskola's youth, J. J. Sederholm was the strong man of geology. The leading expert on East Karelia was Wilhelm Ramsay. The latter had found there a contact between a schist and an overlying quartzite and explained it as an unconformity between an older, Kalevian, and a younger, Jatulian formation. J. J. Sederholm had accepted this division and applied it to the whole of eastern and northern Finland. Sederholm's division was chiefly based upon the grade of metamorphism. He concurred with L. V. Pirsson, who had postulated that older rocks are more likely to be strongly metamorphized than younger ones. The year 1916 found Eskola demonstrating, however, that the so-called Kalevian schist and Jatulian quartzite were really conformable horizons of one formation and that the schistosity of the former was due only to its incompetent character. Thus he reached conclusions similar to those arrived at a year

earlier by his lifetime-long friend Eero Mäkinen in the western part of central Finland. Eskola concluded that the grade of metamorphism of rocks has nothing to do with their age, and he named the whole formation in question *Karelian*, which was known to rest unconformably upon the older Precambrian. This re-interpretation was accepted by Ramsay, but vigorously rejected by Sederholm. Many hot disputes on this and other Precambrian age problems took place between Eskola and Sederholm around 1920. Finally, probably thanks to C. E. Wegmann's intervening with additional evidence, Sederholm gave up. The term »Karelian» was used for the first time in print in 1921 — appearing in the elementary textbook on geology, mineralogy and crystallography written by Eskola. Later on, the new name for the combined formation was generally and everywhere accepted and adopted. Recalling happenings of 1920—1924, Eskola wrote to Hans Cloos in a letter dated 18 IV 1944: »Ich war wirklich für Sederholm ein'malus spiritus', der ihm plagte.»

In his old age, Eskola was obliged to take part in trying to solve a new puzzle connected with the Precambrian stratigraphy of Fennoscandia, and his »Karelian» became involved again. The rocks of western and southern Finland have always been considered as older than those of eastern and northern Finland, where the Karelian formation mainly occurs. The older ones have been subdivided into Svionian and Bothnian, but in 1909 Ramsay combined them into a single Svecofennian or Svecofennidic folded zone. Thus, in the southern part of Fennoscandia there mainly existed older Svecofennidic and in the north and east younger Karelidic formations and folded zones. When in the 1950's, absolute ages became more and more available, it was found — principally due to the work of Russian geologists, notably Polkanov and Gerling, of Leningrad — that both the Svecofennidic and the Karelidic rocks are of more or less the same age, roughly 1800—1900 m.y. Consequently, both supposed orogenies, foldings, may be of the same age — the same orogeny. Both Russian and Finnish younger geologists became more and more skeptical about the correctness of the division into Svecofennidic and Karelidic, especially because the drawing of boundaries between these two had actually never been successful.

Eskola considered the fresh facts very seriously and brought up the possibility of a rejuvenation of Svecofennidic rocks as the result of a younger orogeny, thereby explaining the discrepancy between absolute ages and geology. Furthermore, he challenged the younger geologists to enter a scientific debate with him, which took place in the pages of the newsletter of the Geological Society of Finland, »Geologi», which is mainly written in Finnish. This took place in 1958. After that he became, perhaps, a bit calmer. To the present writer he said, however: »Your arguments cannot stand. It would be easy to reject them.» Still, when Ahti Simonen, one of his students, too, lectured at Göttingen on March 9, 1962, in the presence of Eskola, and mentioned that the

Svecofennides and Kareliides should probably be bound together, Eskola took the floor in opposition contending that any statements in this matter are premature and that we must first learn more about rejuvenation. At any rate, he later became less rigid in this controversy. Just a year before his death, in 1963, there appeared Eskola's last major work on geology, «The Precambrian of Finland.» It was a work written by a man who had celebrated his 80th birthday. In this paper he gave full credit to the stratigraphy of Finland and also discussed the absolute ages mentioned. He still held reservations concerning the possible regeneration of minerals, thus invalidating the absolute age determinations. He did not categorically deny the conclusions of younger men, but: «The number of radioactive datings of Finnish rocks is still too small to allow any sure age classification of the rock systems.» In this Eskola was perfectly right. Still we, or most of us, believe that Svecofennian and Karelian go together, but as yet we have failed to produce any actual proof. We have only much evidence in favour of the new ideas! The question is still open to dispute. In his very last paper, written for the 70th anniversary of Carl W. Correns, and issued in 1964, Eskola agreed, however, to accept both orogenies as simultaneous: «Eine besonders gewaltsame Revolution war die Karelidisch-Svecofennidische Orogenese um 1800 Mill. Jahre vor unserer Zeit. Sie hat die ganze Fenno-skandien und grosse Areale der russischen Platte betroffen und die Bildung der meisten Granite und anderen Eruptivgesteine veranlasst.»

THE EVOLUTION OF LIFE

Since the 1890's in Finland J. J. Sederholm had been an enthusiastic and victorious advocate of the actualistic principle, while in Sweden more sceptical or antiactualistic views lingered to some extent up to the beginning of the 1930's. Eskola also became interested in this problem, and he wrote in the introduction of his paper «Conditions during the Earliest Geological Times as Indicated by the Archaean Rocks» (1932, p. 7): «Some of the old questions still wait to be answered and new questions have been raised: How far can the actualistic mode of origin be traced back in the Archaean series? Of what kind are the oldest rocks? The physical conditions having not been uniform through the ages, but showing cyclic changes following the alternation of orogenic and anorogenic periods, it may be asked for closer details of the characteristics of the climate witnesses, or the indicators of physiographic conditions during the Archaean periods.» Eskola also asked a physicist, Risto Niini, to make a theoretical study to determine whether the oldest Archaean rocks could have originated either before the condensation of water was completed or immediately after. Eskola also discussed the Precambrian graphite-bearing schists, deducing that they must originally have been suba-

quatic sediments, and that organic life may have existed in the waters from which they were deposited. He further suggested that these schists comprise bituminous shales formed in the same way as recent bituminous muds in deep, closed basins.

The discussion on the conditions prevailing during earliest geological times led Eskola to the problems of the evolution of these conditions and of life. Somewhat earlier, in a paper written in Finnish, Eskola had discussed human life (1931, p. 16): »Not only has the evolution of domesticated animals and cultivated plants been guided along entirely new lines, but even wild organic life is under man's direct or indirect control. In quite a special way the consequences of rational control, whether social or individual, influence human evolution. The evolution of civilized man in modern communities has already deviated from natural life. Man has taken the reins of his evolution into his own hands and he cannot possibly any longer withdraw from responsibility for its direction.» Eskola subsequently turned his attention in many connections to a similar matter: Man is influencing the recent geological processes. He does it like an anarchist disturbing normal evolution.

After that Eskola certainly worked a great deal on these questions but wrote very little, and when he did, it was in Finnish. He did a considerable amount of reading during this period, especially in the late 1940's, and he often remarked that it was necessary to read so much though time was running short. He was searching for his own cosmogony, for the truth of the natural world, his picture of the world. »One has to consider the science of life, as well as biology, biochemistry, the laws of heredity and of evolution, geology and astronomy», he wrote in one of his textbooks in Finnish. Sometimes, however, he was in doubt whether these sciences should be studied by him, a geologist. Should not a shoemaker stick to his shoes? In 1952 Eskola was ready to present the results of his search, and he did it in the form of a lecture series for students. This was the subject of his last lectures at the University of Helsinki, in the capacity of ordinary professor of geology and mineralogy. Then he became emeritus. He repeated these lectures in a revised form in 1954 at the Labor Academy, and in the same year published them as a book in the Finnish language. In it: »I tried to include my conception of the world in the form in which it had crystallized in my mind at the end of the seventh decade of my life», as he wrote in the introduction. (In his later days, his daughter tells, Eskola prepared everything clearly in his mind before he started to write). In my opinion it is really a pity that this book has appeared only in Finnish and Swedish. It includes an excellent review of all the sciences necessary for the construction of such a picture as Eskola was searching for, and the reader is struck by the enormous knowledge Eskola had to absorb before being able to write his book. This is not the place to review his picture of the world. Its characteristic feature is humility, which is possible

only to a scientist who has gone through the whole evolution of his learning, has striven to solve the problems of nature and life basing strictly in the light of the sciences but finally halted at the point where knowledge fails to allow further progress: »The scrutiny of nature does not conjure up anything like a 'miracle' deviating from the common laws, nor any 'life force', even if the expediency and harmonious accomodation of different evolutionary directions are most astonishing, as for instance in the case of the development of the faculty of vision. The fact that the features of organic life, if separated from all the other phenomena, do not obey the law of entrophy can be explained on the basis of the character of both these features and entrophy itself.»

»In any case, we can find guidance in Nature. The Almighty Creator has been able to make his laws fit His purposes.»

Eskola closed his book with the following words: »In the future, science will face tasks involving more and more responsibility and ever greater demands. Controlling the energies of Nature will be a comparatively easy task compared with what the future holds in store, that is when man must learn to govern his own nature.»

THE EPIC OF LIFE

Despite the fact that Eskola maintained his vital scientific activity almost to the very last days of his life, and thereby kept himself strict and exact and always prepared to learn new ideas and experiences — even, if necessary, at the cost of changing his own old opinions — he simultaneously delved into philosophical problems, especially the problems of life itself, and into questions to which the sciences failed to provide answers. Eskola was planning to write a work to be titled »The Epic of Life», but he never got the opportunity to realize the project. In a letter to Dr. A. Streckeisen of Bern, who had written a book about the origin of Man, Eskola observed: »Ganz besonders interessierte mich Ihr Manuskript »Vom Ursprung des Menschen«, weil ich etwa seit 2 Jahren mit einem verwandten Thema beschäftigt bin, aber leider bis jetzt nur einige wenige Seiten geschrieben habe. Als Titel schrieb ich 'The Epic of Life.' Das erste Kapitel soll den Ursprung des Lebens behandeln. Zum primitiven Leben gehört u.a. unser *Corycium aenigmaticum*. Die Lebens Entwicklung, und besonders die Bedeutung des Christentums dafür. Zuletzt die Aufblicke für die Zukunft.» At Eskola's funeral, Toivo Tarkka, the officiating minister, concluded that actually »In Quest of a Picture of the World», the book referred to in the foregoing, was in itself such a work. This is a view all of us probably would more or less agree with.

To trace the development of Eskola's philosophy of life, or to discover how he arrived at this stage in his scientific work, is entirely beyond the ca-

capacity of the present writer. Therefore, just a few glimpses across the path of his life must suffice here. Possibly they can give us some hints on these questions.

In his memorial address, Väinö Auer characterized Eskola at the beginning of our century as »an angry young man, who sharply criticized the teaching system of the University at that time.» This may be applied to his attitude toward his own scientific work as well. It is only natural that he simultaneously criticized the church and its limiting influence upon the natural sciences in the past. He joined those who attempted to explain all natural and supernatural problems in the light of scientific laws. And his wife, Mandi Eskola, also a scientist, shared her husband's views. Yet they did not object when in 1935 their daughter decided to rejoin the church. Eskola was quite tolerant of the convictions of other people.

But Eskola grew continuously as a scientist. He constantly wrestled with problems that failed to obey the laws he knew. His son was killed at the front in 1941, and he was profoundly worried about the destiny of his country. This is reflected by his correspondence in 1944 with Hans Cloos, who vigorously tried to cheer Eskola up and bring him back to a more optimistic outlook. In 1947 his beloved wife was stricken with a serious illness and Eskola's personal burden constantly grew heavier. Perhaps these happenings had a bearing on his turning more and more to philosophic questions, perhaps not. He never opened himself in these matters even to his family. But it is equally possible that he also reached the stage as a scientist where he realized the necessity of seeking something transcending the laws he knew from the sciences.

It is almost impossible to say when he went back to the religion. It was his wife who, confined to bed, first wanted to talk with a minister. On Dec. 21, 1950, Eskola's home received a clergyman Sigfrid Sirenus for the first time. He was an elderly Lutheran pastor, who became very close friend of Eskola's. They had intimate and ceaseless conversations, mostly on religious problems. When already mortally sick, Eskola once said: »One must extend a warm hand to one's neighbour.» Ethical responsibility became a mental habit of his. This aspect of Eskola's character became especially conspicuous to most of the Finnish people through the correspondence, published in two books, Eskola and Sirenus carried on with a young man named Rolf Arnkil, who was bedridden and condemned to an early death. This correspondence started in 1960. Arnkil lived as long as he did only because of his extremely strong will, greatly strengthened by his serious attitude to the problems of life. He also drew strength from the encouraging letters he received from Eskola and Sirenus which also probed the depths of these questions. This exchange of letters continued after the death of Sirenus.

As for Eskola himself, these letters at last laid bare his attitude towards the basic questions of life. There is something fateful and dramatic in the

circumstance that Rolf Arnkil died only two days after Eskola, and the last letter he could read was also the last letter dictated by Eskola to his daughter. It contained only two words: »Dear Arnkil». The end of Eskola's life marked the end of the career of a great scientist. He passed away on the 6th of December, 1964, Finland's Independence Day.

The most suitable words to close this memorial, so as simultaneously to give full credit to Eskola as a scientist and as a man, are those expressed by Toivo Tarkka at his funeral service: »It is common for a great man to speak either to a large mass of people or around the globe, but he does it only by confining himself to his own special subjects. Today many scientists from all parts of the world follow Pentti Eskola in their minds to his last resting place. But that he should also be followed by a mass of people who know very little about what Eskola had done as a scientist is much less common, but they remember full of thanksgiving his sincere strivings to find the truth behind the basic issues of life.»

Vladi Marmo

PENTTI ESKOLA

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In some cases the original version of a scientific contribution — paper or lecture — was written in Finnish or Swedish but published later in a major language. In the bibliography only the last-mentioned contribution is listed and a reference to the original version is given in brackets. Those Finnish or Swedish contributions that have no equivalents in a major language are included in the list.

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Marjatta Okko

1

ON THE SCHISTS AND ASSOCIATED INTRUSIVE ROCKS OF THE
VIEREMÄ-KIURUVESI REGION ¹⁾

OBSERVATIONS AND CONCLUSIONS

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ABSTRACT

The area of Vieremä-Kiuruvesi is situated in the border zone between the Karelides and Svecofennides in the North Savo region of Finland. The following rocks have been described as occurring in the region: basal conglomerates, mylonites, quartzites with mica schist intercalations, various schists containing calcic minerals (biotite-plagioclase schist, biotite-plagioclase schist containing garnet porphyroblasts, biotite-plagioclase schist containing hornblende porphyroblasts, biotite-epidote-plagioclase schist containing hornblende porphyroblasts and quartz-bearing amphibolite), oligomictic conglomerate, staurolite mica schist, pseudo-staurolite porphyroblastic mica schist, so-called Kukkomäki conglomerate, mica gneiss, biotite-plagioclase gneiss, hornblende gneiss, cummingtonite-bearing hornblende gneiss, microcline-plagioclase gneiss, graphite-bearing schist, garnet-bearing pyroxene amphibolite, diopside amphibolite, pyroxene-bearing porphyritic granite, pyroxene-hornblende granodiorite, biotite granite, aplite and pegmatite. Nine chemical analyses of the rocks were made, in addition to many analyses of their modal mineral composition and of the optical properties of the minerals. Furthermore, observations concerning the mutual relations of the rocks are presented, opinions are expressed on the genesis of some of the rocks, and a few of the petrological and geological questions bearing on the subject of this paper are discussed.

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INTRODUCTION

The Vieremä-Kiuruvesi schist belt is situated in the North Savo region of Finland. It lies one to two degrees east of Helsinki and on the southern side of the 64 th parallel N. lat. It includes the western part of the commune of Vieremä and the eastern part of the commune of Kiuruvesi.

The author began his geological investigations in this area in the summer of 1957 in response to advice and urging from Professor Pentti Eskola. The field work in North Savo was carried out during the summers of 1957—1962, with the exception of the latter part of the summer of 1960. In the spring of 1958 the Exploration Department of the Outokumpu Company began ore prospecting operations in the commune of Kiuruvesi, and the author joined the geological staff of the company at that time; and, as a consequence of this action, his original research program could be enlarged. The area to be investigated finally included the whole of the commune of Kiuruvesi, the western part of the commune of Vieremä, the southern part of Pyhäntä commune, the southeastern part of Käräsämäki commune and the eastern part of the Pyhäjärvi commune. The area investigated contains a great variety of rocks and exhibits a rich variety of geological phenomena. The expansion of the research program proved in itself to be a good thing, although it did delay the publication of the results. The assistance of the Outokumpu Company has been invaluable. I have had at my beck and call, for example, boys enrolled in the trade school run by the Outokumpu Company, and they have saved me much leg work by seeking and finding outcrops for me. I know that I have benefited much more from my association with the Exploration Department of the Outokumpu Company than it has from my services. As basic reference maps in my own mapping work, I have been able to use so-called »instant» air-photo maps of the entire territory on a scale of 1 : 20 000 and aerial photographs on a scale of 1 : 10 000 of a few of the important target localities.

Figure 1 shows two sketch maps of the Vieremä-Kiuruvesi schist belt and the intrusive rocks belonging to it. One of the sketch maps has been drawn after the map published by Mäkinen (1916) and the other after the one published by Wilkman (1929). The marks identifying the rocks on the maps have been made uniform upon being drawn so that the sketch maps might be easier to compare, one with the other.

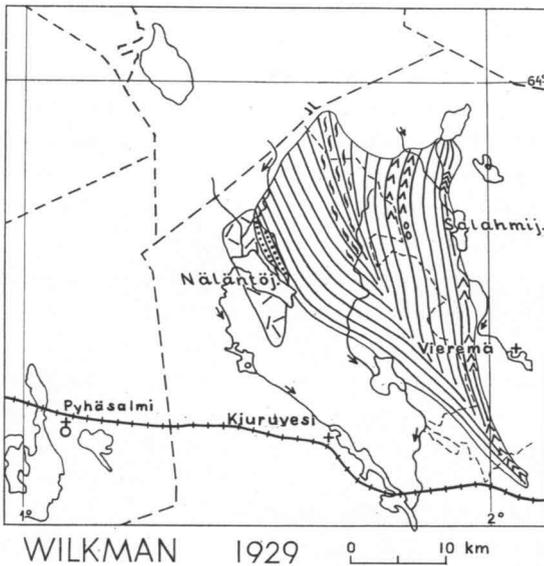
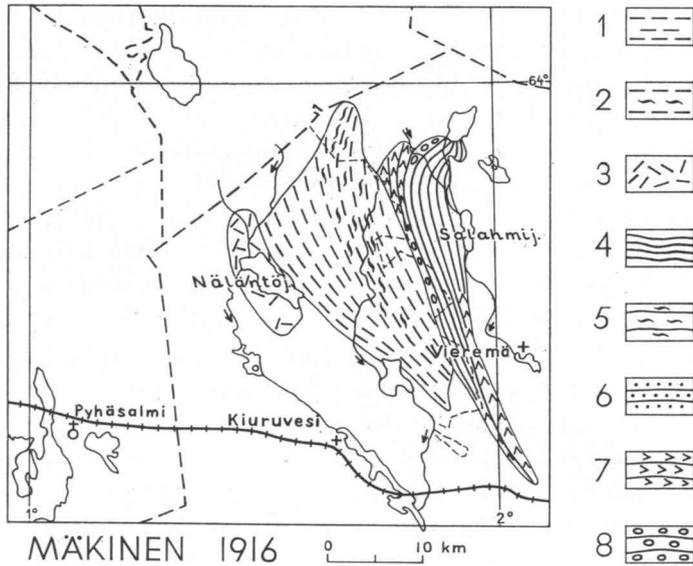


Fig. 1. Sketch maps of the so-called Vieremä-Kiuruvesi schist belt drawn after maps published by Mäkinen (1916) and Wilkman (1929) and reduced to the same scale. 1) Bothnian plagioclase gneiss and mica schist. 2) Bothnian vein gneiss. 3) Post-Bothnian intrusive rock. 4) Kalevalian mica schist. 5) Kalevalian vein gneiss. 6) Kalevalian arkose and quartzite. 7) Kalevalian arkose and augen schist. 8) Kalevalian conglomerate.

According to Mäkinen (1916), the Vieremä-Kiuruvesi schist belt is divided into two parts, which belong to mountain chains of different ages. The part on the side of Näläntöjärvi belongs, according to Mäkinen (1916), among Bothnian mica schists and vein gneisses, which are regarded as included in the Svecofennides, and the part on the Salahmijärvi side among Kalevalian ¹) schists, augen gneisses and conglomerates. It is Mäkinen's view (1916) that the first-mentioned rocks are cut by post-Bothnian granite. According to Wilkman (1929), the entire schist area must be classified among Kalevalian schists, which are younger than the post-Bothnian granite on the northeastern shore of Näläntöjärvi and which have been deposited on top of this granite.

The Kalevalian schists are regarded as belonging among the Karelides. Väyrynen (1954, p. 154) treats the Vieremä-Kiuruvesi schist area in his book as a single formation, joining in this respect Wilkman (1931). He draws a parallel (Väyrynen 1954, p. 153) between this area and the Karelian schists but he points out that these sedimentary formations do not belong among Jatulian types. In the petrographic map of Finland drawn by Simonen (1960 a), the main part of the Vieremä-Kiuruvesi schist belt is marked as phyllite and mica schist and the part Wilkman (1929) designated as arkose and quartzite is marked as quartzite.

On the edge of the schist area on the Näläntöjärvi side, an intrusive rock area has been marked at the point representing Näläntöjärvi (Fig. 1). According to Wilkman (1931, pp. 141 and 142), it contains very many rocks differing in composition and structure: microcline granite, coarse-grained, pyroxene-bearing porphyritic granite, even-grained pyroxene granodiorite, pyroxene granite and potash-natronsyenite.

The present study has, of course, been affected by the literary outpourings in Finland concerning the question as to whether the Svecofennides and the Karelides are of the same age or not. Among papers that either endorse the view that only one mountain chain has existed or argue that there is insufficient evidence to prove the existence of two mountain chains, one might cite the ones by the following: Metzger (1959), Simonen (1960, 1960 a), Schmidt (1960), Mikkola (1961, p. 52), Saksela (1961), Hofmann (1962), Wetherhill *et al.* (1962) and Simonen (1963). The question has not, however, yet been definitively decided, for studies continue to appear in which the existence of two mountain chains is defended. Wegmann (1961), for example, has defended the old view on tectonic grounds, and Eskola (1963) in his extensive work on the Precambrian in Finland has taken an unqualified stand in support of the old view.

¹ The form »Kalevalian» is here used advisedly, instead of the traditional form, »Kalevian», which is an erroneous translation of the original Finnish term *Kalevalainen*. (»Kalevian» would correspond to the Finnish *Kalevainen*, which is non-existent as a geological term.) — The Author.

One purpose of the present study is to present observations made in the schist area of Kiuruvesi and to describe the rocks occurring there. Another purpose is to describe the intrusive rocks occurring at the western edge of the schist area, shed light on their mutual relations and to weigh their age sequence and possible differentiation. Since the schist belt of Vieremä-Kiuruvesi is situated in the border zone between the Karelides and the Svecofenides, certain of the observations included in this investigation are also likely to add pertinent data to the material required to resolve this timely question.

THE SALAHMI REGION

The region of Salahmi discussed in this chapter (Plate I) comprises the eastern margin of the so-called Vieremä-Kiuruvesi schist belt (Fig. 1), with the exception of its extreme south end. The margin of the schist belt runs nearly north-south where it lies within the boundaries of Salahmijärvi. The schist formation is situated west of the margin and the basement complex on the eastern side.

Almost all the schists described here have previously been mentioned as occurring in the Salahmi region (Mäkinen 1916 and Wilkman 1929). The basal conglomerate considered to be best is the one at Kukkomäki, which lies 4.5 km to the east from the center of Salahmijärvi (Fig. 1). The eastern edge of the schist area has been mentioned as having augen schists and arkose schists, and Mäkinen speaks of having found one conglomerate boulder. Mention has further been made of staurolite-mica schists and the occurrence at Metelinmäki, a locality in the Salahmi region, of metabasites (*cf.* p. 18).

BASAL CONGLOMERATE

VALKEISKYLÄ

At Valkeiskylä, in cuts through the bedrock on the new highway from Vieremä to Konolanmäki, a conglomerate is visible for a distance of roughly 350 meters about 4.5 km from the crossroads. The highway intersects the foliation of the conglomerate at an angle of about 20°. The foliation is N25°W, 45°W and the lineation N25°E, 30°S.

This conglomerate contains a rather sparse occurrence of pebbles 0.5 to 5 cm long, the light gray color of which makes them distinguishable in the dark gray matrix. In many cases the pebbles are angular and, to some extent, elongated; and they consist of quartzite and arkosic schists, graphic feldspar, granite and limestone. Potash feldspar porphyroblasts also occur in the form of pebbles with an abundant content of perthite and plagioclase grains, which include sericite as an alteration product. Some of the pebbles appear elongated and among them are broken ones, too.

The grain size of the matrix of the conglomerate is 0.1 to 0.04 mm. It contains over 50 % by volume biotite ($\gamma = 1.643 =$ greenish brown) as well as plagioclase (An_{20}), potash feldspar, quartz, calcite ($\omega = 1.663$), chlorite, (sphene, apatite and opaque minerals). The calcite is frequently observed to occur as narrow fissure veins, which in many cases begin from a limestone pebble. The apatite occurs as long, thin prisms in spots rich in biotite.

WEST OF VIEREMÄ CHURCH

Wilkman (1931, p. 162) in his early day reported that at a distance of some four kilometers to the west of the Vieremä church there occurs some augen schist. During new mapping operations in these parts, four different exposures of clearly stratified conglomerate were found. The structure and composition of the different layers of the conglomerate vary considerably. The strike of the stratification is approximately $N30^{\circ}W$ and the dip is steep. The foliation is NS and vertical. The thickness of the conglomerate formation at right angles to the foliation is over 50 m. See Hausen (1930, p. 42).

The conglomerate contains pebbles consisting of gray, smallgrained gneiss, dark gray, medium-grained rock of dioritic composition, small-grained, light brownish gray granite, graphic granite, quartz, feldspar-bearing pegmatite, vein quartz, quartzite, arkosic schist and phyllite. The size of the pebbles varies from less than 1 cm to over 15 cm. They are mostly subrounded and in many instances elliptical; the phyllite alone occurs as angular fragments and also as long, narrow strips. The pebbles have become somewhat elongated during the process of metamorphosis.

The matrix of the conglomerate contains quartz, feldspar, biotite, calcite, (epidote, hornblende, sphene and a small amount of opaque minerals). The relative amounts of minerals in the matrix vary. Some spots are rich in mica. Calcite occurs in it as veins of microscopic size, while the epidote, sphene and opaque minerals occur as separate grain clusters.

Four different layers can be distinguished in this conglomerate. First, there are layers containing pebbles of quite different sizes and varying composition, phyllite fragments. Second, there are layers in which the matrix consists of arkosic schist of a medium-sized grain and in which there is a sparse scattering of roundish pebbles measuring between two and five centimeters in length. Next, there occurs a grading over into a conglomerate containing a profusion of 2 to 5 cm pebbles; the transition occurs within a span of a couple of centimeters. At the same time there appear strips of phyllite between five and fifteen centimeters long and 0.5 to two centimeters wide. Fig. 2 shows this type. Fourth, the conglomerate contains layers from 20 to 30 cm wide with an abundance of phyllite fragments (Fig. 3).

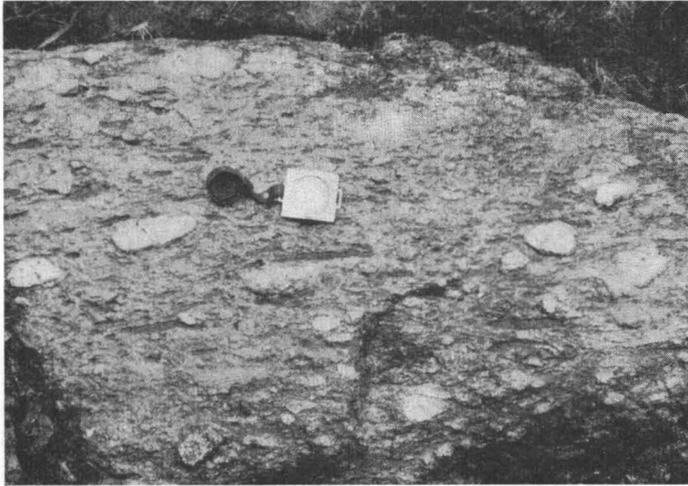


Fig. 2. Basal conglomerate. Some 4 km west of the church at Vieremä.

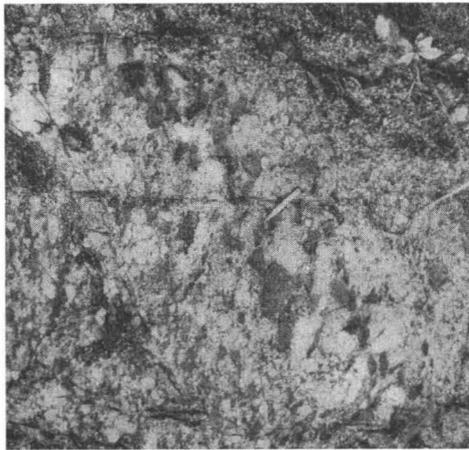


Fig. 3. Layer of basal conglomerate rich in phyllite fragments. Approximately 4 km west of the church at Vieremä.

The phyllite fragments contains about 70 % by volume biotite ($\gamma = 1.652$ = greenish brown), calcite, quartz, (muscovite, feldspar, sphene, epidote, zircon and opaque minerals).

The western part of the conglomerate includes tectonic rocks resembling eye-shaped schist. It would be hard to recognize the rocks as conglomerates in the absence of proximate conglomerates just described.

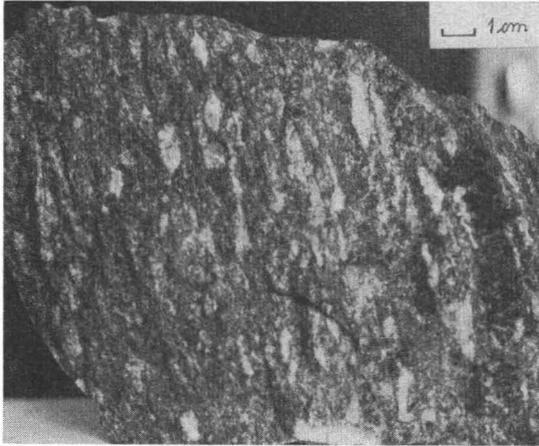


Fig. 4. Basal conglomerate resembling augen schist. Murennusmäki, Salahmi, Vieremä.

MURENNUSMÄKI

Alongside the highway at the eastern edge of Murennusmäki, in the village of Salahmi, basal conglomerate is slightly exposed. Here the conglomerate stratum as a whole is thinner than in the two previously described places. Conglomerate beds a few dozen centimeters thick alternate in it with quartzite layers of approximately the same thickness. The alternating layers occur irregularly, in wedge-like fashion. Tectonic movements were factors in their genesis. Whether any crossbedded features occur in the conglomerate is a question to which one would hesitate to give an answer.

The pebbles in the conglomerate are 0.5 to 2.5 cm long and between 0.5 and 1 cm wide. They are in most cases eye-shaped (Fig. 4) and light gray in color. Viewed under the microscope, the pebbles are rounded or subrounded in shape. They consist of quartzite, arkosic schist, quartz, potash feldspar grains and sericitized and epidotized plagioclase grains. There also occur tiny phyllite fragments.

On the weathering surface the matrix of the conglomerate is yellowish gray and on the fresh surface dark gray. Its structure is blastoclastic and the size of the grain 0.02—0.1 mm. The minerals contained in the matrix are biotite, quartz, feldspar, (muscovite, chlorite, epidote and opaques).

LÄHDEMÄKI

On the eastern and southeastern slopes of Lähdemäki in the village of Nissilä, in the commune of Vieremä, about 200 m from the road leading

from Salahmi to Nissilä toward the west, there is an occurrence of conglomerate containing very small pebbles and consisting in places almost of quartzite. To be sure, in these quartzitic types there are in many cases in addition to the cementing material two clearly distinguishable grain sizes, representing quartz in particular, so that some parts of this rock might also be designated as conglomerate. In a couple of spots in this rock, cross-bedding has been observed. Compared to the conglomerate of Murennusmäki, this rock is poorer in feldspar. The pebbles embedded in it, furthermore, do not represent varieties of rock to any appreciable extent but are composed of individual minerals. The rock is, in addition, mylonitized in many spots. See Hausen (1930, p. 99).

The pebbles range in length between 1 and 7.5 mm, and in mylonitized types they generally appear to be eye-shaped. They are composed of quartz, microcline and plagioclase. Very tiny phyllite fragments also occur in them. Eye-shaped microcline grains appear in the mylonitic types more often than in the other ones, but they suggest porphyroblasts in many instances. Here and there they contain as relicts plagioclase inclusions, the edges of which are no more albitic than the middle portions (Rogers, 1961). In the mylonitic types the large feldspar grains exhibit the effects of rolling and are to some extent broken. Moreover, all the quartz grains have become elongated and twisted, and they occur as curved lenses with a conspicuous undulatory extinction and with a ring of tiny, broken quartz particles encircling them.

The matrix contains quartz, feldspar, biotite, muscovite, (epidote, chlorite, opaques and a slight amount of carbonate). Carbonate also occurs as narrow veins in fissures — *e.g.*, in the fissures of potash feldspar porphyroblasts. Prehnite also occurs as fissure filling. In mylonitized varieties the main portion of the matrix consists of completely re-crystallized quartz and muscovite ($\alpha = 1.566$, $\beta = 1.606$, $\gamma = 1.619$).

BASEMENT COMPLEX

This chapter may surprise the reader by its placing, but the reason is that it concerns only the basement situated immediately to the east of the basal conglomerate.

The strike of the foliation of the basement complex is nearly north-south. Its dip right next to the conglomerate is steep, just as is that of the conglomerate itself. Farther away the dip varies and is frequently gentle. The fold axis is a gentle one, approximately 10° N. On the eastern shore of Salahmijärvi there is a peninsula with a very beautiful, gently dipping fold of large dimensions and very much worth seeing.

According to Simonen (1960 and 1963), the basement complex in eastern Finland contains amphibolites, hornblende gneisses, mica schists, mica gneisses, quartz-feldspar gneisses, gneissose quartz diorites, granodiorites with a cataclastic texture, striped gneisses and migmatites. Eskola (1963) mentions further, as belonging among pre-Karelian cratons, the occurrence of ultramafics, gabbros and potash granites. Both investigators, Simonen (1960) and Eskola (1963), list a number of works dealing with the matter. Suffice it in this connection to cite only a few: Frosterus and Wilkman, 1916; Väyrynen, 1928 and 1933; Preston, 1954 and Matisto, 1958.

The basement complex situated on the eastern side of the basal conglomerate of the Salahmi area contains all the rocks mentioned in the foregoing. Wilkman (1931) describes striped granite gneisses occurring on the eastern side of Salahmijärvi. Without delving deeper into the nature of these rocks, in the present connection, the author nevertheless wishes to point out that some of the outcrops contain basic volcanics, too.

THE MYLONITES OF KARHUMÄKI

On the southern slope of Karhumäki, about 1.5 km southwest of Nissilä, there is a zone of strongly mylonitized rock running N50°E. Also rocks of the basal formation have been observed to occur in it as mylonites. Some of them have become mylonitized to a degree where their minerals have for the most part turned to quartz, feldspar, sericite and chlorite. The structure of such types is augen-gneissoid and totally mylonitic. A portion of these mylonites resemble the mylonitized conglomerates found at Lähdemäki. Evidently, this mylonite zone contains, in addition to the rocks of the mylonitized basal formation, schists and also conglomerates representing the southern side of the schist belt.

THE ALTERNATING FELDSPAR-BEARING QUARTZITES AND MICA SCHISTS OF MURENNUSMÄKI

On the eastern slope of Murennusmäki in the village of Salahmi, the basal conglomerate is overlain by a bed about 200 m thick of feldspar-bearing quartzite and feldspar-bearing mica schist, which generally alternate in layers 0.5 to 5 cm thick (Fig. 5). In addition, the quartzite and the mica schist are stratified: in the quartzite there occur laminae resembling conglomerate and in the mica schist quartzite laminae and quartz-grain lenses. See Hausen (1930, p. 14).

The strike of the stratification and the foliation is N15°W and the dip is vertical. In addition, the mica schist layers are marked by the occurrence of



Fig. 5. Alternating feldspar-bearing quartzite and feldspar-bearing mica schist. The mica schist shows transverse foliation. Murennusmäki, Salahmi, Vieremä.

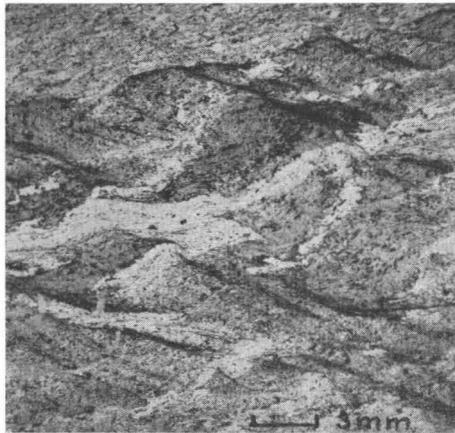


Fig. 6. Feldspar-bearing mica schist, with transverse foliation. Murennusmäki, Salahmi, Vieremä.

transverse foliation in several different directions (*e.g.*, $N35^{\circ}W$, $60^{\circ}W$ and $N10^{\circ}E$, $70^{\circ}E$). Associated with the transverse foliation is pseudobedding (Figs. 5 and 6). The basal direction of the formation is easterly, and it has been measured from the graded bedding (Fig. 7 and 8). No cross-bedding has

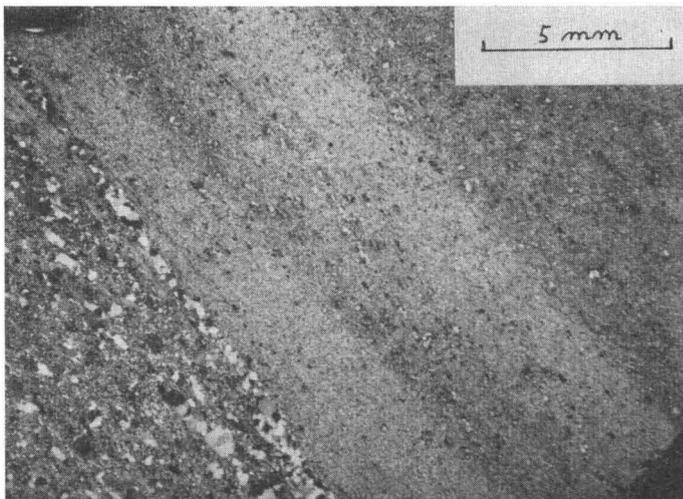


Fig. 7. Feldspar-bearing quartzite. Figure shows basal direction. East is situated on the right. Murennusmäki, Salahmi, Vieremä.

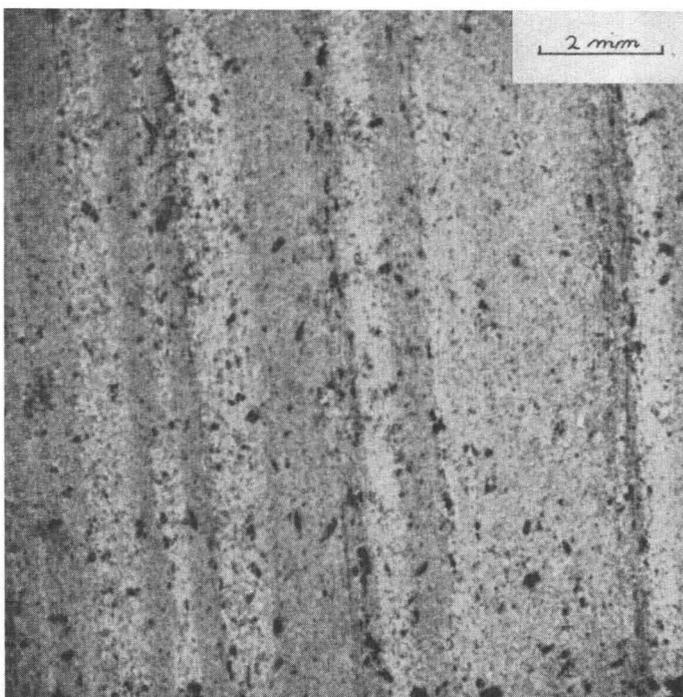


Fig. 8. Mica schist revealing basal trend. East is situated on the left. Murennusmäki, Salahmi, Kiuruvesi.

here been indisputably observed. Spots seemingly possessing this feature have tended to prove to be of tectonic origin. The fold axis here is 15°S , and the lineation, as measured from the longitudinal direction of the minerals, $\text{N}15^{\circ}\text{W}$, 65°S . The pyrrhotite in especial occurs on the black foliation surface of the mica schist as elongated grains as much as 2-mm long and with a metallic luster. It is lamellated in such a way that the schist lineation runs parallel to the intergrowth plane of the pyrrhotite lamellae.

The structure of both the quartzite and the mica schist is blastoclastic. The grain size of the minerals in the quartzite varies between 0.04 and 0.2 mm, while in varieties resembling conglomerate the range is 1.2—0.1 mm and the color light gray. The grain size of the minerals in the mica schist is 0.01 to 0.1 mm and in the quartzitic layers 0.04 to 0.5 mm and the color dark gray. The cement in the quartzite consists of a fine-grained sericite and feldspar mass as well as muscovite, biotite, quartz and pyrrhotite, which occurs as uniform meshes measuring as much as 1 mm in diameter around the grains. Both rocks contain by and large the same minerals, namely, quartz, biotite, muscovite, plagioclase, chlorite, potash feldspar, sphene, garnet, tourmaline, epidote, pyrrhotite, hematite, graphite, goethite, hydro-pyrrhotite and stoichiometric FeS, which occurs as inclusions in the pyrrhotite.

Present in both rocks are anhedral chlorite porphyroblasts, the grain size of which is 0.5 to 1 mm and which do not run parallel to the foliation plane. In spots the potash feldspar and the plagioclase also occur as porphyroblasts. The biotite flakes ($\gamma = 1.646 = \text{brown}$) generally run parallel to the plane of the foliation. Some biotite is also to be noticed in the quartzites. There is a greater amount of muscovite ($\alpha = 1.560$, $\beta = 1.594$, $\gamma = 1.600$) in the quartzites than in the mica schists. It likewise occurs in profusion in the plagioclase as an alteration product. The muscovite flakes do not always run parallel to the foliation plane, either. Apatite is present in greater amounts and as larger grains in the conglomerate-like laminae of the quartzites than elsewhere. The tourmaline, again, favors mica schist; in many places, it occurs in small amounts, but in other spots it forms groups of crystals, including an abundance of prisms a couple of mm long. Garnet is generally met with only in the proximity of lenses rich in calcium minerals.

Meinel (1963) has studied the mode of occurrence, form, internal structure and orientation of the chlorite porphyroblasts contained in the rocks of the Thyninger schist mountain range. The indices of refraction do not, perhaps, suffice to identify the muscovite positively; also the d-values would have to be known (Deer *et. al.*, 1962).

Table 1 presents a chemical analysis of the mica schist and Table 2 one of the quartzite. They do not correspond to typical mica schist or quartzite. Apparently, they contain an overabundance of feldspar, and the quartzites also micas.

Table 1. Chemical composition of the feldspar-bearing mica schist. Murennusmäki, Salahmi, Vieremä. Analyst: A. Heikkinen.

	Weight	Weight		Molecular		Niggli	
	per cent	norm		norm		values	
SiO ₂	63.94	or	23.4	Or	24.5	si	256
TiO ₂	0.85	ab	13.1	Ab	14.5	al	39
Al ₂ O ₃	16.30	an	2.2	An	2.0	fm	41
Fe ₂ O ₃	0.39	Q	29.6	Q	28.6	c	4
FeO	6.17	Σ	68.3	Σ	69.6	alk	16
MnO	0.06					ti	3
MgO	3.20	en	7.9	En	9.2	h	28
CaO	1.03	fs	9.8	Fs	8.6	k	0.64
Na ₂ O	1.56	C	8.7	C	10.0	mg	0.46
K ₂ O	3.93	il	1.7	Il	1.2	o	0.02
P ₂ O ₅	0.09	mt	0.5	Mt	0.4	c/fm	0.11
CO ₂	0.22	ap	0.3	Ap	0.5	qz	+92
H ₂ O+	2.13	cc	0.5	Cc	0.6		
H ₂ O-	0.03	fr	0.2				
F	0.03						
Cl	0.01						
Σ	99.94	Σ	97.9	Σ	100.1		
-O	0.02						
Σ	99.92						

Table 2. Chemical composition of feldspar-bearing quartzite. Murennusmäki, Salahmi, Vieremä. Analyst: A. Heikkinen.

	Weight	Weight		Molecular		Niggli	
	per cent	norm		norm		values	
SiO ₂	62.55	or	18.4	Or	18.5	si	250
TiO ₂	0.82	ab	32.5	Ab	35.0	al	49
Al ₂ O ₃	20.90	an	15.3	An	14.0	fm	15
Fe ₂ O ₃	0.57	Q	19.3	Q	19.1	c	13
FeO	2.16	Σ	85.5	Σ	86.6	alk	23
MnO	0.06					ti	2
MgO	0.91	en	2.3	En	2.6	h	20
CaO	3.06	fs	2.2	Fs	1.4	k	0.35
Na ₂ O	3.86	C	5.6	C	6.9	mg	0.37
K ₂ O	3.10	il	1.5	Il	1.2	o	0.13
P ₂ O ₅	0.20	mt	0.9	Mt	1.0	c/fm	0.89
CO ₂	0.06	ap	0.3	Ap	0.3	qz	+58
H ₂ O+	1.48	cc	0.1	Cc	0.2		
H ₂ O-	0.04						
Σ	99.77	Σ	98.4	Σ	100.2		

OTHER ALTERNATING FELDSPAR-BEARING MICA SCHISTS AND QUARTZITES

Overlying the basal conglomerate situated west of the Vieremä church, there is a formation containing thin, alternating layers of feldspar-bearing mica schists and quartzites; it is thicker than the corresponding formation at Murennusmäki. However, it contains more mica schists than the latter.

Situated between the basal conglomerate of Lähdemäki and the oligomictic conglomerate between Rotimojoki and Luvejoki, there is a formation about 1 km thick, in which very thin mica schist and quartzite layers alternate. This formation is the least metamorphosed in the entire Salahmi area, and the elastic structure of its rocks is still clearly distinguishable. The schists of this formation contain considerably less feldspars than is the case at Murennusmäki — and more mica schists. There are very few quartzite layers and they are extremely thin. These schists in many instances contain a scattering of quartz grains of conspicuously larger size than the minerals in general found in them. The biotite also occurs in places in this way.

SCHISTS CONTAINING CALCIC MINERALS

METELINMÄKI

At Metelinmäki, which is situated immediately west of Murennusmäki, there is a layer about 500 m thick containing schists containing extremely fine-grained, calcic minerals and alternating as thin layers. The thickness of the different schist layers varies from a centimeter to a few meters. On weathered surfaces the layers of finest grain appear to be thinly laminated, whereas on fresh surfaces and in thin section this feature frequently can hardly be detected. In the schists of Metelinmäki there are further some thin quartzite and mica-schist intercalations, particularly in the eastern part of the formation; but mica schists are to be found also at its western edge.

The following types can be distinguished in the schists of Metelinmäki:

- 1) Biotite-plagioclase schist,
- 2) Biotite-plagioclase schist with garnet porphyroblasts,
- 3) Biotite-plagioclase schist with hornblende porphyroblasts,
- 4) Biotite-epidote-plagioclase schist with hornblende porphyroblasts, and
- 5) Quartz-bearing amphibolite.

The hornblende-rich types occur most commonly in the western part of the schist formation, while the garnet-bearing types are more prevalent in the eastern part than elsewhere. The transition between the different types occurs both abruptly and gradually. Types two and three include varieties in which there are both garnet and hornblende porphyroblasts. It was

exceedingly difficult to clear the schists of Metelinmäki completely of lichens and mosses, a circumstance that hampered determining the layer sequence of the various schist types and their situation in relation to each other as well as studying the texture of the schists.

All the rocks included among these schists are full of slickensides and mylonitized zones, along both of which movements have taken place. In addition to the foliation, the directions of movement have been the following, among others: N45°, 60°E, N20°E, vertical and N50°W, 85°W. Movements occurred also after the formation of porphyroblasts, judging by the fact that garnet and hornblende porphyroblasts have broken and their parts shifted in relation to each other. In places the rocks exhibit a beautiful brecciated fabric resulting from the movements, one best seen in thin section.

Various zones of movement can be distinguished even in the same rock through its mineral composition. In the same rock there have been observed shear cleavages formed from partially intergrown quartz grains with almost the identical orientation, biotite-rich, extremely thin shear cleavages intersecting the foliation, and shear zones containing chlorite, epidote and a very fine-grained quartz-potash feldspar mass. Moreover, zones containing principally chlorite and sphene have been noted, as well as shear zones a few centimeters wide containing fine grains of hornblende, epidote and chlorite.

The color of the biotite-plagioclase schist is a brownish black gray, the grain size 0.01—0.3 mm and the fabric blastoclastic. The essential minerals are biotite, plagioclase (An₄₅). Accessory minerals include quartz, potash feldspar, chlorite (partly biotite undergoing alteration into chlorite), epidote, sphene, apatite and, to a slight degree, opaques. The apatite occurs in spots as large, idiomorphic crystals.

The grain size of the biotite-plagioclase schist containing garnet porphyroblasts is 0.02—0.15 mm, the fabric blastoclastic and the color dark gray. In places the schist is laminated, for it is likely to contain chlorite-bearing layers. The reddish-brown garnet porphyroblasts ($n = 1.795$) are idiomorphic and their grain size is 0.2 to 3 mm. In addition, there occur a few chlorite and epidote porphyroblasts here and there. Quartz is present in it as clusters of eye-shaped grains between 0.4 and 0.6 mm long. The minerals contained in the matrix of the schist are biotite ($\gamma = 1.643 =$ dark brown), plagioclase, quartz, chlorite, epidote, apatite, zircon and opaques. In a few types there is a little carbonate. A chemical analysis of the carbonate-free type is presented (Table 3).

The grain size of the biotite-plagioclase schist containing hornblende porphyroblasts is 0.01—0.1 mm, the fabric blastoclastic and the color dark gray. The hornblende porphyroblasts in it

Table 3. Chemical composition of biotite-plagioclase schist with garnet porphyroblasts. Metelinmäki, Salahmi, Vieremä. Analyst: A. Heikkinen.

	Weight		Molecular		Niggli		
	per cent	norm	norm	norm	values		
SiO ₂	56.88				si	176	
TiO ₂	0.79	or	19.5	Or	20.0	al	33
Al ₂ O ₃	18.05	ab	25.7	Ab	28.0	c	16
Fe ₂ O ₃	1.44	an	22.8	An	23.5	alk	16
FeO	6.34	Q	7.7	Q	7.5	ti	2
MnO	0.18	Σ	75.7	Σ	78.8	h	18
MgO	3.18					k	0.42
CaO	4.97	en	7.9	En	9.0	mg	0.42
Na ₂ O	3.05	fs	9.5	Fs	7.2	o	0.10
K ₂ O	3.25	C	1.1	C	1.1	c/fm	0.47
P ₂ O ₅	0.30	il	1.5	Il	1.2	qz	+12
CO ₂	0.0	mt	2.1	Mt	2.0		
H ₂ O+	1.79	ap	0.7	Ap	0.5		
H ₂ O-	0.03						
Σ	100.25	Σ	98.5	Σ	100.0		

Table 4. Chemical analysis of biotite-plagioclase schist containing hornblende porphyroblasts. Metelinmäki, Salahmi, Vieremä. Analyst: A. Heikkinen.

	Weight		Molecular		Niggli		
	per cent	norm	norm	norm	values		
SiO ₂	51.69	or	8.3	Or	8.5	si	141
TiO ₂	1.07	ab	34.1	Ab	37.0	al	31
Al ₂ O ₃	19.09	an	28.4	An	29.0	fm	38
Fe ₂ O ₃	1.46	Σ	70.8	Σ	74.5	c	18
FeO	7.52					alk	13
MnO	0.16	fo	1.0	En	12.0	ti	2
MgO	4.19	fa	1.2	Fs	8.5	h	23
CaO	6.25	en	9.0	C	0.6	k	0.19
Na ₂ O	4.00	fs	9.6	Il	1.6	mg	0.45
K ₂ O	1.45	C	0.5	Mt	2.0	O	0.08
P ₂ O ₅	0.38	il	2.0	Ap	0.8	c/fm	0.48
CO ₂	0.0	mt	2.1			qz	-9
H ₂ O+	2.56	ap	1.0				
H ₂ O-	0.08						
Σ	99.90	Σ	97.2	Σ	100.0		

($\alpha = 1.667 =$ yellowish, $\gamma = 1.690 =$ bluish green, $c \wedge \gamma = 18^\circ$) are black and vary between 0.5 and 2 mm in length, and they contain pale minerals in abundance as inclusions. The schist also has some large quartz grains, which appear to be neither porphyroblasts nor phenocrysts. The minerals of the basic mass of the schist are biotite ($\gamma = 1.632 =$ brown), plagioclase, (quartz, epidote ($\alpha = 1.714$, $\gamma = 1.729$), chlorite, apatite, zircon, chalcopyrite, hematite and ilmenite). But hornblende is altogether missing from the groundmass. This rock has been chemically analyzed (Table 4).



Fig. 9. Hornblende porphyroblasts in the biotite-epidote-plagioclase schist. Around the porphyroblasts is a light gray halo. Metelinmäki, Salahmi, Vieremä.

Table 5. Chemical analysis of biotite-epidote-plagioclase schist containing hornblende porphyroblasts. Metelinmäki, Salahmi, Vieremä. Analyst: A. Heikkinen.

	Weight per cent	Weight norm	Molecular norm	Niggli values
SiO ₂	59.10	or 7.8	Or 8.0	si 191
TiO ₂	0.98	ab 21.0	Ab 23.0	ti 2
Al ₂ O ₃	18.32	an 35.0	An 36.0	h 9
Fe ₂ O ₃	2.04	Q 18.7	Q 18.5	al 35
FeO	3.97	Σ 82.5	Σ 85.5	fm 27
MnO	0.14			c 28
MgO	2.17	wo 1.5	Wo 1.4	alk 10
CaO	8.03	en 5.4	En 6.2	k 0.26
Na ₂ O	2.50	fs 4.0	Fs 2.0	mg 0.39
K ₂ O	1.33	il 1.8	Il 1.4	o 0.19
P ₂ O ₅	0.18	mt 3.0	Mt 3.0	c/fm 1.04
CO ₂	0.0	ap 0.3	Ap 0.3	qz +51
H ₂ O+	0.80	fr 0.2	Fr 0.2	
H ₂ O-	0.11			
F	0.04			
Cl	0.01			
Σ	99.72	Σ 98.7	Σ 100.0	
-O	0.02			
Σ	99.70			

The biotite-epidote-plagioclase schist with hornblende porphyroblasts is blackish gray, its grain size 0.3—0.2 mm, and its structure granoblastic. The hornblende porphyroblasts ($\alpha = 1.670 =$ yellowish green, $\beta = 1.684 =$ green, $\gamma = 1.692 =$ bluish green, $c \wedge \gamma = 20^\circ$, $2V\alpha = 60^\circ\text{—}65^\circ$) are 0.3—10 mm long. They appear rosette in shape and are surrounded usually by a light gray ring free of biotite (Fig. 9). Instead of biotite the ring contains more potash feldspar than occurs elsewhere in the schist. As inclusions the hornblende contains quartz, epidote and plagioclase. The hornblende porphyroblasts are oriented to some extent and there are more of them in certain zones that do not conform to the stratification than elsewhere. The matrix of the schist contains biotite, plagioclase ($\alpha = 1.554$, $\gamma = 1.563$, An_{50}), epidote ($2V\alpha = 86^\circ$), quartz, potash feldspar, sphene, apatite and opaque minerals. The apatite occurs as large, idioblastic crystals (the cross-section of the prism always being 5 mm). There is no hornblende in the groundmass of the schist. The chemical analysis of this schists is presented in Table 5. When this analysis is compared with that in Table 4, one will note that the occurrence of the epidote here is mainly due to the chemical difference between the schists.

The quartz-bearing amphibolite is fine-grained (grain size: 0.3—1 mm) and dark gray. The minerals are hornblende, plagioclase, quartz, apatite, epidote and opaques. The hornblende is idioblastic and it reveals an abundance of inclusions. The plagioclase is zonal ($An_{40\text{—}50}$). The apatite occurs as both large and small grains and the epidote with slickensides.

WEST OF VIEREMÄ CHURCH

Overlying the basal conglomerate and the mica schist formation with quartzite intercalations on top of it west of the parish center of Vieremä, there occur schists containing calcic minerals. This formation is approximately as thick as the formation containing calcareous minerals at Metelinmäki, which, however, exhibits more variety. The latter formation contains a greater number of different rock types and is stratified to a greater extent than the former.

These schists consist mostly of fine-grained amphibolite. A few more acid layers than they are can be distinguished. The plagioclase is zonal, and its composition is $An_{40\text{—}50}$ ($\alpha = 1.548$, $\gamma = 1.563$). The hornblende ($\alpha = 1.659 =$ light green, $\beta = 1.674 =$ green, $\gamma = 1.681 =$ bluish green, $c \wedge \gamma = 20^\circ$) occurs in some instances as a mixture of both small and large grains. Other minerals include quartz, epidote, apatite and opaques.

THE OLIGOMICTIC CONGLOMERATE BETWEEN ROTIMOJOKI AND
LUVEJOKI

Between Rotimojoki and Luvejoki, about 1 km east of Lähdemäki, in the locality of Nissilä, Vieremä commune, there is a zone of oligomictic conglomerate over a kilometer long. The greatest thickness of the zone is about 200 m. The strike of the stratification and foliation of the conglomerate is $N0^{\circ}-30^{\circ}$ W and the dip $70^{\circ}-80^{\circ}$ SW. The basal direction, which has been measured from the bedding, is generally east. The zone exhibits a transverse foliation, which in some places is exceedingly clear. The strike of the transverse foliation is $N30^{\circ}-75^{\circ}$ E and the dip $70^{\circ}-90^{\circ}$ SE. The transverse foliation becomes more distinct as one moves northward. There one meets also with several simultaneous directions of the transverse foliation, and in some cases they are so marked and so numerous that only the lineation of 70° SSW is readily measurable. The last-mentioned phenomenon occurs most clearly in the mica schist intercalations situated in the conglomerate and in the extension of the conglomerate zone on its northern side.

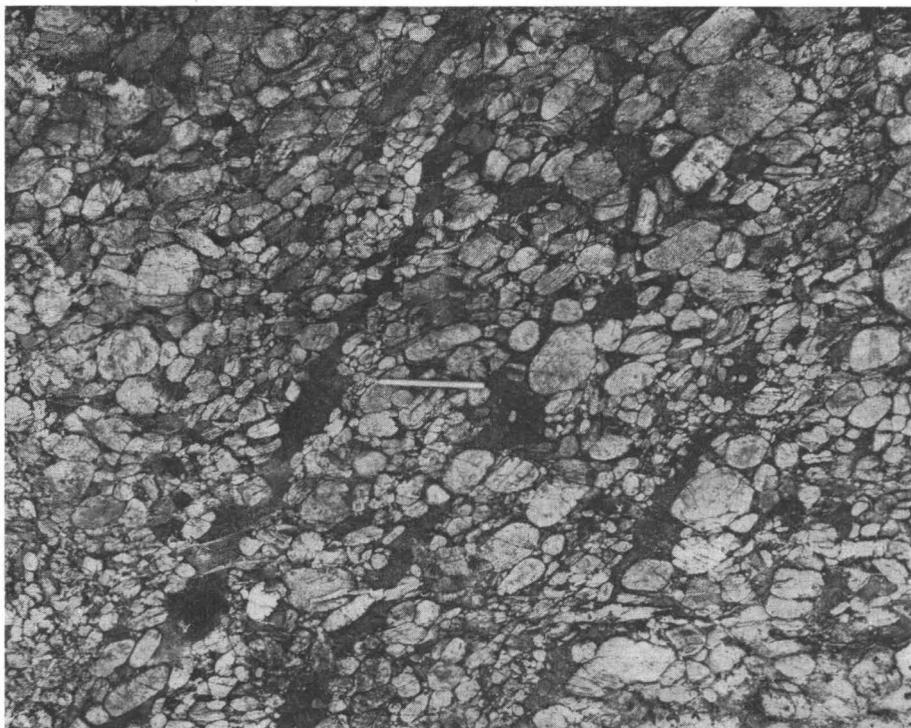


Fig. 10. Oligomictic conglomerate with narrow intercalations of mica schist. The strike of the foliation is almost the same as the position of the lead pencil. Between Rotimojoki and Luvejoki, Nissilä, Vieremä.



Fig. 11. Broad mica schist intercalations in oligomictic conglomerate. Between Rotimojoki and Luvejoki, Nissilä, Vieremä.

The conglomerate is stratified (Figs. 10 and 11). Mica schist layers alternate in it with conglomerate layers. The thickness of the mica schist layers varies between 5 cm and a couple of meters. The conglomerate layers are in general thicker than the former. The conglomerate layers proper also appear to vary in a direction, for instance, perpendicular to the layers. The size of the pebbles, it seems, varies in different parts of the layers, and the situation differs, furthermore, in different layers. Also the mica schist layers exhibit variation. Fine-grained layers alternate with more coarse-grained and sandier layers. The last-mentioned are likely to contain pebbles in the millimeter class composed of quartz and feldspar as well as angular phyllite fragments. The structure of the conglomerate and the mica schist is clastic.

The size of the pebbles in the conglomerate varies between 1 and 15 cm. They are well rounded or rounded and generally slightly elongated. Their long axis runs either parallel to the stratification or has turned in the direction of the strike of the transverse foliation or, then, toward this strike. The pebbles have predominantly evolved out of stratified quartzite. Accordingly, individual pebbles are likely to show bedding (Fig. 12) and to differ in form. Some phyllite fragments likewise occur in this conglomerate.

The minerals present in the pebbles include quartz, biotite, plagioclase, potash feldspar, muscovite, chlorite, sphene, tourmaline, apatite, zircon and opaques. Among the pebbles are quartz-rich, biotite- and quartz-rich, biotite- and muscovite- and quartz-rich, and chlorite- and quartz-rich types. All of

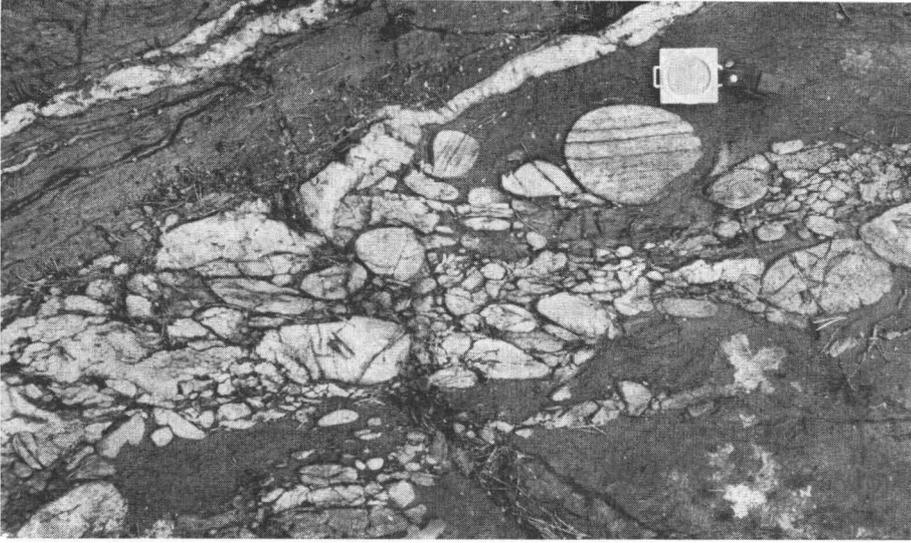


Fig. 12. Oligomictic conglomerate, which contains bedded quartzite pebbles, clusters of pebbles that appear to have broken off some larger body, and pegmatite veins. Between Rotimojoki and Luvejoki, Nissilä, Vieremä.

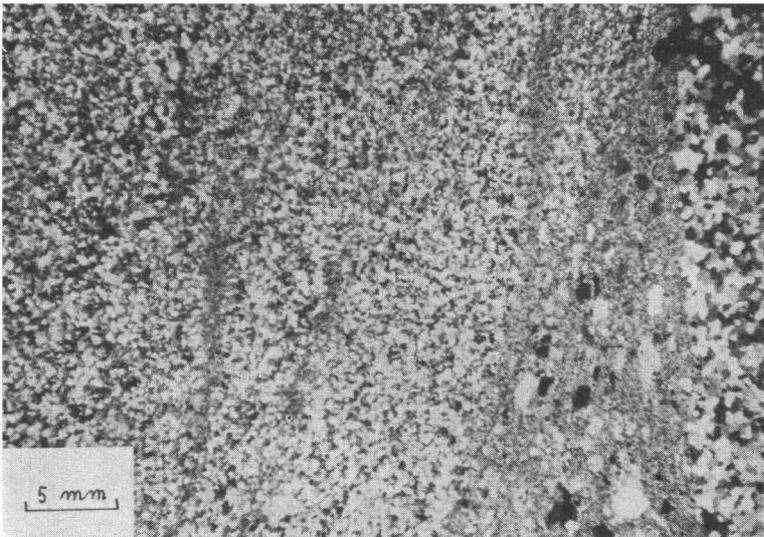


Fig. 13. Pebbles in oligomictic conglomerate. Pebble on right side is larger of grain than the one on the left, which is fractured. The cementing material contains in part larger mineral grains than the pebbles do. Between Rotimojoki and Luvejoki, Nissilä, Vieremä.

them contain feldspars, in some places in very small amounts, in other places in comparative abundance. The grain size of the minerals varies from 0.04 mm to 1 mm. As cementing material, the pebbles contain considerable quantities of quartz and chlorite. They are clastic in structure.

The matrix of the conglomerate consists of the same kind of rock as the mica schist that alternates with the conglomerate layers (Fig. 11). In some cases the cementing material contains larger mineral grains than the pebbles do that are enveloped by it (Fig. 13). The structure of the conglomerate's matrix is similarly clastic, and the grain size of its minerals ranges from 0.01 to 0.2 mm (not including the pebbles).

Near the northern end of the conglomerate zone and near its western edge, there is prominently folded rock in which narrow mica schist layers alternate with narrow quartzite laminae. The quartzite laminae have to some extent been broken into small pieces, which resemble the pebbles of the conglomerate just described (Fig. 14). The material of the quartzite is the same as that of the pebbles. This gave the present author a reason to seek data relating, on the one hand, to the tectonic mode of origin and, on the

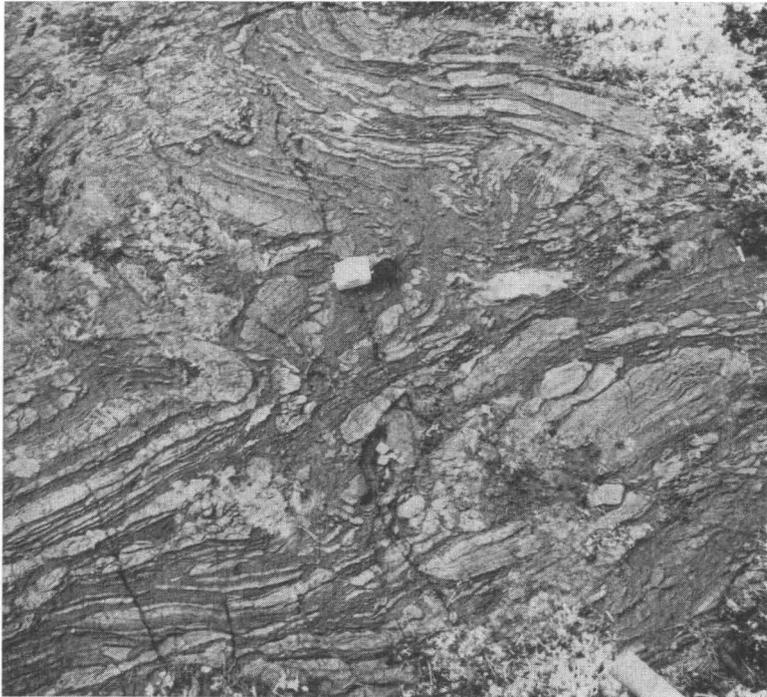


Fig. 14. Alternating folded quartzite and mica schist layers. The quartzite layers have undergone breakage, and a few of the pieces have evolved into forms identical to those of the pebbles in the conglomerate. Between Rotimajoki and Luvejoki, Nissilä, Vieremä.

other, to the normal genesis of this conglomerate. In Finland, Mikkola (1957) has interpreted the oligomictic conglomerate of Taivalkoski situated along the bank of the Kemi river as a tectonic conglomerate or pseudoconglomerate. Generally, the quartzite conglomerates of Finland have been regarded as of sedimentary origin.

The Vieremä conglomerate is situated between two mica schist beds near the northern end of the schist belt, where there is a major fault. Its strike is $N60^{\circ}E$, its dip steep and its axis $15^{\circ}SWW$. The schist beds have bent at their north end toward the east, *i.e.*, the block at the north end of the fault has moved eastward and downward in relation to the segment on the southern side. Other, smaller faults are known on the southern side of the major fault, as, *e.g.*, at the south end of Marttisenjärvi (Lake Marttinen).

With respect to the tectonic mode of origin, the following case can be made:

- 1) The tectonic position of the conglomerate is favorable.
- 2) The strike of the transverse foliation deviates toward the east from the strike of the stratification, and the pebbles embedded in the conglomerate have rolled in that direction. Naturally, they have first been broken off the quartz layers. Figure 15 provides an argument, or at least an assumption,

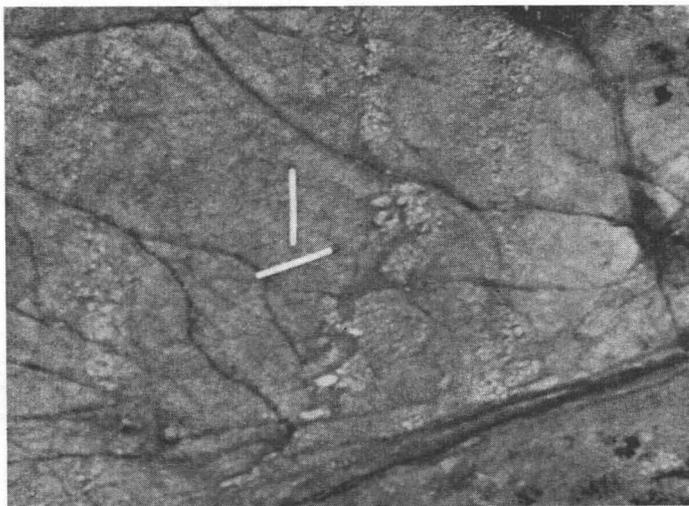


Fig. 15. A quartzite layer in the oligomictic conglomerate. One of the matches runs parallel to the stratification and the other to the foliation. The question arises as to whether the pebbles in the quartzite have evolved through metamorphism or whether the structure was produced during sedimentation. In any case, the quartzite had been bedded before the metamorphism occurred. Between Rotimojoki and Luvejoki, Nissilä, Vieremä.

that such an event has occurred. Gründel and Rösler (1963, p. 1033), among others, have written about such a tectonic genesis of pebbles and the mechanism involved.

3) In places the pebbles may be regarded as having originated from a large body through brecciation, in view of the fact that the edges of the pebbles fit in one another's recesses (Fig. 12). In the photomicrograph (Fig. 16), again, it will be observed how the pebble has been shattered to form smaller pebbles. On the surface of an exposure, one can sometimes see tiny broken pieces around round pebbles — the likelihood being that they were broken off while the pebbles were worn round during movements.

4) A few of the pebbles seen in Fig. 14 must surely have formed by breaking off quartzite layers.

5) There is a lack of evidence of cross-bedding, and the mode of sedimentation is not clear.

The following points of view may be presented in support of the genesis of the conglomerate through a process of sedimentation:

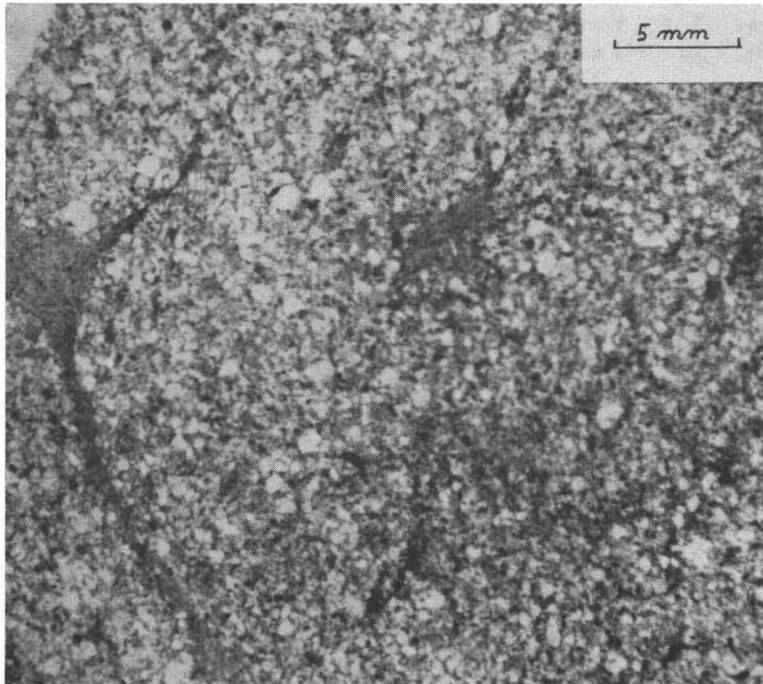


Fig. 16. Part of a large conglomerate pebble in the process of breaking up. Oligomictic conglomerate. 1 Nic. Between Rotimojoki and Luvejoki, Nissilä, Vieremä.

1) The variation in the size of the pebbles in the layers making up the conglomerate in a transverse direction could be construed to signify sorting under sedimentation.

2) The clastic structure of the pebbles and the matrix of the conglomerate (Fig. 17) has been well preserved, and on the whole the conglomerate contains very few marks of deformation. Why should this be if the conglomerate represents a tectonic breccia?

3) A large part of the pebbles in the conglomerate are well rounded, but, on the other hand, very little can be seen of the remains of the rounding process involved in the genesis of pebbles through tectonic deformation. Where would they have vanished?

4) The phenomena of the kind shown in Fig. 14 are rare in the conglomerate zone. The question is raised as to whether they represent only folded and broken intercalations in the conglomerate.

5) The author has had thin sections made of such portions of the conglomerate as contain pebbles that appear to have originated by breaking off a larger body. Fig. 17 illustrates such a thin section — but the pebbles to be

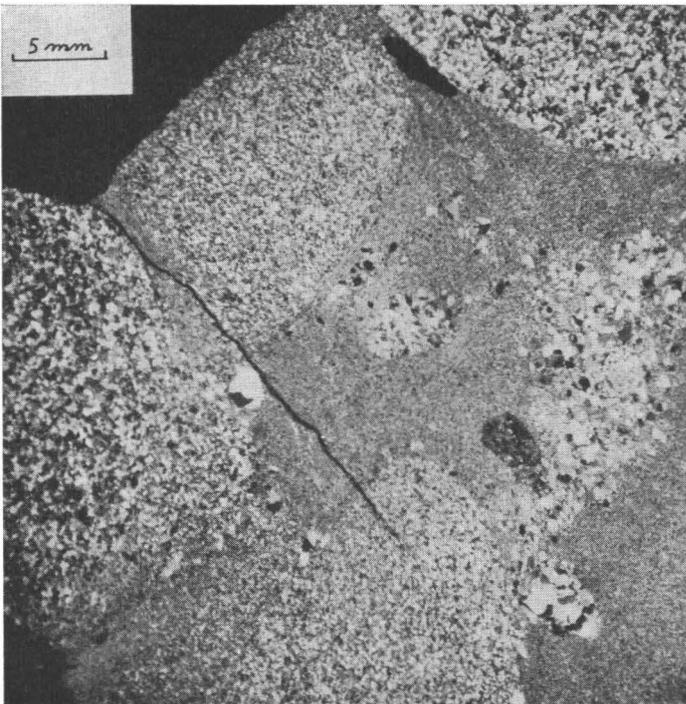


Fig. 17. Different kinds of pebbles in the oligomictic conglomerate. Crossed Nicols. Between Rotimojoki and Luvejoki, Nissilä, Vieremä.

seen in it are so varied that it is difficult to imagine their having broken off the same piece. Similarly, Fig. 13 shows two quite different pebbles side by side, though the one on the right side is, moreover, broken.

In this conglomerate area the quartz layers have folded and broken up, and the broken pieces have taken on the shape of pebbles. Further, the conglomerate contains clusters of pebbles that have formed through breakage, though sedimentation, too, may have had the effect of crowding them close together. Also the pebbles in the conglomerate had rolled in the direction of the transverse foliation and in certain cases even become rounded in the process, while the residue was left next to the resulting pebbles. Still, sufficient evidence is lacking to prove that the whole conglomerate is the product of tectonic brecciation. Phenomena of this kind are likely to be found in any conglomerate whatsoever. Elucidation of the sedimentation process throughout the entire schist area and the mapping of the conglomerate zone with a very small-scale map serving as a guide map could conceivably provide the missing answers to our questions.

Eskola (1941, p. 470) writes about conglomerates of this kind in the commune of Suojärvi and the region of Aunus in Soviet Karelia: »Die Entstehung solcher Quarzgeröllmassen mag schwer erklärbar erscheinen, aber man kennt ja recht Ähnliches in jüngeren Formationen, wie z. B. ein Quarzkonglomerat der Kreideformation bei Washington in Nordamerika, das dem Verf. durch Autopsie bekannt ist.» On the other hand, Horny (1962) includes in his study pictures from central Bohemia in which the harder layers of stratified rock appear to have broken into pebble-like pieces — though these pictures, to be sure, show many concretions, too. Gründel and Rösler (1963) describe the formation of pebbles in the sedimentary Upper Devonian series of Thuringia and comment on the effect of tectonics on the process. Actually, the pebbles consist of limestone, but the composition of the rock is an immaterial consideration in a case where the most significant factor is the differences in hardness between the different layers. It may also be asked whether the modes of formation of auto-conglomerates can be applied to such conglomerates (Hills 1963, p. 18).

In some portions of the conglomerate there are wedge-like quartz veins 0.1 to 3 cm wide and 5 to 100 cm long (Fig. 18). They usually run parallel to the transverse foliation. Some of these veins exist within only a single pebble, some are found only in the mica schist bed and some intersect both the pebbles contained in the conglomerate and its matrix. There also occur a few feldspar-bearing veins (Fig. 12). Since the veins are wedge-shaped — *i.e.*, confined — the material could not have come from outside the conglomerate in this so-called quartzification process. Rather did the process take place wholly within the rock — what we are confronted with is a metamorphism pure and simple.

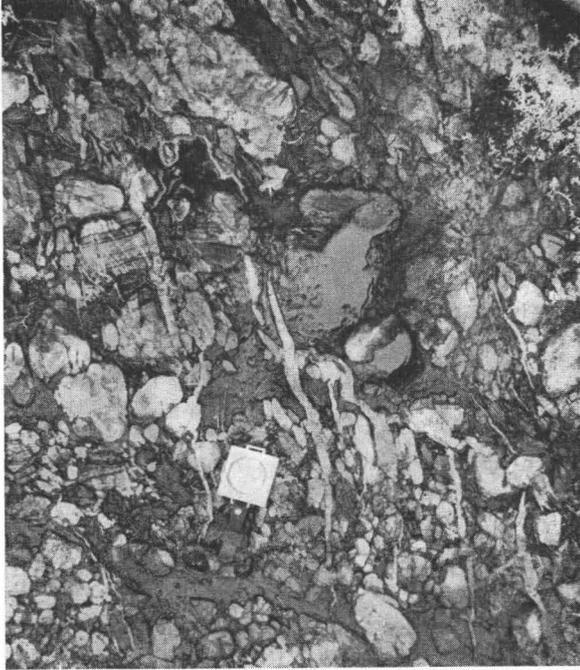


Fig. 18. Lenticular quartz veins in the oligomictic conglomerate. Between Rotimojoki and Luvejoki, Nissilä, Vieremä.

STAUROLITE MICA SCHISTS AND MICA SCHIST WITH PSEUDO-STAUROLITE PORPHYROBLASTS

Staurolite mica schists are to be seen at many points throughout the entire length of the Salahmi schist zone. At the north end of the schist sequence, southwest of Nissilä, there occur the thickest staurolite schist layers, while it is in the area of the village of Valkeiskylä, at the south end, that the thinnest ones are situated. At Kulvemäki, in the village of Salahmi, staurolite mica schists alternate with mica schist layers in a fairly broad zone. Wilkman (1931, p. 200) mentions staurolite mica schists as occurring at Tuliniemi, Salahmijärvi. Further, they occur about 1 km NNW of Tuliniemi and about 1.5 km north of the northwest end of Salahmijärvi. Except for one occurrence, the staurolite mica schists in this area are much alike (Fig. 19). The exception is situated in the last-mentioned locality, that is, about 1.5 km north of the northwest end of Salahmijärvi, where the staurolite porphyroblasts have become transformed in certain portions of the formation into other minerals and now appear as pseudo-porphyroblasts (Fig. 20).

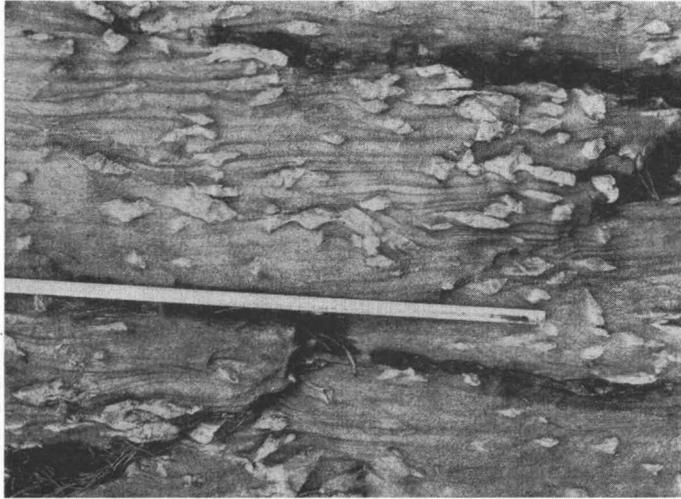


Fig. 19. Staurolite mica schist. Southern edge of Karhumäki, Nissilä, Vieremä.

STAUROLITE MICA SCHISTS

The staurolite mica schists (Fig. 19) are dark gray and fine-grained (0.1 to 1 mm). The quartz grains in particular are in spots larger than the other ones. Here and there the schist is slightly bedded. And it has narrow layers richer in quartz, which are at the same time somewhat coarser of grain than the schist as a whole.

The schist contains yellowish brown staurolite porphyroblasts, which are poikiloblastic. Present as inclusions are quartz, in particular, and also opaque minerals in profusion. The porphyroblasts are usually 3 to 4.5 cm long and 0.8 to 1.5 cm thick. Also smaller grains up to 0.5 cm in length occur, and in some places one can see schist layers in which all the staurolite porphyroblasts are only 0.5 to 1 cm long. The staurolite porphyroblasts appear in part to have become situated parallel to the foliation and to some extent in arbitrary fashion nowise dependent on it. In the former case, the crystal faces of the grains do not seem to be developed so well as they are in the latter case.

Chlorite porphyroblasts occur in nearly all the staurolite mica schists, and in most instances the chlorite flakes do not run parallel to the foliation plane. The length of the chlorite porphyroblasts is between 1 and 3 mm, and they can frequently be observed to be polysynthetically twinned and with a ragged form.

In places in the staurolite mica schist zone there occur layers containing small quartz lenses, which vary in length from a centimeter to several centi-



Fig. 20. Mica schist containing pseudo-staurolite porphyroblasts. About 1.5 km north of NW end of Salahmijärvi, Marttisenjärvi, Vieremä.

seen to have staurolite relicts; instead there have been grain aggregates containing chlorite ($\beta = 1.626$) and muscovite ($\alpha = 1.541$, $\beta = 1.571$, $\gamma = 1.575$). The pseudo-staurolite evidently weathered more readily than the quartz-bearing host rock and now are represented by the pits on the surface of the rock.

In the dark gray layers there are very narrow biotite-rich laminae. The grain size of the minerals in these layers ranges from 0.1 to 0.02 mm, and the schist is blastoclastic in structure. It contains biotite ($\gamma = 1.619 =$ greenish brown, $\alpha =$ nearly colorless) in profusion, quartz, muscovite and, to some extent, feldspar. Accessory minerals are apatite, zircon and opaques.

The grain size found in the gray layers varies between 0.04 and 0.2 mm. The main minerals are quartz, which is present in the greatest amounts, biotite and muscovite. The feldspar content is slight. In addition, there occur zircon, apatite and, very slightly, opaque minerals.

In both schist layers there occur poikiloblastic chlorite porphyroblasts. In the gray schist layers they are distinctly larger (0.4—0.2 mm) than in the dark gray schist layers (0.2—0.1 mm). But the chlorite porphyroblasts are noticeably smaller than the pseudo-staurolite porphyroblasts. In many instances, they are polysynthetically twinned and have a symmetrical extinction in relation to the intergrowth plane of the lamellae. In a few cases, the chlorite flakes have fanned out radially.

Both schist layers have undergone metamorphism, but differently. For instance, the quartz has a much stronger undulatory extinction in the dark gray than moderately gray layers. In the gray schist layers there are lenticular quartz veins a few millimeters thick with trends of N30°W and N60°E, which generally do not extend to the side of the dark gray layer (»en echelon» — cracks). Also the dark gray schist layers contain lenticular, very narrow quartz veins, but their orientation is either N15°E, 70°E or N80°E, 65°W. They do not extend over to the adjacent layer, either. The stratification and foliation of the schist at the point where the foregoing measurements were made are N30°E, 65°E and the lineation is NS, 40°S. See Hausen 1936, p. 25.

The pseudo-staurolite porphyroblasts appear to have rolled somewhat during metamorphism. It is natural to think that they have rolled while they were still in the condition of hard staurolite and not after they had metamorphosed into a soft mineral — chlorite and muscovite. The staurolite porphyroblasts have generally been bigger in the staurolite schists of the Salahmi area than are the pseudo-staurolite porphyroblasts described here (*cf.* Figs. 19 and 20). For some reason, the growth of the staurolite crystals in this schist ceased earlier than elsewhere, though the staurolite schists described all belong to nearly the same horizon. Then the staurolite mineral became unstable in the schist and disappeared, having turned to chlorite and muscovite.

Hietanen (1962) describes staurolite mica schists from a belt near the St. Joe river, Idaho, and pseudomorphs after staurolite found there. The pelitic rocks in northern New Hampshire show, according to Green (1963), biotite or staurolite altering to chlorite, or staurolite armored by andalusite. It is Eskola's observation (1946, p. 300) that staurolite is rich in iron and does not seem to be able to accommodate more than a slight amount of magnesium in its lattice. That magnesium could not replace the Fe²⁺ in staurolite is difficult to understand. According to Turner (1948, p. 82), staurolite appears to be stable over a narrowly limited range of temperature. On the other hand, it is obvious that if a mineral remains in very narrowly limited PT-conditions, it must maintain a nearly constant composition, although one might think differently upon contemplating its formula.

MICA SCHISTS AND MICA GNEISSES

Approximately 5 km southwest of the cross-roads at Karhukangas, Salahmi, there are mica schists, which are situated on the western side of the schists containing calcium-bearing minerals at Metelinmäki. This mica schist is relatively poor in feldspar compared to the mica schists of Murennusmäki, and it is also more thoroughly re-crystallized. The constituent minerals are quartz, biotite ($\gamma = 1.642 =$ brown), muscovite, feldspar, chlorite (porphyroblasts), apatite, zircon and opaques.

Proceeding southwest from Karhukangas, the schist is in places vein-gneissoid and appears to be richer in feldspar. It is mica gneiss.

At Kulvemäki mica schists occur in association with staurolite mica schists as alternating thick and thin beds. They occur in the same way on the southern slope of Karhumäki at Nissilä. Also on the western side of the oligomictic conglomerate at Nissilä one meets with mica schists.

LENSES CONTAINING CALCIC MINERALS IN MICA SCHISTS

Lenses containing calcic minerals have been found here and there as inclusions in the schist of Murennusmäki with alternating layers of quartzite and mica schist and in the mica schist situated about 2 km to the southwest at Karhukangas. They evolved through metamorphism either from calcareous concretions and/or thin calcareous layers by developing boundinage. Fig. 21 shows a portion of one such lens. The lenses are on the whole very much alike in structure.

This lens found about 3.5 km southeast of the cross-roads of Salahmi at Karhukangas is zonal (Fig. 21). The rock situated in the middle of the lens contains porphyroblasts and is fine-grained and light gray in color. Its porphyroblasts are green diopside prisms 1 to 4 mm in length, round garnet grains of a brown color and some black amphibole prisms. Around the mid-portion of the lenses there is a dark gray zone about 2 cm wide, which lacks foliation and which borders rather sharply on its host rock, composed of mica schist. The dark color of the border zone derives from the amphibole present in it. On the other hand, it lacks epidote and — almost — diopside, too, and the garnet content is likewise very scanty. The pale groundmass in the mid-portion of the rock consists of quartz (most), epidote, plagioclase and calcite. In addition to the amphibole, the dark gray ring contains an abundance of quartz and plagioclase. There appears to be more pyrrhotite at the point between the middle portion and the marginal zone than elsewhere. Garnet porphyroblasts are met with also in the mica schist immediately next to the lens. The index of refraction of the biotite (γ) at this point in the mica schist is 1.648 and a short distance farther in the same rock 1.642.

The composition of the plagioclase in the middle of the lens is An_{66} ($\gamma = 1.570$, $\alpha = 1.562$). The garnet poikiloblasts ($n = 1.783$) are indistinctly formed, and they contain considerable amounts of quartz and epidote as well as small amounts of sphene, calcite and amphibole as inclusions. The amphibole ($\alpha = 1.647 =$ light green, $\beta = 1.659 =$ yellowish green, $\gamma = 1.665 =$ blue-green, $c \wedge \gamma = 19^\circ$, $2V\alpha = 76^\circ$) is likewise poikiloblastic. The inclusions in it are the same as in the garnet, but there are more of them. The form of the amphibole porphyroblasts is very indefinite and ragged, and they have conspicuously beautiful, brown pleochroic haloes. The components producing

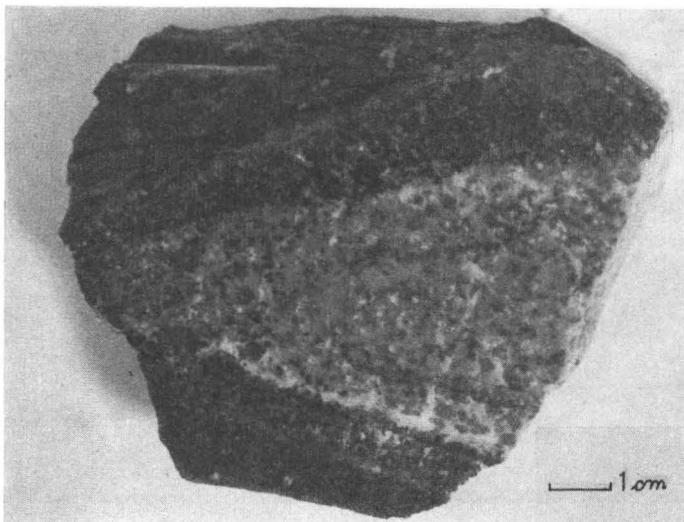


Fig. 21. Zoned lens containing calcareous minerals in mica schist. Karhukangas, Salahmi, Vieremä.

the haloes are so tiny that they could not be identified. The diopside ($\alpha = 1.705$, $\gamma = 1.734$, $c \wedge \gamma = 42^\circ$) occurs in the same way as the amphibole, but it also contains amphibole as inclusions. In places there is a fair amount of calcite ($\omega = 1.664$), and quartz is also present as inclusions. The epidote ($\beta = 1.732$) is likewise poikilitic, but it occurs as rather small, roundish grains. In addition, the lens contains a slight amount of sphene and a few grains of potash feldspar. Besides pyrrhotite, the opaque minerals in it include small amounts of chalcopyrite, magnetite and cobalt pentlandite, which is present as lamellae in the pyrrhotite.

The metamorphism has not advanced so far in the lenses as to make it impossible to see in them still that the zoned structure is largely due to composition differences arising as early as the sedimentation stage. This observation does not deny the possibility that metasomatism might also have occurred between the lens and the host rock, though, to be sure, no positive proof of it has come to light.

THE SCHISTS OF VALKEISKYLÄ

Directly overlying the basal conglomerate of Valkeiskylä is a thin layer of garnet-bearing mica schist. Next follow staurolite mica schist, garnet mica schist and, then, mica schist containing chlorite porphyroblasts. The schist layers here catalogued are only a few meters thick. Farther from them one meets with graphite-bearing mica schist.



Fig. 22. Gneiss from western margin of conglomerate at Kukkomäki. In the upper part of picture, facing east, there occurs augen gneiss with the appearance of a conglomerate. Kukkomäki, Vieremä.

THE CONGLOMERATE OF KUKKOMÄKI

Kukkomäki is situated about 4.5 km to the east from the central part of Salahmijärvi, in the commune of Vieremä. There one can observe strongly tectonized rock resembling conglomerate in a zone over 500 m long running parallel to the foliation.

On both marginal sides of the conglomerate formation there occurs gneiss of the kind shown in Fig. 22, which is stratified and very prominently foliated. In the upper part of Fig. 22, which represents the eastern flank of the formation, there is a layer of augen gneiss with the appearance of a conglomerate. The middle portion of the formation consists of very strongly tectonized conglomerate and/or breccia. Fig. 23 has been taken of the surface of the exposure situated at right angles to the lineation and the foliation, and the »conglomerate» there appears to contain distinct pebbles. Fig. 24, again, was taken of the surface of the outcrop running parallel to the lineation (N10°E, 45°S) but at right angles to the foliation. Both figures (23 and 24) show the same outcrop at adjacent points. The latter picture (Fig. 24) shows an abundance of intersecting shear cleavages, which break up the rock and its various layers into long, wedge-like pieces. These pieces appear to fit into one another and to form layers, and in this picture (Fig. 24) they cannot be easily imagined as elongated pebbles.

The structure of the gneiss and the augen gneiss with the appearance of a conglomerate in the marginal portions of the conglomerate formation is



Fig. 23. Conglomerate and/or breccia? The picture was taken of the surface of the outcrop at right angles to the lineation and foliation. Kukkomäki, Vieremä.

cataclastic. Their principal mineral components are quartz, plagioclase, microcline, biotite, and muscovite. The muscovite occurs as conspicuously large poikiloblasts. The »eyes» in the augen gneiss consist of quartz, quartzite and feldspar-rich quartzite as well as feldspar grains, some of which are porphyroblasts.

The middle portion of the conglomerate formation — the portion designated as conglomerate — contains pebbles consisting, first of all, of the gneiss and augen with the appearance of conglomerate met with in the marginal parts of the same zone as well as tectonic varieties of these constituents. It also contains mica schist fragments, in addition to pebbles of quartz, graphic feldspar and aplitic rock. All are strongly deformed in various



Fig. 24. Conglomerate and/or breccia? The picture was taken of the surface of the outcrop at right angles to the foliation and parallel to the lineation. Kukkomäki, Vieremä.

degrees. The matrix of the conglomerate is made up of various shear zones, which separate these so-called pebbles from one another. The minerals found in the conglomerate are quartz, plagioclase (An_{10}), epidote, sphene, apatite, zircon and opaques. The matrix is in many places richer in mica than the other parts of the rock.

Is this conglomerate a tectonic breccia and/or a true conglomerate? The augen gneiss resembling conglomerate in the marginal zone has structural features that are perhaps distinct relicts of a conglomerate fabric. There may have been layers of more distinct conglomerate structure and with large pebbles in the middle portion of the conglomerate formation than can be seen at present, although no sure evidence to substantiate this view exists.

But the conglomerate formation has undoubtedly also contained layers resembling the rocks of the marginal zone, layers from which very many rocks resembling the large-pebbled conglomerate must have evolved during tectonic movements. It was not until this metamorphism occurred that some of the quartz pebbles, for example, were formed.

Koark (1961) has studied the structure and petrofabrics of the Venna conglomerate. The rock, as he describes it, is very markedly metamorphosed, like the conglomerate of Kukkomäki. The Venna conglomerate, according to Koark (1961), represents a B-conglomerate tectonite with part-areas of predominantly B \perp B fabric. The B-tectonites, he asserts (1961), might be referred to transversal foldings.

The reader probably would like to have an answer to the question as to what conglomerate in the Salahmi area offers a counterpart to this conglomerate and/or breccia. On account of its advanced stage of metamorphism, the question is not easy to answer. The formation may, perhaps, be compared with the basal conglomerate — but also, probably equally well with the oligomictic conglomerate situated between Rotimojoki and Luvejoki. Perhaps it is well to keep in mind that the Kukkomäki conglomerate is flanked on both sides by similar rocks. At least, no essential differences have been noted between the rocks on its eastern and western flanks.

THE NÄLÄNTÖ REGION

The Näläntö region (Fig. 25) comprises the western part of the so-called Kiuruvesi-Vieremä schist belt, and in the middle of this region lies the lake after which it has been named, Näläntöjärvi (Lake Näläntö). The Näläntö region contains many different rocks, including both metamorphic and plutonic varieties. The metamorphic gneisses and schists can be divided handily, according to the grouping of exposures, into three groups: 1) the rocks on the NE-side of Näläntöjärvi, 2) the rocks on the SW-side of the lake, and 3) the rocks situated along the banks of Rikkajoki (Rikka river). The plutonic rocks in the region form two small separate bodies, which are situated on opposite sides of the lake.

METAMORPHIC GNEISSES AND SCHISTS

NE-SIDE OF NÄLÄNTÖJÄRVI

MICA GNEISS

There is mica gneiss on the NE-side of Näläntöjärvi — at the eastern margin of the area encompassed by the map. It is partly garnet-bearing, in places fine of grain (grain size 0.1—0.5 mm), in other places medium-grained (grain size 0.3—1.5 mm). In the latter places there are zones containing small quartz lenses and other zones, containing quartz and quartz-feldspar veins. The mineral composition of the mica gneiss deficient in garnet is as follows, expressed in terms of percentages by volume:

Quartz	30.2
Plagioclase	26.8
Biotite	35.1
Muscovite	1.8
Chlorite	1.1
Apatite	1.3
Opaque minerals	2.8
Other minerals	0.9
Total	100.0

The designation »other minerals» includes sphene, potash feldspar, epidote and zircon.

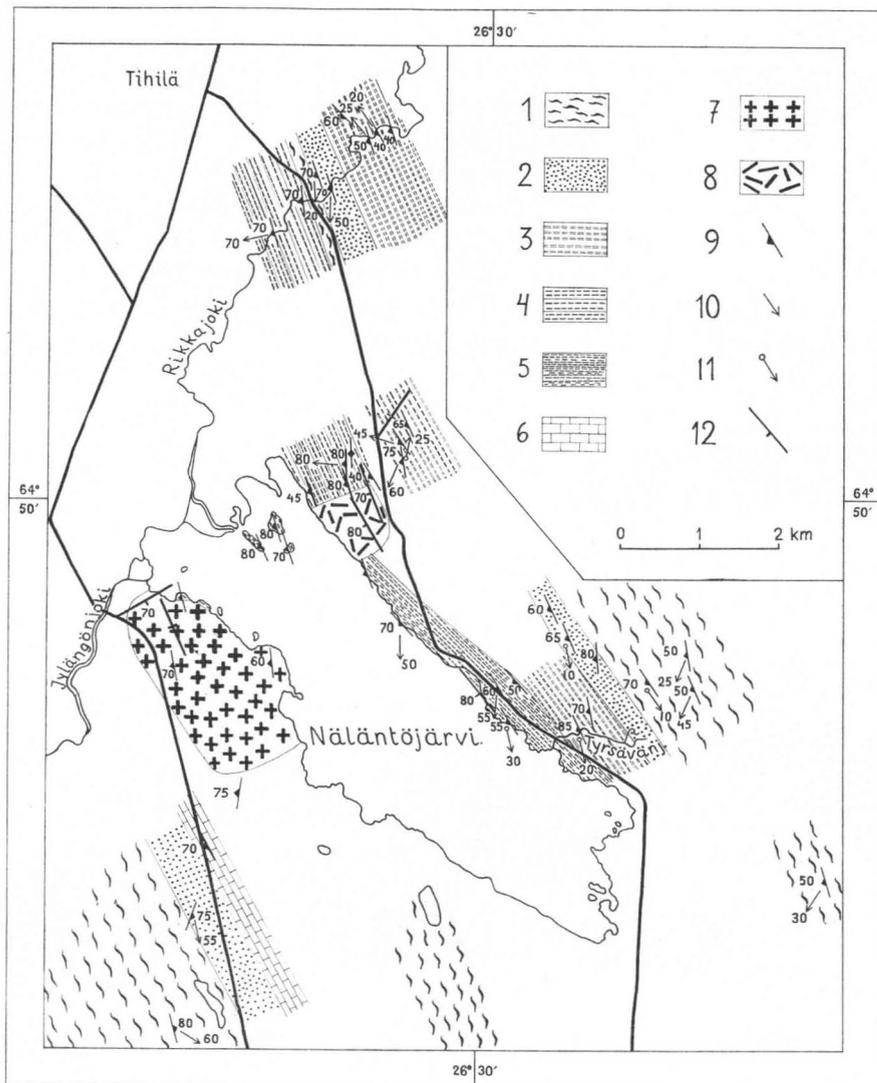


Fig. 25. Nälantö region. 1) Mica gneiss, 2) Amphibolite, diopside amphibolite and garnet-bearing pyroxene amphibolite, 3) Biotite-plagioclase gneiss, 4) Microcline-plagioclase gneiss, 5) Hornblende gneiss and cummingtonite-bearing hornblende gneiss, 6) Graphite-bearing schist, 7) Pyroxene-bearing porphyritic granite and pyroxene-hornblende granodiorite, 8) Biotite granite, 9) Strike and dip of foliation, 10) Lineation, 11) Fold axis and 12) Strike and dip of shear zone.

In medium-grained varieties there is more plagioclase ($a = 1.548$, $\gamma = 1.558$, An_{40}) than in fine-grained ones. Similarly, the content of epidote and sphene is larger in the medium- than in the fine-grained varieties of the rock. The biotite ($\gamma =$ reddish brown) is poikiloblastic in medium-grained types.

AMPHIBOLITE

Between the mica gneiss and the biotite-plagioclase gneiss on the north-eastern side of Näläntöjärvi, there occurs a dark gray, medium-grained, weakly schistose amphibolite, with which fine-grained amphibolite varieties are to be found in association. Particularly in the fine-grained amphibolite, one will observe very dark gray, lenticular hornblende schist inclusions 10 to 70 cm long and 2 to 10 cm wide. This amphibolite contains a scattering of garnet porphyroblasts a couple of centimeters in diameter. The mineral composition of the medium-grained amphibolite, minus the garnet porphyroblasts, is as follows in percentages by volume:

Quartz	14.6
Plagioclase	34.1
Hornblende	29.8
Biotite	14.1
Apatite	3.0
Opaque minerals	2.5
Other minerals	1.9
Total	100.0

This amphibolite contains a relative abundance of biotite and also quartz, and therefore it apparently has a high Niggli *k* value. Leake (1964) writes that amphibolites having low Niggli *k* values are almost certainly igneous in origin, but amphibolites with high Niggli *k* values may be either igneous or sedimentary in origin. The amphibolite under consideration here should perhaps be designated as para-amphibolite.

The composition of the plagioclase is An_{40} ($\alpha = 1.549$, $\gamma = 1.555$). The hornblende ($\alpha = 1.662$ = light green, $\beta = 1.674$ = green, $\gamma = 1.684$ = bluish green, $c \wedge \gamma = 20^\circ$), being poikiloblastic, especially in medium-grained types, and appearing in both types in conjunction with biotite. The garnet ($n = 1.780$) contains quartz in particular as inclusions. There is an abundance of apatite, the idioblastic grains of which are largish (the cross-sections of the prisms: 0.1—0.3 mm). The apatite contains hornblende, among other things, as inclusions. Polished sections have revealed the presence of the following opaque minerals: magnetite, pyrite, chalcopyrite, graphite and hydro-pyrrhotite.

In the hornblende schist the composition of the plagioclase is An_{45} ($\alpha = 1.552$, $\gamma = 1.560$) and the refractive indices of the hornblende are $\alpha = 1.663$, $\beta = 1.679$, $\gamma = 1.684$.

BIOTITE-PLAGIOCLASE GNEISS

Biotite-plagioclase gneiss may be observed on the eastern side of Näläntöjärvi in two different places: at Loutemäki, toward the NE from the biotite

granite body lying on the eastern side of the highway which runs past the eastern end of the lake, as well as along the banks of Tyrsävänjoki northward from the southeastern corner of Näläntöjärvi.

Loutemäki

At Loutemäki there occurs biotite-plagioclase gneiss containing microcline and amphibole. On the weathered surface of exposures the gneiss is yellowish gray, whereas on fresh surfaces it is bluish gray. It is a relatively homogeneous rock. In places it is somewhat brownish gray, in which case it contains more than the usual amount of potash feldspar. Similarly, the degree of metamorphism of the gneiss varies: in spots it has undergone a more advanced stage of mylonitization than in general. Some unmistakable mylonite zones can be seen in the rock. The types rich in potash feldspar are generally more metamorphosed and resemble augen gneiss to a greater extent than the rest do.

The grain size of the minerals composing the biotite-plagioclase gneiss varies from 0.2 mm to 2 mm and most usually ranges from 0.5 to 1 mm. The structure of the gneiss is granoblastic and cataclastic. Evidence of a blastoclastic structure can scarcely be detected in it. It contains numerous microscopically thin shear zones, which run mainly in two directions, cutting across each other at a 45° angle, though, to be sure, they do meander a bit, curving around large feldspar grains. In these zones the mineral grains have been ground small and have to some extent altered into other minerals. Present are, *inter alia*, quartz, microcline, epidote, amphibole and sphene. Table 7 shows the chemical analysis and mineral composition in percentages by volume of the most common biotite-plagioclase gneiss. The minerals contained in it are plagioclase, quartz, microcline, biotite, amphibole, epidote, muscovite, sphene, apatite, chlorite, orthite, hercynite and opaques.

Some of the plagioclase grains ($\alpha = 1.535$, $\beta = 1.539$, $\gamma = 1.544$, An_{15}) have broken and the lamellae of the grains twisted. The grains are in many instances ovoid and their edges have bands formed out of tiny particles. The microcline ($\alpha = 1.520$, $\gamma = 1.526$) contains broad flamelike perthite. The quartz occurs as very tiny grains in shear zones, but as considerably larger grains elsewhere in the rock. In the shear zone the biotite flakes ($\gamma = 1.637 = \text{brown}$, $\alpha = \text{yellowish}$) have partly altered to amphibole, epidote, sphene, muscovite and, sometimes, chlorite. In spots no biotite can be seen at all. It would appear as if the biotite had vanished to make way for epidote and amphibole. The epidote ($\gamma = 1.737$, 15 mol. per cent pistacite component) is present in crushed areas as clusters or rows of grains. The amphibole ($\alpha = 1.652 = \text{nearly colorless}$, $\beta = \text{slightly greenish}$, $\gamma = 1.675 = \text{light bluish green}$, $c \wedge \gamma = 20^\circ$) occurs in many cases as indefinitely shaped grains resembling porphyroblasts. The apatite ($\omega = 1.640$, $\varepsilon = 1.636$; according to

Table 7. Chemical analysis and modal mineral composition of biotite-plagioclase gneiss. Loutemäki, NE-side of Näläntöjärvi, hamlet of Näläntö, Kiuruvesi commune. Analyst: A. Heikkinen.

	Weight per cent	Weight norm	Molecular norm	Niggli values
SiO ₂	68.60	Q 22.1	Q 20.7	si 300
TiO ₂	0.43	or 15.0	Or 15.0	al 41
Al ₂ O ₃	15.8	ab 40.4	Ab 43.0	fm 19
Fe ₂ O ₃	0.60	an 13.3	An 13.5	c 13
FeO	2.15	Σ 90.8	Σ 92.3	alk 27
MnO	0.06			ti 1
MgO	1.29	en 3.2	En 3.6	h 7
CaO	2.85	fs 2.9	Fs 2.2	k 0.38
Na ₂ O	4.80	C 0.3	C 0.3	mg 0.45
K ₂ O	2.55	il 0.8	Il 0.6	o 0.11
P ₂ O ₅	0.13	mt 0.9	Mt 0.8	c/fm 0.72
CO ₂	0.00	ap 0.3	Ap 0.3	qz +92
H ₂ O+	0.50			
H ₂ O-	0.05			
Σ	99.81	Σ 99.2	Σ 100.0	

Modal mineral composition:

Microcline	12.5	Sphene	1.2
Plagioclase	35.5	Chlorite	1.1
Quartz	29.7	Muscovite	0.4
Biotite	8.5	Hercynite	0.4
Amphibole	4.5	Opaque minerals	0.7
Epidote	4.5	Σ	100.0
Apatite	1.2		

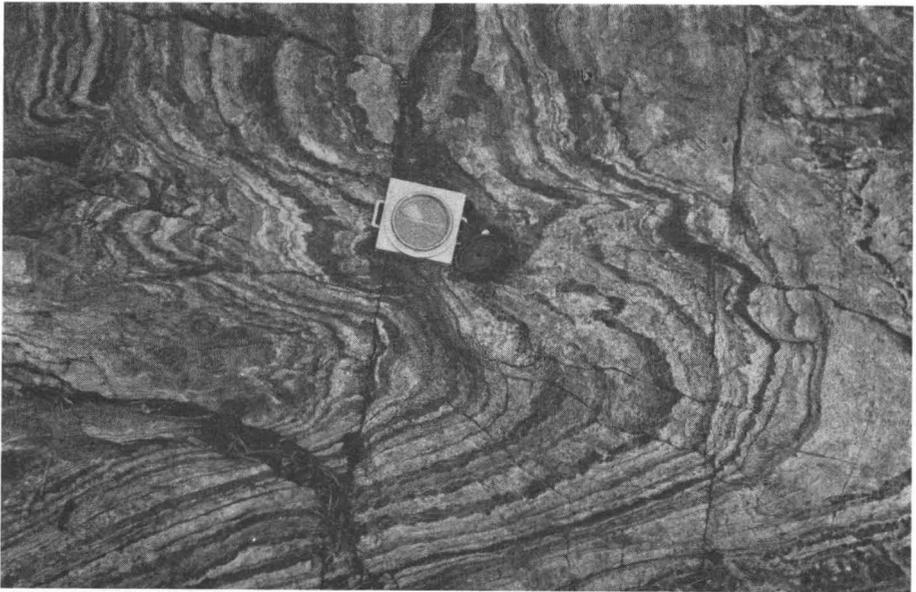


Fig. 26. Quartzite containing intercalations of hornblende schist. The rock is markedly metamorphosed and folded. Northeastern side of Näläntöjärvi, Loutemäki, Näläntö, Kiuruvesi commune.

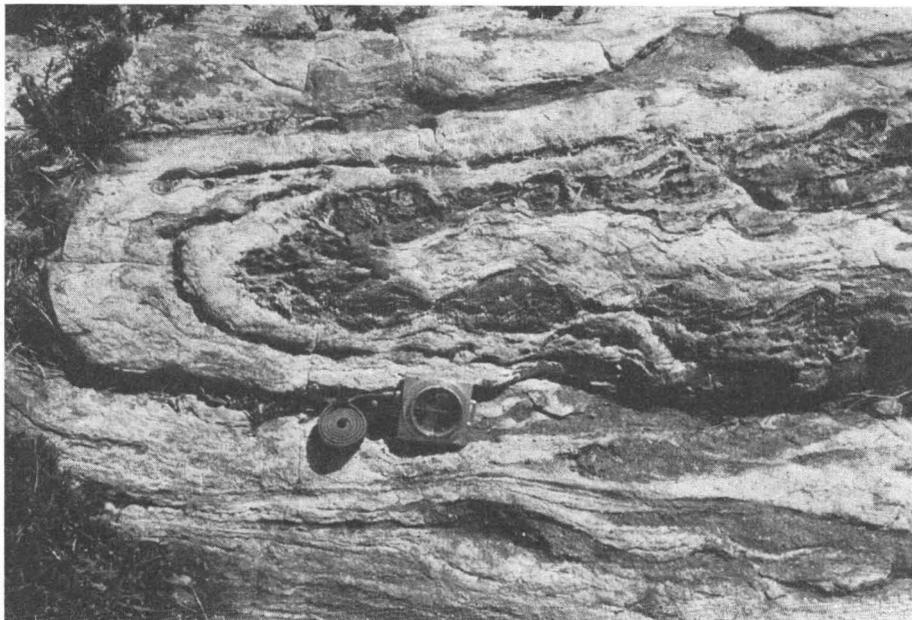


Fig. 27. Top of fold in folded quartzite containing intercalations of hornblende schist. The hornblende schist layers have thickened and also developed a boudinage structure during the folding process. Northeastern side of Nälantöjärvi, Loutemäki, Nälantö, commune of Kiuruvesi.

Tröger (1959), it would be fluorapatite) takes the form of largish grains. The hercynite consists of a few very green, euhedral, isotropic little grains. The opaque minerals are hematite, pyrite, pyrrhotite, magnetite and hydro-pyrrhotite. The hematite has the same kind of lamellae as those described by Ramdohr (1955, p. 690).

In this biotite-plagioclase gneiss, there are lenticular, very light gray quartzite layers with intercalations consisting mainly of hornblende schist. These lenticular quartzite layers are quite highly metamorphosed and folded (Figs. 26 and 27). The thickness of the lenticular layers varies from one to ten meters and the longest one known stretches 150 m in length. The lenses are not all in the same foliation plane, and they could hardly have belonged to the same layer, for examples of them are to be found along both western and eastern margins of Loutemäki. In all likelihood, they belong to at least three different layers, albeit the different lenses resemble each other closely in structure and composition.

The structure of the largest lens, which at the same time is situated farthest to the SW, is shown in Figs. 26 and 27. On its weathered surface, the quartzite is a very light gray, and it contains intercalations, mostly of a green hue. The thickness of the intercalated bands varies between less than 1 cm and as much as 10 cm. They are conspicuously folded (fold axis N5°E,

25°N), and in places they have thickened and become broken. The structure of the quartzite is granoblastic and the size of the quartz grains in it ranges from 0.1 to 1 mm. The quartz grains have an undulatory extinction and are quite definitely oriented. The mineral composition of the quartz grains is as follows, expressed in percentages by volume: Quartz 94.5, feldspars < 0.1, hornblende 3.5, pyroxene 1.5, biotite < 0.1, apatite 0.1 and opaque minerals 0.4. The pyroxene and the hornblende occur as stripes 0.1 to 3 mm thick, which conform to the foliation, invariably contain quartz and generally are not sharply bounded against their country rock but grade over into it by degrees. Both the hornblende and the pyroxene have in many cases been enriched into different stripes. The fine-grained stripes are usually narrower than the large-grained ones. The hornblende ($\alpha = 1.655 =$ colorless, $\beta = 1.660 =$ light greenish, $\gamma = 1.675 =$ light bluish green, $c \wedge \gamma = 17^\circ$, $2V\alpha = 83^\circ$, 40 mol. per cent (Fe²⁺, Mn, Ti)-component) occurs in the stripes as laminae 0.2 to 2 mm long, and pyroxene ($\alpha = 1.668$, $\gamma = 1.696$, $c \wedge \gamma = 39^\circ$) is present as grains of the same size.

On the weathered surface the hornblende-schist intercalations in the quartzite are green and, in rare instances, reddish brown in color, and the grain size of their minerals varies between 0.3 and 1 mm. The main mineral is hornblende. The reddish brown variety also contains a slight amount of pyroxene. In addition, there are small amounts of quartz, apatite, biotite and opaque minerals. The brown hue of a few of the hornblende schist layers is given them by hematite, which envelops all the minerals in the rock, including the quartz and is present in their fissures. The dark minerals in the brown bands are richer in Fe than in the green layers: The composition of the hornblende in the brown layer is 65 mol. per cent (Fe²⁺, Mn, Ti)-component ($\alpha = 1.674 =$ olive green, $\beta =$ green, $\gamma = 1.717$, $c \wedge \gamma = 18^\circ$) and that of the green layer ($\alpha = 1.662$, $\beta = 1.676$, $\gamma = 1.684$, $c \wedge \gamma = 23^\circ$, $2V\alpha = 69^\circ$) 50 mol. per cent (Fe²⁺, Mn, Ti)-component. The following optical properties have been measured from the pyroxene contained in the reddish brown layer: $\alpha = 1.683$, $\beta = 1.689$, $\gamma = 1.715$, $c \wedge \gamma = 41^\circ$. The quartz generally occurs in the hornblende schists as separate, very narrow wedge-like veins.

The genesis of these quartzite lenses is not easy to understand. They are quite strongly tectonic lenses in metamorphic, probably sedimentogeneous gneiss. Perhaps they have first undergone powerful dynamometamorphism and then, in addition, developed a boudinage structure. The quartzite contains no relicts of the original rock. In it has been found, however, one layer 10 cm wide that has the same mineral composition and structure as biotite-plagioclase gneiss, but these quartzites need not have evolved out of such rock. The metamorphosis of the quartzites could scarcely have happened without marked simultaneous metamorphic differentiation, although the

original rock could have been somewhat stratified. In the biotite-plagioclase gneiss the biotite of the shear zones has altered partly into epidote and amphibole. In the diopside amphibolite of Rikkajoki, there exists an example of the banded texture partly produced by metamorphic differentiation, although the primary stratification prevails there.

In the biotite-plagioclase gneiss there are a few veins of crushed rock resembling pegmatite, which run parallel to the schistosity. Some of them consist of pegmatites with an abundant content of plagioclase ($\alpha = 1.535$, $\gamma = 1.543$, An_{15}) and, in addition, small amounts of quartz and potash feldspar, while some of the other ones consist of pegmatite containing abundant microcline ($\alpha = 1.523$, $\alpha = 1.529$) and, in addition, plagioclase, quartz, muscovite and biotite as well. The latter are more markedly metamorphosed than the former. In many cases they almost look like mylonite. There are also pegmatites intersecting the foliation. For example, one consists of pure white quartz. It is 3 m wide and some 50 m long. Its strike is $N45^\circ N$ and its dip vertical. The quartz grains contained in this vein have a prominent lattice orientation and an undulating extinction.

Banks of Tyrsävänjoki

The biotite-plagioclase gneiss situated along the banks of Tyrsävänjoki is rich in quartz, fine-grained (grain size 0.05—0.9 mm, commonly ranging between 0.2 and 0.7 mm), bluish gray, brownish pale gray on the weathered surface. In it occur sparsely scattered light gray, ovoid feldspar grains up to 5 mm in length. Its structure can still be described as blastoclastic. As cementing material it contains quartz, microcline and mica. The foliation of such types is relatively hard to detect. The mineral composition of the quartz-rich biotite-plagioclase gneiss, expressed in terms of percentages by volume, is as follows:

Microcline	4.3
Quartz	51.1
Plagioclase	26.7
Hornblende	1.1
Biotite	12.4
Epidote	1.6
Apatite	0.7
Muscovite	0.7
Opaque minerals	1.0
Other minerals	0.4
Total	100.0

The plagioclase grains (An_{15-25}) are subangular. The quartz grains are small (0.03—0.3 mm), with an undulating extinction, partly intergrown and also elongated. They are oriented in such a way that ϵ' is on the foliation plane. The microcline likewise occurs as small grains. The brown biotite flakes ($\gamma = 1.647$) are not nearly all on the foliation plane. The epidote and the hornblende are xenoblastic. The apatite occurs as small euhedral grains.

This gneiss is in some places more than commonly dynamometamorphic and its structure is cataclastic. Such portions of rock are regularly richer in potash feldspar than is the main type. And in these instances the potash feldspar also has the appearance of porphyroblasts. The apatite in it, moreover, consists of larger grains than normally.

This biotite-plagioclase gneiss further contains a number of dark gray mica schist intercalations (0.5 m thick) as well as pegmatite veins.

HORNBLLENDE GNEISS

On the northeastern shore of Nälantöjärvi, to the south of the biotite granite, there occurs hornblende gneiss of highly varying character. Alternating by layers, there are a very markedly foliated gneiss, a very weakly foliated gneiss, and a gneiss in which biotite is present as lenses 2 cm long, 1 cm wide and 1—2 mm thick. The last-mentioned gneiss has an outward appearance bound to arrest the eye. The amounts of biotite and quartz vary in hornblende gneiss, otherwise the differences between the types are mainly structural. In terms of percentages by volume, the mineral composition of the hornblende gneiss is as follows:

Quartz	6.6
Plagioclase	56.5
Hornblende	16.5
Biotite	11.8
Apatite	2.8
Opaque minerals	3.5
Other minerals	2.3
Total	100.0

The plagioclase ($a = 1.546$, $\gamma = 1.556$, $2V\gamma = 85^\circ$, An_{35-40}) is idioblastic and partly xenoblastic and it occurs as grains 0.3 to 1 mm long. To some extent it has undergone sericitization. The hornblende ($a = 1.651 =$ light green, $\beta = 1.663 =$ green, $\gamma = 1.669 =$ green, $c \wedge \gamma = 21^\circ$, $2Va = 75^\circ$, 33 mol. per cent (Fe^{2+} , Mn, Ti)-component) occurs as prisms 0.3 to 1.5 mm long, among which the large ones are poikiloblastic. The quartz occurs as roundish grains, measuring 0.1 to 0.2 mm in size. The biotite flakes ($\gamma = 1.649 =$ brown) do not run parallel even in lenses. In addition to the biotite, the lenses only contain opaque minerals. The polished section of the hornblende gneiss has revealed the presence of the following opaque min-

erals: pyrrhotite, with cobalt pentlandite, chalcopyrite, pyrites, galena, ilmenite, hematite, magnetite, graphite and hydro-pyrrhotite. Graphite occurs enriched in conjunction with the pyrites and also as lenses in the hornblende.

The hornblende gneiss brings to mind with respect to its mineral composition the amphibolite found on the NE-side of Nälantöjärvi (p. 44), except that it contains much more plagioclase.

MICROCLINE-PLAGIOCLASE GNEISS

Immediately north of the biotite-granite body, there occurs microcline-plagioclase gneiss. In it during the dynamometamorphic process the dark and the light minerals were apparently sorted and separated from each other, and the gneiss took on a somewhat augen-gneissoid appearance. The light minerals in the gneiss form pale brownish »eyes» that in sections perpendicular to the lineation are 1 to 3 mm thick and 5 to 10 mm long and that in sections parallel to the lineation are 1 to 4 cm long. The dark minerals surround these eyes. The grain size of the minerals in the gneiss varies between 0.1 mm and 1.5 mm, the usual size being 0.8 mm. This gneiss is generally rather homogeneous, but at the northern edge of the group of outcrops it has been found to occur as indefinite, fine-grained, gray fragments.

The microcline-plagioclase gneiss is granoblastic and cataclastic. Its constituent minerals are microcline, plagioclase, hornblende, biotite, (sphene, epidote, allanite (?), quartz, apatite, zircon, garnet and opaques). Table 8 presents the mineral composition of the microcline-plagioclase in percentages by volume and its chemical analysis.

The microcline ($a = 1.520$, $\beta = 1.524$, $\gamma = 1.527$) has a distinct cross-hatched structure, and it displaces the plagioclase. In spots one can notice the vanishing plagioclase in the form of streaks resembling perthite included in the microcline. The lamellae of the plagioclase ($\gamma = 1.540$, $a = 1.524$, An_{20}) have twisted and the marginal portions adjacent to the potash feldspar are albitic. Quartz occurs as small grains but only in small amounts. The biotite ($\gamma = 1.658 =$ brownish green, $a =$ yellowish light green) and the hornblende ($a = 1.680 =$ yellowish green, $\beta = 1.690 =$ dark green, $\gamma = 1.695 =$ bluish green, $c \wedge \gamma = 24^\circ$) occur together, and in conjunction with them there is quite an abundance of sphene as euhedral grains. The biotite may be seen invading the plagioclase. The apatite occurs as relatively large grains, like the zircon. The allanite is embedded in the epidote ($a = 1.709$, $\beta = 1.717$, $\gamma = 1.721$).

The microcline-plagioclase gneiss is penetrated by pale reddish brown, fine-grained granitic (partially aplitic) dikes. Fig. 28 shows one such gneiss-cutting dike and a dikelet of it. There is also an occurrence of pegmatite dikes, which in many instances have a considerable abundance of microcline ($a = 1.520$, $\beta = 1.524$, $\gamma = 1.527$). Both types of dikes may often

Table 8. Chemical analysis and modal mineral composition of microcline-plagioclase gneiss. NE-side of Näläntöjärvi, hamlet of Näläntö, Kiuruvesi commune. Analyst: A. Heikkinen.

	Weight		Molecular		Niggli	
	per cent	norm	norm		values	
SiO ₂	59.35	or 38.4	Or 38.0	si	205	
TiO ₂	0.82	ab 35.6	Ab 37.5	al	39	
Al ₂ O ₃	19.1	an 13.9	An 14.0	fm	18	
Fe ₂ O ₃	0.88	Q —	Q 0.1	c	15	
FeO	3.39	Σ 87.9	Σ 89.6	alk	28	
MnO	0.18	wo 1.7	Wo 1.4	ti	2	
MgO	1.00	fo 0.4	Fo —	h	6	
CaO	3.94	fa 0.8	Fa —	co ₂	1	
Na ₂ O	4.20	en 1.9	En 2.8	k	0.50	
K ₂ O	6.45	fs 3.3	Fs 3.0	mg	0.29	
P ₂ O ₅	0.16	il 1.7	Il 1.2	o	0.14	
CO ₂	0.15	mt 1.4	Mt 1.4	c/fm	0.80	
H ₂ O+	0.56	ap 0.3	Ap 0.3	qz	—7	
H ₂ O—	0.07	cc 0.3	Cc 0.4			
Σ	100.29	Σ 99.7	Σ 100.1			

Modal mineral composition:

Microcline	35.9	Apatite	1.3
Plagioclase	34.5	Zircon	0.4
Quartz	0.9	Opaque minerals	0.5
Hornblende	9.5	Other minerals	0.4
Biotite	8.6		
Sphene	4.4	Σ	100.0
Epidote	3.6		



Fig. 28. Fine-grained granite dikelet cuts across microcline-plagioclase gneiss (lower margin of figure). Northern corner of northeastern side of Näläntöjärvi, Näläntö, Kiuruvesi commune.

be observed to run parallel to the foliation, and more of them have been met with close to the biotite granite than farther away. At least to an extent, the dikes are likely to have made their penetration from the granite situated south of the gneiss. This cannot, however, be stated as a certainty insofar as not a single dike has been observed to branch off directly from the granite.

SW-SIDE OF NÄLÄNTÖJÄRVI

CUMMINGTONITE-BEARING HORNBLLENDE GNEISS

On the islands at the northern end of Näläntöjärvi the rock found is mainly cummingtonite-bearing hornblende gneiss. It is slightly banded. Biotite, among other minerals, occurs in it as very narrow bands of short length. Also the hornblende and the cummingtonite as well as the plagioclase have to a certain extent undergone sorting into streaks, which are visible only under the microscope. Mica gneiss is present in places in this gneiss as intercalations. The gneiss is foliated, but the foliation is poorly visible in places with a scanty content of mica. In color the rock is dark gray, and the grain size varies between 0.5 and 1 mm, but the plagioclase, in especial, occurs as much larger grains, which measure up to 5 mm in length.

The mineral content of the cummingtonite-bearing gneiss is as follows, expressed in terms of percentages by volume:

Quartz	1.5
Plagioclase	55.9
Hornblende	12.2
Cummingtonite	8.3
Biotite	14.8
Apatite	0.8
Opaque minerals	3.7
Other minerals	2.8
Total	100.0

The plagioclase ($a = 1.554$, $\beta = 1.559$, $\gamma = 1.563$, An_{52}) is euhedral and it contains some epidote as an alteration product. The subhedral cummingtonite ($a = 1.642 =$ colorless, $\beta = 1.653 =$ light grayish green, $\gamma = 1.672 =$ light bluish green, $c \wedge \gamma = 17^\circ$, $2V\gamma = 81^\circ$, 47 mol.-per cent Fe^{2+} -component) and subhedral hornblende ($a = 1.644 =$ greenish, $\beta = 1.656 =$ light green, $\gamma = 1.663 =$ light bluish green, $c \wedge \gamma = 21^\circ$, $2Va = 80^\circ$, 25 mol.-per cent (Fe^{2+} , Mn, Ti)-component) occur together in many instances as grain clusters. In some cases the cummingtonite is situated in the middle of the grains as well as in such a way that their c-axes run parallel. Hornblende also occurs as poikiloblastic grains larger than usual. Biotite ($\gamma = \beta = 1.623 =$ yellowish brown, $a =$ light yellow) occurs also in conjunction with hornblende and cummingtonite. The apatite grains are large and subhedral or anhedral. They are situated here and there as enriched grain clusters. Slight amounts of chlorite, epidote and muscovite are also to be detected.



Fig. 29. Mica gneiss resembling vein gneiss. Southwestern side of Nälantöjärvi, Remeskylä, commune of Kiuruvesi.

MICA GNEISS

Mica gneiss occurs in two places on the SW-side of Nälantöjärvi — at the SSE-end of the granite body and at the SW-corner of the region covered by the man (Fig. 25).

The mica gneiss at the SW-edge of the region is fine-grained and bluish gray. Here and there it shows a banded structure. In places the rock is rich in feldspar and in other places rich in mica. It is conspicuously metamorphosed. In many instances, this mica gneiss has the appearance of vein gneiss (Fig. 29). The veins are mainly composed of quartz and plagioclase or quartz alone. Very little potash feldspar has been detected in it. Fig. has been taken from the surface of the rock at right angles to the lineation. On surfaces parallel to the lineation, the light veins form long, uniform surfaces.

The chief minerals contained in the mica gneiss are plagioclase (An_{33} , $\gamma = 1.553$, $\alpha = 1.545$), quartz and biotite ($\beta = \gamma = 1.648$). The mica-rich types are in some instances garnet-bearing, though garnet has also been met with in mica-poor types. On the other hand, types are known that lack garnet. There is very little microcline in the mica gneiss. Accessory minerals are muscovite, chlorite, apatite, zircon and opaques.

There is very little mica gneiss visible on the SSE-wide of the granite body between the graphite-bearing schist and the cummingtonite-bearing hornblende gneiss. In the cases met with, it has been quite like the mica gneiss described in the foregoing. Here, however, the mica gneiss evidently contains various other rocks as intercalated layers. Various kinds of rock have been

observed in the intrusion and in the portions of the rock on the eastern edge of the body as fragments. In addition to the mica gneiss and the garnet-bearing mica gneiss, there are also present in it fragmented amphibolite, diopside gneiss, graphite-bearing gneiss and biotite-plagioclase gneiss, *etc.*

There is reason to believe that between the cummingtonite-bearing hornblende gneiss on the islands of Näläntöjärvi and the graphite-bearing schist at the southwestern end of the lake there occur other rocks besides the mica gneiss referred to; but they do not crop out. This may be said even though one does not take into account the gneiss fragments present in the intrusion on the southwestern side of Näläntöjärvi and which apparently originate from the layers between the rocks mentioned.

GRAPHITE-BEARING SCHIST

Graphite-bearing schist occurs between the pyroxene-amphibolite and the mica gneiss on the eastern side. It contains narrow layers of feldspar-rich and mica-poor fine-grained gneiss. In places in the graphite-bearing schist, pyrrhotite grains can be seen even megascopically. The size of the mineral grains in the schist varies between 0.1 and 0.8 mm. Here and there in the rock there occur a few large, poikiloblastic pyroxene and amphibole grains. The graphite-bearing schist is banded. In it generally microscopically thin hornblende- and biotite-rich bands alternate with pyroxene-rich layers. The latter contain quartz and a bit of potash feldspar. According to Marmo (1960), pyroxene-rich layers are always present in amphibole-rich sulphide-graphite schists. The mineral composition of the graphite-bearing schist, in which amphibole-rich and pyroxene-rich layers alternate, is as follows:

Quartz	7.0
Plagioclase	23.9
Pyroxenes	11.6
Hornblende	31.2
Biotite	10.8
Apatite	0.8
Opaque minerals	13.3
Other minerals	1.4
Total	100.0

The composition of the plagioclase contained in the graphite-bearing schist is An_{77} ($a = 1.566$, $\gamma = 1.576$, $2V\gamma = 82^\circ$). According to Tröger (1959), the properties of an amphibole ($2Va = 81^\circ$, $c\wedge\gamma = 17^\circ$, $a = 1.650$ = greenish yellow, β = pale green, $\gamma = 1.671$ = brownish light green) apply better to hornblende than to tremolite. According to Peltola (1960), the amphibole in calcareous black schist is usually tremolite. There is a small amount of hypersthene ($2Va = 55^\circ$, $c\wedge\gamma = 0^\circ$, 35 mol.-per cent Fs-component) present throughout the rock, but it decreases

in the pyroxene-rich thin bands, in which diopside occurs alone. Except in cases marked by the presence of pyroxene, quartz with an undulating extinction is found forming separate streaks. There is profusion of biotite (α = yellowish, $\gamma = \beta = 1.646$ = reddish brown) found in conjunction with amphibole prisms. Sphene occurs as inclusions in the amphibole. There are slight amounts of tiny apatite grains and chlorite. Graphite occurs abundantly as flakes 0.5 mm long, being concentrated in places. In addition to the graphite, the thin sections have been observed to contain the following opaque minerals: pyrrhotite, chalcopyrite, magnetite, sphalerite and hydropyrrhotite.

GARNET-BEARING PYROXENE AMPHIBOLITE

Between the mica gneiss and the graphite-bearing schist situated along the southwestern margin of the region included in the map, there occurs banded garnet-bearing pyroxene amphibolite. In it there alternately occur amphibole-rich and pyroxene-rich layers, which range in breadth from 0.1 mm to 20 mm. Garnet ($n = 1.798$) is present in the rock here and there, in many cases as porphyroblasts the size of the end of a man's thumb. Moreover, the outcrop contains blackish gray hornblende schist lenses 10 to 30 cm long. In spots the pyroxene amphibolite is brecciated by light gray veins of, *inter alia*, calcite and pale amphibole which measure about a centimeter in thickness (Fig. 30). Of the fragments created by brecciation, some have become slightly rolled, although the brecciating material did not probably

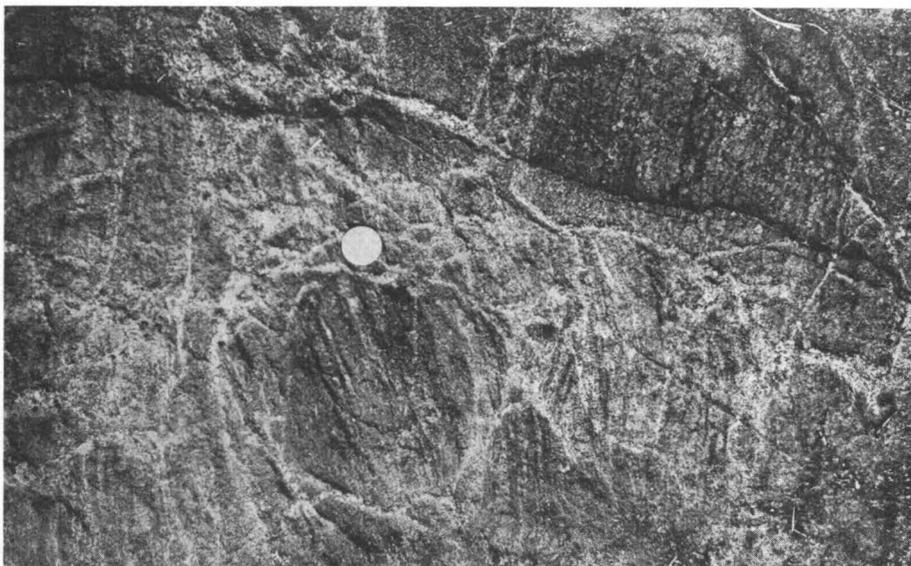


Fig. 30. Brecciated garnet-bearing pyroxene amphibolite, in which fragments have rolled somewhat. Brecciated material probably did not originate outside the amphibolite. Southwestern side of Nälantöjärvi, Remeskylä, Kiuruvesi.

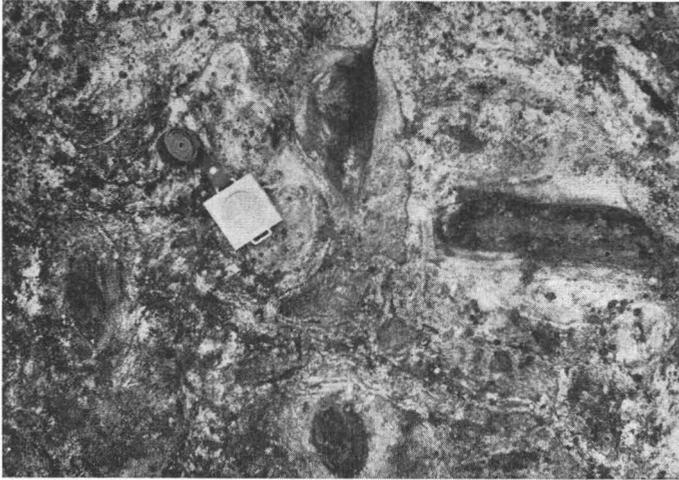


Fig. 31. Rolled pyroxene-gneiss fragments in metamorphic sedimentogeneous gneiss. Osmanki, Kiuruvesi commune.

originate even outside the pyroxene-amphibolite layer ¹⁾. The pyroxene amphibolite further contains grain clusters, in which the grain size is larger than in the host rock and in which there occurs, among other minerals, calcite, pyrrhotite, garnet and pale amphibole. Also granitic veins are to be noticed in the pyroxene amphibolite.

The pyroxene amphibolite is fine-grained (0.2—1 mm) and its structure is nematoblastic. Hypersthene occurs in it as poikiloblastic grains between 2 mm and 10 mm long. It occurs in both pyroxene-rich and amphibole-rich bands. The mineral composition of the rock in percentages by volume as measured from a specimen in which extremely narrow pyroxene-rich and amphibole-rich bands alternate, is as follows:

Plagioclase	34.5
Hypersthene	8.3
Diopside	33.4
Hornblende	17.6
Apatite	1.2
Opaque minerals	3.8
Other minerals	1.2
Total	100.0

¹⁾ This would seem to show that if the fragments in some breccia occur in arbitrary positions, it serves as no unambiguous proof that the brecciating material must be a magmatic rock, as has sometimes been contended. Fig. 31 shows another, similar breccia, in which the fragments are arbitrarily arranged. Its situation lies some distance south of the southern margin of the region in the map (Fig. 25). The brecciating material in it is a metamorphic sedimentogeneous gneiss, which had not originally been a magma, either.

The plagioclase ($a = 1.560$, $\beta = 1.565$, $\gamma = 1.568$, An_{62}) is euhedral. The hypersthene ($a = 1.712$, $\beta = 1.722$, $\gamma = 1.726$, 45 mol.-per cent Fs-component) does not generally occur as small grains, and in certain broad, pyroxene-rich bands there is little of it. The diopside ($a = 1.687$, $\beta = 1.696$, $\gamma = 1.718$, 34 mol.-per cent (Ca, Fe^{2+})-component) occurs both as small euhedral grains and sometimes also as large poikiloblasts. The hornblende ($a = 1.658$ = light green, $\beta = 1.674$ = brownish green, $\gamma = 1.680$ = brownish green, 45 mol. per cent (Fe^{2+} , Mn, Ti)-component) is subhedral and occurs only in some instances as large poikiloblasts. The rock further contains a bit of reddish brown biotite, which in certain narrow, separate zones is present in comparative abundance. The sphene is fairly abundant, but the apatite content is rather slight. In addition, the following opaque minerals have been noted in polished sections: pyrrhotite, ilmenite, chalcopyrite, graphite, hematite, sphalerite, pyrite and cubanite.

BANKS OF RIKKAJOKI

BIOTITE-PLAGIOCLASE GNEISS

Biotite-plagioclase gneiss occurs along the banks of Rikkajoki farthest eastward on the eastern side of the highway bridge. The biotite-plagioclase gneiss is to some extent rich in microcline. In places it contains intercalated layers of mica gneiss and in other places it resembles granite gneisses. In some cases it is likely to be somewhat banded. On occasion one will observe pegmatites associated with it, and in the proximity of the pegmatites it usually has narrow quartz veins running parallel to the foliation. In spots the gneiss nearly has the appearance of mylonite and it is marked by an abundance of shear surfaces trending in different directions.

On fresh surfaces the biotite-plagioclase gneiss is bluish gray and it is fine- or medium-grained. In spots it has light-colored »eyes» two to three mm long. In structure it is granoblastic or cataclastic. The main minerals are plagioclase, quartz, biotite and, in some cases, microcline as well. Accessory minerals are epidote, tremolite, muscovite, chlorite, sphene, apatite, zircon and opaques. The mineral composition of one microcline-rich type of biotite-plagioclase gneiss is as follows, in percentages by volume:

Microcline	20.7
Quartz	38.2
Plagioclase	14.0
Hornblende	0.4
Biotite	16.7
Muscovite	4.5
Epidote	0.7
Sphene	1.0
Chlorite	0.4

Apatite	1.4
Opaque minerals	1.5
Other minerals	0.5
Total	100.0

The composition of the plagioclase in this microcline-rich type is An_{28} ($\alpha = 1.542$, $\gamma = 1.550$). The microcline ($\alpha = 1.521$, $\gamma = 1.529$) has a distinct cross-hatched structure. The edges of the biotite flakes are very ragged and the grains themselves are twisted.

DIOPSIDE AMPHIBOLITE

The diopside amphibolite is situated on the sloping bank of Rikkajoki between the mica gneiss and the biotite-plagioclase gneiss on the western side of the highway bridge.

The diopside amphibolite is blackish gray and streaky. The streaks are a greenish dark gray. The thickness of the streaks varies from a microscopic line to a couple of centimeters. Many of them are lenticular. The average grain size of the rock is 0.4—1 mm, but in the streaks the diopside content, for example, is much greater than in other parts of the rock. In the streaks one can distinguish various types according to the main minerals, although all the stages between them are known. For example, the following principal mineral assemblages have been observed: diopside and plagioclase; plagioclase and epidote; plagioclase, diopside and epidote. In connection with these streaks, quartz lenses, which have less often been encountered elsewhere, occur, being 1—5 mm broad and 2—5 cm long). The quartz grains in them have a very prominent undulating extinction, are orientated and elongated and in spots appear to be intergrown. The mineral composition of this streaky diopside amphibolite is as follows, in percentages by volume:

Quartz	9.0
Plagioclase	20.1
Diopside	14.8
Hornblende	44.6
Epidote	5.0
Sphene	3.5
Tremolite	1.0
Apatite	0.7
Opaque minerals	0.7
Other minerals	0.6
Total	100.0

The hornblende ($2Va = 80^\circ$, 85° (marginal portion), $\alpha = 1.661 =$ light green, $\beta = 1.679 =$ green, $\gamma = 1.683 =$ bluish green, 45—50 mol. per cent (Fe^{2+} , Mn, Ti)-component) is idioblastic.

The diopside ($a = 1.690$, $\beta = 1.698$, $\gamma = 1.720$, $c \wedge \gamma = 42^\circ$, $2V\gamma = 56^\circ$, 35 mol. per cent (Ca, Fe²⁺)-component) occurs as small idioblastic grains, but in its greenish gray streaks it is poikiloblastic, in which case it contains as inclusions hornblende, plagioclase and tremolite. Also plagioclase ($a = 1.551$, $\gamma = 1.559$, $2V\gamma = 77^\circ$, An₄₅) is present as tiny grains, though in the greenish gray streaks it is noticeably larger of grain than elsewhere in the rock. The lamellae of the grains are then twisted and it is appreciably richer in anorthite ($a = 1.558$, $\gamma = 1.566$, An₆₀) than in general. Epidote occurs in the plagioclase as an alteration product. In the greenish gray streaks the epidote ($a = 1.717$, $\beta = 1.724$, $\gamma = 1.730$, 14 mol. per cent pistacite component) consists of large grains with a consistent occurrence of small, colorless, worm-like inclusions (quartz?), which in spots in the middle of the grains can be observed in great profusion while, in many cases, there are none at the margins. Pyrrhotite and sphene are concentrated in the spots rich in epidote and quartz. The diopside amphibolite, furthermore, contains a trifle magnetite, tremolite, reddish brown biotite and apatite.

The streakiness of the diopside amphibolite is largely primary, but especially in the areas rich in epidote and quartz, metamorphic differentiation would seem to contribute to its formation.

The diopside amphibolite contains blackish gray amphibolite and black hornblende schist intercalations, of which the thickest measures 1.5 m. The last-mentioned has small light gray lenses, from which there branch out light gray veins 2 to 3 mm thick, which in some cases run parallel to the foliation and in other cases cut across it. These veins and lenses have been corroded on the surface of the rock into hollows, which, at least to an extent, is due to their carbonate content.

The amphibolite's main minerals are plagioclase (An₄₀₋₄₅) and hornblende ($a = 1.658 =$ light green, $\beta = 1.672 =$ green, $\gamma = 1.678 =$ green, $c \wedge \gamma = 16^\circ$, 45 mol. per cent (Fe²⁺, Mn, Ti)-component). The accessory minerals are sphene (rather abundant), quartz, apatite, epidote, biotite and opaques.

The grain size of the hornblende schist is 1-2 mm. Its principal mineral is a green hornblende ($a = 1.662$, $\beta = 1.678$, $\gamma = 1.689$, $c \wedge \gamma = 20^\circ$). In addition to the hornblende there occurs a small amount of epidotized plagioclase, epidote, sphene, apatite, pyrrhotite, chalcopyrite, pyrite and ilmenite. The light gray lenses and veins in this rock contain diopside, epidote, plagioclase, sphene, tremolite and carbonate.

MICA GNEISS

Mica gneiss occurs underneath the bridge crossing Rikkajoki and also on the eastern side of it between the occurrences of diopside amphibolite and microcline-plagioclase gneiss. Along the western edge of the layer it is reminiscent of vein gneiss: Biotite-rich bands (varying in thickness between 0.5 and 2 mm) alternate with thicker bands rich in feldspar and narrow quartz veins running parallel to the schistosity. The gneiss also contains pegmatite lenses. In the eastern portions of the layer there are varieties of the rock resembling augen gneiss, in addition to fine-grained mica gneiss.

Varieties poor in quartz or in feldspar likewise occur there, just as do layers poor in mica.

The structure of all the gneisses mentioned is granoblastic and cataclastic. The mineral composition of the fine-grained mica gneiss is as follows, in percentages by volume:

Microcline	4.8
Quartz	38.2
Plagioclase	17.6
Biotite	32.9
Chlorite	1.2
Apatite	1.2
Opaque minerals	2.8
Other minerals	1.3
Total	100.0

The plagioclase grains (An_{30}) in these mica gneisses are in many cases badly broken. The quartz is marked by a strong undulatory extinction and its grains frequently observed to be elongated, taking on a lenticular form. In spots the quartz has broken up into very tiny particles, which form narrow bands running through the grains. The biotite flakes ($\gamma = 1.658 =$ dark reddish brown) are badly twisted and also fractured. They are in many cases surrounded by a very narrow ring formed of muscovite grains. The muscovite also occurs as slightly larger grains. The muscovite flakes are not usually situated in line with the foliation plane of the gneiss. Other minerals are sphene, chlorite and zircon. In places the gneiss contains sphene in great profusion.

MICROCLINE-PLAGIOCLASE GNEISS

Microcline-plagioclase gneiss can be seen on the bank of the Rikkajoki immediately to the west of the highway bridge crossing the river for a distance of some 600 m. The gneiss is either a light bluish gray or a reddish brown. It contains black, partly twisted, flaked mineral aggregates less than 1 mm thick, 3—10 mm broad, and 10—30 mm long, which run parallel to the foliation plane. A large proportion of the dark minerals in the gneiss have accumulated in them. Under the microscope the pale minerals can also be seen to be sorted to some extent in the same way apart from each other.

The microcline-plagioclase gneiss is granoblastic and cataclastic. The main minerals in the gneiss are microcline, plagioclase and biotite. The accessory minerals are apatite, sphene, quartz, epidote, chlorite, muscovite, carbonate, magnetite and zircon. Table 9 shows the chemical analysis of the gneiss and the quantity of minerals in percentages by volume.

Table 9. Chemical analysis and modal mineral composition of microcline-plagioclase gneiss. Northern end of Näläntöjärvi, Rikkajoki, hamlet of Näläntö, Kiuruvesi commune. Analyst: A. Heikkinen.

	Weight per cent		Weight norm		Molecular norm		Niggli values	
SiO ₂	61.70	or	41.1	Or	40.5	si	238	
TiO ₂	0.65	ab	45.6	Ab	47.5	al	43	
Al ₂ O ₃	18.76	an	3.6	An	3.5	fm	15	
Fe ₂ O ₃	0.25	Σ	90.3	Σ	91.5	c	5	
FeO	3.08					alk	37	
MnO	0.08	fo	0.8	Fo	0.9	ti	2	
MgO	0.71	fa	2.2	Fa	1.5	h	12	
CaO	1.26	en	0.8	En	0.8	co ₂	2	
Na ₂ O	5.40	fs	1.6	Fs	1.8	k	0.46	
K ₂ O	6.93	C	1.0	C	1.2	mg	0.27	
P ₂ O ₅	0.17	il	1.2	Il	0.8	o	0.06	
CO ₂	0.29	mt	0.5	Mt	0.4	c/fm	0.35	
H ₂ O+	0.51	ap	0.3	Ap	0.3	qz	—10	
H ₂ O—	0.03	cc	0.7	Cc	0.8			
Σ	99.81	Σ	99.2	Σ	100.0			

Modal mineral composition:

Microcline	36.6	Epidote	1.0
Plagioclase	41.4	Carbonate	1.5
Quartz	0.8	Chlorite	1.6
Biotite	9.7	Opaque minerals	1.2
Muscovite	3.5	Zircon and other minerals	0.5
Apatite	0.8	Σ	100.0
Sphene	1.4		

The microcline ($\alpha = 1.519$, $\beta = 1.523$, $\gamma = 1.526$) has a cross-hatched structure and it replaces the plagioclase. The plagioclase ($\alpha = 1.534$, $\gamma = 1.542$, An₁₀) contains as alteration products muscovite, epidote and carbonate. The perthite in the microcline is in certain instances joined together without any visible boundary to the adjacent plagioclase grain, with which it has a simultaneous extinction. The amount of perthite in the microcline varies. In some cases it seems to be lacking even from large grains almost totally, while in other cases it is present in profusion.

The biotite ($\gamma = 1.653$ = brown or greenish brown) is the predominant constituent of the grains producing the black splotches. Associated with it are large grains of apatite and considerable amounts of sphene. There is a higher content of chlorite and epidote in the reddish brown varieties than in the others.

The microcline-plagioclase gneiss contains as inclusions and narrow intercalations a slight amount of smaller-grained, bluish-gray gneiss, which has the same mineral composition as the host rock. But in these gneisses the minerals have not been sorted into separate lenticular groups. The fine-grained gneiss is evidently a relict in the microcline-plagioclase gneiss, which has been formed from it mainly in connection with the dynamic metamorphism.

CONCLUDING REMARKS

How should these three gneiss groups be connected? Mäkinen (1916) and Wilkman (1931) have connected the microcline-plagioclase gneiss situated on the bank of Rikkajoki and at the NNE end of Näläntöjärvi. Petrographically, they are very much alike. Similarly, they have joined the cummingtonite-bearing hornblende gneiss occurring on the islands of Näläntöjärvi and the hornblende gneiss situated on the northeastern shore of this lake. In addition to the content of cummingtonite, they differ from each other with respect to the composition of the plagioclase, but, on the other hand, both formations are rather unhomogeneous. The occurrence of mylonite zones, which are at the same time to some extent fault zones, makes for difficulty in combining the occurrences.

Wilkman (1931) has regarded the gneisses situated on the northeastern side of Näläntöjärvi as Kalevalian, probably in the main on account of the quartzite layers with intercalations of hornblende schist existing in the biotite plagioclase gneiss. But such occurrences are to be found also in southwestern Finland ¹⁾.

PLUTONIC ROCKS

INTRODUCTION

In the proximity of Näläntöjärvi, plutonic rocks occur in two different places: 1) on the southwestern side of Näläntöjärvi, near the northwestern corner of the lake and 2) on the northeastern side of the lake, west of the Tihilä-Tyrsävä highway. In the former locality there is a composite body about a km broad and 2 km long consisting of coarse-grained porphyritic pyroxene-bearing granite and even- and medium-grained pyroxene-hornblende granodiorite. In the latter locality there is a biotite granite body measuring roughly a square kilometer.

It has not yet been possible to divide the rocks in the western body into separate areas on the available 1 : 20 000 — scale map, for they form a kind of breccia. The relatively homogeneous porphyritic granite occurs in greater abundance at the northern end of the body. There is a greater content of non-homogeneous granodiorite than porphyritic granite. The greener and darker varieties of granodiorite occur to a greater extent in the marginal portions of the body and in the proximity of large fragments in the body. These contact varieties are finer of grain and also appear to be poorer in quartz than the other granodiorite types.

The biotite granite body is quite homogeneous, and it contains scarcely any fragments.

The granite body is completely discordant.

Both bodies contain a certain amount of aplites and pegmatites.

¹⁾ Mr. Toivo Mikkola, Phil. Mag., oral communication.

PYROXENE-BEARING PORPHYRITIC GRANITE

The pyroxene-bearing porphyritic granite is a coarse-grained, brownish-gray rock. It contains an abundance of rectangular microcline phenocrysts with more or less rounded corners and measuring between 4.5 and 7 cm in length and between one and three cm in width. In addition, it contains microcline grains one to 2.5 cm long and 0.2 to one cm wide. These grains have a fairly clear orientation (N15°W, vertical), and the mafic minerals in the rock as well as some of the rows of quartz grains twist around the large microcline grains.

The main minerals contained in the pyroxene-bearing porphyritic granite are microcline, quartz, plagioclase, hornblende, augite, hypersthene and biotite. Accessory minerals are zircon, apatite, opaques, monazite, chlorite and iddingsite.

The large microcline grains are frequently twinned, according to Karlsbad's law, and they contain an abundance of other minerals as inclusions, notably plagioclase and quartz. Plagioclase perthite occurs in the microcline ($\alpha = 1.519$, $\gamma = 1.527$) as what looks like strings of pearls and also as thin, straight bands. The latter lie nearly at right angles to the former. The cross-hatched structure of the microcline can be plainly seen. Plagioclase likewise occurs ($\alpha = 1.542$, $\gamma = 1.550$, An₂₅₋₃₀) outside the microcline grains. It is usually albitic against the potash feldspar, which in places contains tiny albite grains. Not infrequently the plagioclase may be seen to contain myrmekite alongside the microcline.

Tuttle (1952) and Tuttle and Bowen (1958) describe phenomena of the kind referred to in the relation of potash feldspar to plagioclase, and they stress the importance of post-solidification recrystallization to account for many of the textures exhibited by the alkali feldspars. Rogers (1961) demonstrates in his extremely interesting paper that, at least in certain cases, albitic rims on more calcic plagioclase grains and tiny, separate, unoriented albite grains have formed through the simultaneous crystallization of albite and potassium feldspar.

Hornblende ($\alpha = 1.691$ = yellowish green, β = green, $\gamma = 1.714$ = dark green, over 75 mol. per cent (Fe²⁺, Mn, Ti)- component) occurs as poikilitic grains and in some cases develops in conjunction with quartz structures resembling myrmekite. It also works itself into cleavages in the feldspars, producing a weed texture. The augite content ($2V\gamma = 52^\circ$, $c \wedge \gamma = 42^\circ$, $\gamma - \alpha = 0.023$, $\alpha = 1.714$, $\gamma = 1.736$) is frequently considerable. Orthopyroxene is also present in the rock ($2V\alpha = 78^\circ$, $\gamma - \alpha = 0.019$, $\alpha = 1.749$, $\gamma = 1.769$, Fs₇₅). The biotite ($\gamma = \beta = 1.691$ = dark brown) is likewise rich in Fe.

Zircon occurs as quite idiomorphic, strikingly large crystals. Apatite is likewise present as large crystals, but it is partly anhedral.

The aplites in the porphyritic granite contain hornblende, monoclinic pyroxene and biotite as mafic minerals.

PYROXENE-HORNBLLENDE GRANODIORITE

Pyroxene-hornblende granodiorites are even-grained rocks with a gray, dark gray, brownish gray or greenish gray color. Potash feldspar is present in them as grains 3 to 7 mm long and 2 to 3 mm broad. Hornblende occurs in them as ragged, indefinitely shaped, poikilitic grains up to 15 mm in diameter, which can be seen on the highly weathered surface of outcrops. The dark gray and greenish varieties of these rocks are usually to be found near the edges of an intrusion. The chemical analysis of pyroxene-hornblende granodiorite (Wilkman 1931, p. 141) corresponds well to the chemical composition of granodiorites (Table 10).

The rock's main minerals are microcline, plagioclase, quartz, hypersthene and hornblende. Accessory and secondary minerals are opaques, iddingsite (Sun 1957, Wilshire 1958 and Brown and Stephen 1955), apatite and zircon. Table 10 further gives the modal mineral composition of the rock.

Table 10. Chemical analysis and modal mineral composition of even- and medium-grained pyroxene-hornblende granodiorite. Northwestern corner of southwestern side of Näläntöjärvi, schoolyard at Remeskylä, commune of Kiuruvesi. Analyst: L. Lokka.

	Weight per cent	Weight norm	Molecular norm	Niggli values
SiO ₂	61.85	Q 14.5	Q 14.0	si 225
TiO ₂	1.88	or 16.7	Or 17.0	al 34
Al ₂ O ₃	15.80	ab 31.4	Ab 34.0	fm 32
Fe ₂ O ₃	0.32	an 18.1	An 18.8	c 15
FeO	8.35	Σ 80.7	Σ 83.8	alk 20
MnO	0.01			ti 5
MgO	1.02	wo 0.4	Wo 0.2	p 1
CaO	3.82	en 2.5	En 2.8	k 0.33
Na ₂ O	3.72	fs 11.9	Fs 10.0	mg 0.17
K ₂ O	2.80	il 3.7	Il 2.8	c/fm 0.46
CO ₂	—	mt 0.5	Mt 0.4	qz +45
H ₂ O	0.44			
Σ	100.01	Σ 99.7	Σ 100.0	
Modal mineral composition:				
Microcline	31.8			
Plagioclase	43.4			
Quartz	11.3			
Hornblende	2.8			
Pyroxenes and iddingsite	8.4			
Biotite	0.5			
		Opaque minerals		1.0
		Apatite		0.7
		Zircon		0.1
				100.0

The *microcline* ($\alpha = 1.522$, $\beta = 1.524$, $\gamma = 1.527$) contains quartz, plagioclase and, in some instances, hypersthene as well as inclusions. At the contact of *plagioclase* ($\alpha = 1.547$, $\beta = 1.552$, $\gamma = 1.554$, An_{35}) and microcline there occurs *myrmekite*. The composition of the plagioclase against the microcline is more albitic up to An_5 than otherwise. Tiny Ab-rich plagioclase grains also are present. The *hypersthene* ($\alpha = 1.731$, $\beta = 1.742$, $\gamma = 1.747$, $\gamma - \alpha = 0.017$, $2V\alpha = 70^\circ$, Fs_{65}) occurs both as poikilitic grains and as clusters of tiny grains. The *hornblende* ($\alpha = 1.671 =$ light green, $\beta = 1.684 =$ green, $\gamma = 1.692 =$ dark green, $c \wedge \gamma = 17^\circ$, 64 mol. per cent (Fe^{''}, Mn, Ti)-component) occurs as independent grains, but also along the margin of hypersthene grains (homoaxially, in such a way that their c-axes run in the same direction). Similarly, it penetrates the feldspars and forms a weed texture. There is a small amount of *biotite* ($\gamma = \beta = 1.679 =$ dark brown, $\alpha =$ light brown) and in many cases it forms in conjunction with quartz a texture resembling that of *myrmekite*.

BIOTITE GRANITE

The biotite granite is generally quite unoriented, but in places it has distinct crush and shear zones. In them the granite possesses a distinctly different texture, one acquired at a later date, such as the kind, shown in Fig. 35, which resembles augen gneiss. In color the biotite granite is a pale reddish brown or, in some cases, a light gray. Especially the types exhibiting shear foliation are apt to be gray. The grains vary in size, by and large, from 0.5 mm to 2 mm, and the rock contains a scattering of potash feldspar crystals 5 to 15 mm long.

The principal minerals contained in biotite granite are potash feldspar, quartz, plagioclase and biotite. Accessory minerals are muscovite, chlorite, pistacite, fluorspar, apatite, opaques and zircon. Table 11 presents a chemical analysis of this granite and the calculations made from a specimen taken from an exposure on the east shore of Näläntöjärvi that had been described as early as 1916 by Mäkinen.

The *potash feldspar* grains ($\alpha = 1.519$, $\beta = 1.522$, $\gamma = 1.525$) generally exhibit to some extent the cross-hatched texture of microcline. This texture is more clearly seen in small than in large grains. Large potash feldspar grains contain quartz and plagioclase, in especial, as inclusions. At the edges of the plagioclase particles *myrmekite* occurs in places with albitic edges. The included quartz grains frequently occur in rows in line with the crystal faces. *Biotite* likewise is present ($\gamma = \beta = 1.657 =$ black-brown, $\alpha =$ light brown) as inclusions, though in slight amounts, in large potash feldspar grains. The included biotites and the biotites against potash feldspar grains in general are in many instances somewhat chloritized. Next to the chloritized biotite one often finds allotriomorphic chlorite.

Table 11. Chemical analysis of biotite granite. NE-side of Näläntöjärvi, hamlet of Näläntö, Kiuruvesi. Analyst: A. Heikkinen.

	Weight per cent	Weight norm	Molecular norm	Niggli values
SiO ₂	71.89	Q 28.9		
TiO ₂	0.36		Q 27.3	si 380
Al ₂ O ₃	14.9	or 27.2	Or 27.5	al 46
Fe ₂ O ₃	0.40	ab 30.4	Ab 33.0	fm 12
FeO	1.71	an 6.7	An 7.0	c 8
MnO	0.05	Σ 93.2	Σ 94.1	alk 34
MgO	0.29			ti 2
CaO	1.33	en 0.7	En 0.8	h 10
Na ₂ O	3.60	fs 2.2	Fs 1.8	k 0.46
K ₂ O	4.65	C 1.5	C 1.6	mg 0.18
P ₂ O ₅	0.07	il 0.8	Il 0.6	o 0.16
CO ₂	0.0	mt 0.7	Mt 0.4	c/fm 0.77
H ₂ O+	0.56	ap —	Ap —	qz +144
H ₂ O—	0.04			
Σ	99.85	Σ 99.1	Σ 100.0	

In the plagioclase ($\gamma = 1.545$, $\alpha = 1.539$, An₂₀), especially in the middle of the grains, there are secondary mineral occurings as inclusions, among them muscovite and, as very tiny grains, epidote. The portions containing secondary minerals border sharply upon the inclusion-free marginal zones, which are richer in Ab than the middle portions of the plagioclase grains. The lamellae of the plagioclase grains are partially twisted in the sheared and crushed portions of the rock. Muscovite occurs not only in the plagioclase but also at the edges of the biotite as flakes lying in many directions. This muscovite, too, crystallized at a comparatively late date.

Fluorspar is frequently met with as allotriomorphic grains near biotite that has altered into muscovite.

THE RELATION BETWEEN PLUTONIC ROCKS AND THEIR ORDER OF ERUPTION

In the intrusion situated on the western side of Näläntöjärvi, the relation between the pyroxene-bearing porphyritic granite and the pyroxene-hornblende granodiorite is to be observed best in the northern portions of the body, that is, the outcrops in the vicinity of the Remeskylä elementary school.

Wilkman (1931, p. 141), describing the exposures in the schoolyard at Remeskylä, writes that the pyroxene-bearing porphyritic granite contains broad, dike-like portions of pyroxene-hornblende granodiorite. He adds, to be sure, that the borders between the different varieties of rock are indistinct and that the transitions appear to be gradual. The present writer has not found evidence of such gradual transitions.



Fig. 32. Edge of fragment of coarse-grained, pyroxene-bearing granite from even- and medium-grained pyroxene-hornblende granodiorite. Near schoolhouse in hamlet of Remeskylä, in the northwestern corner of the southwestern side of Nälantöjärvi, Kiuruvesi commune.

Eastward and also southward from the site referred to, pyroxene-bearing porphyritic granite fragments have been found in the pyroxene-hornblende granodiorite. In Figure 32 one can see the corner of one such fragment. This particular fragment measures about 10×15 meters. It borders quite sharply on its country rock, which consists of pyroxene-hornblende granodiorite. The outline of the angular fragment has been wholly exposed. The mafic minerals of the country rock conform to the shape of the fragment, and this feature appears to indicate some sort of flow structure. The fragment probably was a rigid body when the pyroxene-hornblende granodiorite intruded around it.

When one regards the pyroxene-hornblende granodiorite veins mentioned by Wilkmann (1931, p. 141) and the afore-described pyroxene-bearing porphyritic granite fragments as incontestable proof that the pyroxene-hornblende granodiorite is the younger of the two rocks, then it is easy to interpret also the contact relations of the exposures in the schoolyard at Remeskylä unambiguously: The pyroxene-bearing porphyritic granite in the schoolyard is cut by a fine-grained, light brownish aplite vein. They are both cut by an even-grained pyroxene-hornblende granodiorite (Fig. 33). All three

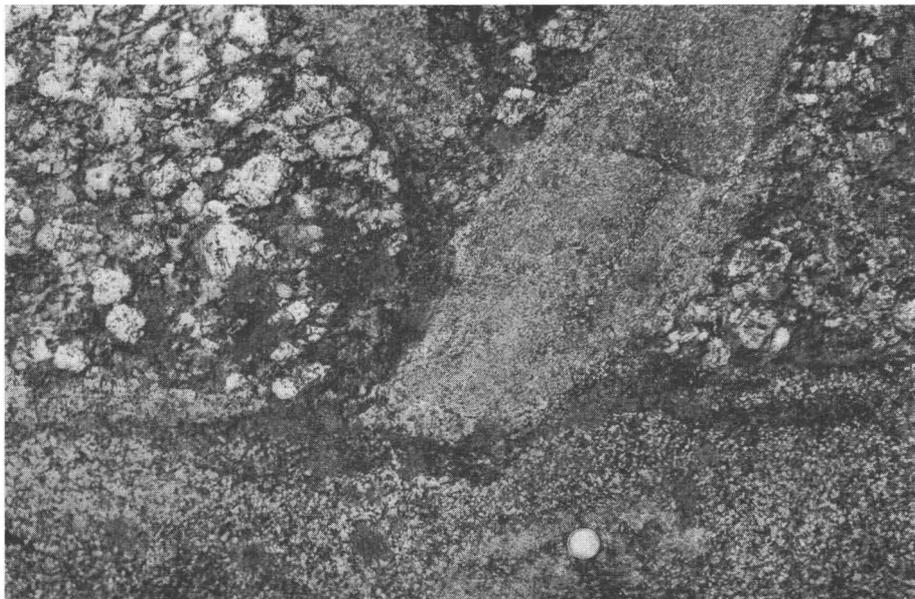


Fig. 33. Coarse-grained, pyroxene-bearing porphyritic granite, containing an aplite vein with an apophysis. The granite and the aplite are cut by even- and medium-grained pyroxene-hornblende granodiorite (lower margin of figure). Northwestern corner of southwestern side of Nälantöjärvi, schoolyard at Remeskylä, Kiuruvesi commune.

rocks are intersected in turn by aplite and pegmatite veins, but these, of course, represent the pegmatites and aplites of the even-grained pyroxene-hornblende granodiorite. Thus, the pyroxene-bearing porphyritic granite and its aplites and pegmatites are older than the even-grained pyroxene-hornblende granodiorite and its aplites and pegmatites.

In this body there must, therefore, have occurred at least two different magma eruptions followed by differentiation *in situ*. Both erupted masses have complete final crystallizations: aplites and pegmatites. The main rock of the earlier eruption, the porphyritic pyroxene-bearing granite, is richer in Si and K than the main rock component of the later eruption, the even-grained pyroxene-hornblende granodiorite.

Whether what is involved consists of successive magmatic pulsations (Case 3), as described by Harry and Richey (1963), or of two independent eruptions in succession — since local co-existence does not suffice as an explanation —, it is not easy to decide. If successive magmatic pulses are the answer to the question, then at least what we apparently have is a reversed differentiation sequence, as in the case of rapakivi. In this body, however, no proof exists, as in rapakivis, of the rocks' belonging to the same

magmatic series. The porphyritic granite is distinctly foliated, and it is oriented in the same way as the metamorphic rocks situated in the area. The pyroxene granodiorite, on the other hand, is generally unoriented, except in the foregoing cases and those described on p. 68. This circumstance suggests that there occurred two separate, possibly independent, magma eruptions. The magmas participating in the separate eruptions (the biotite granite magma as the third) need not even have been produced under similar conditions and in the same way, for granites are created in many different ways (*e.g.*, Bowen 1948, Gilluly 1948, Read 1945, Wegman 1956, Mehnert 1959, Winkler 1961, Saridse *et. al.* 1961, and Marmo 1963).

The intrusion on the eastern side of the lake is a separate one. It does not even touch the intrusion on the western side of the lake; between them are other rocks. On the islands at the northern end of Näläntöjärvi there are metamorphic supracrustal rocks, and the majority of the rocks situated along the shore between the lake and the granite body comprising the intrusion area on the western side are likewise metamorphic supracrustal varieties. Some of these shore rocks and, in general, the exposures on the eastern side of this intrusion area are typical of the border varieties of the granodiorites here. In color they are greenish gray or dark gray and finer of grain than the other granodiorite types present in the intrusion. The phenomenon resembles that found in the boundary zones of rapakivi, as previously reported by the present author (Savolahti 1962, pp. 38—44). The occurrence of such border varieties of rock evidently constitutes the best proof of the separateness of these two granite bodies — especially in the event that the rocks of the area on the western side are older than those of the intrusion on the eastern side.

Of the plutonic rocks found in the environs of Näläntöjärvi, the biotite-granite intrusion is the youngest, also according to earlier views (Wilkman 1931 and Mäkinen 1916). The rocks of the biotite-granite intrusion are less metamorphosed than those of the area on the western side. This intrusion, furthermore, is more massive than the rock on the western side, which contains a few narrow granite veins that look like biotite granite. One of them has even yielded a fragment (30×10 cm) consisting of even-grained pyroxene-hornblende granite. None of the available evidence is conclusive, however, not even the fact just mentioned, for it has not been demonstrated for sure that what we have is actually a vein extending from the biotite-granite area on the eastern side — but no evidence to the contrary exists, either.

Probably, the intrusion on the western side of Näläntöjärvi is older than the one on the eastern side. In the former body, coarse-grained pyroxene-bearing porphyritic granite cuts through the pyroxene-hornblende granodiorite. All three rocks have their own aplite and pegmatite.

According to Wilkman (1931, p. 120), the extension of the intrusion area at the northern end of the lake contains potash-natronsyenite. But this syenite is no magmatic rock; rather is it a conspicuously metamorphic layer belonging to the area's series of metamorphic, supracrustal rocks. On the other hand, this layer is no more metamorphosed, considering the conditions, than the rest of the rocks belonging to the area's supracrustal series.

CRUSH AND SHEAR ZONES IN THE PLUTONIC ROCKS

In all the rocks, including the plutonic rocks, around Nälantöjärvi — coarse-grained pyroxene-bearing porphyritic granite, even-grained pyroxene-hornblende granodiorite and biotite granite — there occur mylonite zones. In general, they trend in two different directions: $N25^{\circ}-15^{\circ}W$, vertical and $N75^{\circ}-55^{\circ}E$, $90^{\circ}-75^{\circ}S$. The first-mentioned direction conforms at the same time to the nearly general foliation of the area, whereas the latter sharply cuts across the foliation. The lineation measured from the latter is $N25^{\circ}W$, $50^{\circ}SE$.

PYROXENE-BEARING PORPHYRITIC GRANITE AND PYROXENE-HORNBLLENDE GRANODIORITE

At the northwestern corner of the intrusion situated on the SW-side of Nälantöjärvi, a certain mylonite zone cutting across the general foliation of the rocks is clearly visible. The trend of this zone is $N70^{\circ}E$, $70^{\circ}S$, and it is about 5 m wide. It intersects both coarse-grained pyroxene-bearing porphyritic granite and even-grained pyroxene-hornblende granodiorite, which shows that it is younger than both the latter rocks.

Immediately north of this mylonite zone, the terrain lies much lower down, marking the beginning of broad, level stretches of bogland. Here one gains the impression that different segments of the earth's crust were situated on opposite sides of the mylonite zone, although this one, visible zone could not, perhaps, be regarded as the cause.

Fig. 34 shows a part of this mylonite zone, and it reveals the structure. The zone contains rocks of different types, which border on each other sharply and alternate as narrow bands. Thus the entire mylonite zone takes on a banded appearance. It contains bands of light gray, fine-grained rock and others of a dark gray shade but likewise fine of grain, both of which contain few or no potash feldspar porphyroblasts. Then there are bands the fine-grained rock of both light and dark gray of which both contain eye-shaped microcline porphyroblasts 2 cm long in abundance. The result is outwardly a beautiful augen gneiss. All the bands have evolved, however, through mylonitization out of the body's granite and granodiorite.

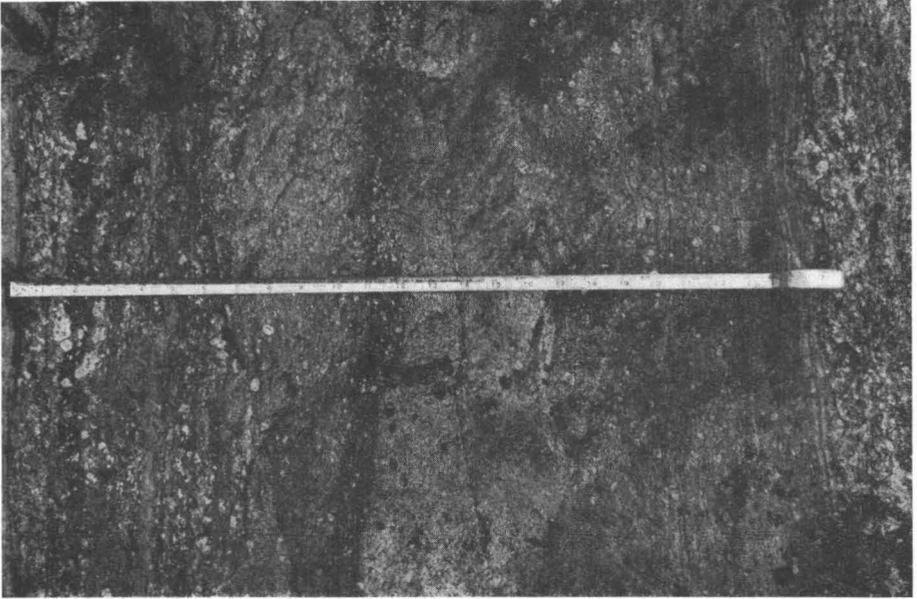


Fig. 34. Mylonitized pyroxene-hornblende granodiorite, which is banded in structure and composition. Northwestern corner of southwestern side of Nälantöjärvi, in the hamlet of Remeskylä, Kiuruvesi commune.

It is biotite that gives the color to the dark gray, fine-grained mylonite. In addition, the principal minerals include quartz, plagioclase (An_{25}) and microcline. The light types are rich in quartz, microcline and plagioclase — sometimes in all three, sometimes with one or two of the minerals missing. In many instances, the quartz grains have grown together to form twisted lenticular stripes conforming to the foliation of the rock. The eye-shaped microcline grains represent either just a single microcline porphyroblast or a cluster of tiny particles. The plagioclase likewise occurs in spots as grain clusters. In places the porphyroblasts have obviously rolled around. The perthite of the microcline porphyroblasts is extraordinary: it is composed of plagioclase blebs rectangular in shape with rounded corners, 0.02 mm thick and 0.05 mm long, which form chains resembling strings of pearls. Also myrmekite occurs in places at the contact between the microcline and the plagioclase. Besides the minerals already mentioned, the mylonite zone contains the following as well: epidote, muscovite, sphene, zircon, opaques (very slightly) and apatite.

This mylonite zone further contains a narrow aplite vein running parallel to the foliation and a wedge-like aplite lens, as well as another aplite vein running $N25^{\circ}W$, the strike of which runs diagonally against the length of

the vein. The mylonite zone also has stretches in which there are so many feldspar porphyroblasts stuck together that the rock resembles crushed and sheared pegmatite.

Such an apparently banded mylonite would seem to have originated when rock within the mylonite zone failed to move during the process of mylonitization simultaneously throughout, with equal force and in the same way. Rather did movements of varying strength and different kinds evidently take place in the several »layers» composing the mylonite zone, movements that are apt to occur either simultaneously or at different stages of the mylonitization process at different places. It is also possible that the mylonite zone contains adjacent rocks produced during several mylonitization processes.

Tuominen (1957) has published a map, based on an exceptional number of geological and geophysical investigations, which depicts the fault system of the Orijärvi area. In a later study (Tuominen 1962, p. 1) he has divided the system into three groups: faults of older origin, those of contemporaneous age, and those of later origin than the regional amphibolite facies. The ones in the first category he refers to as relict faults (Tuominen 1961, p. 512) and reports (Tuominen 1962, p. 2) that these old faults were active in later times.

The intrusion on the northwestern side of Nälantöjärvi has perhaps more mylonite zones trending $N25^{\circ}-15^{\circ}W$, $70^{\circ}-90^{\circ}W$ than in a crosswise direction. In their main features they are like the mylonite zone described in the foregoing, which sharply cuts across the foliation. Considering that the trend of the mylonite zones we have been discussing here very nearly corresponds to the prevailing direction of the foliation in the area, it is tempting in certain cases to contend that they do not represent mylonite zones but rather strongly metamorphosed intercalations, of perhaps sedimentary origin, in the coarse-grained porphyritic pyroxene-bearing granite and the even-grained pyroxene-hornblende granodiorite. Such a contention is likely in some instances to seem natural, if only because this intrusion contains considerable quantities of various fragments of the country rock. But at the same time it would be opportune to state here that the kind of pseudolayer parallel to the foliation shown in Fig. 34 could easily but erroneously be interpreted as a potash-feldspar porphyroblast-bearing granitized layer and the original rock alongside it as a totally granitized product of it. Descriptions also of this kind, purporting to present »examples of the granitization», are not unknown in Finland.

The aplites situated in the crush and shear zones apparently originated during the stage of mylonitization or, perhaps, more accurately stated, late during this stage. In the environs of Nälantöjärvi, plutonic rocks erupted during at least three different periods. Mylonite zones were produced in

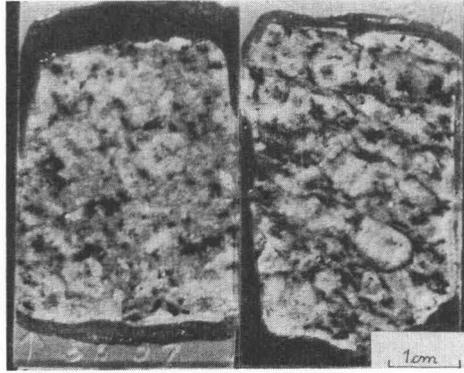


Fig. 35. Biotite granite (on left), augen-gneissoid biotite granite (on right). Nälantö-Kiuruvesi.

connection with at least each of these eruptions, in addition to which such zones could have developed — and presumable did — at other times, too, and for other reasons. For example, the crush and shear zones contained in the pyroxene-bearing porphyritic granite could have evolved during late movements that perhaps followed an eruption, though they might just as well or more readily have been produced by some subsequent movement. In the latter event, the aplites in the crush and shear zones did not evidently crystallize out of the residual magma left over from the intrusion's main crystallization stage but rather formed out of the magmatic liquids resulting from the mylonitization process.

Thus do we find ourselves confronted with the task of distinguishing aplite and pegmatite veins of different ages in the bodies. Each of the three intrusions has its own aplite and pegmatite content, which has crystallized out of residual liquids; and to each mylonitization process there perhaps belongs its own share of aplite and pegmatite. It is interesting to try to discover what differences might exist between the aplites and pegmatites of these different generations.

BIOTITE GRANITE

Mylonite zones situated in the biotite-granite intrusion on the NE-side of Nälantöjärvi were described as early as 1931 by Wilkman (p. 142). According to this investigator, the granite at the boundary of the schist zone in the outcrop at Hukka, toward the north, has undergone powerful pressure and alters on the eastern slope of the outcrop into a schistose mylonite containing microcline eyelets 0.5—1 cm long.

In this biotite granite near the eastern margin of the body, there runs a crush and shear zone about 30 meters wide in the direction N5°W, 70°W.

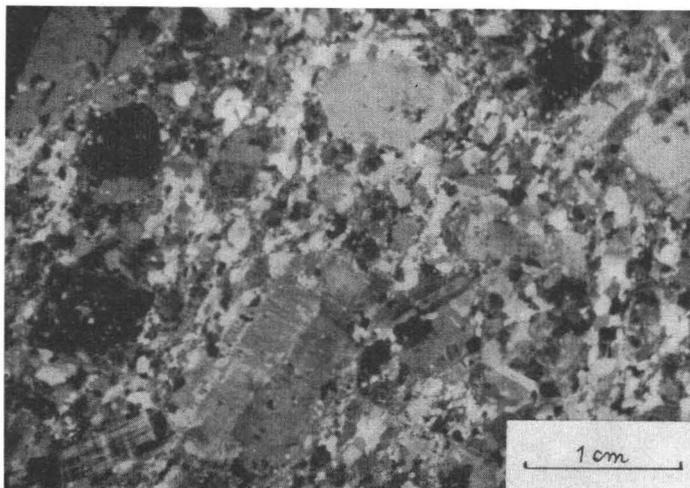


Fig. 36. Slightly sheared biotite granite. Thin section, crossed Nicols. Northeastern side of Nälantöjärvi, hamlet of Nälantö, Kiuruvesi commune.

From it has been measured the lineation $N10^{\circ}W, 30^{\circ}N$. In places the crush-and-shear zone contains layers that can still be positively identified as biotite granite. Fig. 35 shows one such augen-gneissoid rock created out of biotite granite by slight crushing and shearing, the origin of which is clear beyond doubt. This mylonite zone also includes rocks that have been mylonitized to a far more advanced degree, among them being types described in the foregoing.

Proceeding eastward along this body one will encounter more and more mylonite layers resembling highly sheared pegmatite that alternate with layers still recognizable as biotite granite and that border on the biotite granite either sharply or by gradually merging. In some places the sheared pegmatite also appears to cut across mylonites.

Fig. 35 shows slightly sheared biotite granite in this crush-and-shear zone — but the original rock is still easy to recognize. Fig. 35 shows the biotite granite proper. The sheared biotite granite contains an abundance of slickensides running in different directions and cutting across each other, which have curved around feldspar grains. It is in this way that the rock gained its augen-gneissoid appearance. Yet this variety of granite contains the same minerals as does fresh biotite granite, though more chlorite, epidote, sphene and muscovite (?). Furthermore, its biotite has undergone more alteration, there is more myrmekite at the edges of the plagioclase, and the cross-hatched structure of the microcline can be seen much more clearly in the potash feldspar grains (Figs. 36 and 37) than in the case of biotite gran-

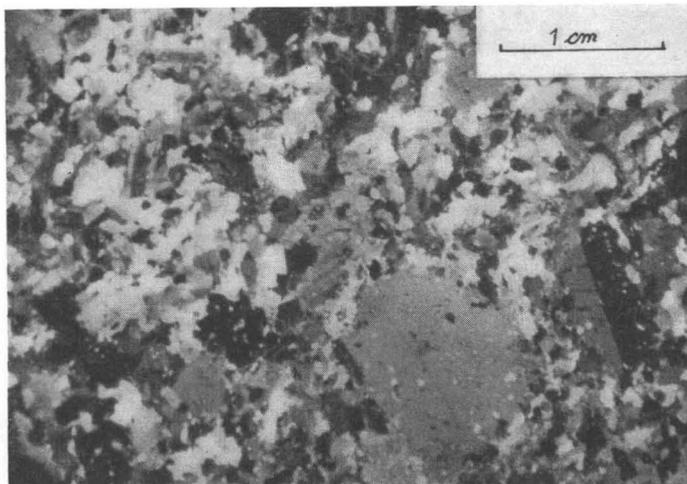


Fig. 37. Biotite granite. Thin section, crossed Nicols. Nälantöjärvi, hamlet of Nälantö, Kiuruvesi.

ites proper. As evidence of mylonitization, it contains narrow shear zones and exhibits a mortar texture. The shear zones are mainly composed of micas and tiny quartz grains. In places they run through feldspar crystals, sundering them.

The distinctly more abundant and more conspicuous occurrence of the cross-hatched microcline texture in the augen biotite granite than in the regular biotite granite appears to indicate that tectonic movements contribute toward promoting the creation of a cross-hatched texture in microcline. Perhaps tectonic movements have the same effect on the growth of the degree of triclinicity in general in potash feldspar. Perhaps partly for this reason the potash feldspar in synkinematic granites is generally microcline. Even according to definition, thoroughgoing movements have occurred in them subsequent to their crystallization. Karamata (1961, p. 129) presents examples of how in granites and other rocks subjected to tectonic pressure, microcline with a cross-hatched structure occurs to a greater extent than in other rocks of otherwise similar character. In Finland Marmo (1955, 1959, 1962 and 1963) has written voluminously about potash feldspars and in so doing has also taken a stand against Laves (1950, 1951, 1952, 1955 and 1960) and his disciples, who have specialized in the study of feldspars.

SUMMARY AND CONCLUSIONS

Two parts of the so-called Vieremä-Kiuruvesi schist belt have been described in the foregoing: the Salahmi area, or the eastern margin of the schist belt, and the Näläntö area, or the western margin of the schist belt. The central portion of the schist belt has been deliberately omitted, and it measures over ten kilometers across. This tract contains many bogs and moraines, but few exposures. Wilkmann (1929) and Mäkinen (1916) had reported a total of about thirty exposures from this tract, and it was on such a basis that they connected the occurrences (Fig. 1). The present author found roughly twice as many outcrops, but none of them provided clear indications as to how a connection of the separate observations might be carried out. For that reason, the writer has deemed it wiser to describe all the observations appearing to be fundamental from each of the border areas regardless of whether they seem to fit into any system or not. A beautifully generalized system could easily cover opposing observations — that is why any attempt at a unified picture should be confined to a separate publication.

In the basal conglomerate at Valkeiskylä, the pebbles are angular, and the cementing material contains much clay and more carbonate than in other basal conglomerates of the Salahmi area. West of the Vieremä church, the basal conglomerate contains a great variety of rocks reduced to pebbles of greatly varying sizes. But the pebbles in this conglomerate are more rounded and the cementing material is sandier than in the conglomerate of Valkeiskylä. Both conglomerate formations are fairly thick, but, in contrast to the Valkeiskylä conglomerate, it is distinctly stratified. The pebbles in the basal conglomerate of Murennusmäki are small and do not contain any large variety of rocks. The conglomerate is distinctly stratified and the thickness of the entire formation is appreciably exceeded by that of the basal conglomerates farther south. The basal conglomerate of Lähde-mäki can almost be designated as a feldspar-bearing quartzite, and it occurs as a relatively thin layer. It has undergone the most far-reaching sorting and, among other features, it has exhibited current bedding.

Overlying the basal conglomerate is a formation with narrow alternating layers of quartzite and mica schist, which exhibit bedding. At Valkeiskylä the formation is rather thin and thickens toward the north, being thickest

at Lähdemäki. Tourmaline is among the minerals met with in these rocks. According to Simonen and Kouvo (1951), the slates of the Tampere area contain between 0.03 % and 1 % B_2O_3 . According to Simonen and Kouvo (1951) as well as Eskola (1963), it could signify deposition in salt water.

Schists containing calcareous minerals occur in the central and southern parts of the Salahmi area, but none has been met with in the northern parts of the area. They contain mica schist and quartzite as intercalations. Whether they also contain volcanogenic rocks, I would not dare to say for sure, as I have discovered no unmistakable structural evidence of such a condition.

The thickest formation of staurolite mica schist alternating in layers with mica schists exists at the northern end of the Salahmi area. From there the staurolite mica schist zone extends all the way to the southern end of the Salahmi area — getting thinner, to be sure, the whole way. In one place the staurolite has turned to chlorite and muscovite after its crystallization.

In this resumé, the changes in the thickness of the formations and the variations in the degree of sorting and the composition of the rocks agree well with the general principles of sedimentation (Scroek 1948, Pettijohn 1949, Dunbar and Rogers 1957, and Krumbein and Sloss 1963).

In the quartzite conglomerate between Luvejoki and Rotimojoki, pebbles were formed apparently to some extent during the movements that occurred after the sedimentation; but the observations so far made do not suffice to support the argument that it is completely of tectonic origin. Karelian formations are known to contain quartzite conglomerates, which many investigators consider to be intercalations and which are of the same description as the ones under discussion and which some observers are inclined to view as tectonic breccias.

On the eastern edge of the Salahmi area lies the so-called Kukkomäki conglomerate. It is very conspicuously deformed and the original structure of the rock is faintly distinguishable. Both sides of it consist of the same kind of rock. That it represents the basal conglomerate as Mäkinen thought (1916), is by no means certain on the basis of the observations made in connection with the present investigation. Wilkman (1929) shifted the base of the schists in the Salahmi area to the northeastern shore of Näläntöjärvi.

The gneisses and schists of the Näläntö region are conspicuously metamorphosed, and they scarcely have any relicts left of their previous structure. It has been related in the foregoing on what basis Mäkinen (1916) and Wilkman (1929) combined them and on what basis the latter (1931) included them among Kalevalian formations. Nothing indicative of a basal conglomerate has, however, been found in the Näläntö region, but, on the other hand, faults trending approximately $N20^\circ W$ and $N65^\circ E$ have come to light. Near the point intersection of the differently trending faults, magmas had

erupted several times in a sequence that does not fit in very well with the classical theory of crystallization differentiation.

Basal conglomerates do occur at the eastern edge of the Salahmi area. Should the basement complex underneath them be compared to the 2 800 million-year-old basement of the Karelian formations or to the Svecofenides? The schists overlying this basal conglomerate have always been counted among Karelian formations (though not representative of their Jatulian type, as Väyrynen (1945) has pointed out). Where is the basal conglomerate situated on the western edge of these schists? It has not been found for sure in the Salahmi area, no sign of it whatsoever has been noticed in the Näläntö area, and it is hardly conceivable that farther west one would have better luck. If the basement at the eastern edge of the Salahmi area represents a counterpart to the basement of the Karelides and if the Karelides and the Svecofenides belong to the same mountain chain, one need not expect to find basal conglomerates along the western margin of the schist area. It can nevertheless be argued that a basal conglomerate has existed in the vicinity and that relicts of it are to be seen, for example, at the western edge of the Salahmi area.

At the northern end of the Salahmi area there is a fault, and there is one also in the northern part of the Näläntö area. These faults may belong together, or, at least, to the same group. Abruptly, in their northern part, these faults take on the same trend and nearly join each other's extension. They belong, at least, to a group of faults running in the same direction. In the Näläntö area, perhaps, there lies a fault slightly farther north. It is further to be noted that the long northern margin of the Vieremä-Kiuruvesi schist area (Fig. 1) nearly coincides with the fault here mentioned and runs in the same direction.

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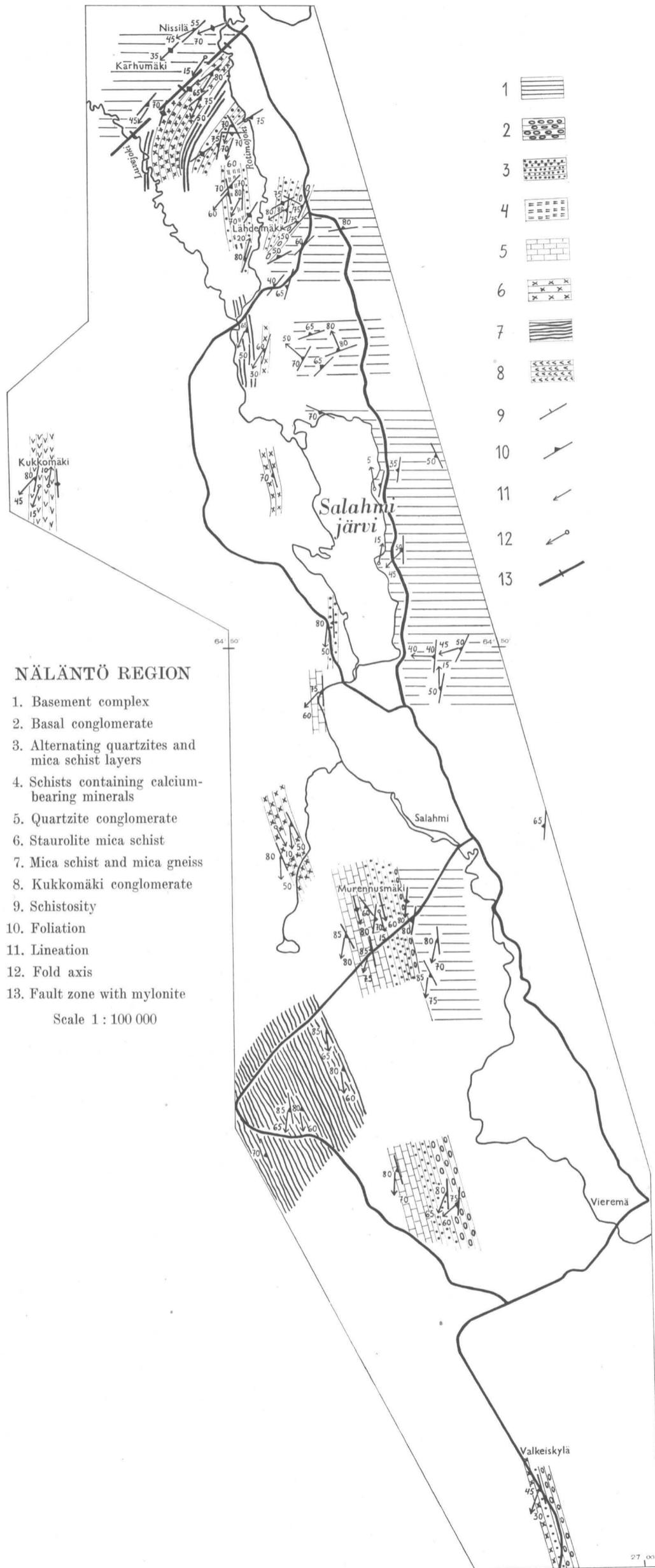
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NÄLÄNTÖ REGION

1. Basement complex
2. Basal conglomerate
3. Alternating quartzites and mica schist layers
4. Schists containing calcium-bearing minerals
5. Quartzite conglomerate
6. Staurolite mica schist
7. Mica schist and mica gneiss
8. Kukkomäki conglomerate
9. Schistosity
10. Foliation
11. Lineation
12. Fold axis
13. Fault zone with mylonite

Scale 1 : 100 000

ON BASALT ROCKS WITH NATIVE IRON IN DISKO,
WEST GREENLAND ¹⁾

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ABSTRACT

The article deals with petrographic features characteristic of basalt rocks containing native iron in Disko island. The association of anorthositic component with these basalts has been discussed as well as the role of reddish spinel as an indicator of iron bearing basalts and related metalliferous formations in Disko area has been emphasized. A new analysis on iron basalt of Asuk area, Disko, has been represented.

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INTRODUCTION

The areas of basalt formations in the Disko-Syartenhuk region, West Greenland (Fig. 1) belong to a greater sphere called the Brito-Arctic basalt province. As proposed in earlier literature (Krueger, 1928; von Wolff, 1931) the eruption of the West Greenlandic basalts was promoted by fissures formed through block-faulting, presumably initiated during the last part of the Cretaceous epoch. The basalt eruptions probably started during the Eocene

¹⁾ Received October 5, 1964

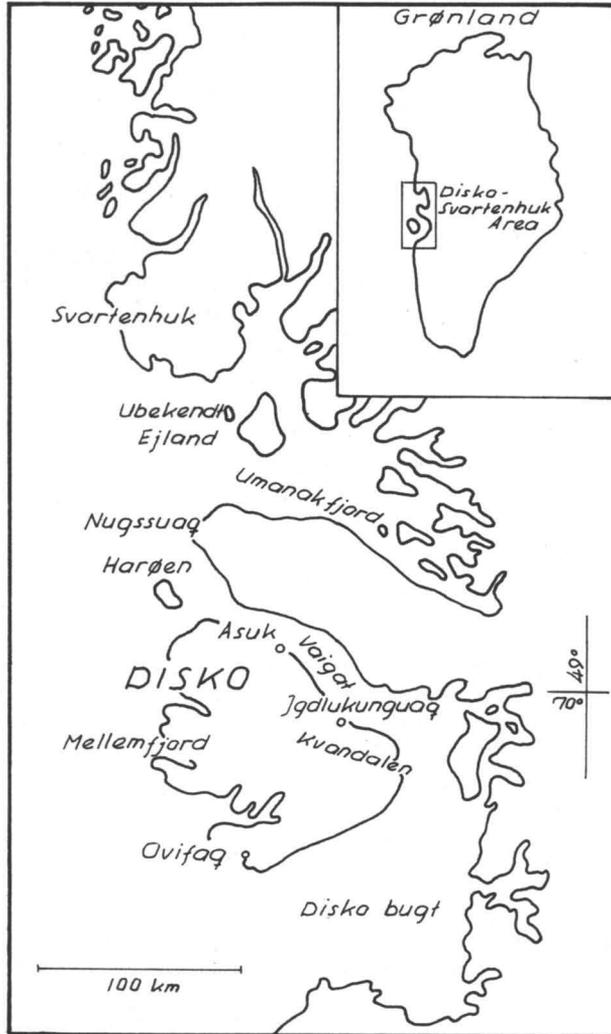


Fig. 1. Sketch map showing the localities mentioned in the text.

but when this activity ended has not been determined. During the whole period of basalt eruptions, the old basement forming the Greenlandic shield was probably continuously uplifted. On Disko island the Precambrian gneisses are exposed in the Godhavn region on the south coast, where the basalt formations seem to occur partly in direct contact with the underlying Precambrian basement. In places, however, there occur thin layers of sedimentary material between the basalt and the old basement formations. The

sediments may be Cretaceous or Early Tertiary in age. Obviously, the basaltic beds in the Godhavn region, where the Precambrian formations crop out, represent the oldest members of the basalt series in the Disko area.

The existence of native iron in loose boulders on Disko island and its surroundings has been known for more than a hundred fifty years, as proved by written reports. This fact was primarily brought to the attention of the scientific world at large, however, by the investigations carried out by A. E. Nordenskiöld at Ovifak, on the SW-coast of Disko island, where, with the assistance of Greenlanders, he found several boulders composed mainly of native iron (1870 and 1872). The largest of the boulders found (25 tons) was delivered to Stockholm, the next in size (8.5 tons) to Copenhagen and one weighing about 4 tons to Helsinki. It is well known that Nordenskiöld thought that the iron boulders at Ovifak originated in a meteoritic shower which fell into the molten basalt during an eruption in Miocene times.

After Nordenskiöld published his findings and conclusions on the native iron of Ovifak, several other investigations were soon initiated. At first, interest was focussed on the question as to whether the iron in the blocks was meteoritic or telluric in origin. After it was observed that native iron also occurs in basaltic beds cropping out in Disko, the question arose as to what processes had led to the formation of the iron in these rocks. This question is still an open one, for divergence of opinion prevails.

The investigations carried out earlier on the iron of Disko by Nauckhoff (1872), Törnebohm (1878), Smith (1879), Steenstrup (1880) and Lorenzen (1880) should be mentioned. The research done by Steenstrup and Lorenzen concentrated especially on the description of the petrographic and analytical features of samples. Later investigations by Löfquist and Benedicks (1941) and by Pauly (1958) have proved of considerable value to the present author. Both of these investigations are especially important in the study of metalliferous phases associated with Disko basalts.

In the years 1872—80 Steenstrup systematically sought to locate all possible occurrences in the Disko-Umanak area bearing evidence of native iron. The number of localities has remained unchanged to date. From a petrographic point of view, Steenstrup (*op. cit.*) gave special attention to observations that a reddish spinel, graphite and coarse-grained calcic plagioclase always occur in association with iron basalts. To some degree these features had already been considered by Törnebohm (*op. cit.*) and they also have an important bearing on the conclusions arrived at by Lorenzen (*op. cit.*). On the basis of the features mentioned, Steenstrup developed the terms »iron basalt» and »graphite basalt» (Steenstrup. 1880).

In the present article, the main emphasis is on the pertinent petrographic features observed in Disko basalts containing native iron. It should be noted, however, that many of the petrographic observations to be discussed in this

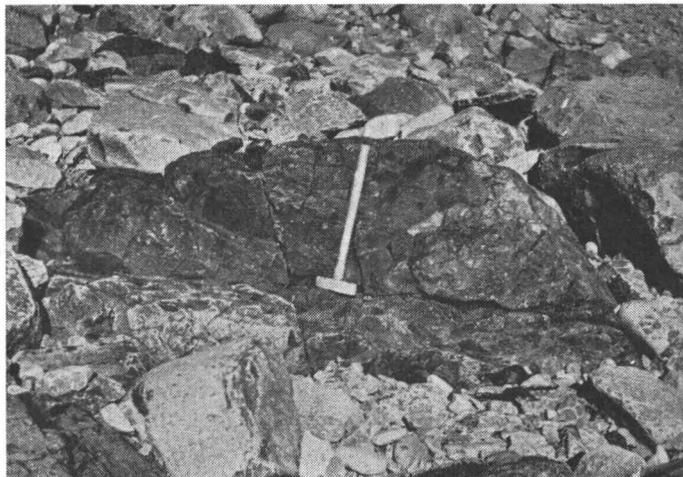


Fig. 2. Rusty sulphide-bearing dyke at Ovifak, Disko.

connection have been known since Steenstrup's investigations. The author came to personal grips with the subject on visits to Disko island in the summers of 1962 and 1963 during the expeditions to West Greenland organized by the Kryolitselskabet Øresund A/S, Copenhagen. A proposal to study the basalts of ENE-Disko was mentioned to the author by Professor Dr. Hans Pauly, at that time chief geologist of the firm mentioned.

The observations described in the present study, which must be considered preliminary in character, are based on material collected by the author during visits to Disko, on samples from the collections of the Kryolitselskabet Øresund A/S and on samples from the collections of the Museum of Geology and Mineralogy, the University of Helsinki.

DESCRIPTION OF THE SAMPLES STUDIED

The Ovifak Area

Nordenskiöld has given a detailed description of this locality, where he made his boulder discoveries in the immediate vicinity of the shoreline. Besides the iron boulders, Nordenskiöld has also described two dykes cropping out close to each other and piercing the cover of basalt boulders accumulated on the shore. One of these dykes is shown in Fig. 2; it is rusty on the surface and includes sulphide ore minerals as unevenly distributed flecks. The other dyke is situated at a distance of c. 5 meters from the dyke shown in Fig. 2 and does not contain visible sulphides.

Three samples from Ovifak, designated as A, B and C, have been consid-

ered. Sample A represents the doleritic dyke without sulphides. Sample B is from the rusty dyke with sulphides. Both of these samples were collected by the author in the summer of 1963. Sample C is from the collections of the University of Helsinki and described only as basalt from Ovifak, and it probably belongs to the primary material delivered by Nordenskiöld.

Sample A is a medium-grained dolerite with an ophitic texture. In appearance, the rock is identical to the doleritic inclusions to be seen in Pl. I, Fig. 1. Plagioclase (An_{60}) and augite ($Z \wedge c = 37^\circ$, $2V_z = 45^\circ$) are the dominant components. The olivine is varyably altered along the cracks or appears as pseudomorphs composed of a brownish iddingsitic-hisingeritic mass. Opaque minerals occur in amounts typical of average dolerite.

The rock of sample B is primarily fine-grained, showing a sub-ophitic texture. There are lesser areas of a coarser-grained portion corresponding in appearance to the dolerite of sample A. Plagioclase (An_{55-60}) is the dominant component in the fine-grained parts. Between the laths there occur small grains of augite and rounded olivine. The opaque material is abundant locally and is composed of pyrrhotite and ilmenite. Native iron was not observed. The pyrrhotite has been altered secondarily and probably contains pentlandite as minute inclusions.

The main part of sample C is composed of doleritic material corresponding to that of sample A. The dolerite, however, has been transected and brecciated by anorthositic material (Pl. I, Fig. 1). The anorthositic portion of the rock contains abundant graphite, small amounts of native iron and a reddish spinel. The plagioclase (An_{60}) is coarse-grained and has strongly resorbed the spinel, as illustrated in Pl. I, Fig. 2.

The Asuk Area

The known formations of iron basalt at Asuk are exposed in a cliff in the immediate vicinity of the Vaigat shore. The samples investigated from this locality consist partly of older samples collected for the Kryolitselskabet Øresund A/S and of new ones taken by the author during a few hours' visit to Asuk in the summer of 1963. All the samples show such a uniformity of properties that a comprehensive petrographic description could be given from any one sample.

Steenstrup has published a detailed description of this locality, which he visited twice (1880). A view of a part of the exposed occurrences is shown in Fig. 3, which also illuminates the stratigraphic section determined as follows:

- Older sedimentary beds (lowermost but not seen in the picture)
- Interbasaltic breccia (partly covered by talus)
- Columnar beds of olivine bearing basalt, c. 15—20 m thick
- Iron basalt (at the top), c. 10—20 m thick



Fig. 3. Exposed basalt cliff at Asuk, Disko. Iron basalt at top. Olivine bearing basalt with excellent columnar jointing.

The total height of the cliff as seen in Fig. 3 is c. 50—60 meters. Inland from the cliff edge, the landscape forms a gently sloping surface for the most part covered by overburden and blocks of basalt up to the beginning of the talus accumulations in front of the escarpment of the basalt plateau. Because of the topography and shortness of time during the visit to Asuk, the upper contact of the iron basalt bed remains to be defined. In earlier reports on the locality it has been suggested that the formation as seen in Fig. 3 is primarily part of a plateau escarpment that slid down to its present position on the shore. This phenomenon is a typical feature along the plateau escarpment elsewhere in Disko and, in the light of the tentative observations of the present author, it seems possible even in the case now under consideration.

It should be pointed out that the iron basalt at Asuk differs clearly on the basis of macroscopic characteristics from the other basaltic rocks in the area. It is a dense, grey, glassy rock, somewhat brittle in appearance. Steenstrup (1880) had noted this difference in the appearance of iron basalt at Asuk. He later found that this type grades into the usual basaltic rocks in the Mellemfjord area and decided that the rock in question is merely a distinct variety of the basaltic series in Disko and should be called simply »iron basalt». In comparison with some typical basalt members from Kvandalen, Disko, the chemical composition of the iron basalt from Asuk shows a distinctly deviating character (Table 1).

The individual samples from Asuk regularly show microscopical properties that, for convenience of description, might be collectively termed as the main material, inclusions and veinous material.

Table 1. Analyses of Disko basalts, Western Greenland

	1		2		3		4		5	
	Wt. %	Cation %	Wt. %	Cation %						
SiO ₂	59.60	56.14	49.80	47.91	48.87	46.39	48.78	47.48	43.70	45.17
TiO ₂	1.09	0.79	1.85	1.39	1.75	1.25	1.95	1.46	1.20	0.93
Al ₂ O ₃	13.80	15.34	13.57	15.38	15.81	17.67	14.74	16.88	13.66	16.63
Fe ₂ O ₃	0.77	0.57	3.28	2.37	3.45	2.51	3.45	2.57	10.86	8.44
FeO	4.85	3.85	9.89	7.98	8.31	6.61	8.56	6.95	1.24	1.05
MnO	0.13	0.11	0.15	0.12	0.14	0.11	0.15	0.12	0.16	0.12
MgO	5.85	8.21	5.98	8.61	6.16	8.72	5.45	7.89	6.59	10.17
CaO	6.68	6.73	10.86	11.21	10.96	11.11	10.56	10.98	11.04	12.22
Na ₂ O	2.10	3.85	2.43	4.51	2.76	5.07	2.63	4.96	2.36	4.71
K ₂ O	1.32	1.58	0.31	0.35	0.25	0.28	0.47	0.53	0.32	0.37
P ₂ O ₅	0.16	0.11	0.22	0.17	0.35	0.28	0.24	0.18	0.22	0.19
CO ₂	0.23	0.34								
H ₂ O+	1.18		0.73		0.65		0.88		6.67	
H ₂ O-	0.25		0.74		0.22		1.50		2.22	
Fe	1.64	1.64								
C	0.15	0.74								
	99.80	100.00	99.81	100.00	99.68	100.00	99.36	100.00	100.24	100.00
Norms										
en	16.42		17.22		17.44		15.78		20.34	
fs	5.76		11.04		8.44		—		—	
hy	—		—		—		8.64		—	
wo	2.50		11.34		8.96		9.96		12.24	
or	7.90		1.75		1.40		2.65		1.85	
ab	19.25		22.55		25.35		24.80		23.55	
an	24.78		26.30		30.80		28.50		28.88	
q	17.60		3.01		0.60		2.41		2.09	
ap	0.29		0.45		0.75		0.48		0.51	
il	1.58		2.78		2.50		2.92		1.86	
mt	0.86		3.56		3.76		3.86		0.72	
hm	—		—		—		—		7.96	
cc	0.68		—		—		—		—	

1. Iron basalt. Asuk, Disko, Greenland. Analyst A. Heikkinen.
2. Doleritic basalt. Kvandalen, Disko, Greenland. Analyst S. Turkka.
3. Tholeiitic basalt. Kvandalen, Disko, Greenland. Analyst S. Turkka.
4. Olivine-bearing basalt. Kvandalen, Disko, Greenland. Analyst S. Turkka.
5. Basalt tuff. Kvandalen, Disko, Greenland. Analyst S. Turkka.

The main material of the samples is a relatively fine-grained rock showing an orthopyric texture (Pl. II, Fig. 1). The dominant components are plagioclase (An₆₅) and an orthopyroxene with a low birefringence. However, large resorbed grains of clinopyroxene may occur occasionally. It should be mentioned that in his study on the Asuk rock, Törnebohm (1878) described it as an enstatite basalt. Lorenzen (1880) did not accept this conclusion and was of the opinion that the pyroxene present is commonly an augite. According to observations made by the author, and on basis of the chemical analysis of the rock (Table 1), it is obvious that Törnebohm was nearly correct in his conclusion.

Ore minerals occur sparsely. The individual grains vary in size from 0.1 to 0.5 mm in diameter. Steenstrup (1880) reports that the biggest ore mineral aggregate observed by him was 14×18 mm in size. The ore material is composed of pyrrhotite, native iron (*i.e.*, the carbon-free cubic form, sementite, etc.) and some magnetite. The pyrrhotite and native iron apparently may exist as separate grains but in most cases these minerals are intergrown. A textural relation often present is illustrated by Fig. 2, Pl. II. The figure shows that the iron components occupy the inner part of a grain while the pyrrhotite forms the outermost rim. Pauly, too, has mentioned this phenomenon in samples from Mellemfjord (*op. cit.*, Fig. 2, Pl. XX).

In addition to the sulphides and native iron, there occur unevenly distributed opaque patches, which are characteristic of the main material in the Asuk samples. The patches are blackish, even under reflected light, and contain very tiny, light inclusions. The dark material of these patches might be composed of hydrous iron oxides while the bright spots have been identified as native iron. Gradations from a grain composed of relatively homogeneous black material to a grain composed chiefly of native iron and surrounded by a narrow rim of dark material can be followed step by step in the polished sections (Pl. III, Fig. 1).

The inclusions present in the Asuk samples are composed of plagioclase, which occurs as resorbed coarse, individual grains or as grain aggregates in the main material (Pl. III, Fig. 2). On the one hand, the inclusions are represented by aggregates of plagioclase accompanied by reddish spinel, iddingsitic pseudomorphs after olivine, glass and ore. The ore material occurs mainly as pyrrhotite, some chalcopyrite, native iron and magnetite. A second type of inclusions is represented by aggregates of coarse-grained plagioclase enclosing abundant opaque material, most probably composed solely of graphite (Pl. IV, Fig. 1). The material of these inclusions is analogous to that of the anorthositic portions described from Ovifak samples.

The veinous material is comparable to the inclusions of the second type and borders sharply on the main material in the Asuk samples (Pl. IV, Fig. 2). Glass is common along the contacts of the two portions. The veinous material is also accompanied by resorbed grains of the reddish spinel. The material mentioned here as veinous may actually represent brecciated fragments in places surrounded by glassy margins. The term veinous as used in this connection merely refers to the mode of occurrence in the thin section.

A petrographic study of the samples from Ovifak and Asuk show that in both instances the graphite-bearing plagioclase rock occurs as an essential component. Lorenzen (1880) especially emphasized that the origin of the graphite-bearing plagioclase rock might have been in a close genetical relationship to the processes producing the native iron. Törnebohm (1878) was

also interested in the rock and proposed that during the eruption of basalts additional material was derived from sedimentary beds, including limestones, carbon-rich mergels, etc., and gave rise to the formation of graphite.

THE REDDISH SPINEL

In agreement with the earlier results (*cf.* Steenstrup, 1880; Lorenzen, 1880) the observations made in the course of the present study have shown that the reddish spinel is consistently incorporated in specimens containing native iron in the Disko area. In some cases, additional green spinel has been observed though not as frequently as the reddish one.

As already emphasized by Steenstrup (*op. cit.*) and Lorenzen (*op. cit.*), the occurrence of the reddish spinel is indeed quite characteristic as regards the mineralogical properties of the iron-bearing formations of the Disko area. For instance, among the systematic samples collected by the present author in the summer of 1962, representing a profile through the basalt formation exposed at Kvandalen, NNE-Disko, the spinel under consideration was never observed and even the native iron was absent in these formations. On the other hand, it is of interest to realize that the reddish spinel was observed by the author from the subsilicic vein adjacent to the nickeliferous pyrrhotite deposit at Igdlukunguaq (Pauly, 1958). This observation is of importance in assuming that the reddish spinel would serve as an indicator of formations genetically related to basalt flows containing native iron (*cf.* Pauly, *op. cit.*)

According to a complete analysis performed in 1901 by Nicolao (see, Bøggild, 1953), the composition of the reddish spinel is as follows:

SiO ₂	0.46 %
Al ₂ O ₃	74.93 »
Fe ₂ O ₃	3.03 »
FeO	4.43 »
Cr ₂ O ₃	0.91 »
MgO	17.83 »
Total	101.61

For the present purpose, an attempt was made to separate the spinel from sample C for analytical determinations. The minute amount of material recovered did not, however, afford more than a provisional run by using the X-ray fluorescence method. The run was made by Mr. A. Hoffren, M.Sc., at the Geological Survey of Finland. Though the material separated was too slight even for the method applied, the results obtained tentatively suggest that the amount of chromium would approximate that of Nicolao's analysis.

On the basis of an uncomplete analysis of the reddish spinel, Lorenzen (1880) likewise observed the presence of chromium and thought that it might give the spinel its characteristic reddish-violet color.

The lattice parameter of the spinel was determined by the author using a Philips X-Ray spectrometer (CuK α -radiation and Ni filter). The measurements on the basis of reflections (400), (311) and (220) gave the result $a_0 = 8.136 \text{ \AA}$. This result suggests a hercynitic composition.

REMARKS AND DISCUSSION OF THE OBSERVATIONS

On the basis of his studies on the Bühl iron, Germany, Eitel (1920) decided that the principal factors promoting the formation of native iron in basalt have been a thermal dissociation of sulphides and the reducing effects of bituminous material mixed with molten basalt. In regard to the formation of iron blocks at Ovifak, Löfquist and Benedicks (1941) similarly assume that the blocks originated from sulphide material by the addition of coal substances and illustrate the processes, which probably took place, by the following reactions:



In regard to the overall processes leading to the formation of the iron blocks at Ovifak, Löfquist and Benedicks (*op. cit.*) suggested the following interpretation: »One of the boulders of the nickeliferous pyrrhotite of the neighbourhood has happened to be enclosed in a crack in which molten basalt has risen, after having absorbed carbon from the coal beds broken through. Under the compined action of the high temperature of magma and its carbon, the sulphides of the pyrrhotite have been dissociated, giving carbonrich iron *i.e.* a natural crude iron containing nickel, being the Ovifak iron».

A few observations carried out under reflected light in the course of the present study indicate that the native iron may, at least in part, have originated from sulphide materials. From this point of view, the pyrrhotite rims as shown in Fig. 2, Pl. II, will be referred to. Pauly (1958) likewise considers that the occurrence of native iron in Disko generally is a consequence of the reduction caused by carboniferous materials incorporated into basalt melts from bituminous beds, which are widely distributed in the Cretaceous and Tertiary sediments of Disko.

On the other hand, the probable mechanism of reduction as such apparently forms only part of the total problem confronted in investigations into the occurrence of native iron in the Disko area. The concept as set forth by

Löfquist and Benedicks (1941), even if correct in regard to the genesis of the blocks at Ovifak, definitely does not explain the genesis of the relatively homogeneous iron basalt at Asuk. This obvious deficiency in the theory of the authors cited is also pointed out by Pauly (*op. cit.*), who states further that the iron blocks at Ovifak are to be considered merely as a special concentration of the usually separate iron droplets in the iron basalt.

One should, indeed, start from the reality that a basalt, like that from Asuk, is a rarity, and that a doleritic rock accompanied by an anorthositic component, as observed from Ovifak, is a unique combination as such. These distinguishing features in certain rocks of the otherwise normally tholeiitic basalt series of Disko formations must be especially emphasized. They probably mean that rather exceptional conditions had prevailed during the eruption of the rocks in question, not only in the near surface zones, but even in the locale of the deep-seated source. In regard to the lastmentioned aspect, changes in water pressure and partial oxygen pressure (*cf.* Muan, 1958; Osborn, 1959) may have been of primary importance. Both of these factors may have a decisive influence on the solid phases finally produced by individual eruptions or intrusions in the course of the magmatic activity in general.

In this connection, suggestions that an increase in water pressure may increase the plagioclase content of residual liquids of tholeiitic source material are of special interest. A condensed and critical review of the matter is given in a recent study by Yoder and Tilley (1962), who conclude that there is substantial evidence of the existence of an anorthositic magma at reasonable temperatures. This estimate seems most applicable to the case now under consideration; the anorthositic portions as described from Disko occurrences are understood to have crystallized from anorthositic melts, which originated and separated at an early stage in the course of crystallization. This assumption makes it understandable why the graphite, obviously derived from bituminous sediments, is abundantly present in the anorthositic portions. By their mode of emplacement the anorthositic melts might have brecciated the older members of the basalt series, as shown by the textures observed in sample C from Ovifak. The resorbed anorthositic inclusions and veinous fragments, which are characteristic of the iron basalt at Asuk, may represent material of a previously solidified anorthositic component.

An increase in P_{O_2} , which may occur as a consequence of dissociation of higher water pressure, first effects the portion of the source material enriched in ferric constituents. It would be expected at the beginning that a change of oxidation affects the olivine, which would then be more or less consumed in reactions resulting in a pyroxene-tholeiitic magma. One of the possible processes to be considered is illustrated by a reaction presented by Muir, Tilley and Scoon (1957):

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Fig. 1. Doleritic inclusion in the anorthosite. Ovivak, Disko, sample C. Thin section. Crossed nic. 10 ×.

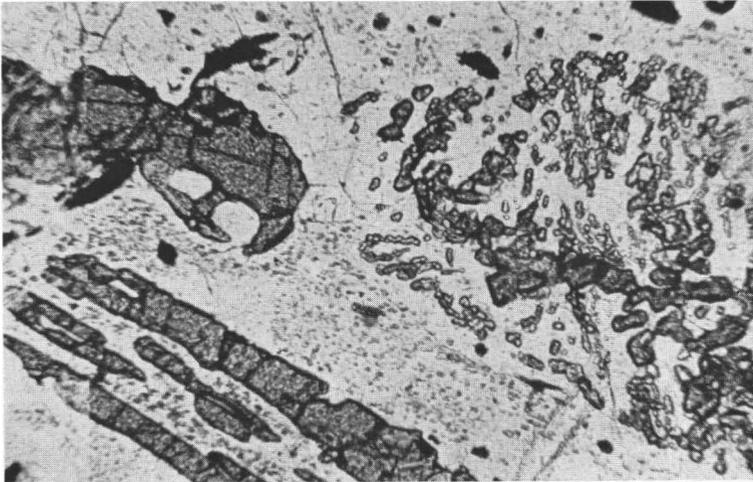


Fig. 2. Resorbed grains of the reddish spinel in the anorthosite. Black flecks are graphite. Ovivak, Disko, sample C. Thin section. One nic. 100 ×.

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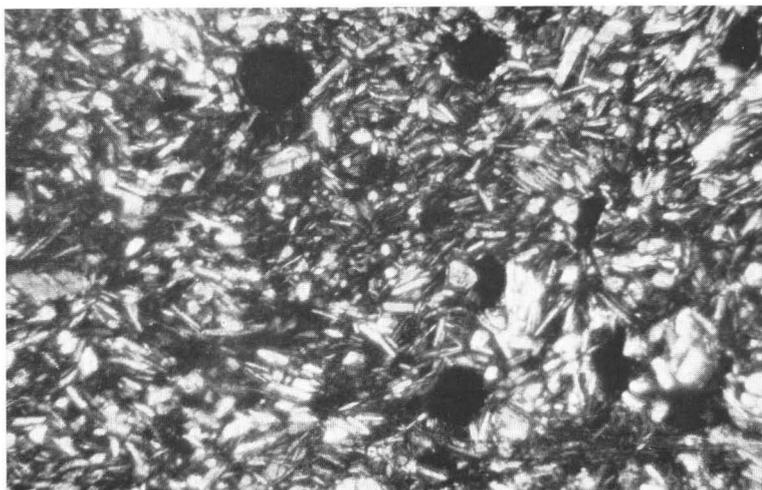


Fig. 1. Iron basalt. Black flecks represent iron. Asuk, Disko. Thin section. One nic. 100 \times .

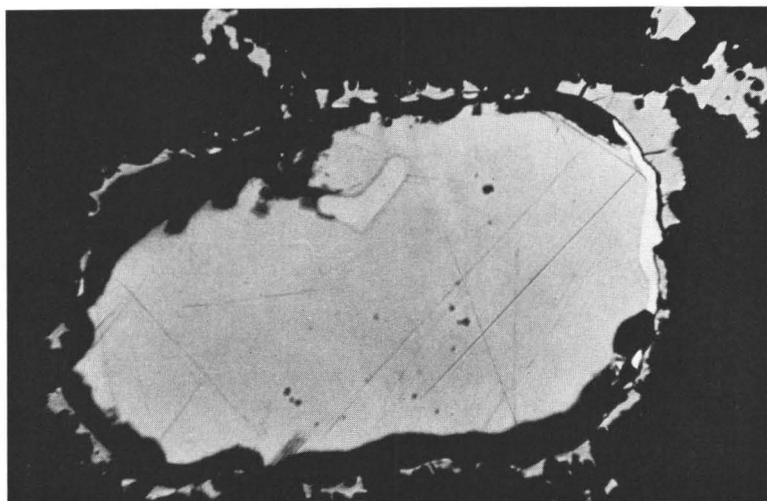


Fig. 2. Grain of iron with a rim of pyrrhotite situated near by. The grain is for the most part composed of cementite (cohenite), though cubic iron occurs as well, usually in kidneyshaped forms in cementite. Iron basalt, Asuk, Disko. Polished section. Oil. im. 320 \times .

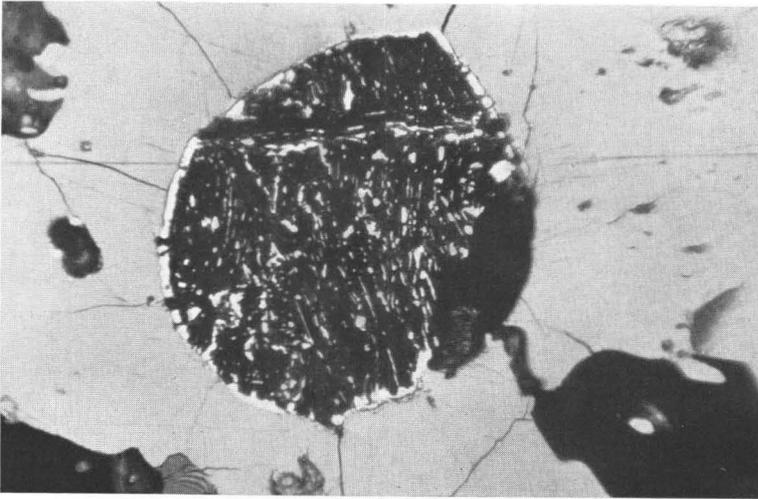


Fig. 1. Blackish patch surrounded by pyrrhotite. Small bright spots in the patch as well as bright portions rimming the patch are iron. Iron basalt, Asuk, Disko. Polished section. Oil im. 320 ×.



Fig. 2. Resorbed aggregate of plagioclase in the basalt. Dark grains in the plagioclase are reddish spinel. Asuk, Disko. Thin section. One nic. 100 ×.

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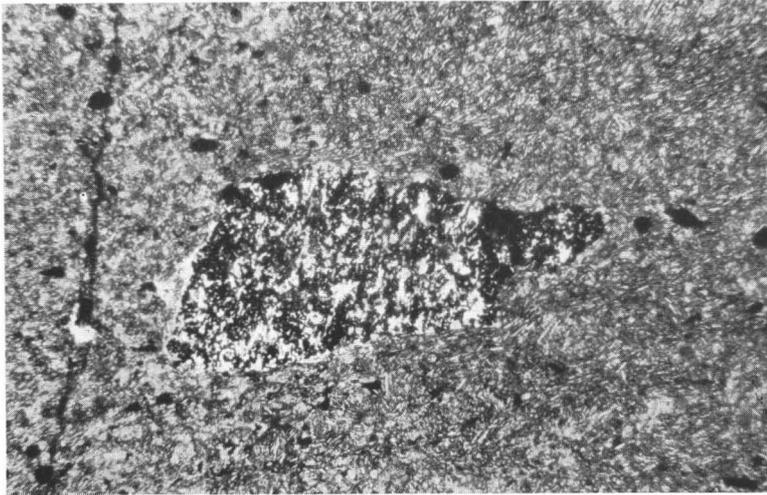


Fig. 1. Grain of plagioclase including abundant graphite. Iron basalt. Asuk, Disko. Thin section. One nic. 40 \times .

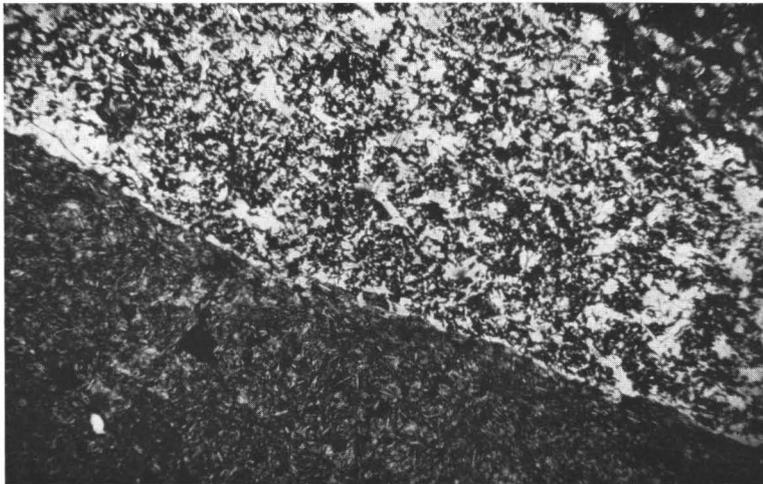


Fig. 2. Lower part of picture shows anorthosite with abundant graphite while upper part is iron basalt. The figure illustrates the anorthositic material referred to as veinous in the text. Asuk, Disko. Thin section. One nic. 40 \times .

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A ZONED SKARN DIKE IN SILVOLA, SOUTHERN FINLAND ¹⁾

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ABSTRACT

A zoned skarn dike situated in limestone is described. The middle of the dike is of scapolite, bordered on both sides by belts of tremolite and farther on by phlogopite. Thus, some elements are also zonally distributed in the dike. The agent causing the formation of the skarn has most likely been a liquid rich in alkalis and volatiles and relatively poor in silica.

INTRODUCTION

During the building of an artificial lake for the purification of water at Silvola, about 13 kilometres NNW of Helsinki, a small occurrence of Precambrian limestone was found. The rock surrounding the limestone deposit is migmatitic gneiss. The limestone was cut by some dikes of potash-rich granite, which are divided from it by a very thin rim of skarn minerals (tremolite, diopside, epidote). At one point, however, a skarn dike proper was found, consisting almost exclusively of skarn minerals and only a few grains of plagioclase and quartz. These point to a granitic composition of the material which caused the formation of the skarn. The thickness of the skarn dike is observed to vary locally between 2 and 10 centimetres.

¹⁾ Received October 5, 1964

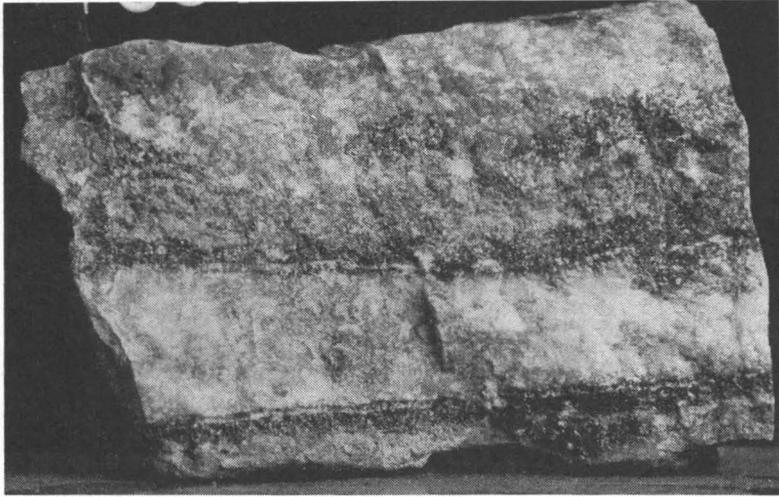


Fig. 1. Skarn dike in the limestone. Silvola, Helsinki. About $\frac{1}{2}$ nat. size.

PETROGRAPHY

The middle portion of the skarn dike (Fig. 1) contains mainly scapolite, in addition to a little plagioclase as well as a few grains of calcite and quartz. The scapolite-rich center is symmetrically bordered on both sides by a zone of amphibole and farther on by a zone rich in mica. The thicknesses of the different zones vary locally. Measured from the center line outwards, the thicknesses of these zones are on the average in the ratio 3 : 1 : 3. These zones are prominent but owing to their coarseness they are not suited to a planimetric analysis of their mineral composition.

The wall rock is a crystalline limestone, which, in addition to the calcite, contains humite minerals (plus some diopside, tremolite and wollastonite) altered near the skarn dike into clinochrysotile. The amount of the humite minerals varies in bands, as is generally the case in the limestones of southern Finland. This banding follows the bedding sometimes observable in the limestone, and hence the banding most likely reveals the primary compositional differences in the limestone.

One extremely humite-rich band was analyzed in part (by A. Heikkinen) and it contained 10.90 per cent SiO_2 , 1.05 per cent FeO , 15.90 per cent MgO , 36.07 per cent CaO , 33.80 per cent CO_2 , and 0.45 per cent F_2 . So we may consider that the composition of the limestone varies between that and pure calcite. On the average, however, bands with only small amounts of silicate minerals or without silicates are more common.

Table 1. Chemical composition¹ (analyst A. Heikkinen) and physical properties of scapolite. Silvola.

Weight %	Mol. prop.	Number of ions on the basis of 12(Si, Al)		Physical data
SiO ₂	45.54	7 579	Si 6.984	} 12.00 Sp.gr. = 2.672 ε = 1.547 ω = 1.582 ω - ε = 0.035 a ₀ = 12.12 ± 0.01 Å c ₀ = 7.55 ± 0.02 Å
TiO ₂	0.00		Al 5.016	
Al ₂ O ₃	25.75	2 526	Ti —	
Fe ₂ O ₃	0.00		Fe ⁺ ₃ —	
FeO	0.25	34	Mg 0.139	
MnO	0.02	3	Fe ⁺ ₃ 0.031	
MgO	0.61	151	Mn 0.003	
CaO	16.73	2 983	Ca 2.749	
Na ₂ O	3.96	639	Na 1.178	
K ₂ O	0.22	23	K 0.042	
P ₂ O ₅	0.20	14	H 1.372	} 4.14
CO ₂	4.33	984	P 0.026	
H ₂ O+	1.34		C 0.907	
H ₂ O+	0.08		S 0.022	
SO ₃	0.19	24	F 0.024	
F ₂	0.05	13	Cl 0.133	} 1.11
Cl ₂	0.51	72		
	99.78		$\frac{100 \times (\text{Ca} + \text{Mg} + \text{Fe} + \text{Mn} + \text{Ti})}{(\text{Na} + \text{K} + \text{Ca} + \text{Mg} + \text{Fe} + \text{Mn} + \text{Ti})} = 70.5$	
—O = F ₂ , Cl ₂	0.13			
Total	99.65			

¹ The analyzed material contained less than 1 per cent of tremolite as admixture.

Table 2. Chemical composition of the skarn dike of Silvola, Helsinki. Analyst A. Heikkinen.

Weight %	Mol. prop.
SiO ₂	35.66
TiO ₂	0.01
Al ₂ O ₃	12.70
Fe ₂ O ₃	0.29
FeO	0.58
MnO	0.10
MgO	16.65
CaO	17.16
Na ₂ O	1.36
K ₂ O	1.88
P ₂ O ₅	0.87
CO ₂	8.72
H ₂ O+	3.57
H ₂ O-	0.18
SO ₃	0.05
F ₂	0.61
Cl ₂	0.12
Total	100.51

In places it seems that near the contact of the skarn dike the limestone is bleached, *i.e.*, there is a smaller amount of humite minerals near the skarn dike than elsewhere in the limestone. Owing to the banded mode of occurrence of the humite minerals in the limestone, this observation is somewhat indistinct. It is quite possible that material from the wall rock near the proper reaction zone, especially MgO and FeO, took part in the skarn reaction.

The scapolite in the skarn dike is megascopically mostly bluish, seldom grey. The largest grains may be up to one inch long. The chemical composition of the scapolite is presented in Table 1. The X-ray determinations show that it contains 69 per cent meionite component and is thus a mizzonite. According to the mean refractive index $[(\omega + \varepsilon)/2 = 1.5645]$, the scapolite should contain (Shaw, 1960, p. 253) 60 per cent meionite, and further, according to the birefringence ($\delta = 0.035$), it should contain (Deer, *etc.*, 1963, Vol. 4, p. 329) 90 per cent meionite.

The middle portion of the skarn dike contains some interstitial grains of plagioclase. According to the X-ray determinations, its An-content is 22 to 28 per cent.

The amphibole is lamellous, greenish in colour, microscopically colourless, monoclinic. The X-ray determination shows it to be tremolite. The Mg-rich component is proved by the refractive indices ($\alpha = 1.609$, $\beta = 1.619$, $\gamma = 1.632$) to amount to (Deer, *etc.*, 1963, Vol. 2, p. 257) more than 90 per cent of its total composition.

The mica scales are characteristically oriented perpendicular to the skarn dike. The mica is brownish in colour and microscopically colourless. The X-ray determination shows it to be phlogopite.

A chemical analysis (Table 2) is presented of a specimen taken across the skarn dike and including the outermost zones of the dike, which are rich in the phlogopite but still contain carbonate. The result of the analysis is somewhat inaccurate because the thickness of the dike and the different zones varies locally. In the petrologic sense, however, the analysis gives a good idea of the compositional change in the skarn process here as well as making it possible to calculate roughly the composition of the material from the outside.

The bulk composition of the skarn dike contains more fluorine than is contained in the scapolite. This shows that fluorine is mostly situated in the phlogopite and the tremolite. On the other hand, the main portion of the chlorine is in the scapolite. The greatest portion of the calcium, the magnesium and the iron most likely derives from the limestone. The content of alkalis is remarkably high, the sodium is mainly contained in the scapolite and the potassium in the phlogopite. Hence, it can be stated that many elements, too, are zonally distributed in this skarn dike.

DISCUSSION

If we seek to determine what portion of the bulk composition (Table 2) represents the material derived from outside to form the skarn in the limestone, we can make some rough presumptions. Most likely all the CO_2 in the limestone has not set in the scapolite but a portion of it has escaped. On the other hand, it is likely that the incoming material, too, has contained a small amount of CO_2 . The Mg-content of the skarn is remarkably high and one may assume it to be derived mainly from the silicate minerals of the limestone (see p. 100). The same may be true also of the iron of the skarn.

In the light of the foregoing we could deduct all the CO_2 and the main portion of the CaO from the bulk composition of the skarn dike. Further, we could take away the greatest portion of the MgO and FeO as well as some of the SiO_2 , which may have been bound in the humite minerals of the wall rock. On good grounds we can, however, conclude that at least the main portion of the silica arrived from outside the rock was bound in the skarn reaction. The fact is that the silica content of the outside matter was rather low and the alumina content relatively high. The amount of the alkalis coming in was relatively high, and in addition the percentage of the potassium was higher than that of the sodium.

On these grounds it can be verified that the matter that entered the limestone was rich in volatiles, including the water, and contained much feldspathic material, and was rather poor in silica.

Landes (1938, p. 388) has demonstrated the formation of scapolite and tremolite in connection with a potash-rich pegmatite phase and, further, the formation of phlogopite during the following hydrothermal phase. Larsen (1942, p. 32) has verified the occurrence of phlogopite as a hydrothermal product in contact-metamorphosed marbles associated with tremolite. In that case the hydrothermal solutions derived from alkalic magmas.

Francis (1958, p. 158) reports a skarn formation caused by metasomatic fluids. In this case also phlogopite and amphibole occur. Such a metasomatism was linked with the neighbouring alkali injection, pegmatite invasion and feldspathization. He (*ibid.*, p. 128) has found tremolite in regionally metamorphosed carbonate rocks. Several other authors have reported the occurrence of tremolite as a product of low-grade metamorphism in silica-bearing dolomites.

According to Shaw (1960, p. 231), the common occurrences of scapolite in contact skarns indicates that scapolite can certainly form at a high temperature and relatively low pressure. Eugster and Prostka (1960) synthesized meionite from SiO_2 , Al_2O_3 and CaCO_3 , at 850°C and a pressure of one atmosphere. Michel-Lévy (1957) has reported a hydrothermal synthesis of tremolite from dolomite heated with SiO_2 and H_2O in the presence of Na_2CO_3

in the temperature range of 400° to 450°C and 270 bars pressure. Grigoriev and Isküll (1937, p. 172) prepared fluor-tremolite by melting the powdered natural amphibole together with 5 per cent CaF_2 in graphite crucibles at 1350°C.

The limestone occurrence of Silvola is situated in an area dominated by migmatitic gneisses. These gneisses are injected by alkali-rich granitic material relatively poor in silica (*cf.* Härme, 1962, p. 11) and most likely a similar injecting material produced the skarn dike presented here.

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INFRARED ABSORPTION OF NEPHELINE ¹⁾

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ABSTRACT

The infrared absorption spectrum of 35 samples of nepheline was recorded, covering practically the whole range of variation of the mineral found in Nature. The spectrophotometer used allows recording in the wavelength range 13—24 μ . The KBr pressed pellet technique was used. The shift of a band numbered III with the composition of the mineral was found useful for estimating the alkali ratio of nepheline. No indication of an Al/Si order-disorder transition similar to that known for the alkali feldspars could be traced. The general appearance of the infrared absorption of the main hexagonal phases of the nepheline-kalsilite system is given.

INTRODUCTION

In connection with more or less systematic studies of other silicate minerals, the infrared absorption of nepheline has been reported by a number of authors. Among the papers dealing with the subject and reporting the data obtained, the most recent have been published by Lyon (1962) and Moenke (1962). The information given is based, however, on only one or a few samples and does not illustrate any further details on the relationship between the infrared absorption and the composition of the mineral. For this reason, the infrared absorption spectrum was recorded for a series of nephelines that covers practically the entire range of variation in the composition of the mineral to be found in Nature. This paper briefly summarizes the results obtained.

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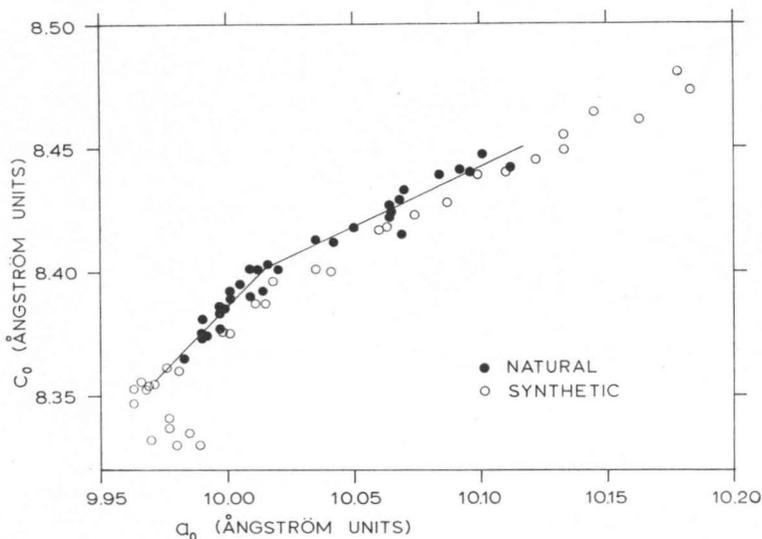


Fig. 1. Unit cell parameters of natural and synthetic nepheline. c_0 plotted against a_0 .

CELL PARAMETERS OF NATURAL NEPHELINE SOLID SOLUTIONS

The unit cell parameters of natural and synthetic nepheline solid solutions are compiled in Fig. 1, plotting c_0 against a_0 . The compilation of this figure is based on data given by Smith and Tuttle (1957), Donnay *et al.* (1959) and Table 1 of this paper. For the sake of clarity, only a part of the data for subpotassic nepheline is included in the figure. All the cell parameters are based on the unit cell of nepheline defined by Hahn and Buerger (1955). The larger true cell postulated by McConnell (1962) was not used.

The presentation of the unit cell data used in Fig. 1 has the advantage of being independent of the chemical composition of the samples. The figure illustrates clearly that there exists a slight discrepancy in cell parameters between natural and synthetic nepheline solid solutions. The least square fit for the relationship between c_0 and a_0 for natural nepheline, as marked in the figure, is given by the following equations:

$$\text{For mediopotassic nepheline: } c_0 = 1.0648 a_0 - 2.262$$

$$\text{For perpotassic nepheline: } c_0 = 0.4867 a_0 + 3.526$$

Without going into the reason for the discrepancy, it may be stated that the existence of such a discrepancy makes the X-ray powder data obtained for synthetic nephelines less suitable in estimating the alkali ratio in natural nepheline.

Table 1. List of nepheline specimens used.

No.	Collection number	Locality	Mode of occurrence	a_0	c_0	V	100 K
				Å	Å	Å ³	K + Na + Ca
1	H 15	Toror Hills, Uganda	Phenocrysts in tinguaitite ...	9.983	8.365	721.9	15'''
2	FEAE 198	Lumbwa-Kericho road, Kenya	Phenocrysts in phonolite ...	9.990	8.373	723.6	17.1'
3	FEAE 136	Toror Hills, Uganda	Phenocrysts in tinguaitite ...	9.990	8.375	723.8	13'''
4	FEAE 170	Oi Esakut, Kenya	Phenocrysts in phonolite ...	9.992	8.374	724.1	17.9''
5	FEAE 134	Toror Hills, Uganda	Phenocrysts in tinguaitite ...	9.990	8.381	724.6	18.0'
6	VM 211	Nyiragongo, Congo	Phenocrysts in intravolcanic dyke	9.997	8.377	724.8	19.7''
7	No.	Mte Somma, Italy	Single crystal	9.997	8.383	725.5	17.7''
8	No.	Horta Velna, Caldas de Monchique, Portugal	Foyaite	9.997	8.383	725.5	19.3''
9	VM 165	Nyiragongo, Congo	Phenocrysts in flow	9.997	8.383	725.5	20.7''
10	VM 241	Nyiragongo, Congo	Phenocrysts in intravolcanic dyke	9.997	8.386	725.8	19.4''
11	5737	Iivaara, Finland	Ijolite	9.999	8.385	726.0	19.8'
12	S 97	Nyiragongo, Congo	Phenocrysts in bomb	10.001	8.389	726.7	21.3''
13	S 90	Baruta, Congo	Phenocrysts in lava	10.001	8.392	727.2	21.0''
14	FEAE 90	Nyiragongo, Congo	Groundmass of lava	10.009	8.390	727.6	25.9''
15	VM 212	Nyiragongo, Congo	Phenocrysts in intravolcanic dyke	10.005	8.395	727.9	22.8''
16	No.	Spitzkop, Transvaal	Foyaite	10.005	8.395	727.9	24.4''
17	VM 578	Nyiragongo, Congo	Phenocrysts in flow	10.014	8.392	728.3	22.1''
18	VM 581	Nyiragongo, Congo	Phenocrysts in flow	10.009	8.401	729.0	26.7''
19	RG 4922	Nyiragongo, Congo	Phenocrysts in flow	10.012	8.401	729.5	27.4''
20	S 76	Nyiragongo, Congo	Phenocrysts in flow	10.016	8.403	730.0	28.4''
21	VM 274	Nyiragongo, Congo	Phenocrysts in intravolcanic dyke	10.020	8.401	730.5	28.5''
22	C 5550	Nyiragongo, Congo	Phenocrysts in flow	10.035	8.413	733.8	31.0'
23	FEAE 49	Nyamununka, Katwe-Kikorongo, Uganda	Groundmass in subvolcanic bomb	10.042	8.412	734.7	33.0'
24	S 88	Baruta, Congo	Phenocrysts in lava	10.050	8.418	736.3	31.1'
25	FEAE 87	Nyiragongo, Congo	Phenocrysts in flow	10.064	8.422	738.8	37.2'
26	VM 220	Nyiragongo, Congo	Phenocrysts in flow	10.065	8.424	739.0	38.1''
27	FEAE 88	Nyiragongo, Congo	Phenocrysts in flow	10.069	8.415	739.0	36.8''
28	FEAE 89	Nyiragongo, Congo	Phenocrysts in flow	10.064	8.427	739.1	36.6'
29	FEAE 92	Nyiragongo, Congo	Phenocrysts in flow	10.068	8.429	739.9	40.6''
30	VM 235	Nyiragongo, Congo	Phenocrysts in flow	10.070	8.433	740.5	36.8''
31	FEAE 86	Nyiragongo, Congo	Phenocrysts in flow	10.084	8.439	743.1	39.5'
32	FEAE 93	Nyiragongo, Congo	Phenocrysts in flow	10.092	8.441	744.5	42.1'
33	S 77	Nyiragongo, Congo	Phenocrysts in flow	10.096	8.440	745.0	42.5''
34	C 9956	Nyiragongo, Congo	Phenocrysts in flow	10.101	8.447	746.4	40.8'
35	No.	Nyiragongo, Congo	Phenocrysts floating in molten lava lake	10.112	8.442	747.5	41.2'

' From chemical analysis.

'' From X-ray powder pattern.

''' Given by Hytönen (1959).

The equations for calculating the composition of nepheline solid solutions from the X-ray powder pattern recommended by Smith and Sahama (1954) are based on data for synthetic nepheline. For the reason mentioned, the equations given by these authors are not used in this paper. Instead, new equations were calculated, based on the 2θ -values for the reflections (21.0) and (20.2) for natural nepheline. The following linear equations, calculated by least square methods, represent the data for natural nepheline solid solutions ($k = 100 \text{ K}/(\text{K} + \text{Na} + \text{Ca})$):

Reflection (21.0):

$$\text{Mediopotassic nepheline: } k = 67.72 - 192.7 (2\theta_{Cu} - 27.00)$$

$$\text{Perpotassic nepheline: } k = 108.94 - 68.03 (2\theta_{Cu} - 26.00)$$

Reflection (20.2):

$$\text{Mediopotassic nepheline: } k = 122.8 - 153.8 (2\theta_{Cu} - 29.00)$$

$$\text{Perpotassic nepheline: } k = 71.8 - 72.0 (2\theta_{Cu} - 29.00)$$

The difference in composition of a particular nepheline obtained by using these equations and, on the other hand, by using the equations given by Smith and Sahama (*op. cit.*) is slight. As has been pointed out by Hytönen (1959), the difference between the 2θ -values for natural and synthetic nephelines becomes significant in the composition range $k = 9-20$. However, the new equations given were used in calculating from X-ray powder patterns the composition of the nephelines listed in Table 1.

MATERIALS

The nepheline specimens that form the material of this study are for the most part the same as the ones used for optical birefringence measurements by Sahama (1962). Because, however, a few of the samples are different and because unit cell data were obtained for all the nephelines used, Table 1 is included. The unit cell parameters were taken from the list published by Donnay *et al.* (1959). The new unit cell data were obtained from X-ray powder pattern using the calculation method presented by these authors. The unit cell volume listed in the seventh column of the table is not that calculable from the values given for a_0 and c_0 but represents a corrected value obtained in the following way. From each point indicating the unit cell dimensions of a particular nepheline in Fig. 1, a line was drawn perpendicular to the line of the least square fit marked in the figure. The cell parameters were read on this least square line and the unit cell volume was calculated from the values for a_0 and c_0 thus found. The unit cell volumes corrected in that way were

used throughout in this study. The values for $100 K/(K+Na+Ca)$ listed in the last column of the table, which are based on the X-ray powder pattern, were calculated using the new equations given in this paper.

EXPERIMENTAL PROCEDURE

The infrared spectra were obtained using a Leitz double beam infrared spectrophotometer with potassium bromide optics. With that optics, the instrument makes it possible to record the spectra in the wavelength range from 13 microns (769 cm^{-1}) to 24 microns (417 cm^{-1}). For wavelengths below this range, the dispersion becomes small. According to Moenke (1962), below $13\ \mu$ the nepheline spectrum shows only one large complex band at about $10\ \mu$. That band was not recorded in this study.

The KBr pressed pellet technique was employed. The sample preparation was made as follows. The nepheline samples were ground in an agate mortar and eluted in water. The fraction of a grain size less than $6\ \mu$ was used. Exactly 3.00 mg of this fine-grained nepheline powder together with 1 000 mg of finely powdered KBr (grade: for infrared spectroscopy) was placed in a bakelite capsule and mixed thoroughly with the aid of a mechanical dental vibrator (Wig-L-Bug) for 5 min. To facilitate mixing, two small steel balls were added in the capsule. Tests made with pure KBr did not reveal any contamination from the bakelite capsule. Of the homogenized mixture 800 mg was placed in a brass ring in the steel pressing device supplied by the company and pressed in vacuum under a pressure of about 10 tons/cm². The clear pellet measured 20 mm in diameter with a thickness of ca. 0.9 mm containing 0.30 % nepheline in KBr.

After some preliminary runs, the following instrumental settings were found most suitable: slit program 5, average speed of recording ca. 7 min./micron, light source amperage 0.9. To ensure the full scale of deflection on the chart, an adjustable wire filter was placed in the reference beam. By means of this filter, the wavelengths of highest transparency, which for nepheline are around $16.5\ \mu$ and $17.7\ \mu$, were adjusted to correspond approximately to 95 % on the chart. The correct electrical adjustment of the instrument was checked at frequent intervals. The wavelength calibration was made by recording the carbon dioxide band at $14.99\ \mu$.

The infrared absorption spectrum of nepheline in the wavelength range in question is illustrated in Fig. 2. The four bands found in this range were numbered in the way indicated in the figure. This numbering is used throughout this paper. The wavelength of band I was measured by reading the wavelengths of both flanks of the band at, say, 40 % transparency and by adopting the mean of the two readings. Because band I is unsymmetric, the wavelength thus obtained does not coincide with the wavelength of maximum

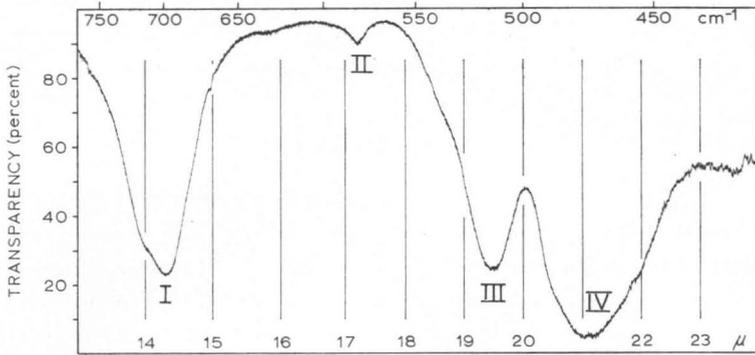


Fig. 2. Tracing of the infrared absorption spectrum of nepheline, illustrating the numbering of the bands used in this paper.

absorption. The wavelengths of the other bands were measured in a similar way. The reproducibility in measuring band III, which is sharp and symmetric, was found to be mostly 0.01μ or better. Bands I and IV being broad and unsymmetric, the reproducibility obtainable in measuring their wavelengths is less good. Because the intensity of band II is low, the measurement of its wavelength necessarily is somewhat inaccurate.

CHARACTERISTICS OF THE SPECTRA

Small portions of nepheline samples Nos. 4 and 29 (Table 1) were fused by heating for 10 min. at $1\ 580^{\circ}$ — $1\ 600^{\circ}\text{C}$ and quenching in water. A microscopic study of the glasses obtained revealed no presence of crystalline remnants. These glasses were used for comparing the infrared absorption with that of the crystalline minerals.

Fig. 3 illustrates the tracing of the infrared absorption of nepheline No. 4 compared with the tracing produced by the fused glass of the same nepheline. As is evident from the figure, bands II and III are characteristic of the crystalline material and are not present in the tracing of the glass. On the other hand, bands I and IV, though broad and less intense, have been preserved in the disordered structure of the glass, band IV, however, shifted towards higher wavelengths. Hence, it is concluded that bands I and IV are produced by vibrations in the $(\text{SiAl})\text{O}_4$ tetrahedra. A similar relationship between the tracings of crystalline and glassy material was observed also for nepheline No. 29.

Fig. 4 illustrates qualitatively the shift of all bands towards higher wavelengths with an increasing potassium content of the mineral. The

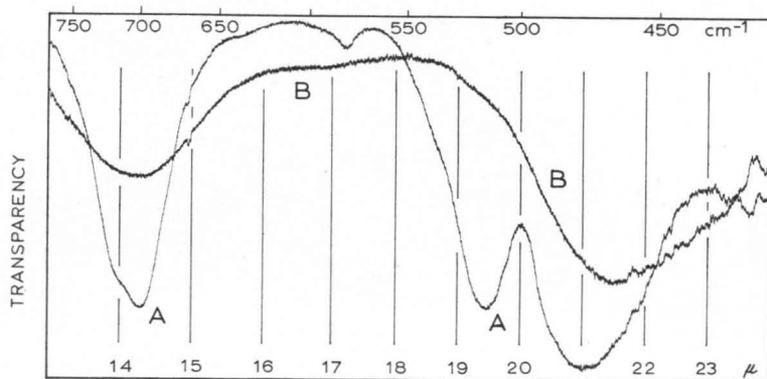


Fig. 3. Tracing of the infrared absorption spectrum of crystalline nepheline No. 4 (A) and of a fused glass of the same nepheline (B).

tracing marked A in the figure represents a mediopotassic nepheline (No. 5 in Table 1) and that marked B corresponds to an extremely potassium-rich perpotassic nepheline (No. 32 in Table 1). The shift is most conspicuous for band III, which lies between 19 μ and 20 μ .

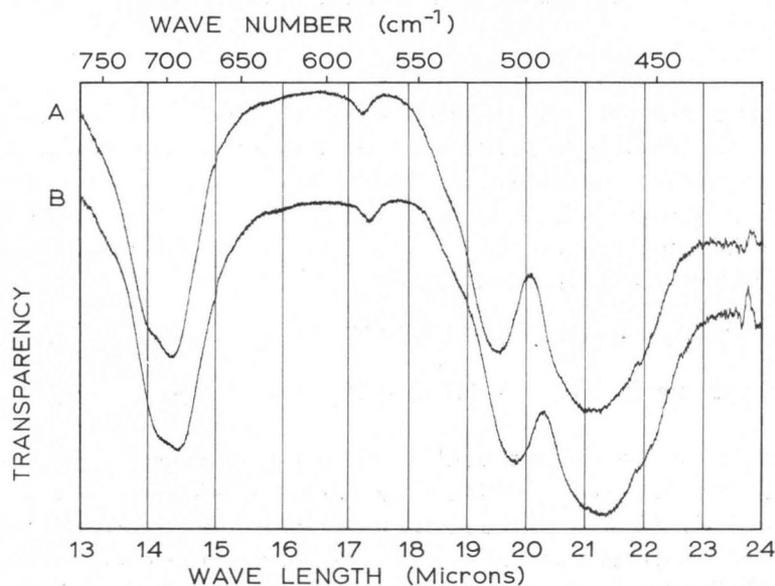


Fig. 4. Tracings of the infrared absorption spectra of mediopotassic nepheline No. 5 (A) and of extremely potassium-rich nepheline No. 32 (B).

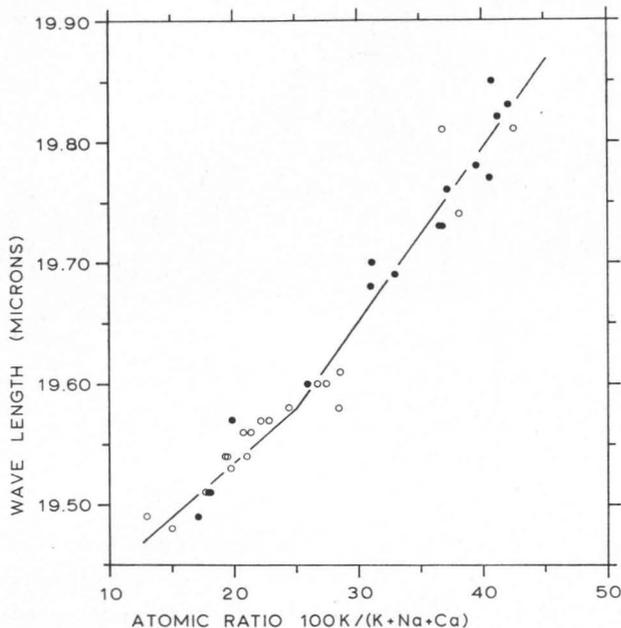


Fig. 5. Wavelengths of band III for nephelines of Table 1 plotted against atomic ratio $100K/(K+Na+Ca)$. Solid circles: composition known from chemical analysis. Open circles: atomic ratio calculated from X-ray powder pattern.

Band III is the only one in this wavelength range that is suitable for determining the alkali ratio in nepheline. In Fig. 5 the wavelengths of band III of the nephelines mentioned in Table 1 have been plotted against the atomic ratios $100K/(K+Na+Ca)$. The least square fit, marked in the figure, of the relationship between the wavelength of band III and the atomic ratio is given by the following equations:

For mediopotassic nepheline: $100K/(K+Na+Ca) = 113.64 \mu - 2200.0$

For perpotassic nepheline: $100K/(K+Na+Ca) = 68.80 \mu - 1322.0$

On calculating these equations, double weight was assigned to the points in the figure for which the composition of the mineral was known from complete chemical analysis. Along with the X-ray method given by Smith and Sahama (1954), these equations may be used for estimating the composition of a natural nepheline by means of its infrared absorption spectrum.

Fig. 6 shows the wavelength of band III plotted against the corrected unit cell volume listed in Table 1. The least square fit of the relationship

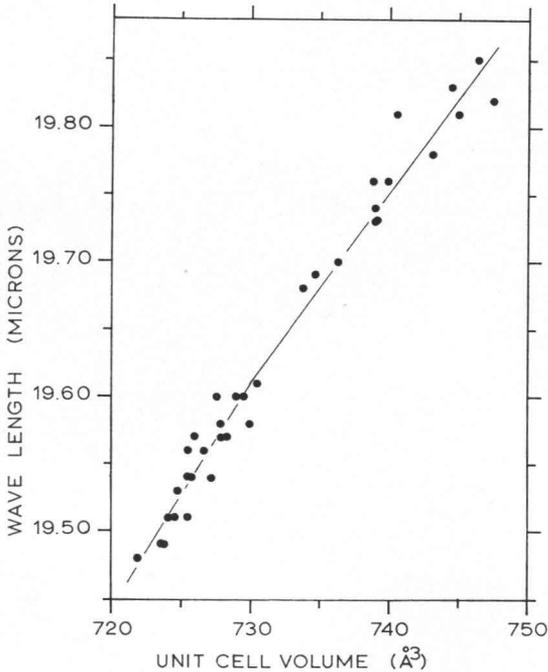


Fig. 6. Wavelengths of band III for nephelines of Table 1 plotted against the corrected unit cell volumes listed in the table.

between the wavelength of band III and the unit cell volume, marked in the figure, is given by the following equations:

$$\text{For mediopotassic nepheline: } \mu = 0.01587 V + 8.02$$

$$\text{For perpotassic nepheline: } \mu = 0.01431 V + 9.16$$

The lines corresponding to these two equations lie very close together. Accordingly, the relationship between the wavelength of band III and the unit cell volume could be given by a single equation as well. The accuracy attainable in measuring both the wavelength of band III and the unit cell volume does not allow a decision as to whether or not there exists such a real difference between the mediopotassic and perpotassic nephelines. As compared with band III, the bands I, II, and IV are less accurately measurable. Therefore, the relationship between the wavelengths of these bands and the unit cell volume is not illustrated here in the way corresponding to Fig. 6. The data available seem to indicate, however, that a singularity exists in the variation of the wavelengths of these bands, especially of band I, with cell

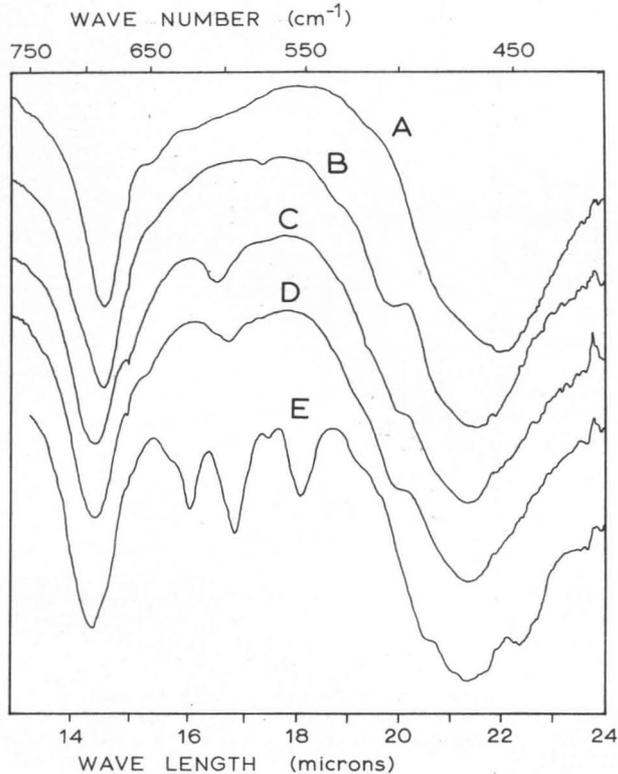


Fig. 7. Tracings of the infrared absorption spectra of the main hexagonal phases in the nephelinekalsilite system.

- A) Kalsilite. Kyambogo crater, Bunyaruguru, Uganda.
- B) Nepheline and kalsilite in exsolution texture. Kabfumu lava, Nyiragongo, Congo.
- C) Tetrakalsilite. Homogenized through heating exsolved nepheline-kalsilite crystals from Kabfumu lava.
- D) Trikalsilite and tetrakalsilite. Heated nepheline-kalsilite crystals from Kabfumu lava.
- E) Kaliophilite. Vesuvius.

volume. For that reason, the two equations for band III are given for medio-potassic and for perpotassic nephelines separately.

As has been recently summarized by Laves and Hafner (1962), the Al/Si order disorder transition in alkali feldspars can readily be detected and studied by means of the infrared absorption method. For nepheline, however, no indication of such a transition could be traced in the infrared absorption. Accordingly, the optical behavior of nepheline found by Sahama (1962) has no counterpart in the infrared absorption of the mineral.

In addition to the infrared absorption of nepheline, that of the other hexagonal phases known in the nepheline-kalsilite system is illustrated in

Fig. 7. All these phases (kalsilite, nepheline, trikalsilite, tetrakalsilite, natural kaliophilite) show the strong bands I and IV. The various phases concerned differ from each other in the situation of the small bands that occur between the two major bands mentioned. In accordance with the tracings reproduced in Fig. 7, the following small bands between band I and band IV will be noted:

Kalsilite:	No bands.
Nepheline:	Bands II and III.
Trikalsilite:	16.7 μ and band III.
Tetrakalsilite:	16.5 μ and band III.
Natural kaliophilite:	16.0 μ : 16.8 μ : 17.4 ₅ μ : 18.1 μ .

These small bands are useful for identifying the phases. It is worth noting that, in the series nepheline-trikalsilite-tetrakalsilite, the small bands located between bands I and III shift towards shorter wavelengths. As could be expected, natural kaliophilite differs markedly from the other phases listed.

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ALKALI POSITION IN THE BERYL STRUCTURE¹⁾

BY

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ABSTRACT

Experimental data are given supporting the proposition that in alkali beryl the position 2a, *i. e.*, in channels between the Si_6O_{18} rings in the plane of Al and Be ions, is occupied both by the alkali ions and the water molecules. Be is possibly substituted by Li.

Chemical and crystallographical data are given for nine beryl specimens, the same specimens being also investigated by the infrared absorption method.

INTRODUCTION

In connection with a study of the polygonal texture of beryl, to be published later by one of the authors (Th. G. S.), *c*-axis precession photographs were taken of a number of beryl specimens from various localities in Brazil, Mozambique, etc. The original aim of the work was to test the uniformity in orientation of the polygons detectable in a clear beryl crystal. During the course of the work, it was found that quite remarkable differences in the relative intensities of certain reflections exist between some of the beryl crystals studied. It seemed of interest to test whether these differences can be quantitatively explained in the light of differences between the beryl specimens with respect to their alkali and water content or whether other structural variations must be sought. Accordingly, the relative intensities

¹⁾ Received November 11, 1964.

of some selected reflections were measured and the results obtained were compared with the intensities calculable on the basis of the beryl structure. This part of the work, done by one of us (A. V.), forms the main substance of this paper. The beryl materials from some Finnish pegmatites used in this study, supplied by one of us (I. H.), will be more thoroughly described by him on a later occasion.

STRUCTURAL CONSIDERATIONS

The crystal structure of beryl was originally determined by Bragg and West (1926). Bergerhoff and Nowacki (1955) showed by electron density maps that the water molecules and a great part of the alkali ions in the beryl structure were in the position $2a$ of Wyckoff's notation, *i.e.*, in the middle of the channels parallel to the c -axis. In addition, they assumed some of the alkali ions to be in the position $4c$, which in the ideal beryl structure is completely occupied by Al ions. Also Frank-Kamenetskii and Sosedko (1958) placed the alkali ions in the centers of the oxygen polyhedra with 24 vertices (position $2a$). Bakakin and Belov (1962), in a thorough study of the crystal chemistry of beryl, proposed the following scheme for the substitution of atoms in the beryl structure:

- Be in $6f$, and in $12l$ if the structure is deficient in Si and Al.
- Si in $12l$, and in $6f$ if silicon is in excess, substituting for Be.
- Al in $4c$, and in $12l$ or in $6f$ if aluminium is in excess, substituting for Si and Be.
- Li a) in Al sites, not substituting for Al but occupying vacancies when there is a deficiency (usually slight) of Al;
 - b) in $6f$, bringing the number of atoms to 6;
 - c) in very rare cases the excess of Li in spaces between the rings Si_6O_{18} .
- Fe^{3+} , Fe^{2+} , Mg, Sc, Mn in $4c$, substituting for Al.
- Na and Ca in $2a$.
- Na exceptionally in the spaces between the rings Si_6O_{18} .
- K and H_2O in $2b$.
- Cs, Rb and sometimes K in channels between the positions $2a$ and $2b$.

Feklichev (1963) placed the alkali ions in the centers of the Si_6O_{18} rings (position $2b$) and the water molecules between the Si_6O_{18} rings in the plane of the Al and Be ions (position $2a$).

All the authors cited place the majority of the alkali ions and water molecules into the channels parallel to the crystallographic c -axis of beryl. The greatest differences between the proposed structure models concern the exact positions of the alkali ions and water molecules in the channels and, on the other hand, the extent to which alkali ions can substitute for the atoms in the basic beryl structure.

The present study gives experimental data supporting the proposition that the position $2a$ — *i.e.*, in the channels between the Si_6O_{18} rings in the

plane of Al and Be ions — is occupied both by alkali ions and water molecules. At least, in the beryl specimens investigated, none of the other sites proposed for the alkali ions and water molecules seem to be occupied by these constituents. The substitution of Be by Li, proposed by Bakakin and Belov (1962), possibly takes place.

MATERIALS

The chemical and other data for the beryl specimens used are compiled in Tables 1 and 2. The unit cell contents given for the alkalis and water are based on 12 silicon atoms per cell. For the beryl specimens listed in Table 2, the SiO_2 content was taken as 63 %. The unit cell dimensions were calculated from powder pattern by means of the reflections $11\bar{2}2$, $21\bar{3}1$ and $21\bar{3}2$. The powder patterns were recorded with the Philips Norelco X-ray diffractometer with silicon as internal standard. The values obtained are believed to be accurate within ± 0.003 Å. The refractive indices of the morganite of Table 1 were measured by applying the prism method.

SINGLE CRYSTAL STUDY

The single crystal study was based on *c*-axis zero-, first-, second-, and third-level precession photographs mainly of the Mozambique morganite (Table 1) and of two colorless Brazilian beryl crystals. The Mozambique morganite was known to be extremely rich in alkalis. No chemical analyses were made of the Brazilian beryl crystals used. As was mentioned in the introduction, *c*-axis precession photographs of a number of beryl crystals from various localities were available. Of these crystals two were selected, both Brazilian, for which the intensities of certain reflections showed most pronounced deviations from those of the Mozambique morganite. It can safely be assumed that these two Brazilian beryl crystals are poor in alkalis.

No differences in the intensities of any reflections between the three beryl crystals were detected in the first-level nor third-level precession photographs. On the other hand, the differences found in the intensities of certain reflections between the Mozambique morganite and the Brazilian beryl crystals are rather great in the zero-level and second-level precession photographs. It was found that, if alkali atoms are introduced in any of the positions $2a$, $2b$, or $4e$ in the beryl structure of space group $P6/mcc$, the change in intensities calculable for the $hki0$ reflections agree with the differences found between the zero-level photographs of the Brazilian and Mozambique beryl crystals. The differences in the intensities of some reflections in the second-level photographs can be explained only by assuming X-ray scattering material in the position $2a$. The reflections $22\bar{4}2$ and $31\bar{4}2$ are sensitive to X-ray

Table 1. Beryl (morganite) from Marropino Mine, Zambezia, Mozambique. Chemical analysis by Oleg v. Knorring.

SiO ₂	62.97 %	Number of alkali ions and water molecules in the unit cell:
TiO ₂	0.00	
BeO	11.56	
Al ₂ O ₃	17.75	
Fe ₂ O ₃	tr.	
MnO	0.01	
MgO	0.00	
CaO	0.12	
Li ₂ O	1.18	
Na ₂ O	0.86	
K ₂ O	0.16	
Cs ₂ O	3.01	
H ₂ O+	2.22	
H ₂ O-	0.02	
Total	99.86	
		$a_0 = 9.216 \text{ \AA}$
		$c_0 = 9.231$
		Sp. gr. = 2.805
		$\varepsilon = 1.593$
		$\omega = 1.585 \pm 0.001$

Table 2. Specimens of beryl from Finnish pegmatites. Chemical analyses by Pentti Ojanperä.

	1	2	3	4	5	6	7	8
Li ₂ O	0.34 %	0.43 %	0.45 %	0.46 %	0.61 %	0.62 %	0.82 %	1.01 %
Na ₂ O	0.82	0.86	0.94	0.91	1.25	1.15	1.21	1.36
K ₂ O	0.06	0.03	0.05	0.06	0.04	0.04	0.08	0.06
Cs ₂ O	0.18	0.14	0.10	0.13	0.16	0.23	0.70	1.74
Rb ₂ O	0.01	0.01	0.01	0.01	0.00	0.01	0.01	0.03
H ₂ O	1.80	1.82	2.07	1.76	2.43	1.85	1.56	0.60
a_0 (Å)	9.213	9.215	9.214	9.215	9.215	9.214	9.214	9.215
c_0 (Å)	9.203	9.208	9.209	9.210	9.216	9.211	9.220	9.230
Number of alkali ions and water molecules in the unit cell								
Na	0.302	0.318	0.348	0.336	0.462	0.426	0.446	0.502
K	0.014	0.007	0.013	0.014	0.009	0.009	0.019	0.014
H ₂ O	1.145	1.157	1.316	1.119	1.545	1.177	0.992	0.381
Li	0.261	0.323	0.346	0.353	0.467	0.474	0.627	0.774
Rb	0.01	0.001	0.001	0.001	—	0.001	0.001	0.004
Cs	0.014	0.011	0.008	0.010	0.013	0.018	0.057	0.142

1. Hunnako, Alavus, Finland.

2—8. Haapaluoma, Peräseinäjoki, Finland.

scattering atoms in the position $2a$. If alkali ions are introduced at $2a$, the reflection $22\bar{4}2$ will increase in intensity while the reflection $31\bar{4}2$ will decrease. The intensity ratio $I_{22\bar{4}2}/I_{31\bar{4}2}$ will thus theoretically give a semi-quantitative estimate of the alkali content in the channels of the beryl structure. For the Mozambique morganite this ratio is approximately 3 : 1, while for the Brazilian beryl crystals it is around 1 : 2 and 2 : 3, respectively.

POWDER X-RAY WORK

In the powder pattern of beryl, $21\bar{3}1$ and $11\bar{2}2$ are represented by rather strong peaks in the front reflection area ($\theta_{\text{CuK}\alpha} = 15.6^\circ$ and 13.7° , respectively). Only the Si ions in position $12f$ and the oxygen ions in $24m$ and $12f$ contribute to the intensity of the $21\bar{3}1$ reflection. Neither Al ions in $4c$ nor Be ions in $6f$ affect this reflection; the same applies to the possible alkali ions or water molecules in $2a$, $2b$, and $4e$. On the other hand, the intensity of $11\bar{2}2$ is sensitive to atoms in the position $2a$, $2b$, and $4e$. In addition, all the atoms of the beryl structure contribute to the intensity of this reflection. Because both reflections correspond in intensity, are very well resolved, and easily recognized in X-ray powder photographs or diffractometer charts, it was interesting to determine whether any conclusions about the alkali content could be drawn on the basis of the intensity ratio of these two reflections.

The coordinates of the atoms in the structure, as listed by Bragg and West (1926) and transformed to correspond to the origin at $6/m$, are virtually the same as those given in later redeterminations by Belov and Matveyeva (1950) and by Bergerhoff and Nowacki (1955). In the present study the coordinates refined by Bergerhoff and Nowacki (*op. cit.*, p. 412, Table I, column 8) are used in calculating the theoretical intensities of the reflections $21\bar{3}1$ and $11\bar{2}2$ for $\text{CuK}\alpha$ -radiation. The atomic scattering factors were taken from the International Tables for X-Ray Crystallography, Vol. III, p. 202-207. The intensities were corrected for the usual Lorentz-polarization factors.

It is evident that the increase in alkali and water contents in the channels of the beryl structure will cause slight changes in the coordinates of other atoms and thus also in the intensities of the reflections $21\bar{3}1$ and $11\bar{2}2$. But because these two reflections are in the front reflection range, the influence on them is necessarily small.

For a beryl free from alkalis and water, the theoretical intensity ratio $I_{21\bar{3}1}/I_{11\bar{2}2}$ is 1.21. For a beryl containing alkali atoms and water molecules in position $2a$, this intensity ratio is calculated from the relationship

$$I_{21\bar{3}1}/I_{11\bar{2}2} = \frac{1.35 \cdot 10^4}{(105.3 + \sum_N f_N v_N)^2}$$

where f_N is the atomic scattering factor of the N^{th} kind of atom in position $2a$ and v_N is the number of N^{th} kind of atoms in the unit cell. This formula was calculated for $\text{CuK}\alpha$ -radiation with f_N at $\sin\theta/\lambda = 0.154$. The calculations are based on the following formula for beryl



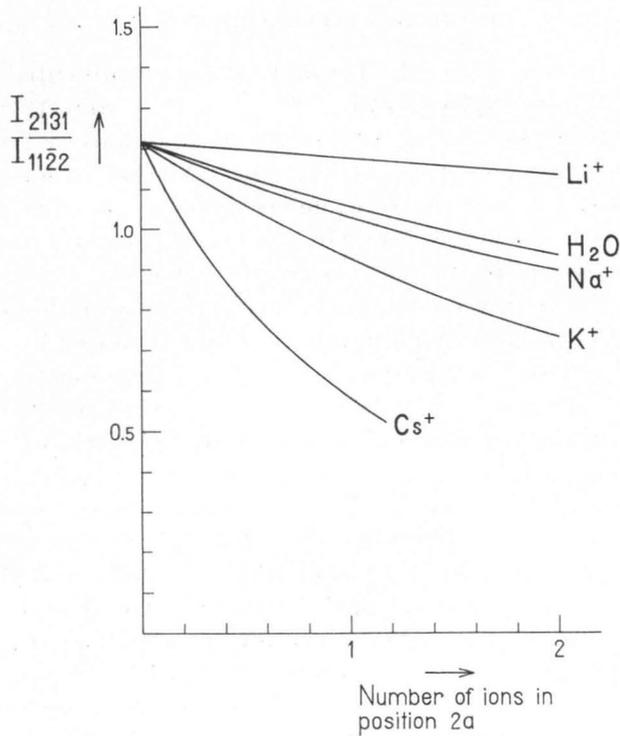


Fig. 1. Theoretical intensity ratio $I_{21\bar{3}1}/I_{11\bar{2}2}$ plotted against the number of alkali ions and water molecules in position $2a$ of the unit cell.

The formula given by Ginzburg (1955)



would have been more exact. The numerical value for ${}^{\text{Be}}F_{11\bar{2}2}$ is so small, however, that a slight deficiency in Be does not affect the result.

Fig. 1 presents diagrammatically the results of the calculation, plotting the intensity ratio $I_{21\bar{3}1}/I_{11\bar{2}2}$ against the number of alkali ions and water molecules in position $2a$. As the diagram shows, the lithium ion contributes very little to this intensity ratio. The effect of the water molecule approximately equals that of the sodium ion. The effect of the caesium ion is quantitatively most important. Even 0.5 Cs per unit cell affects the intensity ratio as much as the theoretically maximum number of two sodium ions.

The formula derived for the intensity ratio $I_{21\bar{3}1}/I_{11\bar{2}2}$ and the numbers of the alkali ions and water molecules in the unit cell listed in Tables 1 and 2 were used in calculating this intensity ratio for the beryl specimens studied.

Table 3. Observed and calculated intensity ratios $I_{21\bar{3}1}/I_{11\bar{2}2}$ for the beryl specimens of Table 1 (No. 9) and 2 (Nos. 1—8). Alkali ions and water molecules assumed to be in position 2a.

Beryl specimen		Observed intensity ratio	Calculated intensity ratio
Hunnakko	No. 1	1.00	0.97
Haapaluoma	» 2	1.00	0.96
»	» 3	1.00	0.94
»	» 4	1.00	0.96
»	» 5	0.97	0.90
»	» 6	0.99	0.94
»	» 7	0.97	0.92
»	» 8	0.94	0.92
Marropino Mine	» 9	0.79	0.76

On the other hand, the same intensity ratios for these beryl samples were measured from powder pattern. The measurements were made by means of the following experimental procedure:

The powdered beryl samples were eluted in water. The fraction of a grain size less than 6μ was adopted. The powder lines $21\bar{3}1$ and $11\bar{2}2$ were recorded with the Philips Norelco diffractometer, with a rotating specimen holder and filtered copper radiation being used. The intensity measurements were made with the automatic counting rate computer, 6400 pulses being counted in steps of 0.02° of the 2θ angle.

Table 3 summarizes the observed and calculated values for the intensity ratio $I_{21\bar{3}1}/I_{11\bar{2}2}$. In the diagram¹ of Fig. 2, the measured intensity ratios are plotted against the calculated ones.

In Fig. 2, the theoretical intensity ratios were calculated according to the following assumptions:

1) All the alkali ions and water molecules are in position 2a (values summarized in Table 3). The results, represented by the squares in the figure, show a fairly good agreement with the solid line.

2) The Li ions are in position 6f, substituting for Be and bringing the number of atoms in this position to 6. The rest of the alkalis and the water molecules are in position 2a. The results, indicated by the triangles, show a slightly better agreement with the solid line than in case 1.

3) The alkali ions in position 2b and the water molecules in 2a. The number of Be ions in the unit cell taken as 6. This assumption was propounded by Feklichev (1963). The discrepancy between the observed and the calculated intensity ratios is clear (solid circles in the figure).

4) The Na and Ca ions are in position 2a; the K ions and the water molecules in 2b; the Cs ions in 4e with $z = 0.68$. The Li ions have been placed

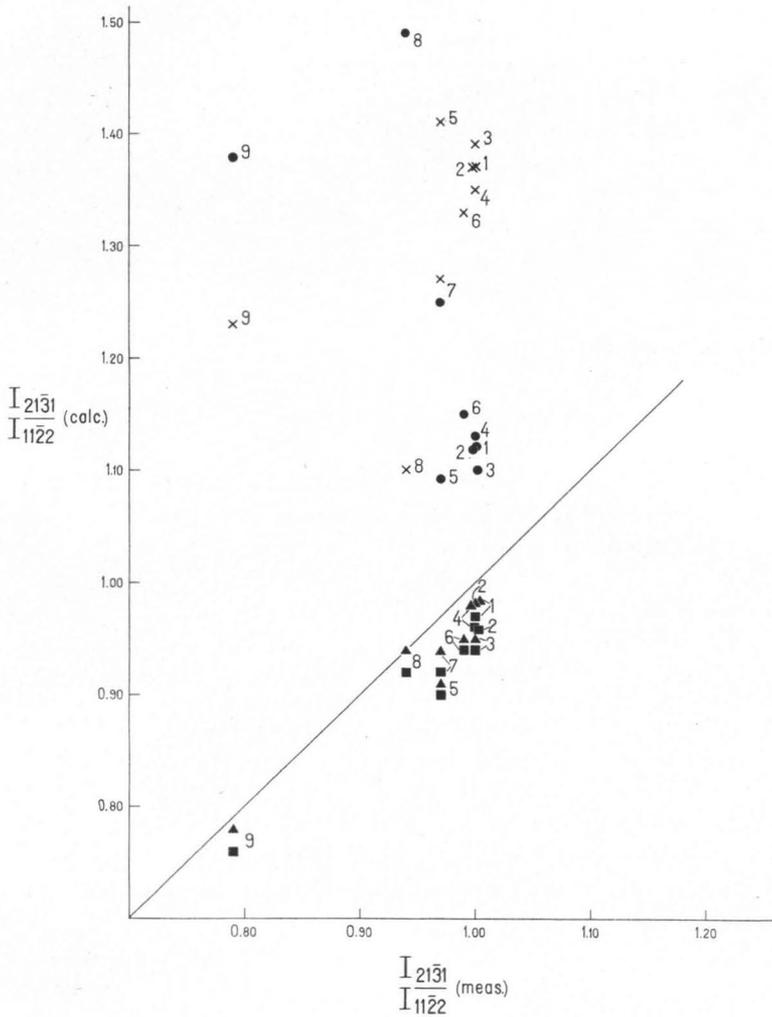


Fig. 2. Measured intensity ratios $I_{21\bar{3}1}/I_{11\bar{2}2}$ plotted against those calculated for the beryl specimens of Tables 1 and 2. The solid line indicates complete agreement of the measured intensity ratios with the calculated ones. Specimens 1—8 refer to the analyses 1—8 in Table 2; No. 9 refers to the beryl in Table 1.

Squares: Assuming alkali ions and water molecules in position $2a$ (values of Table 3).
 Triangles: Assuming Li in position $6f$, bringing the number of Be ions to 6. The rest of the alkalis and H_2O in $2a$.
 Solid circles: Assuming alkali ions in position $2b$ and H_2O in position $2a$.
 Crosses: Assuming Na and Ca in position $2a$; K and H_2O in $2b$; Cs in $4e$.

in 6*f*, bringing the sum Be+Li to 6. Rb was omitted. The agreement between the calculated intensity ratios and the observed ones is poor (crosses in the figure).

If every fourth Al ion is substituted by Li, the intensity ratio will increase by about 14 %. If, on the other hand, every fourth Al ion is replaced by Na, the intensity ratio will increase by only 1 %. If some of the alkalis, especially Li, were to occupy the position 4*c*, which is normally occupied by Al, the results given in this paper would not be valid. In the chemical analyses published by Ginzburg (1955) and by Bakakin and Belov (1962), however, the number of Al ions is usually quite nearly 4, making it less likely for such a substitution to be of any more importance.

The data for the intensity ratio $I_{21\bar{3}1}/I_{11\bar{2}2}$ obtained for the beryl specimens studied indicate that the most probable sites of the alkali ions and water molecules in the beryl structure are 2*a*, *i. e.*, in the channels between the Si_6O_{18} rings in the plane of the Al and Be ions. The greatest part of the Li ions obviously occupies the position 6*f*, bringing the number Be+Li to six. The chemical formulas for beryl compiled by Ginzburg (1955) and by Bakakin and Belov (1962) prove that the number of alkali + H_2O + Be in the unit cell of beryl very seldom exceeds eight (= 2 atoms in position *a* plus 6 atoms in position *f*). This conclusion agrees with the results of the present paper.

INFRARED ABSORPTION

It is of a certain interest to study whether or not the infrared absorption of beryl will be affected by introducing alkali atoms into the channels of the beryl structure. If such an effect is observable, then the infrared absorption method could possibly be used for a quick qualitative test of the alkali content of a particular beryl specimen. Hence, the beryl specimens listed in Tables 1 and 2 were subjected to infrared absorption measurements. The experimental procedure and the apparatus used were the same as those described by Sahama (1965) in connection with the study of the infrared absorption of nepheline.

The tracings of the infrared absorption spectrum obtained with the spectrophotometer will not all be reproduced here. Fig. 3 gives only the tracings of the following three beryl samples: morganite from the Marropino Mine (A), beryl No. 5 from the Haapaluoma pegmatite (B), and beryl from the Hunnako pegmatite (C).

As revealed by the unit cell contents listed in Tables 1 and 2, the Marropino Mine morganite was richest in alkalis and the Hunnako beryl poorest. The Haapaluoma beryl No. 2 lies in its alkali content about in the middle between these two samples.

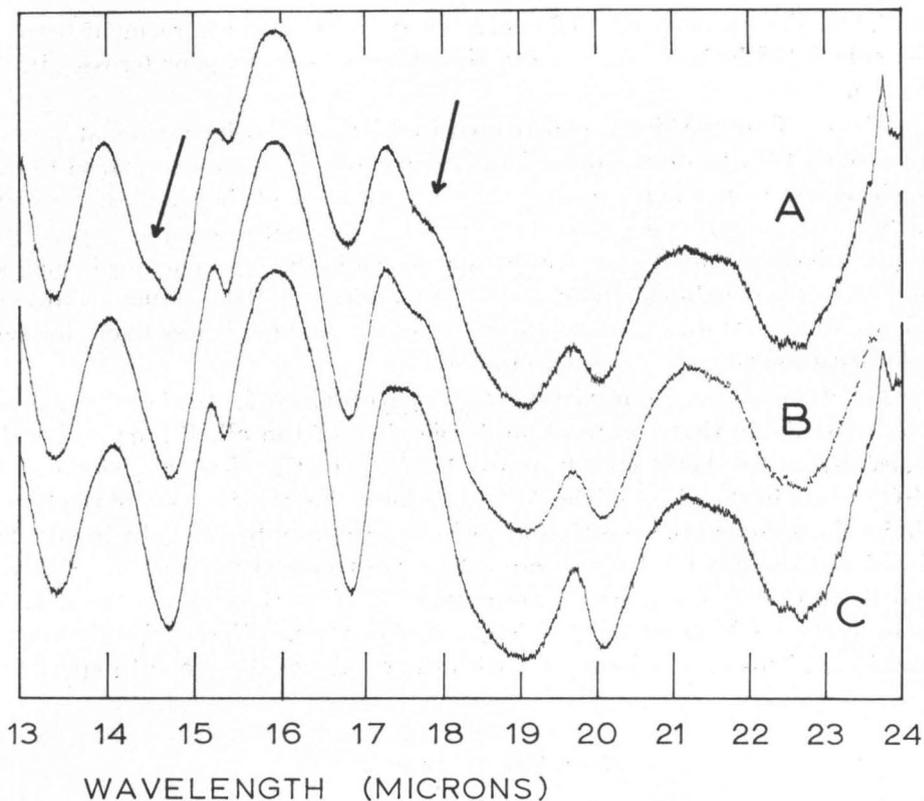


Fig. 3. Tracings of the infrared absorption spectrum of three beryl specimens. A) Morganite from the Marropino Mine. B) Beryl No. 5 from the Haapaluoma pegmatite. C) Beryl from the Hunnakkö pegmatite (No. 1).

In the wavelength range 13—24 μ , there are two places in which the tracings of the infrared absorption spectrum show a regular change in alkali content. Both places — situated at 14.5 μ and 17.7 μ , respectively — are marked with heavy arrows in Fig. 3. For an alkali-rich beryl, the band at 14.7 μ shows a subsidiary band on its short wavelength side. With a decreasing alkali content, this subsidiary band disappears. At ca. 17.7 μ , an alkali-rich beryl shows another subsidiary band that, with a decreasing alkali content, also disappears. This trend, shown by the two subsidiary bands, is quite regular in the series of beryl samples contained in Tables 1 and 2 and is clearly visible in Fig. 3. On the basis of the lack or existence of these two subsidiary bands, the alkali content of the beryl sample can be qualitatively tested. The occurrence of the two bands is independent of the water content of the beryl sample.

Acknowledgements — The authors are indebted to Dr. Oleg v. Knorring from the University of Leeds for making the chemical analysis of the Mozambique morganite as well as to Mr. Pentti Ojanperä, M. A., of the Geological Survey of Finland, for the analyses of the Finnish beryl specimens. Thanks are also due to Mr. Esko Saari, M. A., of the University of Helsinki, for assistance in the Laboratory and to Mr. J. D. Cabral, of the Marropino Mine, Mozambique, for the specimen of morganite.

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ENTWICKLUNG DES MOORES HARJULANSUO ¹⁾

VON

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AUSZUG

Die Entwicklung des Moores Harjulansuo wird mit Hilfe von stratigraphischen Profilen, Pollendiagrammen und Diatomeenanalysen ergründet. Es wird bewiesen, dass die Erosion und die Umlagerung der älteren Schichten, verursacht durch eine Transgression, in der pollenanalytischen Altersbestimmung und in der C¹⁴-Datierung grobe Fehler veranlassen können. Das gilt besonders für Moore, die an einem stark geneigten Uferhang entstanden sind.

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LAGE DES MOORES

Das Moor Harjulansuo liegt etwa 10 km von Jyväskylä nach Norden, nördlich des Sees Tuomiojärvi unweit des Gehöftes Harjula (Abb. 1). Bei Erforschung der Transgression des Vor-Päijänne (R. Aario 1965, S. 131) traten im Aufbau des Moores Einzelheiten hervor, die zu seiner ausführlicheren Untersuchung Anlass gaben.

¹⁾ Eingegangen am 17. Dezember. 1964

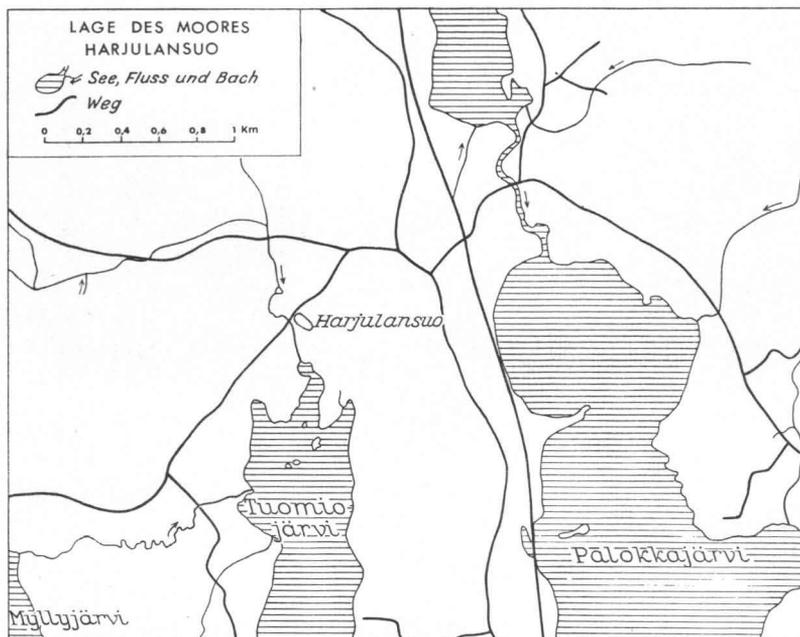


Abb. 1. Lage des Moores Harjulansuo.

Das Moorbecken ist ein steilrandiger Oskessel. Als Gefälle seiner Hänge wurden stellenweise bis über 40° gemessen. Trotz der geringen Grösse des Moores reichten die organogenen Ablagerungen bis in etwa 9 m Tiefe, so dass das Moor zu den tiefsten Finnlands gehört. Das Becken ist im Südwesten, Westen und Nordwesten von einem niedrigen und schmalen, aus Grobsand und Kies bestehenden Rücken begrenzt. In den übrigen Richtungen stösst das Moor gegen höhere Teile des Oses. Die im Nordwesten gelegene Schwelle des Moores liegt in 100.4 m Höhe, nur etwa einen halben Meter höher als seine Oberfläche. Wegen des quer durch den Rücken ausgehobenen Grabens ist die Oberfläche des Moores heutzutage verhältnismässig trocken.

AUFBAU DES MOORES

Zur Klärung des Aufbaues des Moores sind drei Profile (Beilagen I—II und Abb. 3—4) gearbeitet worden, deren Lage wie auch Bohrungspunkte aus der Karte (Abb. 2) hervorgehen.

Bei Profil A-A' (Beilage I und Abb. 3) bildet den Moorgrund auf Grobsand lagernder Ton, der sich bis in die randlichen Moorteile fortsetzt. Ganz am Rande des Moores fehlen jedoch stellenweise die Tonablagerungen, und der

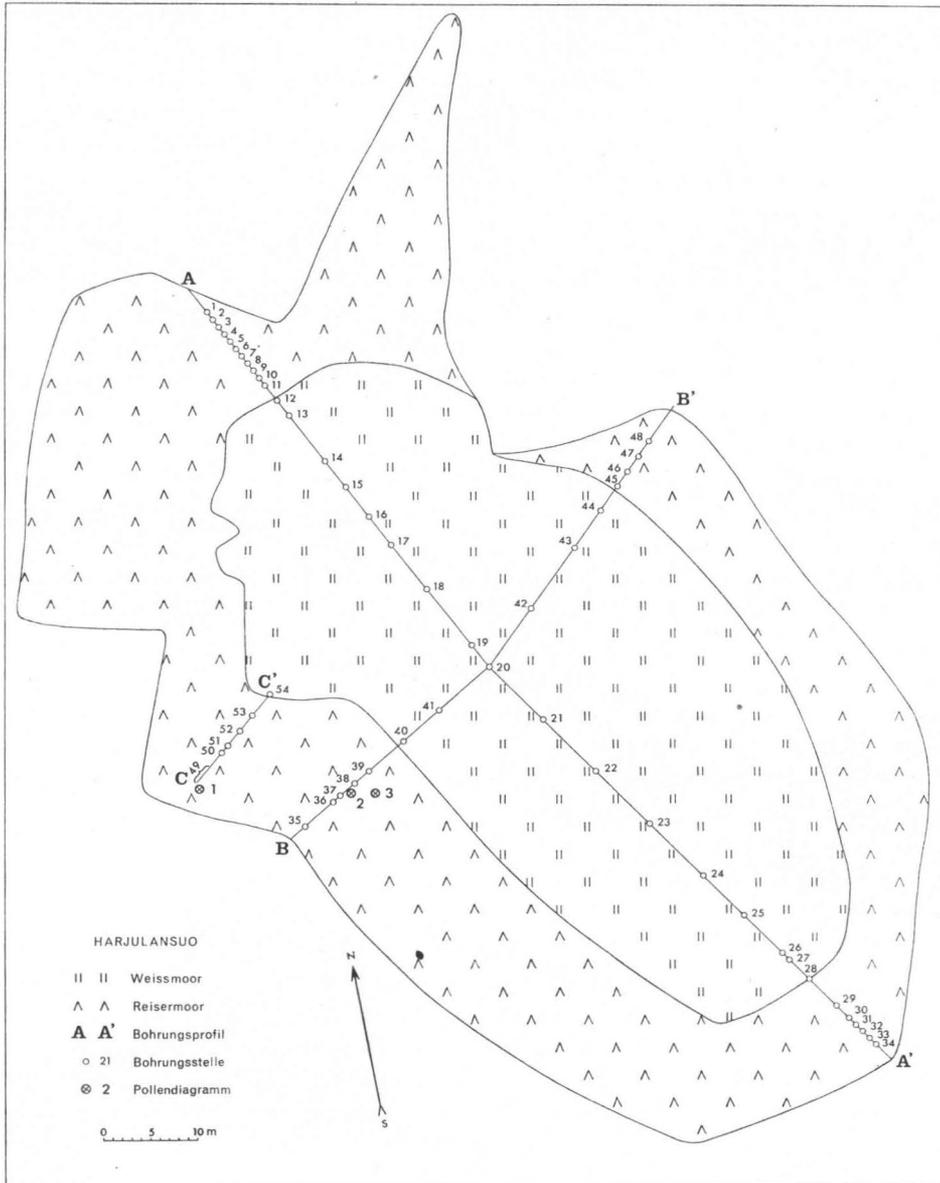


Abb. 2. Das Moor Harjulansuo mit Bohrungsprofilen.

Grund besteht aus Grobsand. An Bohrungstelle 8 enthält die Tonschicht auch Torfstückchen (Abb. 3). In den mittleren Teilen des Moores sind dem Ton dünne Vivianitschichten ein- und aufgelagert, von denen die dicksten

etwa 10 cm stark sind. Auf den Ton folgt in den Randteilen des Moores unmittelbar Torf, weiter nach der Mitte zu Gyttja. Die an den Ton grenzende Gyttjaschicht ist dünn und enthält auch etwas Dy, in den mittleren Teilen des Moores ausserdem Vivianitstreifen. An vielen Stellen ist sie verhärtet, ja sogar mühevoll mit dem Bohrer zu durchdringen. Die Gyttjaschicht ist sowohl von reinem als auch von gyttjadurchsetztem Torf überlagert, den eine auf eine Transgression hinweisende jüngere und dickere Gyttjaschicht bedeckt.

In den mittleren Teilen des Moores umfasst der zwischen Gyttjaschichten lagernde Torf hauptsächlich *Bryales*-Torf. Auch Seggen- und Schachtelhalm-torf kommt verhältnismässig tief vor (Bohrungspunkte 16, 18 und 25). Etwas seitlicher vom Profil ist auch Weissmoortorf anzutreffen (Diagr. 3 S. 142).

In den Randteilen des Moores liegt zwischen den Gyttjaschichten hauptsächlich Seggentorf. Auch *Bryales*-Torf, Holzreste und *Equisetum*-Torf kommen vor, jedoch stets seggenhaltig. An einigen Stellen bleibt diese Torfablagerung ganz aus (Bohrungspunkt 7). Nahe am Rande des Moores setzt sich der auf Grobsand, Ton oder Gyttja lagernde Torf ununterbrochen bis in die Oberflächenteile fort.

Die obere Gyttjaschicht ist in den mittleren Partien des Moores über einen halben Meter mächtig und besteht hauptsächlich aus weicher, brauner, dyhaltiger Gyttja. Näher dem Rande des Moores ist die Gyttjaschicht dünner, und sie enthält einerseits Ton, Gyttjaton und Tongyttja, andererseits auch Torf. Wie auch aus dem den Randteil von Profil A-A' darstellenden Teilprofil (Abb. 3) zu ersehen, kommt der Ton in verschieden grossen Stücken vor. Wegen der wechselnden Grösse der Tonbrocken zwischen Stücken von vielen Zentimetern Dicke und Teilchen von nicht einem Millimeter Stärke (S. 143) haben sie nicht gesondert in die Profile eingetragen werden können.

Die Gyttjaschichten auf dem Torf reichen in der Profilstelle am höchsten bis in ein Niveau von etwa 99 m Meereshöhe, wo sie auskeilen.

Auf die obere Gyttjaschicht folgt eine Torfablagerung, die sowohl in den mittleren als auch in den randlichen Moorteilen hauptsächlich *Carex*- und *Sphagnum*-Torf enthält. Doch kommt in den mittleren Partien auch reichlich *Bryales*-Torf vor. Ein für die randlichen Teile typischer Zug ist die Buntheit der zwischen den *Carex*- und *Sphagnum*-Schichten vertretenen übrigen Torfkomponenten. Die verschiedenartigen Torfbestandteile treten oft in nur kleinräumigen Linsen auf. In den Profilen haben auch diese Linsen ihrer Kleinräumigkeit wegen nicht dargestellt werden können. Bei Bohrungsstelle 6 ist eine Torfgrube ausgestochen, worauf das im Profil vermerkte Wasser beruht.

Bemerkenswert unter den übrigen bei dem zu betrachtenden Längsprofil erscheinenden stratigraphischen Zügen sind die über der oberen zusammenhängenden Gyttjaschicht und von ihr getrennt auftretenden geringmächtigen und kleinräumigen Gyttjalinsen zwischen dem Torf (z. B. Bohrungsstelle 7, 9, 10 und 11).

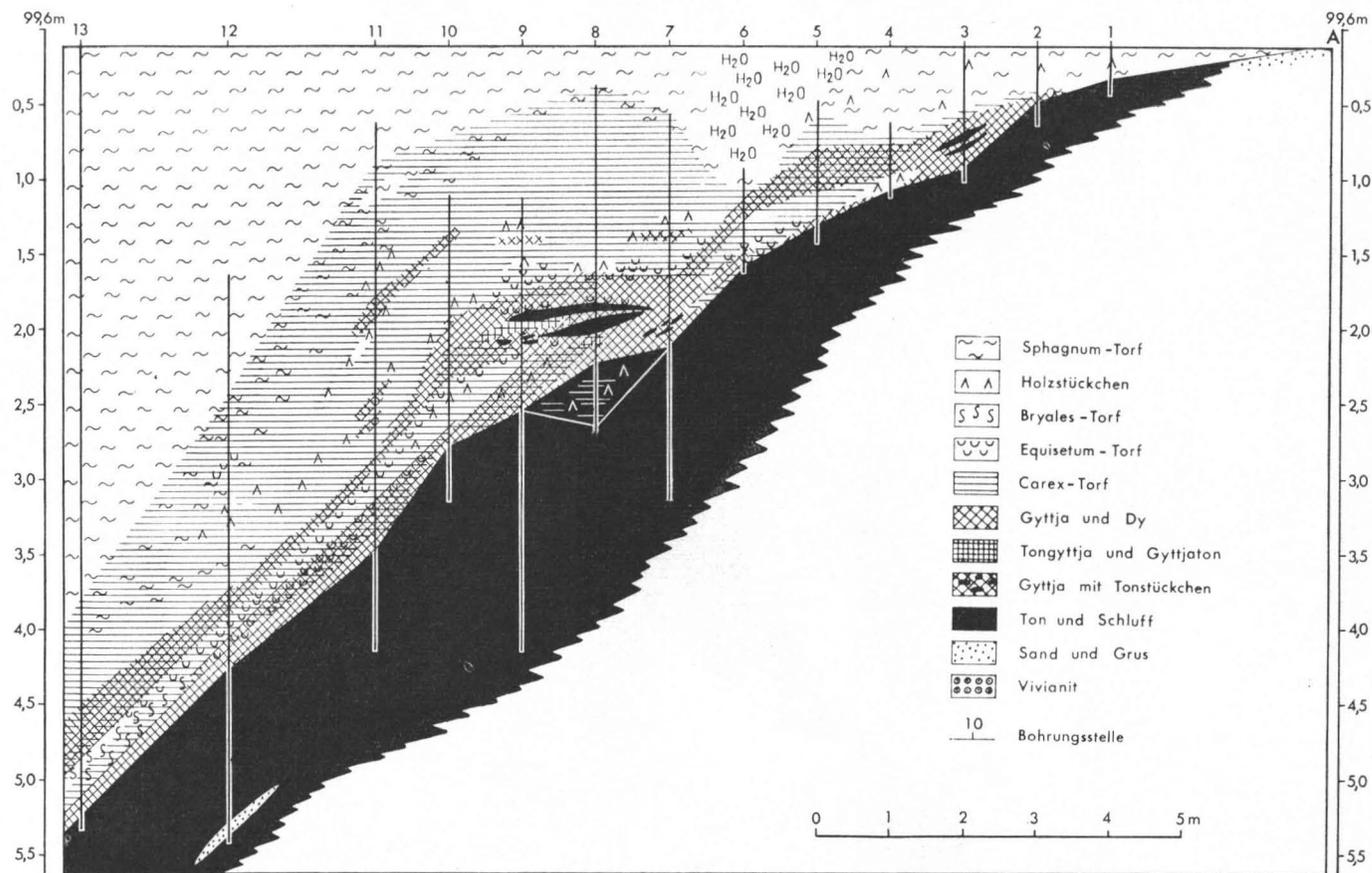


Abb. 3. NW-Ende des Längsprofils A' — A in einem nur zweimal überhöhten Höhenmassstab.

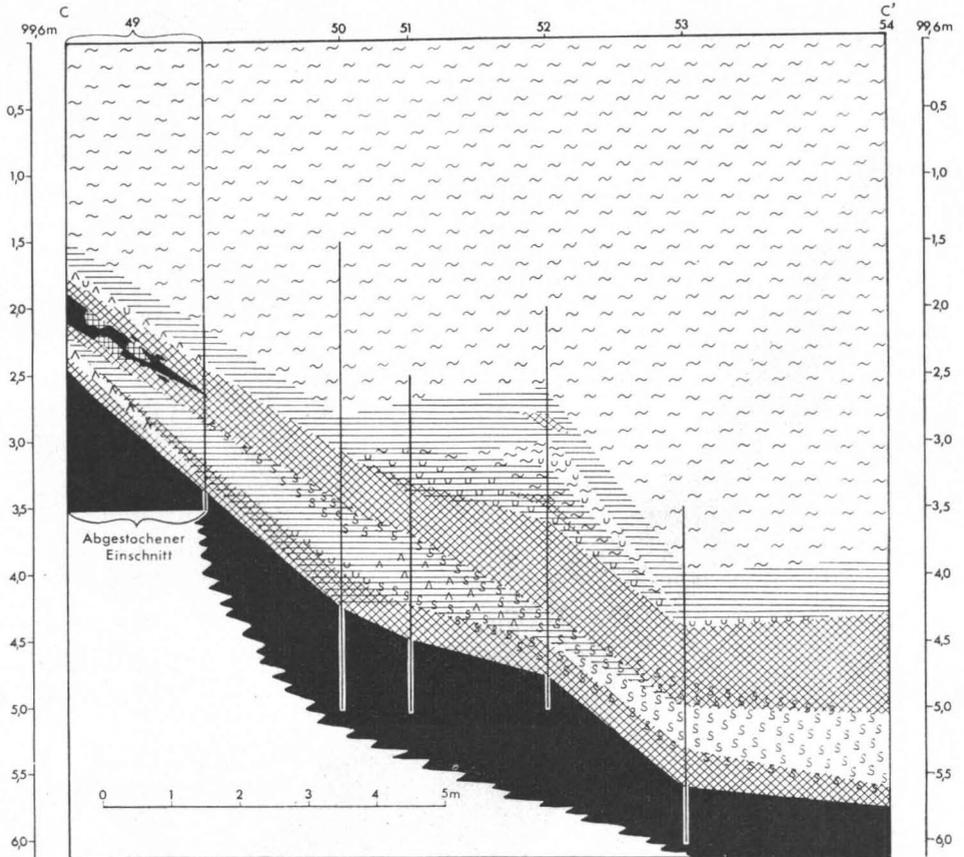


Abb. 4. Querprofil C—C'. Erläuterungen wie bei Abb. 3.

Bei dem Querprofil B-B' (Abb. 2 und Beilage II) treten die gleichen Hauptzüge wie bei dem Profil A-A' auf. Zuunterst liegt Ton, dazwischen und darauf in den mittleren Moorteilen etwas Vivianit. Den Ton überlagert in den mittleren Flächen des Moores eine dünne, oft hartgetrocknete Gyttschicht, die ebenfalls Vivianit enthält. Ihr folgt *Bryales*- oder *Carex*-Torf und darauf nochmals Gyttschicht.

In den mittleren Teilen des Moores besteht der zwischen den Gyttschichten lagernde Torf hauptsächlich aus *Bryales*-Torf, der auf die randlichen Teilen zu in *Carex*-Torf übergeht. *Equisetum*-Torf kommt spärlich vor, wie in dem vorhergehenden Profil.

Die obere Gyttschicht ist an der Profilstelle weich, von brauner Farbe und dyhaltig. Sie setzt sich in dem flacheren südwestlichen Teil des Moores

bis in wenigstens 98.5 m Höhe fort, aber in dem steilen nordwestlichen bis in ein Niveau von nur etwa 96.5 m Höhe.

In den Randpartien des Moores erscheinen in dem hauptsächlich Segge enthaltenden Torf linsenförmig andere Torfarten wie auch kleinere linsenförmige Gyttjaablagerungen (z. B. Bohrungspunkte 38 und 39).

Profil C-C' (Abb. 4) liegt im westlichen Teil des Moores nahe bei Profil B-B'. Die den *Bryales*- und *Carex*-Torf überlagernde, eine Transgression bezeugende Gyttjaschicht ist hier durchweg verhältnismässig mächtig, selbst bei geringmächtigstem Vorkommen nahezu einen halben Meter stark. Dem nordwestlichen Ende des Profils ist mitten in der Gyttja ein ton-, tongyttja- und gyttjatonhaltiger Keil eingelagert. In dem gegrabenen Einschnitt (49) liess sich deutlich erkennen, dass sich der Ton grossenteils in Stücken verschiedenen Umfangs am Orte abgesetzt hatte (vgl. Prof. A-A').

Auch bei diesem Profil begegnen einzelne Gyttjalinsen (Bohrungsstelle 52) oberhalb der einheitlichen oberen Gyttjaschicht im Torfe nahe dem steilen Beckenrand. Das Material dieser Linsen kann randlichen Mooranteilen entstammen, aus denen es durch die im Frühjahr erodierenden Schmelzwasser oder durch Regenwasser hat ausgewaschen werden können. Doch ist es ebenfalls möglich, dass die Gyttjalinsen kurzfristige, bodenfrostbedingte Wasser-schlenken vertreten.

ENTWICKLUNG DES MOORES IM LICHTE VON POLLENCHRONOLOGIE UND DIATOMEENUNTERSUCHUNGEN

Diagramm 1 (Abb. 5) vom Randteil des Moores (Abb. 2) erweist, dass die untere Tonablagerung in der Zeit mit vorherrschender Kiefer vor der Verhäufung der Erle entstanden ist. Auch ist das Abschnüren des Beckens in jener Zeit vor sich gegangen. Ein kleiner NBP-Gipfel, der sich hauptsächlich aus Gramineenpollen zusammensetzt, in ca. 2.2 m Tiefe, mag auf Ufervegetation zurückzuführen sein.

Über Ton in 255 cm (D 1) und 223 cm (D 2) Tiefe sind Diatomeenanalysen angestellt worden (Tab. 1). In der unteren Probe (D 1) überwiegen Süsswasserarten, wie *Fragilaria lapponica*, *Campylodiscus noricus* und verschiedene *Gyrosigma*-Arten. Der *Ancylus*-Artenbestand erweist das Eintreten der Abschnürung vom *Ancylussee* (R. Aario 1965, S. 58), was auch mit den Auffassungen Sauramos (1958 S. 267) und Hyypäs (1963 S. 39 und Beilage I) im Einklang steht. Noch in dem 223 cm tief gelegenen Ton (D 2) finden sich in reichlichem Masse Planktonarten, aber der gesteigerte Flachwasser-Diatomeengehalt erweist, dass eine Verseichtung bereits vor sich gegangen ist. Doch hat sich die Abschnürung offenbar erst beim Absetzen von Tongyttja oder Gyttja abgespielt.

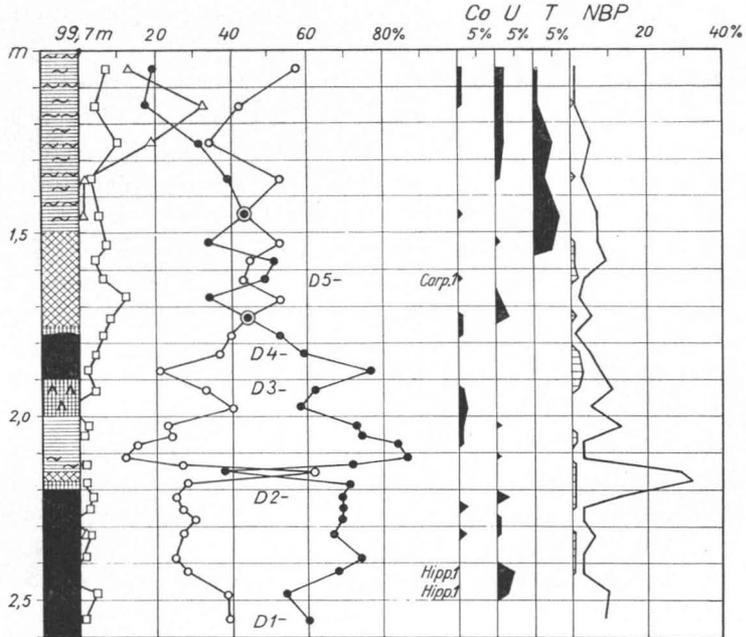


Abb. 5. Pollendiagramm 1. Kräuter- und Zwergstrauchpollen im NBP-Abschnitt schraffiert.

Der nach dem Zuwachsen entstandene Seggentorf entfällt auch auf die Zeit mit vorherrschender Kiefer, desgleichen teilweise noch die den Torf überlagernde Tonschicht. In letzterer liegt jedoch ein ziemlich steiler Abfall der Kiefernkurve, und zugleich beginnt die Erlenkurve langsam zu steigen. Wenigstens der untere Teil des Tones scheint nach dem Diagramm in die der Erlenverhäufung vorausgegangene Zeit zu entfallen. Dagegen gehört die den Ton überlagernde Gytjtjaschicht schon in ihrer Gesamtheit zur erlenreichen, durch vorherrschendes Laubholz gekennzeichneten Zeit, und in ihren oberen Teilen ist auch *Tilia* schon gemein geworden.

Ton und Tongyttja über der Torfablagerung enthalten sowohl Arten saurer Gewässer (D 3 und D 4), wie *Eunotia*- und *Pinnularia*-Arten, als auch für die unter dem Torf vorkommende Tonablagerung kennzeichnende Arten (Tab. 1). In der Gytjtjaprobe (D 5) umfasst der Diatomeengehalt ausschliesslich für Kleingewässer typische Arten.

In Diagramm 2 (Abb. 6), das eine etwas tiefere Stelle des Moores vertritt, reicht die Probenreihe nicht bis in die untere Tonschicht. Auch in dieser Lagerfolge zeigt sich auf dem unteren Torf die auf eine Transgression hinweisende Gytjtjaschicht. Die untere Torfschicht ist an dieser Stelle unverkennbar erst nach der Erlenverhäufung entstanden und ist also bedeutend

Tabelle 1

	D 1 %	D 2 %	D 3 %	D 4 %	D 5 %
<i>Cyclotella Kütziana</i> var. <i>radiosa</i> Fricke					1
<i>Melosira ambigua</i> (Grun.) Müller					4
<i>Melosira granulata</i> (Ehr.) Ralfs	2			3	2
<i>Melosira islandica</i> O. Müll.			2		
<i>Melosira</i> spp.			8	2	8
<i>Melosira</i> spp. fragm.			6	2	3
<i>Achnanthes Östrupi</i> (A. Cl.) Hust.		1			
<i>Amphiprora ornata</i> Bailey	1				
<i>Amphora</i> spp.	2				
<i>Caloneis bacillum</i> (Grun.) Mereschk.	2				1
<i>Caloneis</i> sp.					1
<i>Caloneis</i> sp. fragm.				1	
<i>Campylodiscus hibernicus</i> (Ehr.) Grun.		6	2		
<i>Campylodiscus noricus</i> Ehr.		5			
<i>Campylodiscus noricus</i> fragm.	1				
<i>Campylodiscus</i> spp. fragm.	23	18	4	3	
<i>Cocconeis diminuta</i> Pant.	1				
<i>Cymbella perpusilla</i> A. Cleve			2		
<i>Cymbella sinuata</i> Greg.					1
<i>Cymbella turgida</i> (Greg.) Cleve					2
<i>Cymbella ventricosa</i> Kütz.				1	
<i>Cymbella</i> spp.				2	
<i>Diploneis elliptica</i> (Kütz.) Cleve	3				
<i>Diploneis elliptica</i> var. <i>ladogensis</i> Cleve	3				
<i>Diploneis Mauleri</i> (Brun) Cleve	1			2	
<i>Epithemia turgida</i> (Ehr.) Kütz.		1			1
<i>Epithemia zebra</i> (Ehr.) Kütz.	2				
<i>Epithemia zebra</i> var. <i>porcellus</i> (Kütz.) Grun.	1				
<i>Epithemia</i> sp. fragm.	1				
<i>Eunotia flexuosa</i> (Breb.) Kütz.				2	
<i>Eunotia formica</i> Ehr.					1
<i>Eunotia lunaris</i> (Ehr.) Grun.			2		1
<i>Eunotia pectinalis</i> var. <i>minor</i> (Kütz.) Rabh.			4	1	1
<i>Eunotia pectinalis</i> var. <i>minor</i> f. <i>intermedia</i> Krasske					1
<i>Eunotia pectinalis</i> var. <i>ventralis</i> (Ehr.) Hust.				1	1
<i>Eunotia polyglyphis</i> Grun.					1
<i>Eunotia praerupta</i> Ehr.				2	
<i>Eunotia valida</i> Hust.				2	
<i>Eunotia veneris</i> (Kütz.) O. Müll.			4		
<i>Eunotia</i> spp.			2	8	
<i>Fragilaria brevistriata</i> Grun.					6
<i>Fragilaria construens</i> (Ehr.) Grun.					3
<i>Fragilaria construens</i> var. <i>venter</i> (Ehr.) Grun.			4	3	3
<i>Fragilaria lapponica</i> Grun.	26	53	20	3	8
<i>Fragilaria pinnata</i> Ehr.	5				1
<i>Fragilaria pinnata</i> var. <i>lancectula</i> (Schum.) Hust.		1			1
<i>Fragilaria virescens</i> Ralfs		1			9
<i>Fragilaria virescens</i> var. <i>mesolepta</i> v. Schönfeldt					2
<i>Fragilaria</i> spp.	2				2
<i>Frustulia rhomboides</i> var. <i>amphiplouoides</i> Grun.					1
<i>Gomphonema acuminatum</i> var. <i>coronata</i> (Ehr.) W. Smith			2		1
<i>Gomphonema bohemicum</i> Reichelt et Fricke			4	4	
<i>Gomphonema gracile</i> var. <i>lanceolata</i> (Kütz.) Cleve				1	
<i>Gomphonema</i> spp.				2	
<i>Gomphonema</i> sp. fragm.					2
<i>Gyrosigma acuminatum</i> var. <i>gallica</i> Grun.	5	1			

Tabelle I (Forts.)

	D 1	D 2	D 3	D 4	D 5
<i>Gyrosigma attenuatum</i> (Kütz.) Rabh.		1			
<i>Gyrosigma attenuatum</i> fragm.	2			1	
<i>Gyrosigma Kützingii</i> (Grun.) Cleve	5	2			
<i>Hantzschia amphioxys</i> f. <i>capitata</i> O. Müll.				3	
<i>Hantzschia amphioxys</i> var. <i>maior</i> Grun.					2
<i>Hantzschia</i> sp. fragm.					1
<i>Navicula aboensis</i> Cleve		3	4		1
<i>Navicula americana</i> Ehr.				1	
<i>Navicula Jentzschii</i> Grun.					1
<i>Navicula Järnefeltii</i> Hust.	1				
<i>Navicula menisculus</i> Schum.	1	1			
<i>Navicula mutica</i> Kütz.					1
<i>Navicula pseudoscutiformis</i> Hust.	1				
<i>Navicula radiosa</i> Kütz.					2
<i>Pinnularia brevicostata</i> Cleve				3	
<i>Pinnularia gibba</i> Ehr.				1	1
<i>Pinnularia gibba</i> var. <i>linearis</i> Hust.				3	1
<i>Pinnularia mesolepta</i> (Ehr.) W. Smith					1
<i>Pinnularia nodosa</i> Ehr.					1
<i>Pinnularia subcapitata</i> Greg.					1
<i>Pinnularia viridis</i> (Nitzsch) Ehr.		1		14	4
<i>Pinnularia viridis</i> var. <i>sudetica</i> (Hilse) Hust.			2	2	1
<i>Pinnularia</i> sp.					1
<i>Pinnularia</i> spp. fragm.	1		8	11	5
<i>Rhopalodia gibba</i> (Ehr.) O. Müll.		1			1
<i>Stauroneis anceps</i> Ehr.				2	
<i>Stauroneis anceps</i> f. <i>gracilis</i> (Ehr.) Rabh.				1	
<i>Stauroneis phoenicenteron</i> (Nitzsch) Ehr.			4		
<i>Stauroneis</i> spp. fragm.				6	1
<i>Surirella biseriata</i> Bréb.	1				
<i>Surirella biseriata</i> var. <i>rostrata</i> Schulz		1			
<i>Surirella</i> sp.		1			
<i>Surirella</i> spp. fragm.	5	1			
<i>Symedra ulna</i> var. <i>danica</i> (Kütz.) Grun.		1			
<i>Tabellaria fenestrata</i> (Lyngbye) Kütz.	1		12	6	2
<i>Tabellaria flocculosa</i> (Roth) Kütz.	1		4		1
<i>Tetracyclus emarginatus</i> (Ehr.) W. Smith					1
<i>Tetracyclus lacustris</i> Ralfs				1	1

jünger als im vorhergehenden Diagramm. Die Tonschicht fehlt über dem unteren Torf, das Alter der oberen Gyttjaschicht ist ungefähr dasselbe wie im vorhergehenden Diagramm. Die Verhäufung von Linde vollzieht sich in den oberen Teilen dieser Gyttjaschicht, wie in jenem.

Die in Diagramm 3 (Abb. 7) unterlagernde Tonschicht gehört noch in die Zeit mit vorherrschender Kiefer, aber schon in ihren oberen Teilen sinkt die Kiefernkurve steil ab. Die Erlenverhäufung geschieht im Übergang der dünnen Gyttjaschicht in Torf, so dass also auch an dieser Stelle die untersten Torfschichten des Moores der erlenreichen, laubholzbeherrschten Zeit entstammen. Die auf der unteren Torfschicht lagernde Gyttja scheint in ihren Altersverhältnissen ebenfalls der oberen Gyttjaschicht des vorhergehenden

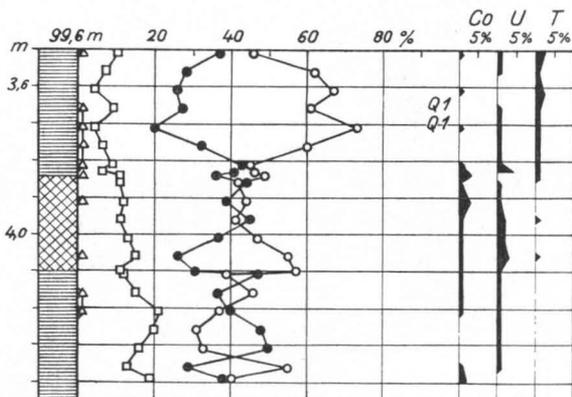


Abb. 6. Pollendiagramm 2.

Diagrammes ungefähr zu entsprechen. Doch ist die Stelle der Verhäufung von *Tilia* in diesem Diagramm nicht ebenso deutlich wie in den beiden vorigen festzulegen, denn vor dem Einsetzen der einheitlichen Kurve sind verstreute Vorkommen reichlich wahrzunehmen.

Ein Vergleich der oben dargestellten Diagramme scheint zu dem Ergebnis zu führen, dass das Moor von zwei Transgressionen überspült worden wäre. Die von der älteren Transgression zeugende Schichtenfolge träte jedoch nur in den randlichen Moorteilen auf, wo wiederum Anzeichen einer jüngeren fehlen. Sowohl die untere Torfschicht als auch der sie überlagernde Ton sind nach dem Diagramm am Rande des Moores deutlich älter als die untere Torfschicht in den beiden übrigen, den mittleren Moorteil vertretenden, Diagrammen.

Die in den Diagrammen 2 und 3 zu sehende Transgression entspricht deutlich der Transgression des Vor-Päijänne, aber die Lage der in Diagramm 1 vorkommenden Transgression scheint in bezug auf die Pollenzonen verhältnismässig genau die gleiche zu sein wie die der Ancylostansgression in der Pollenchronologie Sauramos (1954, S. 202). Dies schiene noch durch das Vorkommen einiger für den Ancylostsee kennzeichnenden Diatomeen in dem der Transgression entsprechenden Ton bestätigt zu werden (D 3 und D 4). Auch stimmt das Ergebnis mit der von Sauramo für dieses Gebiet ermittelten Höhe des Ancylostufers (1958 S. 267) gut überein. Das Diagramm nebst den mit ihm zusammenhängenden Diatomeenbestimmungen scheint also recht überzeugend zu erweisen, dass die den Torf überlagernde Tonschicht durch die Ancylostansgression verursacht worden wäre.

Auf der anderen Seite aber wäre es schwerlich denkbar, dass die Ancylostansgression auch auf so hoher Isobase hätte auftreten können. Ein anderer zu beachtender Umstand besteht darin, dass die den Torf überlagernde

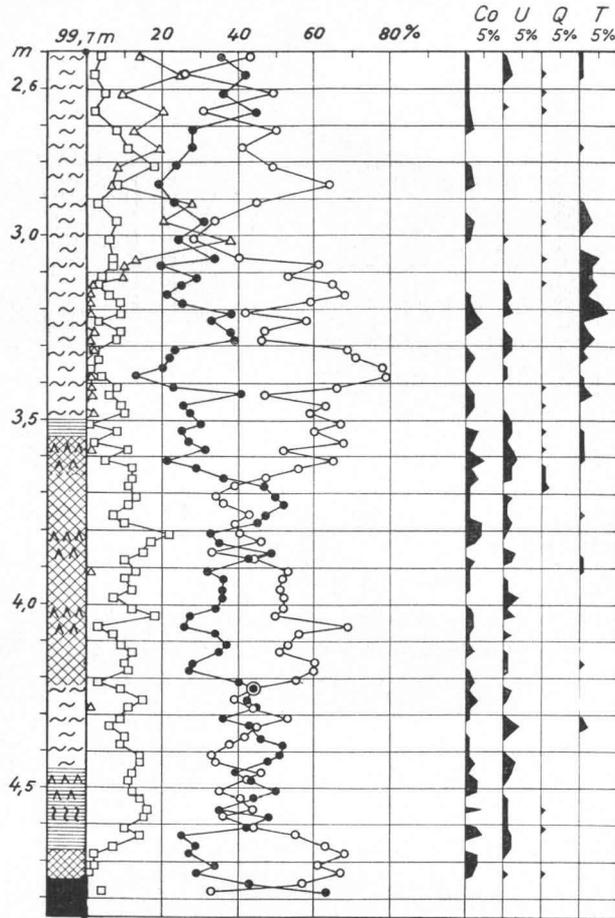


Abb. 7. Pollendiagramm 3.

Gyttjaschicht in Diagramm 1 ungefähr gleichen Alters wie in den beiden anderen Diagrammen ist. Zum mindesten in den Diagrammen 1 und 2 hat die Gytjtjabildung kurz nach der Lindenverhäufung an der Stelle gleichzeitig aufgehört. Dies liesse sich zwar leicht dadurch erklären, dass die Gytjtjaschicht eine jüngere, der Ton und die Tongyttja eine ältere Transgression verträten, obschon dazwischen terrestrische Bildungen fehlen. Die Diskordanz in der Lagerfolge ist jedenfalls offensichtlich.

Aus den oben dargestellten Gründen wurde die Lagerfolge der randlichen Moorteile einer eingehenderen Erforschung unterzogen. Nahe dem Rande des Moores wurde eine in das Profil C-C' eingetragene drei Meter tiefe, 3—4 m lange Grube ausgehoben, die bis in den am Moorgrunde lagernden Ton reicht.

Dabei stellte es sich heraus, dass die auf eine Transgression hinweisende, Ton enthaltende Schicht auf dem älteren Torf steil gegen die Mitte des Moores abfällt und dass sie auf kurzer Strecke in eine Gyttja übergeht, die ebenso beschaffen ist wie die, die in den Diagrammen 2 und 3 auf der älteren Torfschicht vorkommt. Durch nahe beieinander gelegene Bohrungen konnte des weiteren festgestellt werden, dass die zwischen den Torfschichten der Randeile auftretenden Wassersedimente sich mit der in den anderen Diagrammen vorkommenden jünger scheinenden Schicht vereinigen.

In der Wand der ausgehobenen Grube liess sich auch leicht erkennen, dass der Ton keine einheitliche Schicht bildete, sondern der Torf von organogener Gyttja bedeckt war, die deutlich getrennte grössere und kleinere Tonbrocken enthielt. Über die getrockneten Proben konnte ebenfalls festgestellt werden, dass die im Gelände als Tongyttja vermerkte Bodenart in der Tat ein Gemisch von Gyttja und kleinen, durchschnittlich einen Millimeter grossen Tonkörnchen war, beide Bestandteile grossenteils in scharfer Abgrenzung voneinander abgehoben. Die Menge der Tonteilchen vermindert sich im Mittel einerseits aufwärts, anderseits gegen die mittleren Teile des Moores.

Die Beobachtungen in der Grube erweisen, dass im Moore also nur eine Transgression (vgl. R. Aario 1965 S. 134) vorgekommen ist, wie in den Profilen vermerkt. Zugleich machen sie die in Diagramm 1 erscheinenden Unterschiedlichkeiten gegenüber den anderen Diagrammen verständlich. Der den unteren Torf überlagernde Ton muss vom nahen Rande des Moores herrühren und in seinen ursprünglichen Altersverhältnissen dem am Moorgrunde abgesetzten Ton entsprechen. Als der Ton, wahrscheinlich durch Eiswirkung losgelöst, während der Transgression an seine gegenwärtige Stelle übertragen worden ist, hat sich mit ihm auch seine ursprüngliche Diatomeen- und Pollenflora verfrachtet. Somit findet sich unter den Arten saurer Kleingewässer ein ganz andersartiges in grossen Gewässern heimisches Diatomeenmaterial, wie oben dargestellt. Da der am Grunde des Moores lagernde Ancyluston in spärlicher Menge Diatomeen enthält, bleibt der Anteil von Grossgewässerdiatomeen in den Proben D 3 und D 4 weit geringer als der der Formen saurer Kleingewässer, obgleich das Tonmaterial in diesen Proben überwiegt. Im Pollengehalt herrscht anfangs die Pollenflora des am Moorgrunde lagernden Ancylustons vor, aber weiter aufwärts und bei zunehmendem Anteil des Gytjamaterials nimmt die Erlenmenge zu.

In Diagramm 1 ist jedoch auch die am Grunde des Moores vorhandene Torfschicht älter als in den beiden anderen, ja sogar so sehr, dass selbst die obere Kante jener Schicht darin älter ist als die ältesten Teile des Torfes in den Diagrammen 2 und 3. Auch dies ist jedoch nur eine natürliche Folge der Entwicklung des Moores. In der Stelle von Diagramm 1 liegt die untere Grenze des Torfes über 2 m höher als in den beiden anderen Punkten. Die

Torfbildung ist also während der Landhebung früher als in den mittleren Teilen des Moores in Gang gekommen. Auch ist es wohl wahrscheinlich, dass die mittleren Teile des Beckens nicht einmal gleich nach ihrem Auftauchen mit Moorvegetation überzogen worden wären, da die oberen Teile der am Grunde liegenden Gyttja eine zähe, schwer durchsinkbare Schicht enthalten, die auf Abtrocknung hinweist.

Zusammenfassend kann also festgestellt werden, dass die Entwicklung des Moores folgendermassen vor sich gegangen ist:

Während das Land aus dem Ancylussee aufstieg, bildete sich am Ufer des in dem Oskessel entstandenen Weiheres ein Seggenweissmoor, dessen Wachstum jedoch bei weiterem Sinken des Wasserspiegels bald infolge von Abtrocknung halt machte. Beim Absinken des Wassers sind an einigen Stellen am steilen Uferhang Ton- und Torfstücke herabgerutscht und in die einheitliche Tonschicht eingebettet worden. Der Wasserspiegel mag sich längere Zeit in einem Niveau von etwa 97—98 m gehalten haben, da sich am Ufer eine Torfschicht von mindestens 30 cm gebildet hat, während die unteren Teile des Beckens noch bis in eine Meereshöhe von mindestens 96.5 m wasserbedeckt gewesen sind. Es sind eben aus dieser Torfschicht erodierte Stücke, die an Bohrungstelle 8 im Ton als sekundäre Ablagerung vorkommen. Nachdem das ganze Becken aus dem Wasser aufgestiegen war, dürfte sein Grund eine Zeitlang trockengelegen haben, aber bald dann, als die Erle im Gebiet häufig wurde, vermoort sein. Durch die Transgression des Vor-Päijänne stieg das Wasser aufs neue in das Moorbecken und überflutete zuerst die tiefsten Teile. Nachdem die Transgression die ziemlich steilen Randteile des Oskessels erreicht hatte, haben das Eis und vielleicht auch der Wellengang Material der vor der Transgression entstandenen Ton-, Gyttja- und Torfschichten abgetragen und in Ufernähe, aber nicht mehr nennenswert auf tieferen und flacheren mittleren Teilen des Beckens aufgeschüttet. Eine offene Verbindung zum Vor-Päijänne hat sich indessen nicht herausgebildet (R. Aario 1965 S. 132). Das Wasser des Moorbeckens ist also die ganze Zeit durch den trennenden Rücken abgeflossen, wie es sich allgemein bei Osgruben verhält. Nach Verlegung der Abflussrinne des Vor-Päijänne ist das Becken bald, vor der Fichtenverhäufung, wieder abgetrocknet und vermoort.

SCHLUSSFOLGERUNGEN

In dem in einem steilhangigen Oskessel entstandenen Moore Harjulansuo sind die Verhältnisse in dessen verschiedenen Teilen mit Rücksicht auf die Torfbildung ungleichmässig, ausserdem haben sie im Verlaufe der Moorentwicklung in ausserordentlich hohem Masse gewechselt. Aus diesem Grunde sind auch die synchronen Schichten in gewissem Masse faciesreicher als gemeinhin. Auf der anderen Seite braucht eine in ihrer Facies verhältnis-

mässig homogene, einheitlich sich fortsetzende Schicht im allgemeinen nicht synchron zu sein. In normalen Mooren ist eine gewisse am Grunde auftretende Torfartschicht in der Nähe der mittleren Moorteile meistens älter und wird gegen die Ränder immer jünger. In diesem Falle dagegen ist der das obere limnische Sediment unterlagernde Torf am Rande des Moores am ältesten, was auf die frühere Befreiung der randlichen Moorteile vom Ancyclussee zurückzuführen ist. Diese Torfschicht ist am Rande des Moores ebenso alt wie die oberen Teile der am Grunde der mittleren Partien abgesetzten Gytjtja. Derartige vom üblichen abweichende Lagerfolgen geben oft Anlass zu unrichtigen Deutungen. Wenn Pollendiagramme für verschiedene Teile eines und desselben Moores zur Verfügung stehen, lässt sich ein Fehler jedoch im allgemeinen vermeiden.

Mit dem hohen Alter der unteren Torfschicht in den Randteilen des Harjulansuo verbindet sich eine Lücke der Lagerfolge an der Grenze zwischen diesem Torf und den oberen Wassersedimenten. Derartigen Lücken, die durch einen Stillstand des Torfzuwachses und eine damit verbundene oder davon unabhängige Abtragung bedingt sind, ist bei Auslegung der Pollendiagramme allgemein Aufmerksamkeit zugewandt worden. Durch vergleichende Pollenanalysen können sie wohl meistens erklärt werden, und in vielen Fällen kann das Vorhandensein einer solchen schon auf Grund der makroskopischen Betrachtung der Lagerfolge als offensichtlich gelten. Am sichersten lässt sich eine Lücke heute durch C^{14} -Bestimmungen feststellen.

Weniger sind die Schwierigkeiten beachtet worden, die durch Umlagerung einer älteren organogenen oder minerogenen Bodenart veranlasst worden sind, eine Umlagerung, bei der das Material einer alten Schicht in eine jüngere eingemengt oder darauf abgelagert wird. In dem untersuchten Fall ist mit ziemlich rein organogener Gytjtja ein an organischen Stoffen armer Ancyloston vermischt worden, aber bei den Bohrungen war das betreffende umgelagerte Tonmaterial doch schwer festzustellen, weil auch in nächster Umgebung eine entsprechende Tonschicht an gleicher Stelle in der Lagerfolge vorzukommen schien. Ja sogar durch Grabung wäre die Sekundärität schwer nachzuweisen gewesen, da das gesamte Tonmaterial die Gytjtja in Körnchen durchsetzt und somit Tongytjtja gebildet hätte, wie es sich teilweise in dem betreffenden Fall verhalten hat. Auch kann es schwer sein, lediglich durch Bohrung die Torfstücke als sekundäre Bildung festzustellen, soweit sie reichlich vorkommen, so dass sie an allen Bohrungsstellen anzutreffen sind, wodurch sie den Anschein erwecken können, dass sie eine zusammenhängende Schicht bildeten. In der Wand einer Grube ist der Sachverhalt leichter zu erkennen. Ist das an der Stelle sich bildende primäre Material gleicher Art wie das ältere sekundäre, so ist die Umlagerung besonders schwer zu erkennen. Offenbar enthält im Harjulansuo die obere Gytjtjaschicht in ziemlich reichlicher Menge auch aus der unteren Schicht herrührenden Gytjtja.

Die alten umgelagerten Bodenarten bringen die in ihnen enthaltenen Subfossilien mit sich. Auch schon bei einem Auftreten in verhältnismässig geringer Menge können die Diatomeen der alten Ablagerung, die unter anderen Bedingungen gelebt haben, bei Auslegung der Entstehungsverhältnisse der Lagerfolge Fehler verursachen. Pollen kann, soweit er auch in der alten Ablagerung reichlich vorkommt, bei Altersbestimmungen sogar sehr grobe Fehler verursachen, zumal wenn die Lagerfolge Lücken aufweist. Ein gutes Beispiel dafür bietet Diagramm 1.

Sekundäre organogene Bestandteile können auch bei Altersbestimmungen nach der C^{14} -Methode zu Fehlern führen. In Schweden ist sogar die Auffassung vorgebracht worden, dass für einige spätglaziale oder frühe postglaziale Sedimente infolge des in ihnen enthaltenen prä- oder interglazialen Materials ein zu hohes Alter erhalten worden sei (Hörnsten 1964). In den in Rede stehenden Zeiten ist die Menge des im Becken selbst gebildeten organischen Materials noch sehr gering gewesen, und das sekundäre hat somit sogar den Hauptteil des sich absetzenden organischen Stoffes ausmachen können. In den jüngeren Ablagerungen kann der relative Anteil des interglazialen Materials jedoch im allgemeinen so gering bleiben, dass er beim Festlegen des Ablagerungszeitpunktes nicht mehr stört.

In normalen Fällen hat der grösste Teil Finnlands einmal der Wirkung der Küstenagenzien unterstanden, so dass in den Grundschichten des Moores der Einfluss des Diatomeen- oder des Pollengehaltes der älteren Ablagerungen zu spüren sein kann. Im allgemeinen ist der dadurch entstandene Fehler jedoch gering gewesen, weil Gegenstand der Umlagerung solche minerogenen Bodenarten gewesen sind, deren Diatomeen- und Pollengehalt im allgemeinen niedrig ist.

Im Transgressionsfall wird das Gebiet dem Einfluss der Küstenagenzien aufs neue ausgesetzt, und soweit der Wasserspiegel nach der Transgression, wie meistens, sinkt, auch noch ein drittes Mal. Ausserdem wendet sich die Abtragung jetzt auch den weichen organogenen Massen zu, die sich leicht erodieren lassen und einen beträchtlichen Fossilgehalt umfassen.

An den Ufern offener Wasserflächen leistet hauptsächlich der Wellengang die Abtragung, in den Hintergründen der Buchten und in kleinen Weihern ist das Eis ein wichtigerer Faktor. In beiden Fällen sind die Stellen mit steilen Ufern der Abtragung stärker ausgesetzt. An den unter dem ziemlich steilen Uferhang gelegenen flacheren Stellen scheint im Harjulsuo die Aufschüttung zu überwiegen.

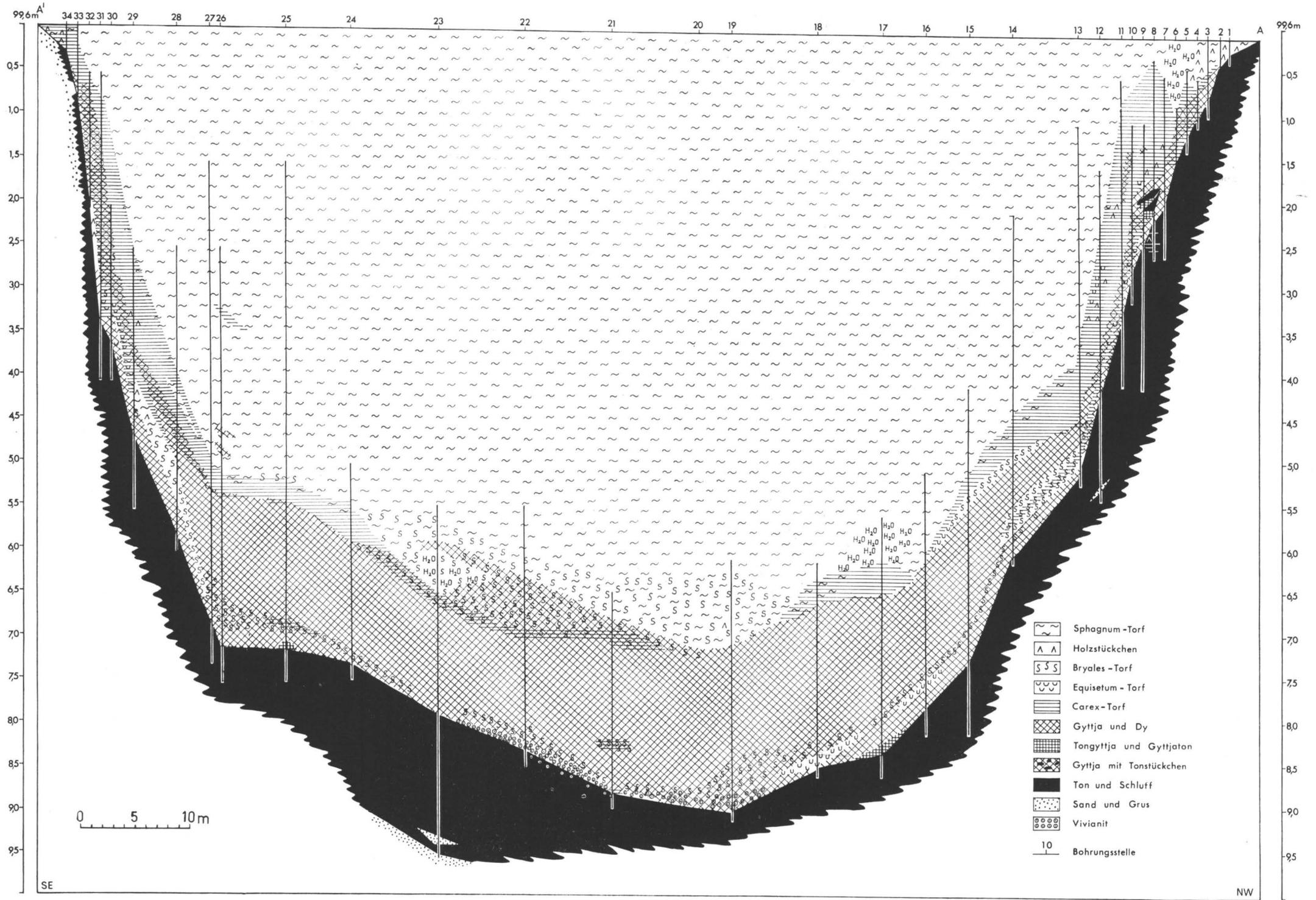
Bei Erforschung der Transgression wären also die für die Datierung zu benutzenden Probenreihen am besten den mittleren Teilen des Moores, wo sein Grund im allgemeinen flach ist, oder solchen Stellen seiner Randteile, wo sein Grund flach ansteigt, zu entnehmen, so dass sowohl die Abtragung als auch entsprechend das Ablagern alten Materials geringer sind.

Bei genauen Altersbestimmungen, wie bei C^{14} -Datierungen, ist einerseits bei Lücken in der Lagerfolge, andererseits bei umgelagertem Material ganz besondere Vorsicht geboten. Da Lücken sich ganz vorwiegend an Grenzstellen zwischen limnischen und terrestrischen Ablagerungen einzustellen pflegen, wären für jeden Transgressionsfall wenigstens vier Proben zu untersuchen, eine von der oberen Grenze des unteren Torfes, die zwei folgenden von der oberen und der unteren Grenze der auf die Transgression hinweisenden limnischen Bildung und die vierte von der unteren Grenze des oberen Torfes. Da das Vorhandensein sekundär abgelagerten Materials schwer festzustellen ist, wären die Probenreihen am besten an zwei verschiedenen Stellen zu erhohren. Die Kostspieligkeit der C^{14} -Bestimmungen bereitet jedoch vorläufig der Beschaffung hinreichender Gewissheit Schwierigkeiten. Zum mindesten die eine Probenreihe sollte herrühren von einer Stelle, wo auf Grund des oben Dargestellten das Vorliegen sekundärer Ablagerungen weniger wahrscheinlich ist. Die Forderung, die Proben für C^{14} -Bestimmungen durch Ausgraben zu entnehmen, führt bedauerlicherweise leicht dazu, die Proben an niedrigstmöglicher Stelle auszuwählen, die oft gerade zu einer Abtragungs- und Ablagerungszone gehört.

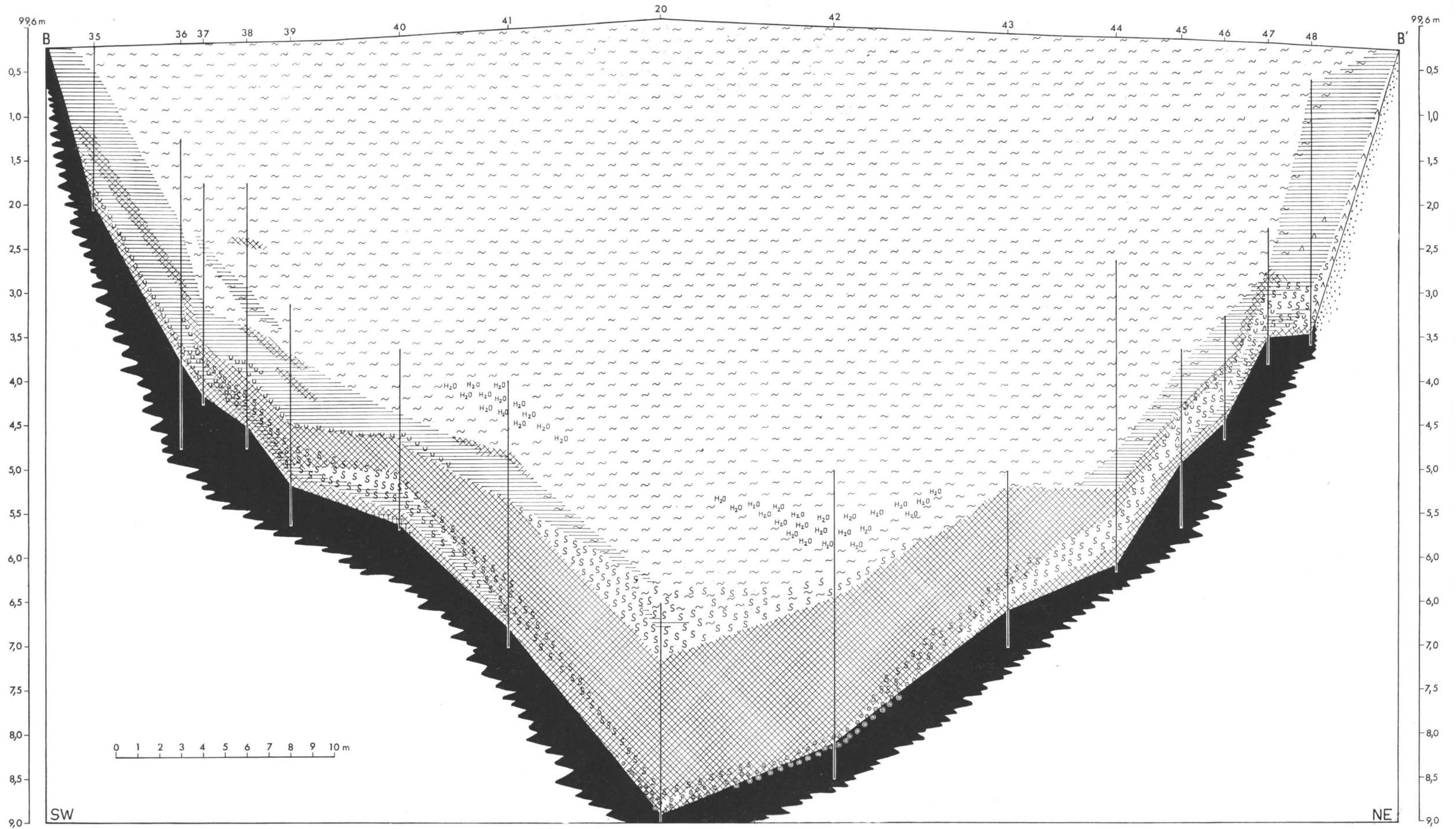
Die sichersten C^{14} -Bestimmungen lassen sich wohl für Torf gewinnen, soweit er mit Gewissheit als autochthon hat festgestellt werden können. Dagegen ist z. B. Akkumulationstorf (R. Aario 1965, S. 178) für diesen Zweck ungeeignet, da in ihm jüngere und ältere Bestandteile miteinander vermengt vorkommen. Bei limnischen Sedimenten ist die Möglichkeit zum Vorhandensein sekundär abgesetzten Materials grösser als in autochthonem Torf, und in mächtigen Gyttyaschichten vermengen bekanntlich auch Gasauftriebe in gewissem Masse die Schichten.

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Risto Aario: Entwicklung des Moores Harjulansuo



Risto Aario: Entwicklung des Moores Harjulansuo

A RESTUDY OF »KARAMSINITE»¹⁾

BY

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ABSTRACT

Two specimens of the doubtful KCaMg-silicate »karamsinite» have been investigated. One proved to be tremolite asbestos mixed with malachite, the other palygorskite.

INTRODUCTION

The mineral known as »karamsinite» appeared in the literature in 1860 in C. F. Rammelsberg's *Handbuch d. Mineralchemie* (p. 776), on the strength of a report orally communicated by A. E. Nordenskiöld. No description to complement the chemical analysis made by Thoreld is given. The mineral was supposed to have been found in Finland. Since that time karamsinite has never been described. All mineralogical handbooks refer to the few lines in Rammelsberg's book only. In, *e.g.* Dana's 6th ed, p. 1039, the chemical analysis is given. In Hey's *Chemical Index of Minerals*, 1955, p. 91, karamsinite is listed as a doubtful species with the formula $\text{KMgCaSi}_4\text{O}_{12}\text{OH}$, plus small amounts (5 %) of CuO , MnO , Fe_2O_3 , and Al_2O_3 . Strunz (1957), p. 384, also refers to karamsinite as CaKMg-silicate and an incompletely defined mineral.

The present authors obtained two specimens listed as »karamsinite» from the mineralogical collection of the University of Helsinki. According to the labels, both specimens are from Nizhne-Tagilsk, in the Urals. One of them, in addition, is more exactly traced to the Karamsin quarry and bears the note »Nordenskiöldska samlingen» (Nordenskiöld collection) on the label.

¹⁾ Received December 31, 1964.

Table 1. X-ray powder data ¹⁾ for the karamsinite specimens from Nizhne-Tagilsk, Ural Mts, No. 1 is tremolite No. 2 palygorskite.

No. 1. Tremolite		No. 2. Palygorskite	
<i>d</i> (Å)	<i>I</i>	<i>d</i> (Å)	<i>I</i>
9.0	w	10.3	vvs
8.4	vs	6.30	s
5.1	w	5.35	m
4.8	w	4.43	s
4.51	m	4.10	m
4.20	m	3.63	w
3.26	s	3.42	vw
3.13	vs	3.20	m
2.93	m	3.16	s
2.80	m	2.97	vw B
2.70	m	2.72	vw
2.59	w	2.61	m
2.38	w	2.57	m B
2.32	w	2.50	m
2.16	w	2.12	w
2.01	w	2.11	w
1.89	m	1.66	vw
1.81	w		
1.64	w		

vvs = very, very strong, vs = very strong, s = strong, m = medium, w = weak, vw = very weak, B = broad.

1. Karamsinite, Karamsin quarry, Nizhne-Tagilsk, Ural Mts. Nordenskiöld collection.

2. Karamsinite, Nizhne-Tagilsk. A. Gadolin's collection.

¹⁾ Uncalibrated Norelco diffractometer charts. Cu-radiation, Ni-filter.

The other specimen originates from A. Gadolin's collection. Since A. E. Nordenskiöld spent the period 1853—1854 in the Tagilsk area collecting mineral specimens, it is highly probable that the material from the Karamsin quarry (belonging to the Demidoff mine) in the University collection, which material came from the Nordenskiöld collection, represents that investigated by Nordenskiöld. Considering Nordenskiöld's contribution to mineralogy, one would expect him to have described the mineral himself if he had regarded it as a new mineral species.

SAMPLE DESCRIPTION

The sample from Nordenskiöld collection ¹⁾ consists of large garnet crystals, measuring up to 3 cm in diameter. The garnet ($a_0 = 11.98 \text{ \AA}$,

¹⁾ No. 3396 from the collection of the Department of Geology and Mineralogy of the University of Helsinki, labelled as Karamsinit, Karamsins gr. (= gruva = quarry), Nischnij Tagil, Ural, Nordenskiöldska samlingen (= Nordenskiöld collection). Another label for the same specimen: Grossularia med (= with) Karamsinit, Ural, Nischnij Tagil, Karamsin's gr.

Table 2. Chemical analyses of Karamsinite and Tremolite

	1.	2.
SiO ₂	51.53	51.3
Al ₂ O ₃	3.20	1.5
Fe ₂ O ₃	5.98	3.2 (total)
MnO	4.62	3.2
MgO	6.86	21.9
CaO	13.05	11.5
Na ₂ O	—	0.1 ¹⁾
K ₂ O	10.08	0.3 ¹⁾
CuO	2.32	—
H ₂ O	1.59	—

1. Karamsinite. Rammelsberg, Handbuch d. Mineralchemie, p. 776. Anal. Thoreld.
2. Tremolite, Karamsin quarry, Nizhne-Tagilsk, Ural Mts. Sample from Nordenskiöld collection. Anal. Pentti Ojanperä.

1) Semiquantitative determination from 6.7 mg of material.

$n_{Na} = 1.832$) is in many cases covered by small crystals (0.2—2 mm), in places rosettes, of white tremolite asbestos. In addition to these two minerals, certain other minerals are also present in the specimen, but in minor amounts. The tremolite tends to be covered by a film of dark brown material, which is evidently a mixture of chlorite and iron hydroxides¹⁾. In places, the tremolite is greenish in colour, owing to a light green powdery material, which also fills some of the small cavities in the specimen. The green powdery material is amorphous to X-rays. Spectrochemical analysis revealed the material to be quite rich in Cu. In places, the greenish material grades over to crystalline malachite (as determined by X-ray powder methods). Many of the walls of the small cavities (0.1—1 cm) are covered by small tremolite crystals, sometimes mixed with iron hydroxides. In addition to the minerals mentioned, the microscopical investigation revealed abundant chlorite and small amounts of diopside ($2V\gamma = 58^\circ$, $c\wedge\gamma = 40^\circ$). The mineral identified as tremolite ($n_a = 1.613$, $n_y = 1.634$, $2V\alpha = 76^\circ$, $c\wedge\gamma = 19^\circ$; X-ray powder data in Table 1) is evidently the same mineral as Nordenskiöld's karamsinite. The original chemical analysis of karamsinite (No. 1 in Table 2) is presented with a new partial semimicrochemical analysis of tremolite (No. 2, Table 2). A small amount of tremolite was hand-picked under a binocular and chemically analyzed for the most important oxides. During the separation of the tremolite, it became clear that it was practically impossible to obtain completely pure material. The tremolite was slightly contaminated by the Cu-rich amorphous material (malachite). This is evidently the reason for the high CuO content of karamsinite. The SiO₂

¹⁾ Only X-ray powder photographs and qualitative spectrochemical analyses were made with other minerals than tremolite and garnet.

contents agree very well. No new determination for water was made because the H₂O content, 1.59 per cent, in karamsinite agrees very well with that in the minerals of the amphibole group. The MnO content in karamsinite is very high for a tremolite. The new chemical analysis gives a value of 3.2 per cent MnO, which is also unusually high. The only actual discrepancy between the two analyses is the low MgO and high K₂O content in karamsinite and vice versa in tremolite.

The chemical analysis of karamsinite fits the structure formula of tremolite well provided that two Mg ions be replaced by two K ions. Taking everything into consideration, it seems clear that

- in the original analysis, Mg was incompletely precipitated, the remainder being included in K,
- the CuO content is due to the contamination of amorphous malachite,
- karamsinite is actually tremolite.

The specimen from A. Gadolin's collection ¹⁾ is 2 cm in diameter, grayish white in colour. The density is very low. Under the microscope the mineral looks like cardboard, however, resembling tremolite-asbestos of sample No. 3396 in appearance. Spectrochemical analysis revealed Mg, Al, and Si as the major components, with a little Fe and Ca and 0.1 per cent Cu. The X-ray powder data (Table 1, No. 2) show the mineral to be palygorskite. DTA analysis of the mineral indicated endothermic breaks at 160°, 300°, 520° C. and exothermic ones at 350° and 980° C., which all fit those reported for palygorskites.

In chemical composition palygorskites differ so much from karamsinite that the mineral investigated by Nordenskiöld certainly could not be palygorskite.

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¹⁾ According to the label the sample is: Karamsinite aus Nischnetagilsk, A. Gadolin's Collection. No sample number is given.

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PALEOMAGNETISM OF THE DIKE SYSTEMS IN FINLAND ¹⁾

I. REMANENT MAGNETIZATION OF JOTNIAN OLIVINE
DOLERITES IN SOUTHWESTERN FINLAND

BY

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ABSTRACT

The direction and strength of the remanent magnetization of Jotnian dolerites is determined. The sampling technique, instruments used and results obtained are reviewed and discussed.

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INTRODUCTION

The magnetization of rocks consists of two main components. The magnetic anomalies in the earth's magnetic field are usually assumed to be produced by the induced magnetization of rocks. The remanent component

¹⁾ Received January 25, 1965

of magnetization can be quite strong, however, and even stronger than the induced magnetization (Nagata, 1961). While the induced magnetization is usually measured directly on the field, the remanent magnetization is best determined from oriented rock specimens in the laboratory. Its strength and direction can give valuable information concerning the genesis and history of the rocks. The knowledge concerning the history of the earth as whole has been greatly augmented by paleomagnetic investigations carried out in different parts of world.

In igneous rocks, the primary remanent magnetism was achieved when the rock material first cooled in the earth's magnetic field to and below the Curie temperature. Thus the local direction of the earth's field was recorded on the rock as thermoremanent magnetism, TRM (Neel, 1955). In some cases the TRM seems to be reversed with respect to the earth's magnetic field.

In sedimentary rocks, the magnetic field of the earth is fixed as a statistic orientation of the magnetic mineral grains during the sedimentation (depositional remanent magnetization) or more commonly as chemical remanent magnetization, during the diagenesis of the sediment. It is possible to reconstruct the orientation of the earth's magnetic field at the time of the genesis of the rock by measuring the direction of the remanent magnetization of the igneous or sedimentary rock samples. In many cases, however, the primary magnetic direction will be changed or masked by later magnetization during the later history of the rock. There are many different ways in which rocks are remagnetized (see Cox and Doell, 1960). As pointed out by Graham and others (1957), one of the most important ways in which the remagnetization of igneous rocks takes place is magnetostriction — or the magnetizing effect of stress. Quite commonly late chemical changes, like weathering, introduce secondary remanent magnetization in rocks (Nagata, 1961).

Bulk remanent magnetization, which is measurable with a sensitive magnetometer, may thus be composed of many components of different origin. Usually, we are interested only in the most primary magnetization of a rock. Fortunately, later magnetization can in some cases be removed by »cleaning» rock specimens by means of partial demagnetization. This is possible when the primary magnetization is more stable than the later components. This demagnetization is done in an alternating magnetic field of sufficient strength (Brynjolfsson, 1957; As and Zijdeveld, 1958; and Creer, 1959), or by heating the specimens in a magnetically zero field to an elevated temperature (Irving and others, 1961; and Chamalaun and Creer 1964).

A large number of published works on the subject of paleomagnetism, reviewed recently by Cox and Doell (1960), show that the direction of magnetization in rocks is not parallel to the earth's present field. Consequently, a wandering of the magnetic pole was postulated. The paleomagnetic data

available at present indicate a clear wandering of the pole during the greatest part of the earth's history. As pointed out by Runcorn (1956), the wandering of the magnetic pole and the variation of the geomagnetic field in time make possible a new method of dating rocks. Classification, grouping and correlation of rocks can, consequently, also be done by means of remanent magnetization, as shown for example by Manwaring (1963) and by McElhinny and Opdyke (1964). Strangway (1964) used remanent magnetization measurements for dike classification.

The age of the main part of the metamorphic rocks in Finland is, according to Kouvo (1958), about 1800 mega years. An older gneiss complex in northern and eastern Finland has an age of 2300 mega years or more (Wetherill and others, 1962). Younger than these are rapakivi granite (1640 m.y.), Jotnian sandstone and dolerite dikes and sills. Dolerites are of several different types and of different ages. The oldest of these penetrate only the old gneiss granite basement in the northern and eastern part of the country. The olivine dolerites in the Satakunta region are the youngest and penetrate the Jotnian sandstones. According to Savolahti (1964), the age of two dolerite samples was determined by the K^{40} — A^{40} method in Freiburg. The samples originate from Sorkka, near the town of Rauma and from the small island of Säppi. They gave a maximum age of about 970 m.y. These determinations are much lower than the figure of 1330 m.y. obtained in Leningrad for a olivine dolerite sample from Satakunta (A. Kahma, personal communication). It is to be noted that Sorkka, according to Laitakari (1925), and Säppi according to Inkinen (1963), are the only localities where dolerite is observed to occur horizontally.

In addition to the Jotnian dolerites, there occur in the Finnish Precambrian several dolerite dike systems, the relative age of which is not known. The scope of the present work is to find whether the directions of the remanent magnetization of the Jotnian olivine dolerites form a uniform system and to test whether this magnetic orientation differs from the directions measured for other rocks of the area. If this proves to be the case, new trials will be made to classify the different dolerite and other anorogenic dike rocks by means of paleomagnetic measurements. Puranen (1960) has made measurements of the remanent magnetization of some Finnish rocks. It is not known, however, if the remanent component of the different dike rocks is strong and stable enough to allow measurements of the primary TRM.

Radioactive dating of Precambrian rocks can usually be made to an accuracy of about 10 %. Consequently, Finnish rocks with an age difference of less than 150 m.y. cannot easily be distinguished by this method. If the average rate of movement of the geomagnetic pole was of the same order of magnitude (about 5° per 10 m.y.) as during the Permian (Creer and others, 1957), the difference in magnetic orientation could possibly be detected.

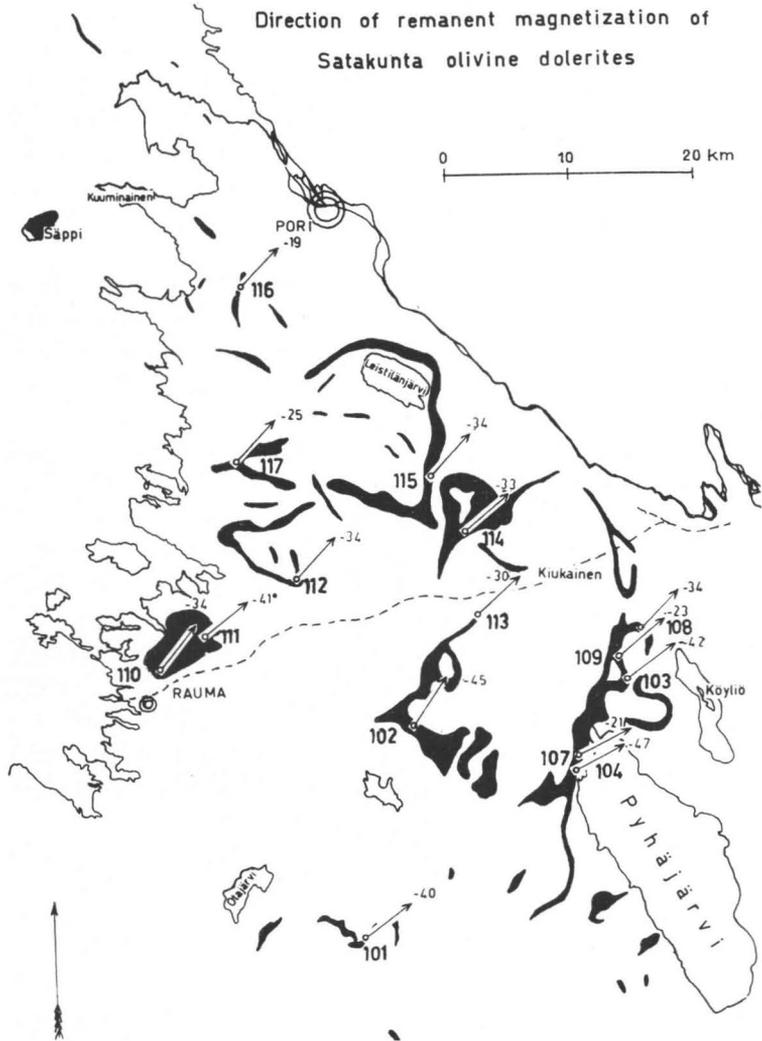


Fig. 1. Regional distribution and direction of the remanent magnetization of the Satakunta olivine dolerites. Dikes (black) drawn according to Laitakari (1925), Kahma (1951), and aeromagnetic maps.

THE SAMPLES

The mode of occurrence and the areal distribution of the Satakunta olivine dolerites (diabases) is described in detail by Laitakari (1925), Kahma (1951) and Eskola (1963). The rock occurs in the rapakivi and sandstone area in Satakunta, S. W. Finland (see Fig.1) as broad vertical dikes. In the western

Table 1. Mineral composition (mode) of some of the dolerite samples measured

Sample No.	101.1	102.1	110.1	112.1	115.1
Plagioclase (An ~55)	55.0	52.3	58.5	56.5	49.9
Olivine (Fa 40)	8.6	15.7	18.7	12.6	14.1
Augite	16.2	12.0	6.3	8.3	26.2
Magnetite	3.9	6.7	3.5	4.2	4.6
Biotite, chlorite and serpentine	11.2	9.6	6.3	8.6	2.4
Accessories	5.1	3.7	6.7	9.8	2.8
	100.0	100.0	100.0	100.0	100.0

part of the area, dolerite occurs also horizontally in the Precambrian migmatites (Inkinen, 1963). Oriented rock samples were taken of all the different types of dikes and sills. Since some of the dikes are connected with a negative and some with a positive anomaly on the aeromagnetic map, both types of dikes were sampled.

The oriented samples were collected as 32 mm cores of a portable diamond drill made for this purpose by the Finnish diamond tool manufacturer L. A. Levanto Oy, Helsinki. All the samples were drilled vertically and north, south and east directions were marked on the side of the core before its extraction as clear scratches with a hard metal tool. The geographic north was found with a simplified solar measurement, except when stubborn clouds made it impossible and a declination corrected magnetic orientation had to be used. The north direction was transferred on the sample with the aid of a conventional cannon sight (azimuth clinometer). The mechanical accuracy of this orientation technique is estimated to be about 1°.

The samples were numbered according to the collection sites, starting with 101. In case several samples were collected from the same locality, they were distinguished by the first decimal (101.1 and 101.2). For measurements, the samples were cut to specimens 32 mm long and the upper surface was polished for microscopic examination. The specimens cut were numbered with the second decimal (101.11, 101.12, 101.13, etc.) starting from the topmost.

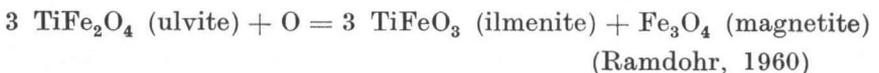
The grain size of the sampled dolerite is mainly about 1-4 mm. At the chilled margins and in narrow dikes, the rock is much finer and sometimes aphanitic. Table 1 gives the mineral composition of some samples used for the magnetic measurements. For the chemical composition of the Jotnian dolerite see Eskola (1963). High temperature plagioclase, An₍₄₀₋₆₅₎, olivine (Fa₄₀) and augitic pyroxene are the main constituents of the rock. Plagioclase in many cases forms the largest crystals in the rock. The water-bearing minerals biotite, chlorite, and serpentine are to a great extent alteration products of

Table 2. The amount of corroded magnetite grains in some of the specimens studied

Specimen No.	Decl.	Incl.	<u>corroded magnetite grains</u> total of magnetite grains
101.11 topmost	41°	-22°	.32
101.12	36°	-20°	.25
101.13	50°	-16°	.20
101.21 topmost	35°	+ 4°	.51
101.22	43°	- 9°	.14
101.23	41°	-11°	.16
102.11 topmost	34°	+ 6°	.11
102.12	26°	- 9°	.08
102.13	25°	-14°	.06
103.11 topmost	51°	-22°	.05
103.12	50°	-21°	.15
103.13	51°	-24°	.14
103.21 topmost	42°	+ 9°	.23
103.22	54°	-17°	.22
103.23	51°	-22°	.21

olivine and augite. Epidote and sericite occur in small quantities as alterations of plagioclase. As other accessories there are apatite and sulphide and oxide minerals. Magnetite is the principal magnetic mineral in the rock. It contains frequently observed but not very numerous lamellae of ilmenite, which also occurs as rare independent grains.

The magnetite shows a submicroscopic exsolution of ulvite in the same manner as in the post-Jotnian dolerite in the type locality of ulvöspinel (Mogensen, 1946). In all the specimens examined in polished sections, some of the magnetite grains are corroded. Table 2 shows that the number of pitted grains is greatest in the topmost specimens. Ilmenite seems to be most abundant in the corroded parts of the magnetite. It forms a beautiful lamellar intergrowth texture with a magnetite that has partially been dissolved out of the rock (Fig. 2). This alteration occurred evidently in connection with the oxidative weathering in the surface layer of the rock:



The ilmenite so formed has grown on the previously formed exsolution lamellae of ilmenite, thus causing the increase of the ilmenite in the corroded parts of the magnetite. The oxidation of ulvöspinel can possibly pass to maghemite:



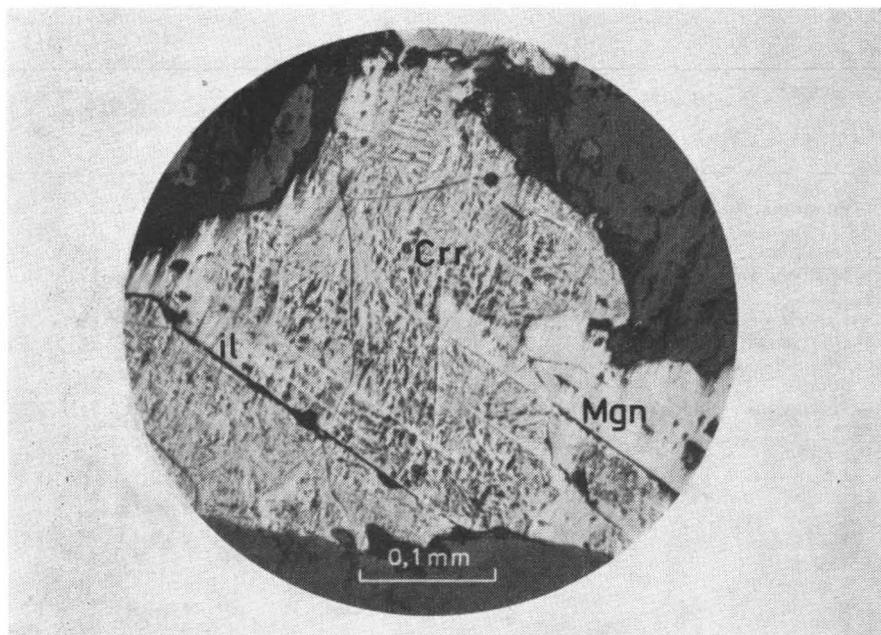


Fig. 2. Correded magnetite grain in olivine dolerite from Suontaka, Laitila. Mgn = unaltered magnetite, il = ilmenite, Crr = correded magnetite, Reflected light.

THE TECHNIQUE OF MAGNETIC MEASUREMENTS

The direction of magnetization of the specimens was evaluated with an astatic magnetometer having two Alcomax IV magnets, 15 cm apart, built according to advice given by Blackett (1952). To facilitate the measurements, the instrument was made highly astatic but not very sensitive, with 18-micron tungsten wire being used for suspending. The period of the instrument was 10 sec. For the measurements the rock specimen was inserted in the centre of an aluminium cube $35 \times 35 \times 35$ mm, which was placed under the magnetometer, and the strength of the magnetic vector was determined in 12 positions of the cube to eliminate the effect of the induced magnetization and that of the magnetic anisotropy in the specimen. The azimuth angle D from north to east, the inclination I (positive when down and negative when up) and the strength of the remanent magnetization were calculated from these measurements. The instrument was calibrated with a coil of the size and shape of the rock specimens used and inserted in the place of the specimen.

The direction of the magnetization of the whole sample was computed from the individual specimen measurements by applying Fisher's (1953) statistical method for compiling vector data. This same method was used in computing the mean values in Tables 3, 4, 5 and 6.

Table 3. Directions of remanent magnetization of Satakunta dolerites measured before demagnetization procedures

Sample No.	Locality			Magnetic direction					Intensity c.g.s./cm ³	Virtual ¹⁾ Pole position	
	Name	Lat.	Long.	Decl.	Incl.	α_{95}	k	N		Lat.	Long.
101.1	Suontaka, Laitila	60.9°N	21.8°E	43°	-19°	6°	167	3	15·10 ⁻⁴	12°N	158°E
101.2	»	»	»	40°	-6°	10°	65	3	13·10 ⁻⁴	19°N	159°E
102.1	Rahilainen, Lappi	61.1°N	»	29°	-6°	11°	59	3	14·10 ⁻⁴	22°N	171°E
103.1	Kyöpeli, Eura . .	»	22.2°E	51°	-23°	8°	100	3	15·10 ⁻⁴	7°N	153°E
103.2	»	»	»	48°	-10°	18°	18	3	15·10 ⁻⁴	15°N	152°E
104.3	Kiperinoja, Eura	»	22.1°E	51°	-35°	—	—	2	22·10 ⁻⁴	0°N	155°E
107.1	Kauttua, Eura . .	»	»	60°	+26°	5°	161	6	19·10 ⁻⁴	26°N	132°E
108.1	Ilmijärvi, Köyliö	61.2°N	22.2°E	42°	-21°	5°	163	6	20·10 ⁻⁴	11°N	160°E
109.1	Sorkkinen, Eura	»	22.1°E	27°	+1°	30°	3	7	24·10 ⁻⁴	26°N	172°E
110.1	Sorkka, Rauma	»	21.5°E	37°	-17°	4°	138	9	10·10 ⁻⁴	14°N	163°E
110.2	»	»	»	38°	-16°	5°	103	8	11·10 ⁻⁴	14°N	162°E
111.1	Kaaronjärvi, Rauma	»	21.6°E	49°	-21°	6°	200	5	12·10 ⁻⁴	8°N	153°E
112.1	Lavila, Eurajoki	»	21.7°E	42°	-26°	3°	224	8	14·10 ⁻⁴	8°N	161°E
113.1	Panelia, Kiukainen	»	22.0°E	42°	-15°	3°	282	8	20·10 ⁻⁴	14°N	159°E
114.1	Laihia, Kiukainen	»	22.1°E	47°	-23°	6°	131	5	21·10 ⁻⁴	8°N	156°E
115.1	Leistilä, Nakkila	61.5°N	22.0°E	40°	-22°	4°	179	8	15·10 ⁻⁴	11°N	152°E
116.1	Viasvesi, Pori . . .	61.4°N	21.7°E	35°	-2°	9°	30	8	15·10 ⁻⁴	22°N	164°E
117.1	Peränkylä, Luvia	»	»	42°	-19°	4°	150	8	11·10 ⁻⁴	12°N	162°E
Mean of all samples		61.2°N	22.0°E	42°	-14°	62°	28	18*	$\delta m = 6.2^\circ$ $\delta p = 3.1^\circ$	14°N	158°E

α_{95} = circle of confidence with a probability of 95 per cent

k = precision parameter (N-1/N-R)

N = number of specimens measured (* of samples compiled)

¹⁾ Negative pole according to Runcorn's nation

DIRECT MEASUREMENTS

The samples collected were stored in the laboratory in an upright position and in the original north-south orientation. This was done to keep the remagnetization after sampling as low as possible. The specimens were measured first without any cleaning treatment as soon as possible. Table 3 lists the collection sites, measured directions of remanent magnetization, the intensity data and the calculated virtual pole positions of 18 samples. All together 103 specimens cut from the Jotnian olivine dolerite dikes were measured.

The declination does not vary greatly from one sample to another. Stranway (1964) has observed that the magnetization of the dolerite dikes follows the trend of the dikes. This cannot be seen in the present area, as shown by Fig. 1. No difference in magnetic orientation can be observed between the different types of intrusives, either.

The angle of inclination is negative for most of the specimens measured. The angle is often smallest or positive for the topmost specimens. This is seen in Table 2, which lists the magnetic directions and the ratio of the corroded magnetite grains to the total number of magnetite in some specimens. The positive correlation between the amount of corrosion and the inclination is clear.

It seems evident that primary magnetization in the rock is masked with a secondary component created through the chemical weathering of the dolerite near the surface of the rock. Efforts were made to avoid the secondary component of magnetization by drilling deeper sample cores, but it was soon observed that the samples never became quite »clean» at greater depths. Therefore, some other method of cleaning should be found. The secondary magnetization is evidently connected with the oxidation of magnetite and ulvöspinel. Among possible oxidation products, only ilmenite and, obviously, maghemite and goethite are to be observed. Maghemite has an inverse spinel structure similar to that of magnetite and is ferrimagnetic. Ilmenite is also ferrimagnetic if it contains a small excess of Fe. Both minerals, however, have low coercive forces and should consequently be easily demagnetized in an alternating magnetic field treatment. Maghemite, on the other hand, is metastable and inverts to hematite at a relatively low temperature (Nagata, 1961).

DEMAGNETIZATION WITH ALTERNATING CURRENT

To remove the secondary magnetization of the specimens, an a.c. method was used according to the principles described by Creer (1959). The specimens were demagnetized in the centre of a Helmholtz coil, inside a 10-cm long coil with an internal diameter of 9 cm, in series with a 110- μ F condenser system and capable of a 1000-oersteds peak field at 10 amperes current. The specimen was rotated simultaneously about two perpendicular axes in the manner described by Creer (1959). The rotation ratio was about 1:1.1 and the slower speed 78 revolutions a minute.

It was observed that the specimens easily became randomly magnetized during the demagnetization process in the manner reported by As and Zijdeveld (1958) because of an uneven decrease in the demagnetizing current. An electrolytic resistance of copper sulphate solution was used, but difficulties appeared in reducing the current smoothly enough near the zero end of the current. The most suitable arrangement was found to be a slow substitution of the copper sulphate solution in the resistance cell by tap water, which in turn was replaced by distilled water. Samples 101.1—104.3 were demagnetized stepwise in the experiments in alternating magnetic fields up to 700 oersteds to test the stability of the magnetization. A stable magneti-

Table 4. Directions of remanent magnetization of Satakunta olivine dolerites measured after demagnetization in alternating magnetic field

Sample No.	Magnetic direction					Intensity
	Decl.	Incl.	α_{95}	k	N	c.g.s./cm ³
101.1	54°	-40°	4°	417	3	4·10 ⁻⁴
101.2	49°	-27°	6°	156	3	3 »
102.1	34°	-45°	19°	18	3	7 »
103.1	50°	-38°	2°	1 818	3	6 »
103.2	57°	-46°	9°	85	3	5 »
104.3	58°	-47°	—	—	2	8 »
107.1	61°	-14°	9°	76	3	4 »
108.1	42°	-37°	4°	488	3	15 »
109.1	49°	-21°	12°	36	4	5 »
110.1	35°	-34°	8°	56	5	5 »
110.2	36°	-34°	6°	141	4	4 »
111.1	51°	-41°	10°	65	3	5 »
112.1	43°	-35°	7°	102	4	6 »
113.1	43°	-32°	4°	248	4	8 »
114.1	50°	-36°	2°	1 540	3	5 »
115.1	40°	-39°	3°	435	4	5 »
116.1	36°	-24°	9°	55	4	7 »
117.1	41°	-29°	7°	98	4	7 »
Mean of all samples	46°	-37°	4.7°	50	18	

Virtual pole position: latitude 0°N, longitude 160°E, $\delta m = 5.4^\circ$, $\delta p = 3.1^\circ$

zation was found in the specimens after treatment in a field of about 200 oersteds (peak). The rest of the samples were demagnetized in the field of 200 oersteds. The directions measured after a.c. treatment are listed in Table 4.

DEMAGNETIZATION BY THERMAL TREATMENT

To check the results obtained by a.c. demagnetization independently, heat treatment was applied in the case of every second specimen of the series 107—117. Thermal treatment for magnetic cleaning of the unstable component had been used earlier by Cox (1957), Irving and others (1961) and Chamalaun and Creer (1964). The procedure has proved successful in the case of sedimentary rocks, but igneous specimens can be cleaned by heating as well. It was found by trial that the unstable magnetization could be washed out from Jotnian dolerite specimens by keeping them at 300—400°C for a period of few hours. Different arrangements was tried but heating the specimens while immersed in heavy cylinder oil (Cylesso 300) for at least 3 hours at 300°C was the most successful. The heating was done with a special gas burner in a zero field inside an Helmholtz coil system.

Table 5 lists the directions of magnetization measured in the samples after heating. Some of the specimens were unfortunately destroyed by thermal experiments conducted at too high a temperature. It was found

Table 5. Directions of remanent magnetization of Satakunta olivine dolerites measured after thermal demagnetization at 300°C

Sample No.	Magnetic direction					Intensity
	Decl.	Incl.	a_{95}	k	N	c.g.s./cm _s
107.1	56°	—28°	11°	53	3	7·10 ⁻⁴
108.1	40°	—32°	5°	267	3	15 »
109.1	50°	—21°	5°	303	3	13 »
110.1	42°	—32°	19°	14	4	5 »
110.2	36°	—30°	10°	66	3	6 »
111.1	destroyed in thermal experiments					
112.1	42°	—32°	4°	323	4	9 »
113.1	43°	—28°	5°	242	4	13 »
114.1	47°	—29°	—	—	2	14 »
115.1	42°	—29°	2°	789	4	10 »
116.1	47°	—14°	7°	100	4	6 »
117.1	42°	—21°	9°	66	4	7 »
Mean of all samples	44°	—27°	3.9°	119	11	

Virtual pole position: latitude 7°N, longitude 159°E, $\delta m = 4.2^\circ$, $\delta p = 2.2^\circ$

during the heating experiments that the Curie point of the specimens is at 550—560°C, indicating about 0.05 mole fraction of ulvöspinel in solid solution in magnetite. This is in close agreement with the immiscibility cap suggested for the system by Basta (Nagata, 1960).

The directions revealed by the thermal treatment are highly similar to those found after a.c. demagnetization. It is evident that both methods clean the specimens for the same type of soft secondary magnetization. It is interesting to observe, however, that the intensity after a.c. demagnetization in a 200-oersteds peak field is less than the intensity found after thermal treatment. This means that the stable component of magnetization, which is not very hard, as a whole, is more easily destroyed by an alternative magnetic field of 200 oersteds than by heating at 300°C. The thermal method of demagnetization would, consequently, be safer than the a.c. method. The thermal treatment consumes more time, though, since a heating and cooling cycle takes at least about 5 hours, whereas an a.c. demagnetization is performed in fifteen minutes. On the other hand, 6—8 specimens can be heated simultaneously.

DISCUSSION OF THE RESULTS

The directions of magnetization measured after a.c. and thermal demagnetization and the calculated positions of the virtual magnetic poles are quite close to each other. The corresponding ovals of confidence overlap to a great extent. They might, consequently, be considered to represent directions belonging to the same category of observations, and be compiled as in Table 6.

Table 6. Directions of remanent magnetization of Satakunta olivine dolerites measured after removal of the secondary magnetization by a.c. or by thermal treatment

Sample No.	Locality			Magnetic direction					Virtual pole position	
	Name	Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.
101.1	Suontaka, Laitila	60.9°N	21.8°E	54°	-40°	4°	417	3	4°S	153°E
101.2	» »	»	»	49°	-27°	6°	156	3	6°N	155°E
102.1	Rahilainen, Lappi	61.1°N	21.8°E	34°	-45°	19°	18	3	2°S	171°E
103.1	Kyöveli, Eura	»	22.2°E	50°	-38°	2°	1 818	3	1°S	157°E
103.2	» »	»	»	57°	-46°	9°	85	3	12°S	153°E
104.3	Kiperinoja, Eura	»	22.1°E	58°	-47°	—	—	2	11°S	153°E
107.1	Kauttua, Eura	»	»	59°	-21°	10°	31	6	5°N	144°E
108.1	Ilmajärvi, Köyliö	61.2°N	22.2°E	41°	-34°	4°	260	6	4°N	164°E
109.1	Sorkkinen, Eura	»	22.1°E	49°	-23°	5°	128	7	3°N	147°E
110.1	Sorkka, Rauma	»	21.5°E	39°	-36°	9°	27	9	5°N	165°E
110.2	» »	»	»	36°	-32°	5°	100	8	7°N	167°E
111.1	Kaaronjärvi, Rauma	»	21.6°E	51°	-41°	10°	65	3	4°S	158°E
112.1	Lavila, Eurajoki	»	21.7°E	43°	-34°	4°	165	8	4°N	162°E
113.1	Panelia, Kiukainen	»	22.0°E	43°	-30°	3°	254	8	6°N	161°E
114.1	Laihia, Kiukainen	»	22.1°E	49°	-33°	4°	272	5	2°N	156°E
115.1	Leistilä, Nakkila	61.5°N	22.0°E	41°	-34°	4°	182	8	4°N	163°E
116.1	Viasvesi, Pori	61.4°N	21.7°E	41°	-19°	7°	46	8	12°N	153°E
117.1	Peränkylä, Luvia	»	»	41°	-25°	6°	71	8	9°N	161°E
Mean of all samples		61.2°N	22.0°E	46°	-34°	43°	60	18	2°N	158°E

$$\delta m = 4.8^\circ \quad \delta p = 2.8^\circ$$

All the samples measured reveal a highly similar magnetic orientation. The horizontal sill at Sorkka (samples 110.1, 110.2, and 111.1) does not differ from the vertical dikes. The great difference in age (Savolahti, 1964), mentioned earlier, is therefore quite strange. The age of 1 330 m.y. fits the feological evidence better.

Fig. 3. shows the virtual pole positions and the ovals of confidence around the poles before and after cleaning off the secondary magnetization. The partial demagnetization did not greatly change the position of the virtual pole but the size of the oval of confidence was reduced considerably. The position of the magnetic pole calculated lies between the poles determined from European Precambrian rocks by Creer (1957) and by Irving and Runcorn (1957) and calculated from the measurements of early Paleozoic European rocks by Cox and Doell (1960).

For comparison, the direction of the remanent magnetization of two other rocks from southwestern Finland was measured:

- 1) Fine-grained quartz dolerite dike, Makarla, Piikkiö (latitude 60.45°N, longitude 22.54°E)
Declination 23°, inclination + 7°, $\alpha_{95} = 16^\circ$, k = 20, N = 4
Virtual pole position: latitude 30°N, longitude 175°E



Fig. 3. Virtual magnetic pole positions and the ovals of confidence according to the measurements of the magnetic directions of the Satakunta olivine dolerites. o = pole position according to the measurements before magnetic cleaning, ● = pole position calculated after demagnetization.

2) Quartz diorite, Röntämäki, Rytö, Maaria (latitude 60.50°N , longitude 22.30°E)

Declination 232° , inclination $+11^{\circ}$, $\alpha_{59} = 30^{\circ}$, $k = 4$, $N = 6$

Virtual pole position: latitude 13°S , longitude 32°W

The dolerite measured is similar to the diabase dikes in SW-Finland described by Laitakeri (1921), Pehrman (1933), and Metzger (1945). These dikes belong to a large dolerite system somewhat older than the Jotnian dolerites (Eskola, 1963).

The quartz diorite from Röntämäki belongs to the second intrusive series of Hietanen (1947) of the same age as the main metamorphic Svecofennides in Finland.

Both measurements listed were made after demagnetization of the secondary magnetization in a manner similar to that used in the case of the Jotnian dolerites. The directions and the pole positions calculated differ considerably from those obtained for the Jotnian dolerites. Much more work, however, is needed to ascertain whether the differences are significant. The results indicate, however, that it is worth extending the magnetic measurements started to embrace the different dike systems in Finland. The older dolerite dikes have undergone more alterations than the well preserved

Jotnian dolerites and the cleaning off of the secondary magnetization will evidently be a more difficult task. When finegrained and unmetamorphosed samples can be collected, reliable results might be expected.

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NOTES ON COLUMBITE AND TAPIOLITE FROM THE PEGMATITE
AREA OF TAMMELA AND SOMERO IN SW FINLAND ¹⁾

BY

ATSO VORMA

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ABSTRACT

Five specimens of columbite from the Tammela—Somero pegmatite area have been investigated. All proved to be in a disordered state. Appropriate heat treatment produces the ordered state. During the transformation a monoclinic FeNbO_4 -like phase develops in addition to the ordered columbite.

Two specimens of tapiolite from the Tammela pegmatite are described. Both are of the type characterized by a monorutile structure. Heat treatment causes the rutile structure to turn into a trirutile one.

X-ray fluorescence analyses of both the columbite and the tapiolite are made.

INTRODUCTION

In conjunction with a study (to be published soon by the present author) on ainalite, the tantalian cassiterite found at Penikoja, Somero, SW Finland, new data for the associated columbite was recorded. In association with the ainalite occurring at Sukula, Tammela, not only columbite, but also tapiolite was encountered. This is the same pegmatite area from which A. E. Norden-skiöld in 1863 described the mineral for the first time and applied to it the name tapiolite. It is thus quite justified to give X-ray data for this tapiolite.

Most of the ainalite specimens from the collection of the University of Helsinki contain columbite (*e.g.*, Nos. 439, 440, 443, 444, 446 from Sukula,

¹⁾ Received February 12, 1965.

Tammela and No. 4561 from Penikoja, Somero) and some tapiolite (*e.g.*, No. 445 from Sukula) as inclusions, while one specimen labelled as ainalite (No. 441 from Sukula) was pure tapiolite. In specimen No. 4561 columbite occurs as rather large inclusions in ainalite. This columbite was selected for a more detailed study. In addition, four other columbite specimens from the Tammela—Somero pegmatite area were investigated in connection with this study.

COLUMBITE

In the ainalite specimen No 4561 columbite occurs as inclusions, with the largest grains measuring up to 5 mm. A small crystal fragment of the mineral was separated from a polished section under a reflection microscope. From 0.7 mg of material a semiquantitative X-ray fluorescence analysis was made for the Nb/Ta ratio. Another fragment of the same grain was separated for single crystal X-ray study. Precession photographs showed that the unit cell of the mineral in question is a subcell of columbite with similar parameters except for b , which is one-third of the b axis of columbite. The same crystal fragment was heated in air for 2.5 hours at 900°C and a new set of precession photographs was taken. The b parameter now turned out to be 14.15 Å in accordance with the ordered columbite structure.

Nickel *et al.* (1963) suggested the name pseudo-ixiolite for the disordered columbite-tantalite, which after appropriate heat treatment would change to ordered columbite-tantalite. The name ixiolite was redefined and it refers to the disordered form of columbite structure, which after the heat treatment would give the X-ray powder pattern of olovontantalite.

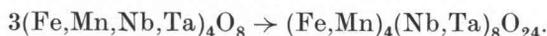
According to the nomenclature of Nickel *et al.* (*op.cit.*, p. 975—977) the mineral from specimen No. 4561 is pseudo-ixiolite. A small reduction in cell volume (the unit cell of columbite is compared with three unit cells of pseudo-ixiolite) during heating (Table 1) also supports the view of Nickel *et al.* that there is an order-disorder relation between the heated and unheated material.

In the precession photographs of heated material the diffraction spots corresponding to columbite predominate. In addition, there exists another system of reflections almost coinciding with those of columbite, the d -spacings, however, being slightly smaller and the systematic extinctions being different from those of columbite. The extinctions of columbite pattern are consistent with the space group $Pcan$ (No. 60). The cell size of the other phase is 5.64 Å ($\parallel a$ -axis of columbite), 4.66 Å ($\parallel b$ -axis of columbite), and 5.02 Å ($\parallel c$ -axis of columbite). This corresponds exactly with the monoclinic (ordered?) modification of $FeNbO_4$ described by Roth and Waring in 1964. The monoclinic $FeNbO_4$ is stable under 1085°C and has a very slight monoclinic distortion of the ixiolite structure type. Roth and Waring give the space

group $P2_1/a$ (No. 14) and the cell size as $a = 5.001 \text{ \AA}$, $b = 5.620 \text{ \AA}$, $c = 4.653 \text{ \AA}$ and $\beta = 89.84^\circ$.

As seen from the precession photographs (Plate I) the most indicative reflection of the additional phase in the photographs recorded by Mo-radiation is the 001 reflection. This corresponds to the 030 reflection of columbite structure. In columbite the 030 reflection is, however, ruled out (Space group $Pcan$). In the photographs recorded by Cu-radiation the splitting of reflections between the orthorhombic and monoclinic phases is clear.

Pseudo-ixiolite may be formulated as $(\text{Fe}, \text{Mn}, \text{Nb}, \text{Ta})_4\text{O}_8$, which represents a disordered structure. After appropriate heat treatment an ordered columbite structure would result



This requires the ratio $(\text{Fe} + \text{Mn}) : (\text{Nb} + \text{Ta}) = 1:2$. If, on the other hand, the ratio is more than 0.5 and ordering takes place, part of the pseudo-ixiolite would presumably be transformed to the monoclinic FeNbO_4 -phase.

Specimen No. 1077 (3) from the collection of the University of Helsinki was compared with the pseudo-ixiolite in specimen No. 4561. Contained in a glass tube, the sample comprised five crystals (with a diameter of 1–2 mm each) labelled »columbite, Penikoja, Tammela, T:hus län». It thus originates from the same small pegmatite quarry from which the ainalite specimen No. 4561 originates. One of the crystals was mounted on a goniometer head and a -axis precession photographs were made. Owing to the large size of the crystal ($> 1 \text{ mm}$) and the resultant strong absorption, combined with the inaccuracy of the effective distance between the crystal and the film, the b and c parameters derived (4.68 \AA and 4.98 \AA respectively) are certainly too small. The result, however, indicated a disordered columbite structure. Another crystal was cut under the microscope into three pieces and an X-ray powder photograph taken of one fraction. The powder pattern indicated the disordered form. The cell parameters derived from the powder pattern are given in the Table 1. The remaining two fractions were heated 3 hours at 900°C in air. A new powder photograph was recorded showing a clear decrease in the d -spacings compared with those of unheated material. In addition, the reflection lines were broader than those of unheated material. A set of precession photographs from the third fraction (also heated) showed that there are two phases present in the heated material (Plate II). The predominating phase was columbite, the other being structurally similar to the monoclinic FeNbO_4 .

During a visit to the Penikoja quarry the author managed to find still another crystal of columbite, which according to the X-ray powder data also proved to be in the disordered state. In addition, a specimen labelled as yttrotantalite from Torro, Tammela (No. 7244 in the collection of the

Table 1. Unit cell size of disordered columbite (pseudo-ixiolite), ordered columbite, and the monoclinic FeNbO₄-like phase produced by heating disordered columbite. Penikoja, Somero, Finland.

Unheated		Heated in air at 900°C for 2.5 hrs.		Heated in air at 900°C for 3 hrs.	
PSEUDO-IXIOLITE	PSEUDO-IXIOLITE	COLUMBITE	MONOCL. FeNbO ₄ -like phase	COLUMBITE	MONOCL. FeNbO ₄ -like phase
4561	1077	4561	4561	1077	1077
$a = 5.72 \text{ \AA}$ $b = 4.72$	$a = 5.74 \text{ \AA}$ $b = 4.73$	$a = 5.70 \text{ \AA}$ $b = 14.15$ (3×4.72) $c = 5.05$ $V = 3 \times 135.8 \text{ \AA}^3$	$b = 5.64 \text{ \AA}$ $c = 4.66$ $a = 5.02$ $V = 131.9 \text{ \AA}^3$	$a = 5.72 \text{ \AA}$ $b = 14.28$ (3×4.76) $c = 5.04$ $V = 3 \times 137.2 \text{ \AA}^2$	not det. $c = 4.64 \text{ \AA}$ $a = 5.04$ not det.
Precession photos MoK α	Powder photos CuK α	Precession photos MoK α	Precession photos MoK α , CuK α	Precession photos MoK α , CuK α	Precession photos MoK α , CuK α

Institute of Technology at Otaniemi, Finland) from the same pegmatite area where also Penikoja is situated, proved to be disordered columbite with inclusions of samarskite. The columbite from the Laurinmäki quarry, which is likewise situated in the Torro pegmatite area, is also, according to the X-ray powder data, in the disordered state (see chemical analyses: Mäkinen, 1913, p. 87). Nordenskiöld (1855, p. 162) mentions that in association with the ainalite of Penikoja, Somero, there occurs a tantalate, which has crystallized in the orthorhombic system. Later (1863, pp. 448, 451) he demonstrated the mineral to be columbite. According to Nordenskiöld columbite forms only small gray-black prisms covered on the surface by »metal ocher». The density is given as 5.61. The chemical analysis published by Nordenskiöld (*op. cit.*, p. 451) represents the composition of partly changed and disintegrated columbite: Nb₂O₅ (includes also Ta₂O₅) 81.70 %, SnO 1.87 %, CuO 0.28 %, Al₂O₃ 2.71 %, Fe₂O₃ 9.26 % and Mn₂O₃ 3.91 %, sum 99.73 %. Nordenskiöld writes referring to the analysis: »... jag förmodar derföre, att det ursprungligen i mineralet såsom oxidul ingående jernet och manganet redan blifvit till större delen syrsatt till jern- och manganoxid. Lerjorden torde härröra från främmande inblandningar.»¹⁾

During this study X-ray fluorescence analyses were also made of the disordered columbites Nos. 4561 and 1077 from Penikoja. Owing to the small amount of material only the ratios Nb₂O₅:Ta₂O₅ and FeO:MnO (in per cent by weight) were determined. For specimen No. 1077 the ratio Nb₂O₅:

¹⁾ »... I therefore assume that the iron and manganese originally contained in this mineral in the lower oxidized state has already for the most part turned into ferric oxide and manganese oxide. The alumina is probably derived from impurities.»

$Ta_2O_5 = 19:1$ and $FeO:MnO = 4:1$, while for the other specimen the ratio $Nb_2O_5:Ta_2O_5 > 18:1$.

Åmark (1941), in her treatment of the minerals ainalite and ixiolite, describes also specimen 4561 (labelled as ainalite) from Penikoja (*op. cit.*, pp. 296, 298) and according to X-ray powder photographs interprets the ainalite to be a mixture of cassiterite and columbite. However, she does not mention that as early as 1855 Nordenskiöld (p. 162) had written on the ainalite of Penikoja: »Den utgöres ej af ett, utan tvenne mineralspecies, af hvilka den ena är kristalliseradt i det tetragonala systemet och tyckes vara en mycket tantalhaltig cassiterit, det andra åter är kristalliseradt i det rhombiska systemet och utgör en verklig tantalitart Namnet ainalite torde vara lämpligast att behålla för det tetragonala mineralet.»¹⁾

As pointed out, Nordenskiöld later (1863) identified the orthorhombic mineral as columbite.

TAPIOLITE

Tapiolite No. 441 from Sukula, Tammela, was subjected to a more detailed study. The specimen consists mainly of potash feldspar, lepidolite and plagioclase. Imbedded in the lepidolite is a tapiolite crystal about 25 mm in diameter. A polished section revealed no opaque minerals as inclusion. A small crystal fragment of tapiolite was crushed and sieved. The small amount of silicate minerals present was separated in Clerici solution. Under a binocular microscope no impurities were found after two treatments with Clerici solution. This fraction was analyzed by X-ray fluorescence methods (Table 2).

The tapiolite from Sukula was investigated by X-ray powder and single crystal methods. The X-ray powder diagram revealed that the tapiolite from Sukula had the rutile structure, the d -spacings and intensities being comparable with those of tapiolite from Skogböle, Finland, described by Hutton in 1958. A heat treatment of one hour at 800°C produced the trirutile structure. Thus the tapiolite from Sukula behaved exactly like that described by Hutton. Hutton (*op. cit.*, p. 116) interprets this as indicative of an order-disorder relationship for the tetragonal compound $Fe(Ta,Nb)_2O_6$. Hutchinson (1955) dealt with the tapiolite occurring at Ross Lake, Northwest Territories, interpreting the rutile structure as disordered and the trirutile structure as an ordered form. Hutton's observation that when heated the tapiolite (ordered form) from Skogböle had denser packing than when unheated (disordered) strongly supports this idea. The unit cell dimensions for ordered and disor-

¹⁾ »This is not a single material but consists of two different varieties of minerals, one of which crystallizes according to the tetragonal system and appears to be a cassiterite with a high tantalum content, while the other mineral crystallizes according to the orthorhombic system and represents a true variety of tantalite. The name »ainalite» is probably best applicable to the tetragonal mineral.»

Table 2. Chemical composition of tapiolites from Sukula, Tammela, Finland.

	1	2
Ta ₂ O ₅	72	73.91
Nb ₂ O ₅	10	11.22
SnO ₂	1	0.48
FeO	13	14.47
MnO	1	0.81
	97	100.89

1. No. 441. Anal. V. Hoffren (semiquantitative X-ray fluorescence analysis).
2. Rammelsberg 1871 (see Laitakari 1921, p. 8).

dered Skogböle tapiolite, as determined by Hutton, are $a = 4.752 \text{ \AA}$, $c = 9.204 \text{ \AA}$, and $a = 4.753 \text{ \AA}$ and $c = 3.092 \text{ \AA}$ respectively. The unit cell dimensions of Kulmala tapiolite were determined by silicon-calibrated X-ray diffraction diagrams from both heated and unheated specimens. The cell constants derived are: $a = 4.747 \pm 0.001 \text{ \AA}$, $c = 9.183 \pm 0.003 \text{ \AA}$, and $a = 4.753 \pm 0.001$ and $c = 3.084 \pm 0.001 \text{ \AA}$ respectively. The cell volume of ordered Skogböle tapiolite according to Hutton (*op. cit.*, p. 117) is 207.840 \AA^3 while three times the cell volume of the disordered variety is 209.599 \AA^3 , the difference being 1.7_6 \AA^3 . The corresponding figures for Kulmala tapiolite are 206.9_3 \AA^3 and 209.0_1 \AA^3 , giving a difference of 2.0_8 \AA^3 , or slightly more than that of Skogböle tapiolite.

A single crystal study of Kulmala tapiolite was made with unheated material by using a Buerger precession camera to record the a -axis zero-level and first-level photographs as well as the c -axis zero-, first-, second-, and third-level photographs (with $c = 9.25 \text{ \AA}$). The photographs revealed that the structure really was of the monorutile type. Very strongly overexposed photographs (first- and second-level c -axis prec. phot.) showed some diffuse spots which conformed with the trirutile structure and indicated the presence of domains where some degree of order exists. Excepting the diffuse reflections the systematic extinctions were in accordance with those belonging to the space group $P4_2/mnm$ (No. 136). The same crystal fragment used for the aforementioned precession photographs was heated in air at 800°C for one hour. A new set of precession photographs revealed that the reflections belonging to the trirutile structure were enhanced. Systematic extinctions indicated again the space group $P4_2/mnm$.

Tapiolite from Sukula, Tammela, has been previously analysed by Rammelsberg (Table 2). The old analysis is highly similar to the new one. It is suprising how poor in Sn the tapiolite from Sukula is, especially in view of the fact that cassiterite is commonly present in Sukula pegmatite.

During this study another sample, labelled adelpholite ¹⁾ originating from the Torro pegmatite area in Tammela, was also identified as tapiolite. The X-ray powder data of this specimen was, as in the foregoing cases, that of the rutile structure type.

SUMMARY

All the columbite specimens so far investigated from the Torro area are in the disordered state (pseudo-ixiolite). Appropriate heat treatment produces the ordered state. During the transformation, however, a monoclinic FeNbO_4 -like phase develops in addition to the columbite. The monoclinic phase apparently is isostructural with the wolframite structure (Roth and Waring 1964, p. 245). The formation of the monoclinic FeNbO_4 -like phase during heating of disordered columbite is considered by the author to be due to the deviation of the ratio $(\text{Fe} + \text{Mn}) : (\text{Nb} + \text{Ta})$ from 1:2.

The two tapiolite specimens investigated in this study — one from Sukula, Tammela, the other from Torro, Tammela — are both of the monorutile structure type. Heated to 800°C for one hour, the disordered rutile structure undergoes transformation to the ordered trirutile structure.

A question naturally arises. What causes all the tapiolites and columbites from the Tammela—Somero pegmatites so far investigated to be in the disordered state? One explanation may be the slight radioactivity of the occurrences. Lima de Faria, 1964, (p. 38) states: »The process of metamictization is such that an ordered structure transforms to a completely disordered or amorphous state.»

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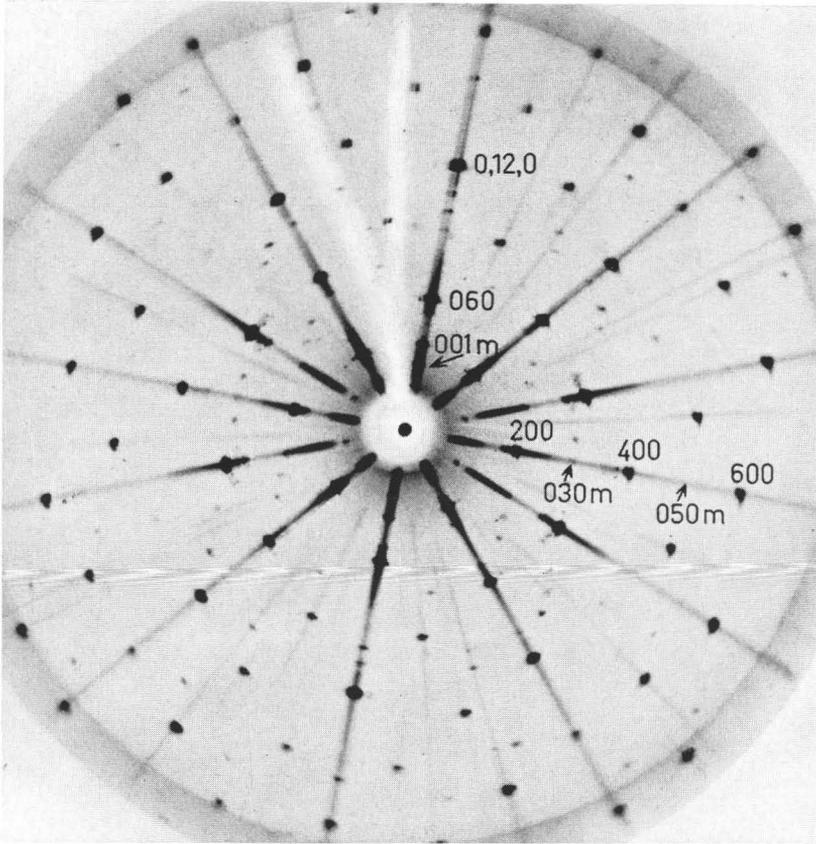


Fig. 1. z -axis, zero-level precession photograph of columbite from Penikoja, Somero (No. 4561). Heated in air at 900°C for 2.5 hrs. Natural size, $F = 6.0$ cm, $\mu = 30^{\circ}$, $\text{MoK}\alpha$ radiation, strongly overexposed. The reflections indicated by m belong to the monoclinic FeNbO_4 -like phase.

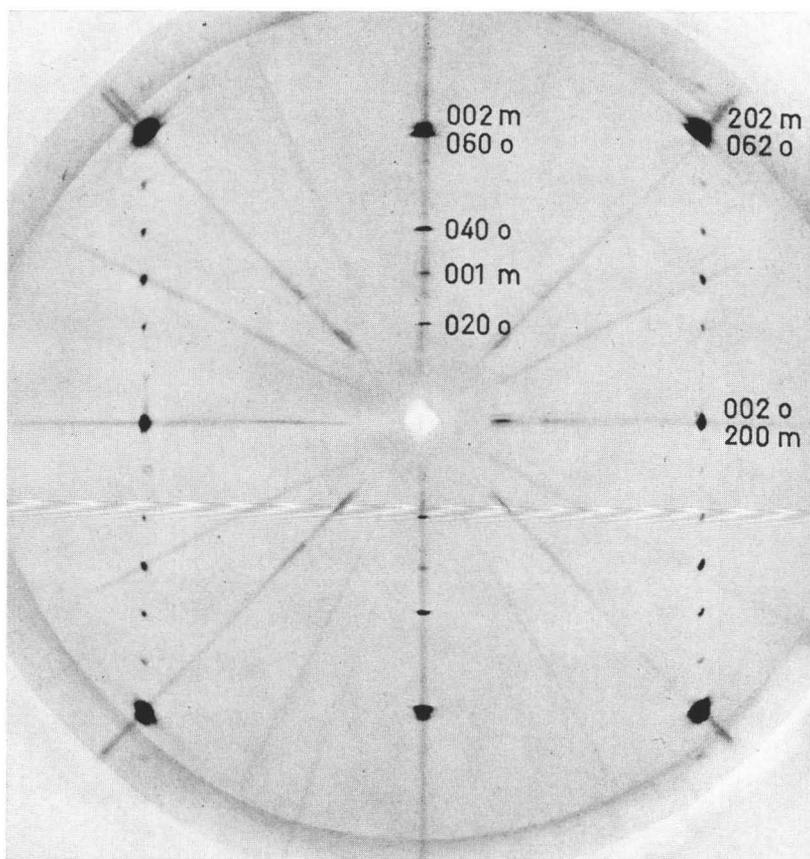


Fig. 2. *x*-axis, zero-level precession photograph of columbite from Penikoja, Somero (No. 1077). Heated in air at 900°C for 3 hrs. Natural size, $F = 6.0$ cm, $\mu = 30^\circ$, $\text{CuK}\alpha$ radiation. Reflections belonging to the columbite phase indicated by *o*, those belonging to the monoclinic FeNbO_4 -like phase indicated by *m*.

ANATEXIS IN THE BARTICA ASSEMBLAGE, BRITISH GUIANA ¹⁾

BY

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ABSTRACT

Geologic and petrochemical data indicate that within the Bartica Assemblage granitization, mobilization and anatexis are the results of a tectonically controlled extension of the metamorphic-metasomatic processes (mainly responsible for the formation of the associated gneisses) with a potash metasomatism now dominant. The resultant granites are distinguished as synkinematic or late kinematic. A synkinematic granite has, in general, acquired the petrochemical characteristics of a late kinematic granite by granitization. These granites, however, cannot be classified by petrochemical criteria alone and reveal several anomalous features with respect to granites in Finland and West Africa. Geologic evidence of anatexis is supported by radiometric data.

INTRODUCTION

The structure and metamorphic history of the Bartica Assemblage have been previously outlined (Cannon, 1962, 1964). In brief, the Bartica Assemblage is a complex of gneisses and subordinate amphibolites on the northern margin of the Guiana Shield; the following major mappable units have been established: amphibolites and granulites, hornblende-biotite gneisses, biotite gneisses, porphyroblastic gneisses and muscovite-biotite granites. The first recognizable episode of deformation and synchronous metamorphism was succeeded by a two-phase metasomatic episode (involving successive additions of Na and K) of the same general cycle. The grade of metamorphism was mainly within the amphibolite facies, locally reaching the granulite

¹⁾ Received February 26, 1965.

facies boundary, and also reaching the lower epidote-amphibolite facies during the retrogressive metasomatic phase. A second period of deformation later than the above phenomena is also recognized.

Anatexis within the above framework has been briefly sketched (*op. cit.*). The present paper examines in more detail (including three new chemical analyses) the geologic and petrochemical data which have led the writer to postulate anatexis as the end-product of the above metamorphic-metasomatic processes and, in particular, the bearing of this data on the problem of synkinematic and late kinematic granites. The present study was stimulated by the published work of Marmo (1955, 1955 a, 1956).

TERMINOLOGY

At the risk of seeming pedantic the writer proposes to describe how the following common (though controversial) terms are used in this paper. Granitization is used as defined by Read (1944, p. 47): »the process by which solid rocks are converted to rocks of granitic character without passing through a magmatic stage». Magma is considered a fluid which has been formed (by any method) from which an igneous-like rock has crystallized (*cf.* Read, 1943, p. 76; Barth, 1952, p. 62). Anatexis is »the process of forming new magma from solid rocks» (Read, 1944, p. 64), whilst mobilization is used as »an ordinary English word with its ordinary meaning, namely, making movable» (*op. cit.*, p. 88). Synkinematic and late kinematic are relative terms within the broad framework of an orogeny; the former is used for those granites formed at an early stage and the latter for those formed at a later stage (*cf.* Eskola, 1932; Marmo, 1955 a).

GRANITES IN THE BARTICA ASSEMBLAGE

Two main types of granite occur within the Bartica Assemblage: muscovite-biotite granites and biotite granites. The latter do not form mappable areas.

MUSCOVITE-BIOTITE GRANITES

These granites form small, irregular bodies within the gneisses, the only mappable area being the Kartabu Granite (Cannon, 1961). The latter is readily accessible (Fig. 1) and the following data have been derived from the area shown. The granite consists of two distinct types which can be conveniently described as a mafic facies and a leucocratic facies; the former is dominant and defines the outcrop of the Kartabu Granite.

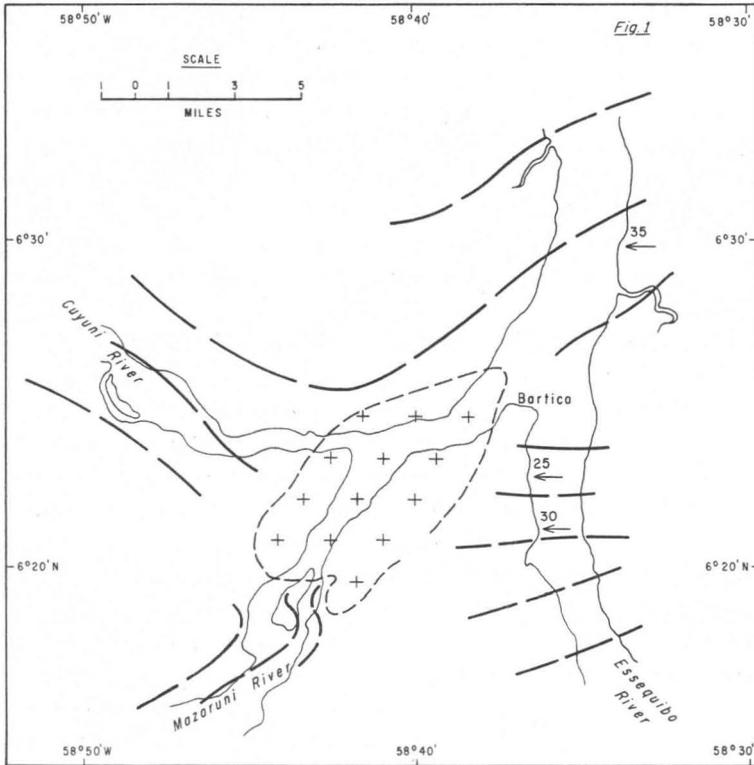


Fig. 1. Location of Kartabu Granite (+), generalized foliation directions of surrounding gneisses and amphibolites are shown, arrows indicate direction and plunge of fold axes of second deformation.

The mafic facies is distinctly gneissose, with conspicuous biotite and muscovite and appears to pass gradually into biotite gneisses to the northeast (Fig. 1). To the northwest, however, hornblende-biotite gneisses occur striking approximately at right angles to the granite margin here, suggesting probable mobilization of the granite mass in this area. However, the typical structures of the first deformation of the gneisses of the Bartica Assemblage (Cannon, 1962) occur within the mafic facies, though less clearly defined. In the gneisses of the Bartica Assemblage the banding and foliation are secondary phenomena defined during the first recognisable deformation; these structures are partially granitized within the mafic facies. Granite grades into gneiss and *vice versa*, a common though important phenomenon as Barth (1952, p. 243) has emphasized. The mafic facies, therefore, is clearly a partly granitized gneiss: a mobilized granite-gneiss.

The leucocratic facies is subordinate and forms irregular »patches» within the mafic facies and, more importantly, intrusive bodies which are confined to the outcrop distribution of the mafic facies. The intrusive bodies are either vertical dikes a few feet wide or larger, inclined bodies tens of feet wide. The dikes are usually uniform, white rocks with distinctive muscovite and subordinate biotite. The larger bodies occasionally have a foliation defined by abundant muscovite, minor biotite and irregular, parallel pegmatites, all of which are parallel to the plane of discordance between the two facies. This uniform, planar foliation is very distinct from the complex banding of the mafic facies.

The foliation patterns of the two facies of the Kartabu Granite have proved irregular and inconclusive in relation to the surrounding gneiss patterns and they have, therefore, been ignored in figure 1. Both facies are noteworthy for red garnets which have been determined within the spessartite range by x-ray methods (R. Bradshaw, written communication). The Kartabu Granite and its associated pegmatites also have a unique mineral suite which includes, in addition to garnet, molybdenite, apatite, and meta-autunite (J. E. T. Horne, written communication); beryl has been recorded but not confirmed by the present writer.

Petrology — Both facies are coarse-grained and consist of muscovite, biotite, plagioclase, microcline and garnet with subordinate accessory minerals which include epidote, apatite, sphene and iron oxides. As noted above, biotite is very subordinate in the leucocratic facies but conspicuous in the mafic facies and is often mantled by muscovite. The plagioclase in both facies is near the albite-oligoclase boundary. U-stage measurements using the methods of Rittmann (1929), Turner (1947) and more recently Slemmons (1962) show a variation between An_6 and An_{14} . Significantly, the average of determinations in the mafic facies is An_{11} and in the leucocratic facies An_8 . The irregular transitional or oscillatory zoning which is occasionally developed in the plagioclases of the gneisses of the Bartica Assemblage (Cannon, 1964) has been observed only in the mafic facies. Microcline occurs with replacive relationships to the plagioclase in both facies but is conspicuously porphyroblastic only in the mafic facies. In the latter, the microcline varies from small, interstitial areas to large porphyroblasts lobing into and replacing the plagioclase. A more sodic margin in the plagioclase against the microcline is often noteworthy, as in the postulated metasomatic replacement phenomena in the surrounding gneisses (Cannon, 1962). The microcline has an irregular development of the typical cross-hatching and may also be microperthitic. Quartz is conspicuous and occurs in two generations; quartz is enclosed by porphyroblastic plagioclase and also lobes into the later microcline.

Petrographic modal analyses show a varying composition between adamellite and granodiorite. However, macroscopic modes using large,

polished, stained (for plagioclase and microcline distinction) slabs confirm that both types can be classified as true granites.

BIOTITE GRANITES

The biotite granites occur as dikes intruding all the rock-types of the Bartica Assemblage except the muscovite-biotite granites. The dikes vary from several feet wide to about two hundred feet and have sharp margins against the surrounding rocks. The smaller dikes may also have pegmatitic margins or a diffuse foliation parallel to their walls and may also contain rotated blocks of gneiss (Cannon, 1962, text-fig. 2). These dikes are not confined to larger areas of synkinematic biotite granites, though such areas may occur within the biotite gneisses so far undetected because of poor exposure.

The granites are coarse-grained, leucocratic (grey or pink) rocks with biotite forming scattered aggregates. Occasional dikes have conspicuous magnetite.

Petrology — The granites are petrologically simple. They consist of large plates of saussuritized plagioclase (An_{15} — An_{24}) and microcline. A replacive relationship of the latter to the former occurs. Biotite forms irregular patches and is occasionally mantled by muscovite whilst strained quartz forms large areas. No petrologic differences between the grey and the pink granites have been observed and the pink coloration is confined to the microcline.

DISCUSSION OF GEOLOGIC DATA

MUSCOVITE-BIOTITE GRANITES

Mafic facies — The field evidence suggests that the mafic facies of the Kartabu Granite is formed by granitization of the gneisses of the Bartica Assemblage. The abrupt passage between hornblende-biotite gneisses in the northwest of the granite (Fig. 1) suggests that some movement of the mass (*i. e.* mobilization) has occurred. The preservation of relict structures within the mafic facies also indicates that a considerable cohesion was preserved during mobilization.

Such movement can be partly explained in terms of increasing chemical mobility. The latter phenomenon has been previously noted in connection with the formation of the surrounding gneisses (Cannon, 1962, p. 169). As the granitization of the gneisses is postulated as an extension of the metamorphic-metasomatic processes previously outlined (*op. cit.*), a concomitant increase in chemical mobility seems reasonable.

The sketch map (Fig. 1) suggests that the location of the Kartabu Granite may be related to the second deformation. There is confirmatory evidence in that structures within the mafic facies appear to be granitized relicts of both the first and the second deformations, the fold styles now being more fluid (*cf.* Reynolds, 1958, p. 382). The second deformation of the Bartica Assemblage is everywhere notable for the fact that no axial-plane structures are produced (Cannon, 1964). Such fold styles surround the Kartabu Granite, again suggesting this mass has been mobilized. It is, therefore, tentatively suggested that an extension of the metamorphic- metasomatic processes has resulted in chemical mobility. This mobility combined with favourable structural controls resulted in mobilization of the granite and its upward movement. The similarity of the age determinations considered later of both the Kartabu Granite and the gneisses of the Bartica Assemblage provides additional support for the above generalizations.

Metamorphism and metasomatism, therefore, have apparently led to granitization and mobilization, a well known conclusion previously suggested by many writers. McGregor and Wi'son (1939), for example, postulated two main processes in the granitization of pre-existing rocks (a) metasomatism and (b) mechanical penetration by magma (*op. cit.*, p. 212). Further, as the formation of the mafic facies is apparently linked to the deformation of the Bartica Assemblage it can be considered synkinematic in relation to the later leucocratic dikes which cross-cut structures of the first and second periods of deformation within the mafic facies.

Leucocratic facies — The leucocratic muscovite-biotite granites occur as dike-like bodies petrographically similar to the mafic facies, both being particularly noteworthy for red garnets which do not occur in the surrounding gneisses. The fact that the dikes have been recorded only within the outcrop area of the mafic facies is particularly important and clearly indicates a genetic relationship with the mafic facies. They have sharp margins against the above-mentioned structures, and are more uniform than the surrounding mafic facies which together with their planar foliation (and concomitant pegmatite development), suggest that these dikes were fluid when intruded and represent magma formed by a further extension of the metamorphic-metasomatic processes outlined previously. A late kinematic origin for these granite dikes is proposed.

BIOTITE GRANITES

Approximately vertical dikes of biotite granite occur throughout the Bartica Assemblage. They are not associated with any synkinematic granite areas though such areas may occur. Similar granite dikes occur in British Guiana associated with Younger Granites which are usually associated with

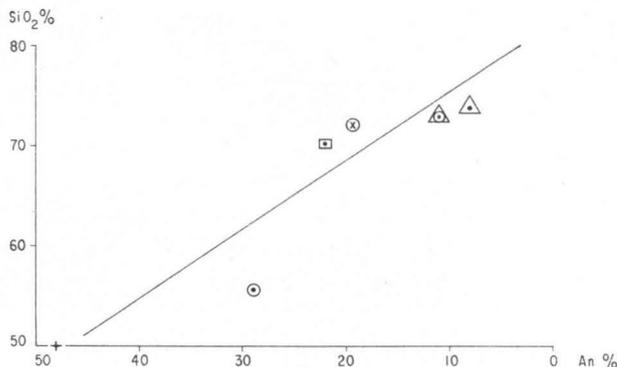


Fig. 2. Average anorthite content of plagioclases of rock-types of the Bartica Assemblage (symbols as in Fig. 3) is plotted against average weight per cent of SiO_2 for same rock-types, (analyses in Table 1 and Cannon, 1964).

gold. The present dikes are not associated with larger granites and little gold occurs throughout the whole of the Bartica Assemblage.

The sharp margins, occasionally pegmatitic or foliated, together with rotated blocks of host gneiss within the dikes suggest that the granite was fluid when intruded. The granites are mainly discordant to the structures of the second deformation and are also cross-cut by pegmatites similar to those associated with the gneisses but have not, themselves, been observed intruding the muscovite-biotite granites.

It seems, therefore, that magma has again been formed within the Bartica Assemblage apparently before the muscovite-biotite granites. The dikes have the characteristics of late kinematic intrusives but appear earlier than both the synkinematic and late kinematic facies of the Kartabu Granite. Clearly it is the terminology that is at fault and this problem is considered later.

DISCUSSION OF PETROCHEMICAL DATA

PLAGIOCLASE

The progressive metamorphic episode and succeeding retrogressive metamorphic episode within the Bartica Assemblage are reflected in the anorthite content of the plagioclase. This is discussed in more detail elsewhere (Cannon, 1964). There is an approximate «trend» within the rock-types of the Bartica Assemblage which extends to the granites discussed in the present paper. This is clearly shown in Fig. 2. It is fully realised that any such rock-types would probably show a similar relationship, but the rock-types shown are an integral part of the Bartica Assemblage and the plagioclases reflect their

adjustment to the same metamorphic and metasomatic episodes. A »trend» is, therefore, valid.

Granitization, therefore, is reflected by a decreasing anorthite content of the plagioclases. Marmo (1955 a) considers albitic plagioclase characteristic of late kinematic granites. This is valid for the leucocratic dikes though it may be noted that the plagioclase of the mafic facies is also albitic. The biotite granite dikes, however, typically contain oligoclase. Epidote, which Marmo (*op. cit.*) considers often characteristic of late kinematic granites, is subordinate in the granites described here but characteristic of the gneisses of the Bartica Assemblage.

The suggestion made previously that the biotite granite dikes may be related to synkinematic granite areas within biotite gneisses is pertinent in Fig. 2. A similar relationship occurs between the biotite granites and biotite gneisses and the leucocratic and mafic facies of the Kartabu Granite.

The zoning (transitional and occasionally oscillatory) of plagioclases throughout all the gneisses and also the mafic facies of the Kartabu Granite is an interesting feature. No such zoning has been recorded in the leucocratic facies of the latter granite. A further confirmation is made, therefore, that a similarity of conditions obtains between the gneisses and the mafic facies during their formation and that these conditions changed in the leucocratic facies, where the formation of magma is postulated.

The plagioclases in both facies usually exhibit lamellar A-twins or simple C-twins (*cf.* Gorai, 1951). This is considered by Tobi (1962) characteristic of magmatic rocks, a conclusion apparently valid only for the leucocratic facies. However, Tobi (*op. cit.*, p. 268) also points out that the twin pattern may become magmatic if the rock is soaked by a melt.

MICROCLINE

Marmo (1955 a, 1956) has pointed out the differing mode of occurrence of microcline in synkinematic and late kinematic granites. In the former, the microcline occurs interstitially or in a replacive relationship to the plagioclase which may be accompanied by the formation of myrmekite; in the latter, the quartz, plagioclase and microcline have all formed together. This distinction is not clearly defined in the Kartabu Granite where there is an obvious replacive relationship between microcline and plagioclase in both facies, although the true porphyroblastic nature of the former occurs only in the mafic facies.

The writer has suggested elsewhere (Cannon, 1964 a) that the potash feldspar of the porphyroblastic gneisses of the Bartica Assemblage was deposited as microcline. The porphyroblastic nature of the microcline throughout the gneisses and the mafic facies of the Kartabu Granite is readily explicable

Table 1

	1	2	3	4	5
SiO ₂	72.27	72.25	73.93	73.26	72.08
Al ₂ O ₃	14.01	15.40	14.37	14.58	14.65
Fe ₂ O ₃	1.95	0.40	0.52	0.50	0.30
FeO	0.60	0.99	0.29	0.18	1.17
MgO	0.84	0.22	0.26	0.22	0.30
CaO	1.15	0.55	1.12	0.62	0.92
Na ₂ O	2.70	3.63	4.20	4.75	4.50
K ₂ O	5.03	6.12	4.65	5.25	5.50
MnO	0.27	0.08	0.03	0.13	0.13
TiO ₂	0.58	0.09	0.08	0.03	0.20
P ₂ O ₅	0.04	0.09	0.04	0.10	0.08
H ₂ O+	} 0.52	0.14	0.57	0.23	0.25
H ₂ O-		0.01	0.05	0.06	0.01
	99.96	99.97	100.11	99.91	100.09
K ₂ O:Na ₂ O	1.86	1.69	1.1	1.1	1.2
Niggli Values:					
al	44	49	47.5	47	45
fm	18.5	8	6	5.5	9
c	6.5	3	7	3.5	5
alk	31	40	39.5	44	41
k	0.55	0.53	0.42	0.42	0.45
mg	0.36	0.21	0.37	0.34	0.25
qz	161	128	158	127	111
si	385	388	416	403	375
alk:c	4.8	13.3	5.6	12.6	8.2
No. of cations in Stand- ard Cell of Barth:	95.5	95.4	96.1	95.8	96.7
Rock No.	—	F 1376 A	F 212	F 1374	F 1372 A
Location:	Kalacoon Quarry	Kartabu Quarry	Barakara Quarry	Kartabu Quarry	Ampa Bay Quarry
Rock-type:	Average composition of quarry rock	Mafic facies, Kartabu Granite	Leucocratic facies, Kartabu Granite	Leucocratic facies, Kartabu Granite	Biotite granite
Analyst:	J.B. Harrison (Harrison, 1934)	M. D. Hope	W.H. Bennett (cf. Cannon, 1961)	M. D. Hope	M. D. Hope

in terms of a metasomatic phase (Cannon, 1962). This porphyroblastic habit is lost in the leucocratic facies (because of anatexis) but the replacive habit of microcline is still manifest.

The microcline in the biotite granites also has an obvious replacive relationship with the plagioclase and, in that respect, is also different from the late kinematic granites described by Marmo (1955 a).

ANALYTICAL DATA

Table 1 shows the chemical analyses of the granites considered in this paper with the Niggli values (al, fm, c, and alk) of the same analyses and also the number of cations in the standard cell (of 160 oxygen ions) according to Barth (1952, p. 82). The values of $K_2O:Na_2O$ and alk:c are often considered pertinent and are also shown.

Marmo (1955) has outlined two groups in the West African granites corresponding in their petrochemistry to the synkinematic and late kinematic Svecofennide granites. The synkinematic group is characterised by alk:c < 3, fm is variable but usually between 15 and 30, $6 < \text{alk} < 11$, $K_2O:Na_2O < 1$ and the number of cations in the Barth standard cell is usually more than 95 (*op. cit.*, p. 416). These granites, however, are usually granodioritic.

In both facies of the Kartabu Granite alk:c > 3, fm variable but usually low (only the bulk analysis > 10) and alk is high (> 30). The values of $K_2O:Na_2O$ are more instructive: all values are > 1 but those of the leucocratic granite and the biotite granite are significantly lower than the mafic granite. Similarly, k-values of the former are lower than the latter. These values are in marked contrast to Marmo's data above; it may also be noted that Backlund (1943) pointed out that the Urgranites of Fennoskandia, the Older Granites, have a lower k-value than the Younger Granites.

The soda-potash diagram (Fig. 3) is more pertinent to the present problem. There is clearly an increase in soda of all the rock-types relative to the amphibolite but generally the soda content is fairly uniform, *i. e.*, if the amphibolites are ignored fluctuations of soda content occur in all the rock-types about a mean of a little over 3 %. The porphyroblastic nature of the plagioclases in the gneisses, in addition to their increasing soda content during granitization, obviously indicate a movement of soda; figure 3 suggests that this movement may be a redistribution. An increase in potash accompanying granitization seems clear (Fig. 3), although the leucocratic facies is not the richest in potash. The writer would prefer to use the graphs to support the broad generalization of an increase in potash (of the Kartabu Granite in relation to the associated gneisses) until more data are available; the above k-values suggest, however, that actual anatexis here involves a decrease in potash. In the Slieve Gullion area Reynolds (1958, p. 387) states that »potash metasomatism culminated in the actual melting of the chemically altered rocks». It is also of interest to note the relation of the major gneiss groups of the Bartica Assemblage with respect to the synkinematic granites of Simonen (1948) — point A in figure 3, and the relation of the granites of the present paper to the late or post-kinematic granites—point B.

The late kinematic group of Marmo (1955) is richer in alkalis, alumina and silica than the synkinematic group and the number of cations in the

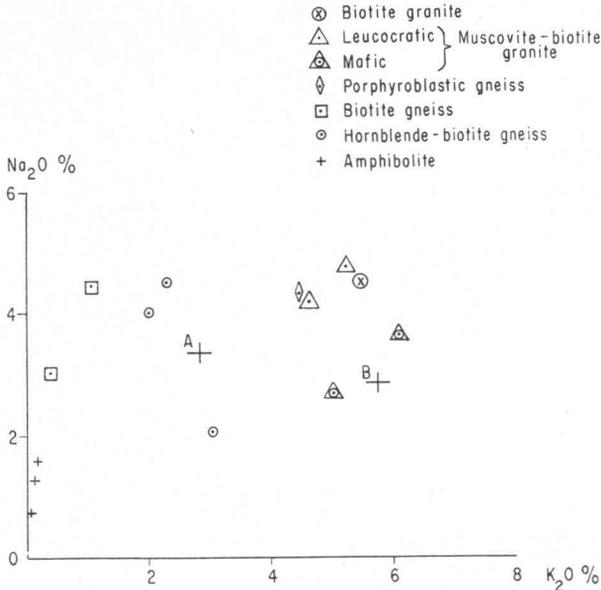


Fig. 3. Soda-potash diagram for rock-types of Bartica Assemblage. A: synkinematic granites and B: late or post-kinematic granites (Simonen, 1948) — taken from Dearnley (1963, Fig. 7).

Barth standard cell is less than 95. The present analyses show no such clear distinction. The number of cations in the standard cell remains reasonably constant, though slightly higher for the leucocratic facies and the biotite granite. In all the granites, however, the number of cations in the standard is greater than 95. The standard cells also show that the most significant differences between the mafic and leucocratic facies are a gain of Na and a loss of K and Mg for the latter compared with the former.

RADIOMETRIC DATA

All the age determinations so far available for the Bartica Assemblage are shown in Table 2. The Bartica Assemblage occupies an area of some 3250 square miles and more data are obviously required before any detailed synthesis can be attempted. If the experimental uncertainty of each determination in Table 2 is considered it is apparent that all the ages are similar. It is logical to assume that for the gneisses the age determinations reflect the last major tectonic episode (*cf.* Stockwell, p. 124 in Lowdon *et al.*, 1963): the second deformation of the Bartica Assemblage. The similarity of the ages of the gneisses and both facies of the Kartabu Granite, therefore, support the

Table 2

Specimen No.	Rock-type	Location	Mineral	Method	Age (10 ⁶ years)
F 1725	Mafic facies, Kartabu Granite	6°22'N 58°42'W	Muscovite	K-Ar	1960 ± 60
F 1374 A	Leucocratic facies, Kartabu Granite	ditto	Muscovite	K-Ar	1870 ± 55
F 1720	Porphyroblastic gneiss	6°10'N 58°35'W	Biotite	K-Ar	1780 ± 55
F 1373 A	Augen gneiss	6°31'N 58°34'W	Biotite	K-Ar	1840 ± 50
M 552	Hornblende- biotite gneiss	6°01'N 59°15'W	Biotite	K-Ar	1875 ± 55

All determinations by N. J. Snelling of the Age Determination Unit, Overseas Geological Surveys at the Department of Geology and Mineralogy in the University of Oxford.

previous evidence that a tectonic control during the second deformation was responsible for the formation of the granite.

SYNKINEMATIC AND LATE KINEMATIC GRANITES

As Reynolds (1958, p. 384) has remarked »the granite problem is a tectonic problem». The writer originally considered that the Kartabu Granite had been emplaced within the axial belt of an antiform in infrastructural gneisses (Cannon, 1961). This conclusion may appear a little premature in a forest covered terrain. It seems clear, however, that a complex, major antiform occurs in the area and is indicated by the plunges of the folds of the second deformation (Fig. 1). The location of the granite and the presence of granitized folds (of the first and probably of the second deformation) indicate a probable control by the second deformation which formed a structural locus (*cf.* Fig. 1).

However, it has previously been established that the granitization is an extension of the metamorphic-metasomatic phenomena of the *first* deformation. It is possible, therefore, that the second deformation provided a locus for the phenomena already initiated. The folding would localise the influx of potash mainly responsible for the granitization and eventual anatexis. This provides an explanation of the confinement of the leucocratic dikes to the area of the mafic facies. The wholesale movement of the granite mass has been noted, this would indicate its formation at a lower structural level and later emplacement upwards. Eskola (1956) has suggested that the source of potash (in his ideal granites) be related to the upward migration of light atoms in the earth's crust.

The possibility of an unrelated leucocratic granite »soaking» the gneisses of the Bartica Assemblage, granitizing the surrounding rocks (to produce the mafic facies) and eventually intruding them has also been considered. This can be rejected for two reasons. Firstly, the »trends» shown in figures 2 and 3 are manifest throughout the Bartica Assemblage and are not confined to the muscovite-biotite granite periphery, *i. e.*, the Kartabu Granite is the culmination of these trends, not the cause. Secondly, the similarity of the age determinations of both facies of the Kartabu Granite when compared with the gneisses of the Bartica Assemblage over a wide area, including one specimen (M 552 in Table 2) nearly 4-5 miles from the Kartabu Granite.

The formation of the Kartabu Granite was probably initiated early in the main tectonic-metamorphic episode and the later leucocratic dikes suggest that the relative terms synkinematic and late kinematic for the two facies are valid. Marmo (1955 a, p. 431—432) listed the main macroscopic differences between synkinematic and late kinematic granites. Most of these differences are applicable to the two facies of the Kartabu Granite except that the mafic facies is not granodioritic. As Marmo (*op. cit.*) has noted, most synkinematic granites are granodiorites. Such granodiorites are well known within the geosynclinal deposits of the Barama-Mazaruni Assemblage in British Guiana. The writer has mapped granodiorites in the middle Essequibo River area which are notable for peripheral mantles of a more potash-rich adamellite characterized by late porphyroblastic microcline. Reynolds (1958, p. 385) considers similar granites located in the Transition zone of Wegmann (1935). Clearly there are synkinematic granites and synkinematic granites.

The mafic facies of the Kartabu Granite has been derived from synkinematic (often granodioritic) gneisses and eventual anatexis is postulated above. Whilst the intrusive, fluid nature of the emplacement of the biotite granite dikes has been outlined their origin is still obscure. As pointed out above, they are apparently slightly older than the e m p l a c e m e n t of the Kartabu Granite. However, the terms late kinematic and synkinematic are relative to the orogeny and the dikes, therefore, can be considered late kinematic on the evidence outlined above. There is moreover a tenuous indication that these dikes may be related to the biotite gneisses of the Bartica Assemblage.

The relations of al and si of the analyses of Table 1 are shown in Fig. 4. Significantly, all the present granites fall within the field of the West African granites which correspond to the Svecofennide late kinematic granites and also within the actual field of the Svecofennian late kinematic granites. All the present granites fall outside or towards the edge of the field of the Swedish »urgranites» even though the k -values of the present late kinematic granites are similar to the latter.

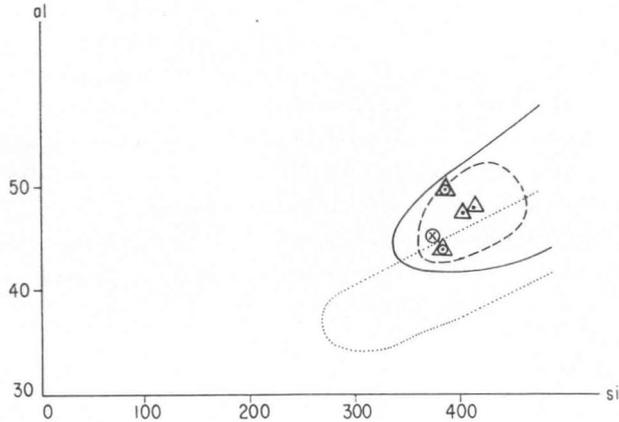


Fig. 4. Relations of al and si for granites of Table 1, symbols as in Fig. 3. Full line shows West African granites which approach in chemical composition the Svecofennide late kinematic granites, the stippled line indicates the field of the Svecofennide late kinematic granites according to Simonen (1948), thinly stippled line shows the field of Swedish »urgranites» (also from Simonen, 1948) — all these lines taken from Marmo (1955, fig. 3).

The geologic data have established the differences between the synkinematic and late kinematic granites of the present paper. The petrochemical data are variable but strongly suggest that the synkinematic mafic facies of the Kartabu Granite has acquired a chemical composition similar to the late kinematic granites, presumably because of its granitization (*cf.* Marmo, 1955, p. 416).

SUMMARY AND CONCLUSIONS

The extension of the metamorphic-metasomatic processes mainly responsible for the formation of the gneisses of the Bartica Assemblage has led to the development of synkinematic muscovite-biotite granites, in particular the Kartabu Granite. Further extension of these processes (within which a potash metasomatism appears dominant) has resulted in mobilization and anatexis within the area of the latter granite. The resulting magma has intruded the synkinematic granite mainly in the form of dikes which are considered late kinematic. Petrochemical data confirm the postulated granitization and anatexis as an extension of the above metamorphic-metasomatic processes. Granitization was apparently initiated during the first major deformation of the gneisses but confined by the second deformation; these processes, operating at another structural level, resulted in the locali-

zation of potash and subsequent mobilization (upwards) of the granite and eventual anatexis within this body.

Potash metasomatism (structurally confined) therefore appears responsible for granitization and anatexis.

Biotite granite dikes, of late kinematic aspect, intrude all the rock-types of the Bartica Assemblage, except the muscovite-biotite granites, and suggest a further (earlier?) period of anatexis. These dikes are not confined to any synkinematic granites though they may be related to such areas within the gneisses.

The macroscopic differences between the present synkinematic and late kinematic granites are broadly similar to those outlined by Marmo (1955 a). The petrochemical data considered above, however, differ from similar granites in West Africa and Finland but in general support Marmo's (1955) suggestion that some synkinematic granites may acquire (by granitization) chemical affinities to late kinematic granites. Notable anomalies (*cf.* Marmo, 1955) occurring within the present granites are: (a) the replacive nature (in all the granites) of microcline towards plagioclase, it may be noted, however, its mode of occurrence in the synkinematic granite is markedly porphyroblastic, (b) the lower *k*-value of the late kinematic granite with respect to the synkinematic granite, (c) oligoclase within the biotite granites and (d) the slight decrease in volume (more cations in the standard cell) of late kinematic granites with respect to the synkinematic granite. The number of cations is > 95 for all the present granites, but the similarity of all the numbers (*cf.* Table 1) indicates that the small differences are probably not diagnostic.

Precambrian granites within the Bartica Assemblage, therefore, cannot be classified by petrochemical criteria alone. The Kartabu Granite provides geologic evidence of anatexis which is supported by radiometric data.

The similarity of the geological setting of northern British Guiana and West Africa has been previously noted (McConnell *et al.*, 1964). It should be emphasised, therefore, that the writer is considering here a special case of anatexis within the Bartica Assemblage. A comparison between the Precambrian granites of British Guiana and West Africa would be of considerable interest. However, petrochemical data for the Younger Granite Group (*op. cit.*) in British Guiana are sparse, though both synkinematic and late kinematic granites occur.

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THE MINERALOGY AND GEOCHEMISTRY OF THE YLÖJÄRVI
Cu-W DEPOSIT, SOUTHWEST FINLAND: BISMUTH-
BEARING APATITE ¹⁾

BY

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ABSTRACT

Hydroxy-fluorapatite in the Ylöjärvi deposit contains up to 1150 p.p.m. bismuth, suggesting that substitution of Bi^{3+} for Ca^{2+} in this mineral may have a significant influence on the distribution of bismuth in post-magmatic differentiates and, perhaps, in magmatic rocks.

Goldschmidt (1954) suggested that diadochic substitution of Bi^{3+} (0.96 Å: Smith, 1963) for Ca^{2+} (0.99 Å) in apatite, and, to a lesser extent, in other calcium-bearing minerals may account for much of the bismuth in magmatic rocks, but no reference was made to quantitative analytical data. Similarly, Rankama & Sahama (1950) and Smith (1963) stress the probable importance of apatite in the geochemistry of bismuth without recording supporting data. Few authors have detected bismuth in apatite, but Volborth (1954) and Brasseur *et al.* (1962) report minor concentrations of this element in apatites from the Viitaniemi Li-pegmatite, Finland, and from pegmatitic rocks in the eastern Congo, respectively. Brooks *et al.* (1960) and Brooks and Ahrens (1961) found no correlation between the Bi and Ca contents of various igneous, metamorphic, and sedimentary rocks, but suggested that traces of Bi (0.66 p.p.m.) in a carbonatite might reflect a Bi^{3+} — Ca^{2+} substitution. No enrichment of Bi could be confirmed in wollastonite, actinolite, apatite, or marble (Brooks & Ahrens, 1961), but no data were given for separated minerals.

¹⁾ Received March 1, 1965.

Coarsely-crystalline apatite is a characteristic accessory gangue mineral in the hydrothermal, epigenetic Ylöjärvi copper-tungsten deposit, near Tampere, southwest Finland. The apatite forms white acicular euhedra (up to 3 cm in length) intimately associated with tourmaline in the matrix of a breccia pipe cutting Bothnian meta-volcanics (Himmi, 1960; Clark, 1964 a). Complete chemical analyses of the apatite are not available, but fluorine occurs in trace amounts in apatite-bearing breccia (unpublished reports), while chlorine was not detected by X-ray fluorescence analysis (lower limit of detection; 0.75 weight per cent). The refractive indices ($\omega = 1.646-1.650$), cell dimensions, and c/a ratios (Brasseur, 1950) suggest that the apatite is a hydroxyl-rich member of the hydroxyapatite-fluorapatite series (Table 1). On the basis of the compositions of co-existing arsenopyrite and pyrrhotite it has been tentatively estimated that the apatite crystallized at an approximate temperature of $550^{\circ}-600^{\circ}\text{C}$, under confining pressures in the range 2000—3000 bars (Clark, 1964 b, c).

Selected minor elements were determined in 15 apatites from the deposit, using a progressive addition X-ray fluorescence method. The apatite concentrates were separated by hand-picking under a binocular microscope, and, with the exception of rare, minute uraninite inclusions, contained no significant impurities. Yttrium and strontium were detected in all specimens, and manganese, silver, and bismuth were confirmed in several of the apatites. The sensitivity of the analytical method for bismuth is 400 p.p.m., and this element may occur in appreciable concentrations in the remaining samples. The analytical data are presented in Table 1.

DISCUSSION

Native bismuth, bismuthinite, and matildite are rare constituents of the deposit, and bismuth occurs as a minor element in arsenopyrite (Clark, 1965; in press) and galena. With the exception of arsenopyrite, however, these minerals were not associated with the analysed apatites, and it is suggested that the bismuth detected in the apatite occurs as a true minor element in the structure of that mineral. The available data indicate that bismuth is preferentially concentrated in apatite in the lower parts of orebody IV. Yttrium and strontium, which are probably camouflaged by Ca^{2+} in the apatite, show a similar enrichment between the 330 m and 375 m levels of orebody IV; the parallel concentration of the three minor elements supporting the proposed occurrence of bismuth in the apatite structure. The bismuth contents of the Ylöjärvi apatites are probably too low to have a measurable effect on the cell dimensions and optical properties of the mineral. The coarse apatite crystals in the breccia matrix show biaxial-positive and biaxial-negative

Table 1. Minor elements contents and X-ray data for apatites, Ylöjärvi

Specimen no. and level*	orebody	concentrations (p.p.m.)					c/a (± 0.001)	a	c
		Yt	Sr	Mn	Ag	Bi		Å	
Ap 130 (1)	IV	110	30	n. d.	n. d.	tr.	0.730		
Ap 240 (3)	I	410	120	tr.	tr.	n. d.	0.730		
Ap 253 (1a)	III/IV	270	90	tr.	40	n. d.	0.732		
Ap 285 (1a)	IV	330	100	n. d.	40	n. d.	0.731	9.39 ₇	6.86 ₃
Ap 307 (1a)	III	410	130	tr.	tr.	r. d.	0.731		
Ap 330 (3)	I	720	250	130	tr.	n. d.	0.731		
Ap 330 (4)	I	710	240	120	60	n. d.	0.733		
Ap 349 (1)	IV	620	80	tr.	30	1 150	0.730	9.40 ₂	6.87 ₂
Ap 358 (1a) ¹	IV	720	260	tr.	60	600	—		
Ap 358 (1b)		750	200	n. d.	tr.	660	0.732		
Ap 358 (2a)		670	210	160	r. d.	450	—		
Ap 358 (2b)		630	—	n. d.	n. d.	510	0.731		
Ap 358 (6)	IV	1 020	390	n. d.	n. d.	1 090	0.730		
Ap 358 (12)	IV	1 370	430	n. d.	tr.	520	—		
Ap 358 (16)	IV	1 170	450	tr.	n. d.	tr.	—		
Ap 375 (7)	IV	430	150	tr.	n. d.	n. d.	—		
Ap 375 (8)	IV	110	30	130	n. d.	400	0.731		

* Depth in m. below +11 m.

¹ Specimens from same hand-specimen bracketed.

tr. trace

n. d. not detected

— not measured

Sensitivities: Yt, Sr, Ag: 20 p. p. m.; Mn: 80 p.p.m.; Bi: 400 p.p.m.

Elements not detected include: Ba, Cl, Cr, K, Mg, Mo, Ni, Pb, Rb, Si.

($2V_{\max}$ 15° and 12°) in addition to uniaxial-negative optics, individual grains occasionally consisting of uniaxial cores surrounded by biaxial-positive or -negative mantles. No relationship, however, has been found between the optical properties of the apatite and the minor element contents.

Yttrium and strontium occur in all the analysed apatites, while the breccia as a whole shows a significant enrichment in the former element. Analyses of calcium-bearing minerals (epidote, carbonates, and hornblende) in the country rocks of the deposit, and in the matrix of the breccias (scheelite, apatite, ankerite, and calcite) have shown that, whereas strontium is present in appreciable concentrations in the mineral phases which crystallized during the pre-mineralization retrograde metamorphism and wall-rock alteration, yttrium is markedly enriched in the breccias but occurs only in very minor amounts in the country rocks. Yttrium appears to have been present in appreciable concentrations throughout the period of hydrothermal ore deposition, and has been detected in calcite and ankerite occurring in low-temperature veins in the late-stage fault zones which displace the breccia bodies. Similar Yt:Sr ratios are shown by all the Ca-minerals in the breccias, although each element is preferentially enriched in the apatite.

Manganese is present in very minor concentrations in the Ylöjärvi apatites, reflecting the hydrothermal rather than pegmatitic nature of the ore fluids. This element was moderately enriched in Ca-minerals in retrograde metamorphic and wall-rock alteration periods preceding ore deposition, but was depleted in the ore fluids.

Silver occurs as a minor element in several of the analysed apatites, but the mode of occurrence of this element is unknown. Silver and bismuth do not co-vary, however, and it is considered that inclusions of argentite, native silver, or matildite were not responsible for the observed silver contents.

Ringwood (1955) has proposed that the high electronegativity of Bi^{3+} (1.8), as compared to that of Ca^{2+} (1.0), would inhibit the incorporation of bismuth in early-magmatic calcium-bearing minerals, resulting in the concentration of this element in late-magmatic fractions. Few reliable quantitative data on the bismuth contents of igneous rocks are available, but discrete bismuth minerals are certainly almost wholly confined to post-magmatic pegmatitic and hydrothermal differentiates. The presence of appreciable bismuth in the hydrothermal Ylöjärvi apatite demonstrates that incorporation of this element, presumably by diadochic substitution of Bi^{3+} for Ca^{2+} , may take place even in the presence of coexisting bismuth minerals. It should be noted, however, that bismuth was not detected by the author in fluorapatite from the Boliden deposit, Skellefte, Sweden,¹ or in several generations of apatite from the Viitaniemi pegmatite, Eräjärvi,² although discrete bismuth minerals occur in minor amounts in both environments. It is suggested that, whereas in many areas apatite in magmatic and post-magmatic rocks may prove to contain negligible amounts of bismuth, the present data lend some support to Goldschmidt's proposal, which should be evaluated by systematic analysis of apatites in various rock types.

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¹ Specimen kindly provided by Dr. E. Grip, Bolidens Gruvaktiebolag.

² Specimens provided by the Smithsonian Institution, Washington, through the generosity of Dr. P. Desautels.

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ON ADELPHOLITE AND ITS RELATION TO THE MINERALS OF
THE YTTROTANTALITE — SAMARSKITE SERIES ¹⁾

BY

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ABSTRACT

New data on the doubtful mineral adelpholite indicate that it is samarskite.

X-ray powder diffraction methods are used to investigate the phase changes in heated adelpholite. The chemical analyses are carried out by the XRF-analytical technique.

INTRODUCTION

The name adelpholite was given by N. Nordenskiöld (see A. E. Nordenskiöld, 1863 a, p. 452; 1864, p. 615) to a hydrated niobate from a small pegmatite quarry at Laurimäki, Torro, Tammela, Finland. According to the description of the mineral (A. E. Nordenskiöld, 1855, p. 87 and 1863 b, p. 144) it has been found only from two pegmatite quarries at Torro (Laurimäki and Rajamäki). An allied mineral has been mentioned from the adjacent Penikoja quarry at Somero and from the pegmatite of Rosendahl, Björkboda in Kemiö. The latter, however, has been shown to be malacon (A. E. Nordenskiöld, 1863 a, p. 452—453; 1864, p. 615; Dana 6th ed., p. 486).

According to the oldest description of adelpholite (Nordenskiöld, 1855, p. 87), which also seems to be the only authentic one concerning this mineral, adelpholite is probably a niobate-tantalate of iron and manganese, containing 41.8 % of metallic acids ((Nb,Ta)₂O₅?), 9.7 % H₂O, and small amounts of SnO₂, and 1.3 % SiO₂, the last being probably due to impurities. The mineral

¹⁾ Received March 16, 1965.

ought to be tetragonal, angles and crystal forms, however, being undetermined. $H = 3.5-4.5$. Streak white or yellowish-white. Fracture conchoidal. Transparent in thin splinters. Luster greasy. Color brownish-yellow, brown to black. Specific gravity 3.8 or probably more since it was impossible to obtain pure material for the determination.

Adelpholite occurs, according to A. E. Nordenskiöld, as small crystals embedded in feldspar. The mineral is associated with beryl and tantalite (columbite).

All the subsequent papers on the minerals of the Tammela pegmatites refer to the above mentioned paper of A. E. Nordenskiöld when describing adelpholite (*e.g.*, Mäkinen, 1912, p. 85). In Dana's System of Mineralogy, 7th ed., p. 762, Palache, Berman, and Frondel regard adelpholite as an altered fergusonite, on page 778 they, however, state that the mineral is possibly an altered mossite.

The present paper gives additional data on adelpholite and shows that the name as a mineral species name has no justification. The mineralogical and crystallographical part of this paper is by A. V., the chemical one by V. H.

SAMPLE DESCRIPTION

The authors had the following specimens at their disposal.

No. A 1094 from the collection of the Dept. of Geology and Mineralogy of the University of Helsinki. The mineral was labelled as Mossite (Adelpholite), locality unknown, a gift from Nordenskiöld. The sample consists of a greenish-black crystal fragment weighting *c.* 3 gm imbedded in potash feldspar. The mineral is transparent in thin splinters. In powder the color is light greenish-gray. The density, measured by Kraus-Jolly balance, is 5.27. Fracture conchoidal. Only three crystal faces were developed and the corresponding φ and ϱ angles measured by a contact goniometer are

(hkl)	φ	ϱ
(001)	—	0°
(011)	0°	48°
(100)	90°	90°

wich faces are indexed in accordance with the axial cross adopted for yttrio-tantalite

Specimen No. 1066 from the same collection is labelled adelpholite from Torro, Tammela, Tavastehus län. It consists of a black tapiolite crystal measuring 6—8 mm in diameter. The physical properties do not correspond to those of adelpholite. X-ray powder data showed the mineral to be a disordered tapiolite (monorutile structure). The tapiolite and columbite from the Torro area are described in a separate paper (Vorma 1965).

Samples 1—5/A. V./Torro/-64 collected by one of the authors (A. V.) from the Laurimäki quarry, Torro, Tammela, consist of coarse potash feldspar. Imbedded in the feldspar there are scattered crystals of beryl associated with disordered columbite (pseudo-ixiolite). In columbite there is sometimes intergrown, sometimes also occurring as individual crystals in potash feldspar, a brownish-yellow — yellowish-red, occasionally dark brown, almost black, mineral, the properties of which correspond closely to those of adelpholite. The fracture is conchoidal, transparent in thin splinters, the luster greasy. The crystal system could not be determined from the morphology. The specific gravity 4.40—4.45, determined with hot clerici solution, is slightly higher than that reported for adelpholite. This may be due to impurities in the old determination (see p. 202) or to a different degree of hydration. The Laurimäki quarry, from where the original adelpholite was described, is only a few meters in diameter and a couple of meters deep. It is situated in a 2—3 m broad granite pegmatite dike cutting diorite. In view of the small size of the Laurimäki quarry, the five specimens found during a few hours, which contain a mineral corresponding to adelpholite in description, and the fact that no other mineral that could be confused with adelpholite has been previously described from the quarry, it can be taken for granted that the mineral in samples 1—5/A. V./Torro/-64 is the same mineral as the one named adelpholite by Nordenskiöld.

Sample No. 7244 (1861) from the collection of the Geology Department of the Institute of Technology, Otaniemi, Finland, is labelled as yttrotantalite from the »green emerald quarry», Torro, Tammela. Additional note on the label: »has not all its characters for blow pipe». The sample contains a greenish beryl crystal with inclusions (diameter 1—5 mm) of disordered columbite. In the columbite the same yellowish-red mineral as in specimens 1—5/A. V./Torro/-64 occurs occasionally as inclusions (less than 1 mm in diameter).

In the literature the only reference to yttrotantalite in Finland is that of M. J. Durocher 1849 (p. 337), who refers to Nordenskiöld having found in Tammela *de l'ytthro-tantale uranifère*. Neither N. Nordenskiöld nor A. E. Nordenskiöld have ever described any yttrotantalite from Finland. It seems probable that this mineral has later been regarded by N. Nordenskiöld as well as by A. E. Nordenskiöld as being a new mineral species for which they used the name adelpholite.

CHEMICAL DATA

According to all known analytical methods the total quantitative chemical analysis of the metamict minerals is difficult and time consuming. It is even more difficult when the volume of the sample is small, as it is mostly. As to

Table 1. Comparison of the chemical analyses of a manganocolumbite (N:o Rb 75/61) analysed by the wet chemical method (by P. Ojanperä) column 1, and by the XRF-method (by V. Hoffrén) column 2.

	1	2
SnO ₂	0.33	—
Nb ₂ O ₅	37.65	40
Ta ₂ O ₅	43.45	45
FeO	1.73	2
MnO	14.50	15
TiO ₂	1.08	1
(Σc) *	1.53	
	100.27	103

*) (Σc) = (SiO₂ = 0.18, Al₂O₃ = 0.17, MgO = 0.04, CaO = 0.22, H₂O+ = 0.28, H₂O- = 0.02, NiO = 0.00, WO₃ = 0.60, U₃O₈ = 0.02) = 1.53 %.

the metamict minerals, a semiquantitative chemical analysis gives in most cases sufficiently accurate results, which can be used together with the crystallographic data for determination of the species of the mineral in question.

The semiquantitative chemical analyses in the following tables are carried out with the Philips XRF-spectrograph using a W-target tube, LiF-crystal and the scintillation and proportional counters.

The analyses were performed using the chart-driving technique. The spectrum was taken from TiK α to SnK α and the instrumental settings were so adjusted that the instrument recorded all the elements heavier than titanium from 1 per cent upwards.

The technique of analysis follows the same line as that of Lima de Faria's (1964).

A fine-ground sample (a few hundred milligrams of the mineral) is packed tightly in an aluminium ring having an inside dia. of 5 mm and a height of 3 mm. The sample ring is then placed into the sample holder with a 5 mm dia. Mylar window on the bottom. The spectrum is recorded and the heights of the peaks are measured.

The XRF-analyses are always based on the standards which have to be similar to the analysed material. For the metamict minerals it is not possible to obtain sufficient amounts of analysed minerals for the standards. However, in this case (metamict minerals) it is possible to use standards which are made by mixing carefully together the pure oxides of the elements in question. Although the standards made in this way do not exactly correspond to the state of the elements in the metamict minerals they are relatively close to

Table 2. Semiquantitative XRF-analyses of the adelpholites. Anal. V. Hoffrén.

Sample	SnO ₂	Nb ₂ O ₅	Ta ₂ O ₅	Y ₂ O ₃	U ₃ O ₈	ThO ₂	PbO	ZrO ₂	SrO	Yb ₂ O ₃	Er ₂ O ₃	FeO	MnO	TiO ₂	Σ
1	2	28	21	5	13	6	1	1	1	1	—	4	—	—	83
2	—	2	50	8	2	—	1	1	—	4	1	3	—	—	72

1. Adelpholite (= samarskite), Laurinmäki, Torro, Tammela.

2. Adelpholite (= yttrotantalite), A 1094.

The consistence of the elements lighter than titanium was controlled with an optical spectrograph by A. Löfgren. The optical spectral analyses showed both specimens to contain (in addition to the aforementioned) Ca, Si, Mg, Mn, Ti, and a small amount of Be.

them. This establishes the basis for the use of the XRF-method in this kind of analysis.

Owing to the very variable composition of metamict minerals, one has in fact to make a standard for almost every sample of this kind of mineral.

The time required to analyse one sample by the described method varies from 3 to 5 hours. The accuracy of the method is dependent on the concentration of the elements. The relative error is about 10 per cent at a concentration range from 5 to 50 per cent and about 20 per cent for lower concentrations.

An important feature in this analysing technique is that after the analysis all the original material is still left in its original form — only fine-ground. This is a notable advantage in the research on rare minerals.

To control the usability of the described method a sample of known composition (manganocolumbite N:o Rb 75/61) was analysed. The results are shown in Table 1.

Table 2 gives the semiquantitative X-ray fluorescence analyses of the brownish-yellow adelpholite from Laurinmäki (1—5/A. V./Torro/-64) and that of the greenish-black adelpholite sample No. A 1094. The chemical composition corresponds in the first case to that of samarskite and in the latter case to that of yttrotantalite. In Dana's 7th ed., adelpholite was included with the fergusonites. According to Butler and Hall (1960), no members in the fergusonite — formanite series have been found with atomic ratio Nb:Ta between 5:1 and 1:5. If this is so, the brownish-yellow material (with Nb:Ta = 2:1) presumably is not fergusonite. On the other hand, in the samarskite — yttrotantalite series there is a very wide variation for the atomic ratio Nb:Ta. Many samarskites with approximately the same Nb:Ta ratio as in the Laurinmäki adelpholite have been described (Lima de Faria, 1964, Table 25, columns 20 and 24; Butler and Hall, 1960, p. 397). The high U-content (13 % U₃O₈) of the Laurinmäki adelpholite is in accordance with the data for samarskites. The corresponding figure for fergusonites varies

between 1 and 4 %, occasionally reaching 8 % (Butler and Hall, *op. cit.*, p. 393). The Y-content (5 % Y_2O_3) also is consistent with that of the samarskite-yttrotantalite series. In the fergusonite-formanite series the figure is usually between 20 and 30 %, in the samarskite-yttrotantalite series only 2–15 %.

The high Ta_2O_5 -content (50 %) in the black adelpholite (No. A 1094) is consistent with the Ta_2O_5 contents both in formanite and yttrotantalite. The low Y-content (8 % Y_2O_3) however, is more consistent with the chemical composition of yttrotantalite. As will be seen later, the behavior of the black adelpholite during heating corresponds exactly to that of yttrotantalite. Thus we are justified in regarding the mineral as being yttrotantalite.

HEAT TREATMENT AND X-RAY DATA FOR ADELPHOLITE

Lima de Faria (1964) gives the chemical composition and X-ray powder data for 23 samarskite specimens. Heated in air at 700°C for three hours a cubic phase with $a = 5.16 \text{ \AA}$ was usually produced, occasionally also the *B*-phase (UTa_2O_8) was developed. Heated in air at 1000°C for one hour in addition to the same cubic phase and the *B*-phase, the F_I -phase (corresponding to the monoclinic $YNbO_4$) was often developed. In addition, the X-ray powder data indicate unidentified lines as well as in one case the η -phase. Of the samarskites investigated by Lima de Faria only Nos. 20, 22 and 23 (Table 25 of Lima de Faria) are comparable in their Nb:Ta ratio to the Laurinmäki adelpholite. Specimen No. 23 is also similar to adelpholite in respect of its U content. However, its Y content deviates very much from that of the Laurinmäki adelpholite. Specimen No. 20 has a very similar Y content to that of adelpholite. Samarskite identical in composition with the Laurinmäki adelpholite does not occur in Table 25 of Lima de Faria. Next in similarity of composition with adelpholite is sample No. 20¹). As to the X-ray data of this samarskite, the unheated material gives unidentified lines according to Lima de Faria. Heated in air at 700°C for three hours the mineral gives lines indicating the cubic structure ($a = 5.16 \text{ \AA}$, disordered pyrochlore-microlite structure?) and additional lines which could not be identified. Heated in air at 1000°C for one hour the F_I -phase was developed in addition to the cubic one. Neither the η -phase nor the *B*-phase were produced.

The brownish-yellow adelpholite (No. 1—5/A. V./Torro/-64) is amorphous to X-rays, occasionally, however, very faint diffuse reflexions are present indicating the presence of the pyrochlore-microlite structure (Table 3). An X-ray photograph of a stationary crystal fragment showed the material with the pyrochlore-microlite structure to be a cryptocrystalline aggregate. This

¹) Labelled originally as wilkite, Impilahti, Finland. University of Helsinki, No. 4675.

indicates that the cubic structure cannot be the original one. While all the specimens of the brownish-yellow adelpholite are from the waste rock that was quarried by heating and exfoliating the rock, it may be possible that some of the specimens have undergone an artificial recrystallization. Another possibility would be natural pyrochloritization. Heated in air at 700°C for three hours the reflexions of the cubic phase ($a = 10.38 \text{ \AA}$) are enhanced (Table 3). Simultaneously some unidentified reflexions develop. Heated in air at 1000°C for one hour the cubic phase is still the prevailing one. In addition, other reflexions were developed. Almost all of these could be indexed in a monoclinic cell using the unit cell constants $a = 5.29 \text{ \AA}$, $b = 11.10 \text{ \AA}$, $c = 5.01 \text{ \AA}$, and $\beta = 95^\circ$. This apparently corresponds to the F_I (monoclinic YNbO_4) or F_{II} (monoclinic YTaO_4) phase of Lima de Faria (1964, pp. 37—39), or to the solid solution of these phases. Ferguson (1957) gives the unit cell dimensions of YTaO_4 as $a = 5.34 \text{ \AA}$, $b = 10.94 \text{ \AA}$, $c = 5.07 \text{ \AA}$ and $\beta = 95.3^\circ$. Gorshevskaya, Sidorenko, and Smorchkov (1961) give the corresponding figures of YNbO_4 as 5.30 \AA , 10.92 \AA , 5.06 \AA , and $\beta = 94^\circ 12'$. In addition, many rare earth niobates and tantalates have the same structure (Komkov, 1959; Stubican, 1964; Gingerich and Bair, 1964). Evidently the monoclinic $(\text{Y,RE})(\text{Nb,Ta})\text{O}_4$ (β -fergusonite like structure) is a crystallization product of the brownish-yellow adelpholite heated in air at 1000°C for one hour.

In addition to the brownish-yellow, and in places reddish-brown, adelpholite, one crystal measuring up to some few mm of dark brown, to the naked eye almost black, adelpholite was found from the Laurinmäki quarry occurring together with the brownish-yellow material and disordered columbite. In places the dark brown material changes to brownish-yellow indicating that the different color variants of the mineral represent only different degrees of alteration. The dark brown adelpholite is amorphous to X-rays. Heated at 700°C for three hours the same cubic phase was developed as above (Table 3). Some additional reflexions which could not be interpreted were present in the corresponding powder photograph. Heated at 1000°C for one hour the monoclinic YTaO_4 -like phase was again developed. The corresponding X-ray powder photograph is almost identical with that of the brownish-yellow adelpholite.

In the literature the available X-ray data for a samarskite with a composition similar to that of the Laurinmäki adelpholite is too scanty for one to judge whether the results of the heating tests of adelpholite are normal or anomalous for a samarskite. At least they are anomalous in the absence of the β -phase (see Table 24 of Lima de Faria, 1964). However, if the data on the adelpholite are compared to those of the samarskite No. 20 in Table 24 of Lima de Faria, the samarskite being next in similarity of composition to the Laurinmäki adelpholite, it is seen that the results are very similar. The only

Table 3. X-ray powder data of adelpholites. The data are on unheated and heated

Cubic phase	Adelpholite ¹ Laurinmäki, Torro, Tammela, Finland					
	Original		Heated in air at 700°C for 3 hours		Heated in air at 1000°C for 1 hour	
<i>h k l</i>	<i>d</i> (Å)	<i>I</i>	<i>d</i> (Å)	<i>I</i>	<i>d</i> (Å)	<i>I</i>
1 1 1	6.00	vw d	6.02 5.32	3 1	5.98	1
0 2 2			3.66	< 1 d		
1 1 3					3.58 3.12	1 2
2 2 2	2.94	s b	3.03 2.95	4 10 b	2.971 2.907	10 b 2
	2.772	vw d	2.778	1	2.775	< 1 d
0 0 4	2.566	w d	2.578	3	2.656 2.580 2.516 2.446	< 1 d 3 < 1 1
	2.227	vw d	2.204	< 1 d	2.328 2.181	< 1 d < 1 d
1 1 5, 3 3 3	1.995 1.893	vw d vw d	1.979 1.875	< 1 d < 1 d	2.052 1.978 1.902	< 1 < 1 < 1
0 4 4	1.812	m d	1.824 1.738	5 1 d	1.866 1.827 1.742	< 1 6 < 1 d
	1.689	w d	1.686	1 d	1.703 1.688	< 1 < 1
					1.641	< 1
2 2 6	1.554	m d	1.558 1.506	5 < 1 d	1.556 1.513	6 < 1
4 4 4	1.493	vw d	1.492	1.5	1.494	1
1 1 7, 1 5 5			1.458	< 1 d	1.456 1.432	< 1 1
3 5 5, 1 3 7			1.403 1.348	< 1 d < 1 d		
0 0 8	1.285	vw d	1.293	2	1.350 1.2945 1.2228	< 1 1 < 1
2 6 6	1.180	w d	1.187	3	1.1874	3
0 4 8	1.150	w d	1.157	3	1.1581	3
			1.116	< 1 d	1.1132 1.0820	< 1 d < 1 d
4 4 8	1.053	w d	1.058	3	1.0568	3
6 6 6992	w d	.998	3	1.0173 .9976	< 1 d 3
0 8 8916	1 d	.9162 .9066 .8964	1 < 1 d < 1 d

materials the heat treatments being three hours at 700°C and one hour at 1000°C.

Adelpholite ² Laurinmäki, Torro, Tammela, Finland				Adelpholite ² No. A 1094				
Heated in air at 700°C for 3 hours		Heated in air at 1000°C for 1 hour		Original	Heated in air at 700°C for 3 hours		Heated in air at 1000°C for 1 hour	
<i>d</i> (Å)	<i>I</i>	<i>d</i> (Å)	<i>I</i>	<i>d</i> (Å) <i>I</i>	<i>d</i> (Å) <i>I</i>	<i>d</i> (Å)	<i>I</i>	
6.02	1	5.98	2	6.0 vw d	5.98	3	5.95	3
					4.75	1	5.5	≪ 1
							4.76	3
							3.75	3
		3.59	2					
		3.13	3	3.11 w b	3.11	< 1 d	3.12	3
3.09	< 1							
2.986	10	2.992	10	2.97 m b	2.98	10 b	2.94	10 b
		2.917	3					
		2.758	1				2.760	1 d
		2.650	≪ 1				2.626	< 1 d
2.587	4	2.592	4	2.60 w b	2.586	5	2.572	2
		2.517	< 1					
		2.457	< 1				2.449	< 1 d
		2.338	≪ 1				2.348	1
		2.174	≪ 1				2.212	1
							2.157	1
		2.065	≪ 1				2.057	≪ 1 d
1.987	≪ 1	1.995	≪ 1		1.986	< 1 d	1.991	≪ 1 d
		1.908	1				1.906	1 d
		1.870	≪ 1				1.857	1
1.831	6	1.833	6	1.83 w b	1.823	5	1.820	4
		1.743	< 1 d				1.736	1
		1.707	< 1					
		1.692	< 1				1.697	1
		{1.646	< 1 d}				1.634	< 1 d
		{1.630	< 1 d}				1.586	2
1.560	6	1.565	6	1.56 vw d	1.559	5	1.551	4
		1.515	< 1				1.516	≪ 1
1.494	2	1.495	2		1.490	2	1.486	1
1.451	< 1	1.456	≪ 1				1.457	≪ 1
		1.432	1				1.442	1
1.342	≪ 1 d	1.355	< 1		1.347	< 1	1.341	< 1 d
1.295	1	1.297	2		1.290	1	1.288	1
		1.220	< 1				1.2488	≪ 1
							1.2075	< 1 d
1.186	3	1.189	3		1.184	3	1.1807	3
1.157	3	1.159	3		1.154	3	1.1525	2
							1.1307	≪ 1
		1.117	≪ 1 d				1.1063	≪ 1
		1.084	≪ 1 d				1.0872	1 d
1.057	2	1.058	3		1.053	3	1.0517	2
.9964	2	.9975	3		.995	3	.9920	3
							.9563	1 d
.9140	< 1 d	.9166	2		.912	1 d	.9102	< 1 d

Table 3. (Continued).

Cubic phase	Adelpholite ¹ Laurinmäki, Torro, Tammela, Finland					
	Original		Heated in air at 700°C for 3 hours		Heated in air at 1000°C for 1 hour	
<i>h k l</i>	<i>d</i> (Å)	<i>I</i>	<i>d</i> (Å)	<i>I</i>	<i>d</i> (Å)	<i>I</i>
2 6 10872	w d	.876	3 b	.8763	3
0 0 12, 4 8 8864	2 b	.8644	2
					.8354	< 1 d
0 4 12820	3 b	.8199	2
					.8071	< 1 d
6 6 10792	3 b	.7909	2
4 4 12783	2 b	.7821	2
<i>a</i> of the cubic phase	10.3—10.4 Å		10.38 Å		10.38 Å	

Intensities by visual estimation; s = strong, m = medium, w = weak, vw = very weak, vvw = very, very weak, d = diffuse, b = broad.

1. Brownish-yellow adelpholite (samarskite). Some specimens of the adelpholite from Laurinmäki were amorphous to x-rays, some, however, gave faint diffuse lines when very long exposure times were used. Heated to 700°C the powder data of the different color variants in yellowish-brown — brownish-yellow or even reddish-yellow were identical. The same applies to the powder data of the materials heated to 1000°C.

conflicting fact is that the *c*-phase of the Impilahti samarskite (No. 20) has been interpreted as representing a disordered pyrochlore-microlite structure ($a = 5.16 \text{ \AA}$), while the corresponding phase of the Laurinmäki adelpholite represents the ordered structure ($a = 10.38 \text{ \AA}$). It is also worth emphasizing that the heat treatments carried out by Lima de Faria were done on powdered material both in air and in nitrogen. In the case of the Laurinmäki adelpholite the heating has been done in air using crystal fragments. Evidently the oxidization is restricted to the surface of the sample in the latter case. This entitles us to compare our results to those obtained by Lima de Faria when heating mineral powder in N_2 . In the case of the black adelpholite specimen No. A 1094, which in fact is yttrantalite, the phases which develop during heating in air at 700°C and 1000°C correspond to those obtained by Lima de Faria when heating Ytterby yttrantalite in N_2 . According to Lima de Faria (*op. cit.*, p. 39), a cubic phase ($a = 10.4 \text{ \AA}$) with the pyrochlore-microlite structure and the F_{II} -phase (monocl. $YTaO_4$) develop in heated yttrantalite. X-ray powder data, in addition, indicate some unidentified lines. The unheated adelpholite No. A 1094 gives very faint diffuse reflexions corresponding to the pyrochlore-microlite structure (Table 3). Heated in air at 700°C for three hours the cubic phase with $a = 10.33 \text{ \AA}$ is produced. Possibly the F_{II} -phase is present, the corresponding reflexions, however, are very

Adelpholite ² Laurinmäki, Torro, Tammela, Finland				Adelpholite ³ No. A 1094					
Heated in air at 700°C for 3 hours		Heated in air at 1000°C for 1 hour		Original		Heated in air at 700°C for 3 hours		Heated in air at 1000°C for 1 hour	
<i>d</i> (Å)	<i>I</i>	<i>d</i> (Å)	<i>I</i>	<i>d</i> (Å)	<i>I</i>	<i>d</i> (Å)	<i>I</i>	<i>d</i> (Å)	<i>I</i>
.8755	2 d	.8767	4			.873	2 b	.8719	3 b
.8628	1 d	.8642	3			.860	1 d	.8595	2 b
								.8400	< 1 d
.8187	2 d	.8204	3			.817	1 d	.8157	2 b
								.7978	< 1 d
.7904	2 d	.7913	3			.788	1 d	.8772	1 b
.7802	1 d	.7824	3					.7789	1 b
10.36 Å		10.38 Å		10.3—10.4 Å		10.33 Å		10.32 Å	

2. Dark brown adelpholite (samarskite), almost black to the naked eye. Amorphous to x-rays. Heated to 700°C only the cubic phase develops well. Heated to 1000°C the powder data are almost identical with those of the yellowish material.
3. Black adelpholite (yttrotantalite). Locality unknown. Unheated material gives very faint diffuse lines using the Debye-Scherrer method and long exposure time. Recorded with Norelco diffractometer the powder data indicate an almost amorphous state: only at 2.97 Å is there a broad diffuse maximum.

faint. The powder data of the adelpholite heated in air to 1000°C are very similar to those of the Ytterby yttrotantalite heated by Lima de Faria at 1000°C in N₂ for one hour. In both cases the lines of the cubic phase, those of the F_{II} -phase and some unidentified lines are present.

For the sake of comparison the heating tests in air were carried out with the Ytterby yttrotantalite crystal fragment (No. 4304 from the collection of the Dept. of Geology and Mineralogy of the University of Helsinki) and the corresponding X-ray powder photographs were compared with those of the adelpholite No. A 1094. The photographs were identical.

If the X-ray data of Laurinmäki samarskite and yttrotantalite No. 1094 are compared (Table 3), a surprising similarity becomes apparent. Both have a cubic phase with $a = 10.3\text{--}10.4$ Å, a monoclinic F_I - or F_{II} -phase, and some unidentified lines. This similarity is evidently due to the high Ta₂O₅ content in samarskite.

In this connection it is worth while recalling the recrystallization phases of the fergusonite-formanite series, when they have undergone the same heat treatment. According to Lima de Faria (*op. cit.*, pp. 37—39) the fergusonites develop an f -phase (tetragonal YNbO₄) if heated to 700°C for three hours. If heated at 1000°C the tetragonal phase usually turns into the monoclinic F_I -phase. If formanite is heated at 700°C for three hours, a c -phase with

$a = 5.16 \text{ \AA}$ is developed. Heated to 1000°C the F_{II} -phase becomes the prevailing one, while the c -phase still continues. Yttrotantalite and formanite have thus, after being heated to 700°C , a cubic recrystallization product. That of yttrotantalite seems to be an ordered pyrochlore-microlite structure, that of formanite a disordered one. Heated to 1000°C both produce the F_{II} -phase. The facts that point to the adelpholite No. 1094 being yttrotantalite rather than formanite are:

- the c -phase has $a = 10.3 \text{ \AA}$
- identical heating tests carried out to Ytterby yttrotantalite gave identical results
- the Y content is more consistent with yttrotantalite than with formanite
- the three crystal faces developed could easily be indexed using the axial cross of yttrotantalite.

SUMMARY

It has been shown that the mineral adelpholite described by A. E. Nordenskiöld in 1855 from the Tammela pegmatites is in fact a mineral belonging to the samarskite-yttrotantalite series. The atomic ratio Nb:Ta approximates 2:1. The occurrence of different color variants points to the possible differences in the degree of hydration of the mineral concerned. The water content, 9.7 %, is very similar to that in strongly altered samarskites, *e. g.*, in ampangabeites (Wambeke, 1960). In its chemical composition the Laurinmäki samarskite is comparable to those from Impilahti described as wiikites. Sample No. A 1094, which according to the museum label is adelpholite, also belongs to the samarskite-yttrotantalite series. However, since the properties of the sample do not correspond to those of adelpholite as described by A. E. Nordenskiöld, and since there is no mention of the locality where the material is derived from, it cannot be maintained that the name adelpholite would also apply to the Ta-rich end, yttrotantalite, of the samarskite-yttrotantalite series.

In consequence it can be stated that the name adelpholite has no reality and should be included among the discredited mineral names as one of the numerous synonyms for samarskite. Another interesting point is that the samarskite occurrence in question is the first one discovered within the present borders of Finland.

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DIE FICHTENVERHÄUFIGUNG IM LICHTE VON
C¹⁴-BESTIMMUNGEN UND DIE ALTERSVERHÄLTNISSE DER
FINNISCHEN POLLENZONEN ¹⁾

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AUSZUG

In der Untersuchung wird herausgestellt, dass das Häufigwerden der Fichte (*Picea abies*) in Finnland von Osten her metachron vor sich gegangen ist. Schon mit theoretischer Begründung kann ebenfalls als unwahrscheinlich erwiesen werden, dass die Grenzen der Pollenzonen auch nur in allen Teilen Südfinnlands gleichen Alters wären, und in zahlreichen konkreten Einzelfällen hat die Ungleichaltrigkeit festgestellt werden können.

INHALT

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DIE FICHTENVERHÄUFIGUNG IM LICHTE VON C¹⁴-BESTIMMUNGEN
DIE MOORE MUURASSUO IN PIHTIPUDAS UND SÄRKIJÄRVENSUO
IN KALKKINEN

Unter Anwendung der Transgressionskulmination des Vor-Päijänne als Vergleichsniveau hat festgestellt werden können, dass die Fichte im

¹⁾ Eingegangen den 31. März. 1965.

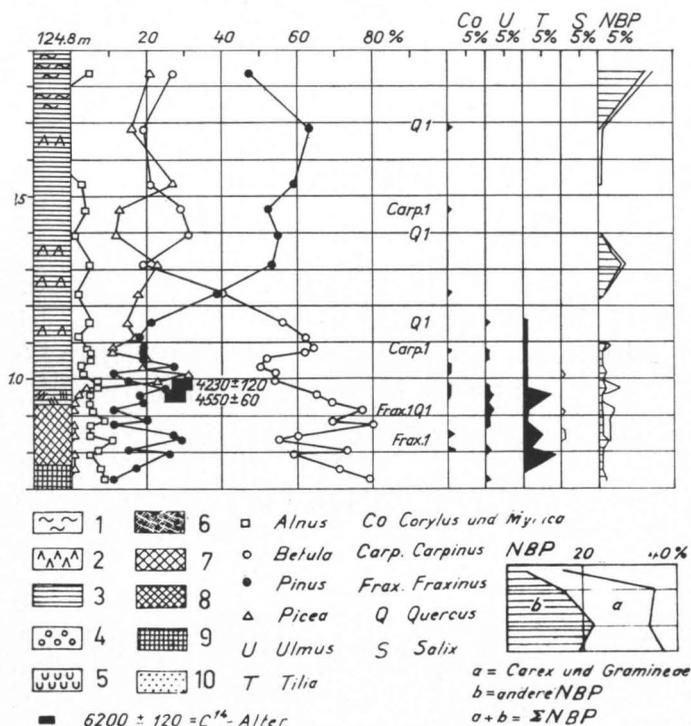


Abb. 1. Pollendiagramm des Moores Muurassuo in Pihtipudas.
 1. Sphagnumtorf, 2. Holzstückchen, 3. Carexortf, 4. Phragmitestorf,
 5. Equisetumtorf, 6. Akkumulationstorf, 7. Grobgyttja, 8. Gytjtja
 i. allg., 9. Tongytjtja und Gytjtaton, 10. Sand.

Seebecken nördlich unverkennbar später als südlich häufig geworden ist (R. Aario 1965 a, S. 185). Für eine genauere Bestimmung der Altersverhältnisse sind das Muurassuo von Pihtipudas und das Särkijärvensuo von Kalkkinnen danach Radiocarbon-Datierungen unterzogen worden. Die Probenreihen wurden durch Ausgraben entnommen. Die für die Isotopenbestimmungen zu benutzenden Proben wurden unmittelbar danach in 105°C Wärme getrocknet und darauf in Aluminiumpapier aufbewahrt. Die C¹⁴-Bestimmungen wurden im Radiocarbon-Laboratorium der Yale University ausgeführt.

Das Moor Muurassuo liegt an der von Muurasjärvi nach der Eisenbahnstation Muuras führenden Strasse südseits der Wegscheide von Laurila. Das neue Pollendiagramm vom Muurassuo (Abb. 1) ähnelt dem zuvor von mir über dasselbe Moor erarbeiteten Diagramm (ibid. S. 92), und die Fichtenverhäufung erscheint auch in ihrer Neuerfassung als steiler Anstieg.

Das C¹⁴-Alter der Fichtenverhäufung wurde anhand zweier aneinander grenzenden Proben bestimmt. Die obere enthielt an Fichtenpollen an

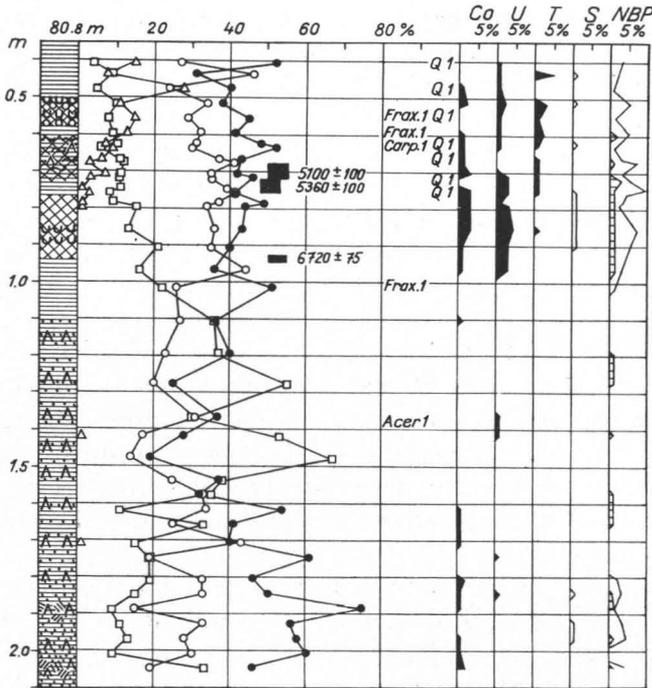


Abb. 2. Pollendiagramm des Moores Särkijärvensuo in Kalkkinen.

ihrer Unterfläche 0—2 % und an ihrer oberen 1—3 %. Die obere Probe barg an ihrer Unterfläche 6—10 % und an ihrer oberen über 20 % Fichtenpollen. Die Prozentsätze ergaben sich durch Untersuchung vieler Präparate für die gereinigten Flächen der Proben. Als C^{14} -Alter der unteren Probe (Y-1442) stellte sich 4550 ± 60 J. vor heute (1950) heraus und als das der oberen (Y-1443) 4230 ± 120 J.v.h. Die Proben bestanden aus Torf, und die Ablagerung liess keinerlei Störung erkennen, so dass die Proben zum mindesten keine solche Menge sekundären Materials enthalten dürften, dass sie das Ergebnis beeinflusst hätte.

Der für das Muurassuo ermittelte Zeitpunkt der Fichtenverhäufung steht im Einklang mit der für das — einige Kilometer weiter nördlich gelegene — Pikku-Ollovensuo herausgestellten C^{14} -Datierung (M-1523) von 4750 ± 180 J.v.h. für ein der Fichtenverhäufung zeitlich etwas vorausliegendes Niveau (ibid. S. 110).

Das Särkijärvensuo liegt in Kalkkinen, östlich der Einbuchtung Vehkaverkonpohja, am Ufer des Sees Särkijärvi, von dem oben beschriebenen Moor rd. 250 km nach Süden (ibid. S. 168). Die C^{14} -Bestimmungen beziehen sich

auf dieselbe ergrabene Probenreihe wie das zuvor von mir veröffentlichte Pollendiagramm sowie die damit zusammenhängende vom C-¹⁴-Laboratorium der Universität Heidelberg geleistete Datierung der Transgression des Vor-Päijänne (ibid. S. 168).

Auch für diese Lagerfolge wurde das Alter der Fichtenverhäufung auf Grund zweier Proben bestimmt. In der unteren Probe (Y-1444) erschien an Fichtenpollen sowohl an der unteren als auch an der oberen Fläche 1—3 %. In der oberen Probe (Y-1445) fanden sich deren an der unteren Fläche 5—9 % und an der oberen 5—10 %. Der Trend der Fichtenkurve ist gerade zwischen diesen beiden Proben am stärksten steigend, wenn auch nicht gleich stark wie im Muurassuo von Pihtipudas. Als C¹⁴-Alter der unteren Probe ergab sich 5360 ± 100 J.v.h.¹⁾ und als das der oberen 5100 ± 100 J.v.h. In den Proben war das Torfmaterial auch mit etwas Gyttja durchsetzt, die sekundäre Bestandteile enthalten konnte (u.a. R. Aario 1965 b). Da die im autochthonen Torf abgesetzte Gyttja wohl mindestens nicht viel Sekundärmaterial enthält und ihr Anteil am Torf gering ist, so ist die Fehlermöglichkeit immerhin als unbedeutend anzusehen.

Der beträchtliche Altersunterschied zwischen den C¹⁴-Datierungen der Fichtenverhäufung von Kalkkinen und Pihtipudas, nahezu 1000 J., entspricht der zuvor von mir dargestellten — auf einen Vergleich von Pollenchronologie und Entwicklung des Vor-Päijänne gegründeten — Auffassung von der Metachronie der Fichtenverhäufung in den verschiedenen Teilen des Seebeckens.

DIE ISOCHRONEN DER FICHTENVERHÄUFIGUNG

Einschliesslich der obigen Bestimmungen stehen augenblicklich aus Finnland über die Fichtenverhäufung folgende C¹⁴-Datierungen zur Verfügung:

U-231	1740 ± 120 J.v.h.	Fries 1961, 1963
U-232	2050 ± 120 J.v.h.	» » »
U-233	2530 ± 90 J.v.h.	» » »
U-235	2080 ± 90 J.v.h.	» 1963
Su- 6	4100 ± 100 J.v.h.	Salmi 1962
Su-10	4700 ± 100 J.v.h.	
Su-17	4350 ± 130 J.v.h.	
Su-25	3250 ± 140 J.v.h.	

¹⁾ Das für die untere Probe (Y-1444) ermittelte Alter ist ein neuer Grenzwert für das Transgressionsmaximum des Alt-Päijänne. Da dieses nach der betreffenden Datierung deutlich älter als 5360 Jahre ist und da es nach der früher ausgeführten Analyse unverkennbar jünger als 6200 Jahre ausfällt (R. Aario 1965a, S. 155), dürfte das zuvor von mir dargestellte Alter, 5800 J. (ibid. S. 183), mit rd. 200 Jahren Genauigkeit als zutreffend angesehen werden können.

Su-29	4400 ± 130 J.v.h.	Salmi 1963a
Su-45	4470 ± 130 J.v.h.	» 1963b
Y-1442	4550 ± 60 J.v.h.	
Y-1443	4230 ± 120 J.v.h.	
Y-1444	5360 ± 100 J.v.h.	
Y-1445	5100 ± 100 J.v.h.	
I - 680	5340 ± 190 J.v.h.	Tolonen 1963

Nur die Altersbestimmungen aus Åland, Pihlpuudas und Kalkkinen gründen sich auf mehr als eine C^{14} -Bestimmung. Bei den Bestimmungen von Åland bestehen geringe Widersprüche (Fries 1961, 1963, S. 118), aber aller Wahrscheinlichkeit nach lautet das Alter der Fichtenverhäufung auf rd. 2000 J.v.h. (Fries 1963).

Salmis C^{14} -Bestimmung aus Nurmo (Su-45) (Salmi 1963b, S. 86) konnte leider nicht verwertet werden, da die pollenchronologische Lage der Bestimmung in Ermangelung eines Pollendiagrammes nicht eindeutig gegeben worden ist.

Von den zwei Altersbestimmungen aus Pello ist Su-10 schon von Anfang an als zu alt angesehen worden (Hyypä, Hoffren, Isola 1962). In dieser Untersuchung habe ich den Wert der Kontrollbestimmung (Su-25) benutzt, der nach Hyypä zuverlässig ist und mit der in Rovaniemi ausgeführten Bestimmung (Su-17) in Einklang steht. In Ermangelung sonstiger Datierung von Fichtenverhäufung in Nordfinnland bleibt jedoch selbst das Festlegen eines summarischen Verlaufes der Isochronen in diesen Gebieten vorläufig unsicher.

Die obengenannten Beobachtungsstellen sind in Abb. 3 angegeben, in der auch die Analysenummern der besagten C^{14} -Datierungen vermerkt sind. Gemäss den Beobachtungspunkten haben in die Karte die ungefähren Isochronen eingezeichnet werden können. Die Küstengebiete, die erst nach der Fichtenverhäufung aus dem Meere aufgetaucht sind, sind bei Eintragung der Isochronen nicht berücksichtigt worden.

Da die Anzahl der Bestimmungen noch bescheiden ist und einige Datierungen vielleicht noch ungenau sind, ist es wahrscheinlich, dass sich der Verlauf der Kurven mit zunehmendem Material in gewissem Masse verändern wird, aber der von der Karte gegebene Gesamtüberblick dürfte als zutreffend angesehen werden können.

Um die Karte zu prüfen, sind auch Vergleiche mit solchen C^{14} -Bestimmungen angestellt worden, die nicht mit der Fichtenverhäufung verknüpft sind, ihr aber zeitlich nahekommen. Innerhalb der Fehlergrenzen des Verfahrens stehen sie mit den gezeichneten Isochronen in Einklang.

Aus der Karte geht hervor, dass die Fichtenverhäufung in Finnland metachron ist, gleicherweise wie in Mitteleuropa (u.a. Firbas 1949). In

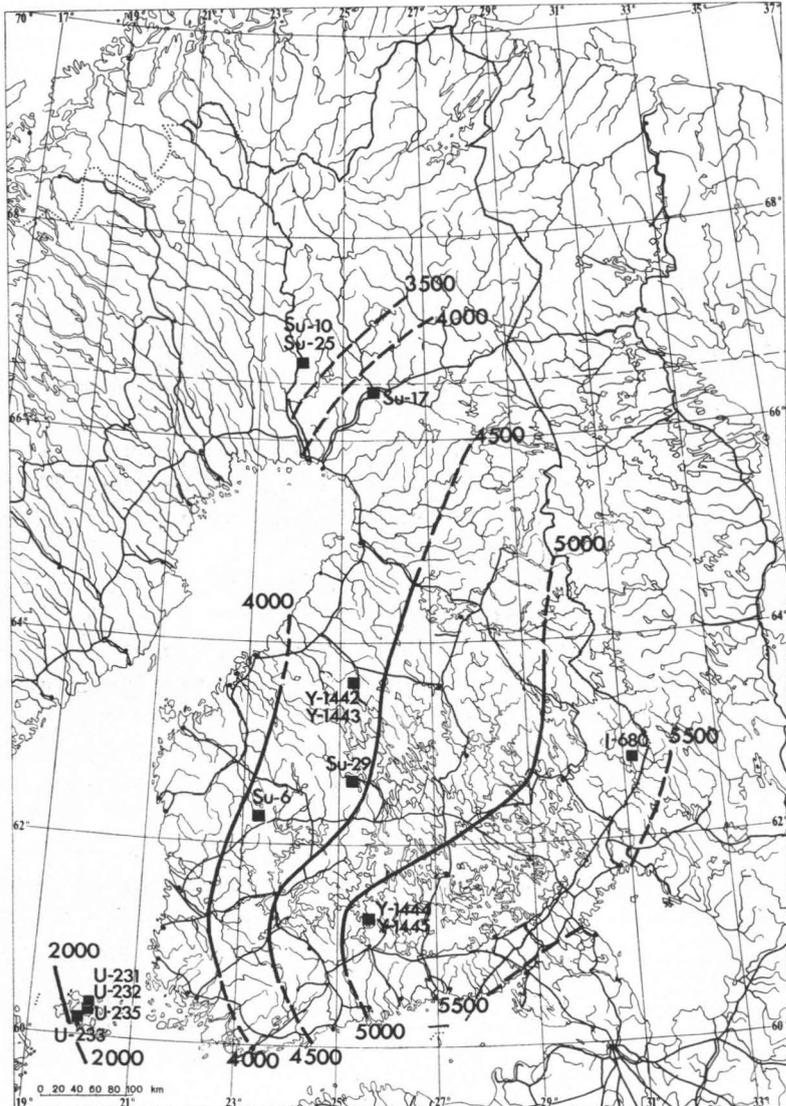


Abb. 3. Die C^{14} -Isochronen der Fichtenverhäufung mit den Analysennummern der C^{14} -Bestimmungen.

Westfinnland haben die C^{14} -Bestimmungen für die Fichtenverhäufung ungefähr das gleiche Alter ermittelt, wie Virkkala es schon 1950 für dasselbe Gebiet durch einen Vergleich von Uferschiebungen und Waldgeschichte herausgestellt hat, aber weiter östlich ist die Fichte älter.

In der Isochronenkarte treten dieselben Züge hervor wie in der von Donner über die Ankunft der Fichte dargestellten Karte (1963, S. 19), die zeigt, dass diese Holzart früher häufiger geworden ist in den Gebieten — wie in den Südteilen der Finnischen Seenplatte —, in denen sie auch zeitiger eingetroffen ist.

Schon lange hat man um die Metachronie der Einwanderung von Fichte gewusst (u.a. Auer 1928, Hyyppä 1932a, 1933, Lumiala 1939, Sauramo 1941, 1958, Donner 1957, 1963), aber ihr Erlangen von Häufigkeit in Südfinnland hat allgemein als synchron, als Folgeerscheinung einer jähen, starken Klimaveränderung gegolten. Doch hat man erkannt, dass in den östlichsten und südöstlichsten Teilen Finnlands die Fichtenverhäufung früher (u.a. Hyyppä 1932a, b, 1937, Donner 1957, Vasari 1962, Tolonen 1963) und auf Åland später (u.a. Backman 1934, Fries 1961, 1963) eingetreten ist.

Als Erklärung der vermeintlichen Synchronie der Fichtenverhäufung in Südfinnland ist vorgebracht worden, dass die Fichte schon lange vor ihrer Verhäufung im gesamten Gebiet hier und da in versprengten Vorkommen erschienen sei. Nachdem die klimatischen Verhältnissen plötzlich für Fichte günstig geworden seien, habe ihr Allgemeinwerden so gut wie gleichzeitig vor sich gehen können, ohne dass eine biologische Wanderung der Pflanze viel Zeit beansprucht hätte. Die Plötzlichkeit der Klimawandlung folgerte man hauptsächlich aus dem steilen Anstieg der Fichtenkurve in den Pollendiagrammen.

Da der Zeitpunkt der Fichtenverhäufung in den verschiedenen Gebieten durch Isochronen dargestellt werden kann und da der Zeitunterschied selbst unter Ausschluss des Schärenhofes über 2000 Jahre ausmacht, kann es sich immerhin nicht um eine jähe, durchgreifende Klimaumstellung handeln. Während sich das Klima allmählich in eine für die Fichte günstige Richtung wandelte, stellten sich die für ein häufiges Vorkommen geeigneten Verhältnisse naturgemäss zuerst in der Nähe des vorherigen Verbreitungsgebietes ein und erweiterte sich bei fortwährender Veränderung allmählich das für die Fichte günstige Klimagebiet. Ihr Vordringen folgte dem Ausdehnen des ihr zusagenden Klimaraumes. Der steile Anstieg in der Kurve der Fichte mag darauf zurückzuführen sein, dass diese als gut Schatten vertragende Art nach einem für sie vorteilhaften Klimawechsel die ihr angemessenen Standorte von Birke und Kiefer, beides mehr auf Licht angewiesene Arten, schnell zu beziehen vermocht hat.

Als ein im Verhäufigen zeitliche Unterschiede verursachender Faktor ist oft auch die von biologischer Ausbreitung beanspruchte Zeit angeführt worden. In diesem Falle hat er kaum einen grossen Einfluss ausgeübt in Anbetracht der genannten zahlreichen versprengten Fichtenvorkommen, wie es sie vor Verhäufung der Art gegeben hat. Derartige Versprengungen von Fichte bestehen auch heute noch nördlich des Raumes ihrer reichlichen

Verbreitung, u.a. in Inari-Lappland und Petsamo. Zudem weiss man, dass viele Arten aus ihrem eigentlichen Wohngebiet auch auf entlegene, ihnen zuträgliche Standorte vorstossen können, ohne dass sie in den Zwischengebieten aufträten. Derartiges ist auch bei Nadelhölzern festgestellt worden (L. Aario 1940, S. 18).

Die Beschaffenheit des in der Fichtenverhäufung sich auswirkenden Klimawandels ist noch einigermassen unklar. Die gegenwärtige Verbreitung der Fichte entspricht einem kühl kontinentalen Klima. Am ausgeprägtesten vertreten ein kaltwinterliches kontinentales Klima in Finnland die Gebiete Savukoski—Sodankylä südlich der Wasserscheide Nordlapplands (Atlas of Finland 1960, 5). Aus diesem Gebiet liegen jedoch keine Daten über die Fichtenverhäufung vor. Kaum dürfte sie sehr früh gewesen sein, da das Gebiet Grenzgegend der heutigen Verbreitung von Fichte darstellt. Ein anderer kaltwinterlicher kontinentaler Raum liegt in den Gegenden von Ilomantsi und Pielisjärvi an der Ostgrenze Finnlands. Auf dieses Gebiet bezieht sich die C¹⁴-Bestimmung I-680, die eine verhältnismässig frühe Verhäufungszeit gibt. Bedauerlicherweise liegen für nahe gelegene Gegenden keine C¹⁴-Bestimmungen vor, so dass ein Verhältnis der Isochronen zu dem Verlauf der die Kontinentalität angegebenden Isarithmen nicht erkennbar ist. Doch ist bekannt, dass in jener Gegend von ausgehender spätglazialer Zeit an Fichte verhältnismässig ständig vorkommt, obschon sich ein deutliches Verhäufigen erst in der zweiten Hälfte der postglazialen Wärmezeit vollzieht.

Die Karelische Landenge sowie die Umgebung der Salpausselkä-Landrücken in den Ostteilen Finnlands vertreten ein warm kontinentales Klima (Atlas of Finland 1960, 5). In diesem Gebiet dringen die Arten eines warm kontinentalen Klimas weit nach Westen vor (Kujala 1964). Da auch die Fichte in diesem Gebiet am schnellsten vorgestossen ist (Abb. 3), wäre es denkbar, dass sie ebenfalls eine gleichartige Kontinentalität bevorzugte. So aber verhält es sich kaum. Eher ist anzunehmen, dass die grossen Seen der Seenplatte, die in der Litorinazeit am weitesten ausgedehnt gewesen sind, das Vordringen der Fichte verzögert haben, teils als Hindernisse biologischer Ausbreitung, vor allem aber durch Beeinflussung des örtlichen Klimas.

Nach Enqvist (1933, S. 150) erfordert die Fichte eine hinreichend lange winterliche Frostzeit. Dies erklärt wohl wenigstens zum Teil den Verlauf der Fichtenverhäufung in Finnland. In der warmen Litorinazeit waren in den jenseits der Ostgrenze Finnlands gelegenen kontinentalen Räumen die Winter kalt genug, aber auf der finnischen Seite im allgemeinen nicht. Während die Temperatur langsam sank und der Temperaturunterschied zwischen Sommer und Winter möglicherweise zunahm, stellte sich in stets umfangreicheren Gebieten eine genügend lange und hinreichend strenge Frostperiode ein. Im maritimen südwestlichen Schärenhof wurde dies erst um die Zeitwende erreicht.

Nach H. J. und S. V. Katz (1953) hat die Fichte erst häufig werden können, als im Boden ein gewisser Auslaugungsgrad erreicht worden war. In Finnland hat dies gleichwohl kaum eine grosse Rolle gespielt, da sich die Fichte auch gegenwärtig schnell auf unausgewachsenen, erst aus dem Meere aufgestiegenen Boden ausgebreitet hat.

DIE ALTERSVERHÄLTNISSE DER FINNISCHEN POLLENZONEN

Da die Fichtenverhäufung, die als eines der wichtigsten und in Südfinnland am sichersten synchronen Leitniveaus der Pollendiagramme angesehen worden ist, sich nun als metachron erwiesen hat, gibt dies Anlass dazu, auch die übrigen Anhaltspunkte der finnischen Pollendiagramme und die Grundlagen ihrer Konnektierung zu prüfen.

In der von Sauramo durchgeführten Zoneneinteilung der süd- und westfinnischen Pollendiagramme ist jede der neun mit waldgeschichtlicher Begründung unterschiedenen Zonen als Entsprechung einer bestimmten klimahistorischen Periode oder eines ihrer Teile festgelegt (u.a. 1949). Als Zonengrenzen sind grossenteils die gleichen in den Diagrammen hervortretenden Züge wie bei den mitteleuropäischen Zoneneinteilungen benutzt (u.a. Jessen 1935, Firbas 1949). Da die Auffassung von der Synchronie der klimahistorischen Zeiträume vorherrschend gewesen war, verband sich mit Sauramos Zoneneinteilung die Auffassung von der Gleichaltrigkeit der als Zonengrenzen benutzten Anhaltspunkte im grössten Teil Finnlands und in Mitteleuropa. Diese Vorstellung äussert sich nicht nur in der Fassung der Zonengliederung, sondern auch in ihren verschiedenen Anwendungen und in den Versuchen, die Pollenchronologie an die gegenwärtige nach Jahren zählende Zeitrechnung zu binden. Sauramos Auffassung von der Gleichaltrigkeit gleichartiger Pollenzonen gründet sich hauptsächlich darauf, dass die Veränderungen der Vegetation in hohem Masse Spiegelungen synchroner Klimawandlungen und somit auch selbst synchron wären. Die auf die Ostseestadien gegründete Landhebungschronologie, die Tonchronologie und zum Teil vielleicht auch die archäologische Chronologie hätten Kontrollmöglichkeiten zu bieten vermocht, aber sie sind in Finnland nicht hinreichend ausgewertet worden.

Sauramos Zoneneinteilung hat danach bei der Bearbeitung der finnischen Pollendiagramme eine umfassende Anwendung gefunden. Doch ist Sauramos Umreissung der Pollenzonen nicht immer genau befolgt worden, wodurch etwas Verwirrung angerichtet worden ist.

Theoretisch betrachtet wird bei einer Klimaveränderung ein gewisser in Vorhandensein und Wettbewerbsfähigkeit einer Holzart bedeutsamer Schwellenwert nur auf gleicher Isarithme des betreffenden Klimafaktors

gleichzeitig überschritten. In der senkrecht gegen diese Zone verlaufenden Richtung wird der Schwellenwert ungleichzeitig erreicht. Somit ist auch die entsprechende Wandlung in den Pollendiagrammen nur in der Richtung der betreffenden Isarithme zeitgleich, sonst metachron. Da beim Klima gleichzeitig in mehr als einem Faktor Veränderungen vor sich gehen können und da auch die edaphischen Faktoren die Entwicklung beeinflussen, wird ebenfalls die Entwicklung der Wälder entsprechend komplizierter. Doch ist es klar, dass bei andauernd gleich ausgerichteter Klimaveränderung alle durch sie bedingten Grenzen metachron sein müssen. Je langsamer die Klimawandlung abläuft, um so schmaler ist das Gebiet, in dem die ihr entsprechende Grenzstelle in den Pollendiagrammen im grossen ganzen als synchron erachtet werden kann.

Sauramos Auffassung, die Pollenzonen seien auch in weit voneinander entfernt gelegenen Gegenden gleich alt (1949, 1958), setzt also voraus, dass die klimatischen Veränderungen sehr schnell und stark gewesen wären, wodurch sich auch eine gleichmässige Vegetation so gut wie gleichzeitig in weite Räume hätte ausbreiten können. In erster Linie auf derartiges weisen die oft scharfen Grenzen der Moorbodenarten sowie die in den Pollendiagrammen vielfach zu erkennenden ziemlich jähen Veränderungen hin. Doch sind diese bei weitem nicht immer auf die Plötzlichkeit eines Klimawechsels zurückzuführen (S. 221). Obgleich die Schnelligkeit der Wandlung in einigen Fällen hat beträchtlich sein können (S. 227), hat sich die von Sauramos Konnektionen vorausgesetzte explosionsartige Geschwindigkeit (Sauramo 1947) doch nicht hinreichend begründen lassen. Dagegen liegen zahlreiche Beobachtungen vor, die erweisen, dass eine die Waldgrenze oder auch die Vorkommensgrenze einer bestimmten Art bezeichnende Stelle in den Pollendiagrammen nicht immer in verhältnismässig nahe beieinander gelegenen Gebieten gleichaltrig zu sein braucht (u.a. L. Aario 1943, 1944 a, b, Faegri & Iversen 1964, R. Aario 1965 a). Auch wären explosionsartige Klimaveränderungen geophysikalisch schwer zu erklären.

Solche in der Zusammensetzung der Wälder vor sich gegangenen Veränderungen, die auf eine in gleicher Richtung fortgesetzte Klimaentwicklung zurückzuführen sind, können also nur in beschränkten Gebieten als ungefähr synchron angesehen werden. Der in einem bestimmten walddeschichtlichen Vorgang bestehende Altersunterschied in verschiedenen Gebieten ist in jedem Falle gesondert festzulegen, und wenn sichere Altersbestimmungen in hinreichender Menge vorliegen, kann eine Isochronenkarte ausgearbeitet werden. Dann kann auch die Geschwindigkeit der Entwicklung festgestellt werden.

Tritt in der Klimaentwicklung eine Wendung ein und bewirkt diese Kulminationen im Verlauf der Pollenkurven, so können derartige Maxima und Minima im Prinzip auch in ausgedehnten Gebieten zum mindesten in grossen Zügen als synchron angesehen werden (u.a. Firbas 1954). Praktisch sind

solche in den Pollendiagrammen jedoch oft schwer festzustellen, so dass auch mit ihrer Hilfe in sehr umfangreichen Gebieten im allgemeinen kaum sichere Konnektionen vorgenommen werden können.

Wendet man den Rhythmus des Wechsels der Warm-Kalt-Elemente beim Konnektieren der Diagramme an, so hat man sich ausserdem daran zu erinnern, dass ein und derselbe klimatische Vorgang in den Pollendiagrammen verschiedener Gebiete auf unterschiedliche Weise hervortreten kann (u.a. L. Aario 1939, 1940, 1943, 1944 a, b, Faegri & Iversen 1964, S. 111), ein Umstand, der nicht immer in hinreichendem Masse berücksichtigt worden ist. Einige Vorgänge mögen in gewissen Gebieten sogar ganz ohne Widerspiegelung in den Pollendiagrammen bleiben. Ein und derselbe Temperaturanstieg, der in der Birkenzone nahe dem Gebiet der Kiefernwälder eine starke Kiefernverhäufung bewirkt, verursacht weder eine Wandlung in den Pollenverhältnissen im zentralen Kiefernraum noch eine deutlich auszulegende Veränderung in weit von der Waldgrenze abgelegener Tundra. Desgleichen wirken Temperatursenkungen, die im Grenzgebiet deutlich hervortreten, nicht gleicherweise in den zentralen Gebieten derselben Holzart. Die Maxima und Minima des in den Diagrammen der verschiedenen Gebiete zu erkennenden gleichartigen Warm-Kalt-Rhythmus können infolgedessen teils synchron, teils von verschiedenem Alter sein. Dies betrifft insbesondere die Schwankungen von kurzer Dauer, aber bei Ausführung von Konnektionen weit voneinander entfernter Gebiete ist diese Möglichkeit auch bei der Deutung des in den Diagrammen hervortretenden Grossrhythmus zu berücksichtigen.

Im folgenden suche ich im Lichte der obendargestellten Prinzipien die finnischen Pollenzonen und ihre Altersverhältnisse zu betrachten.

Abb. 4 gibt eine schematische Darstellung der Pollenverhältnisse in einigen der ältesten in Finnland untersuchten Wassersedimente.

Da die Pollen krautartiger Pflanzen nicht dazu neigen, sich als Fernflug anzureichern und auch nicht reichlich im Sekundärpollengehalt vorkommen (Iversen 1936), vertreten ihre in der schematischen Darstellung wiedergegebenen Maxima offenbar eine Tundravegetation nahe gelegener Gebiete¹⁾. Die in geringer Menge vorhandenen Pollen des Eichenmischwaldes sind Sekundärpollen (Iversen 1936) oder durch Fernflug verfrachtet.

Den zwischen den NBP-reichen Zonen lagernden Teil kennzeichnen Spärlichkeit der NBP, ein Baumpollengehalt mit oft vorherrschender Kiefer sowie eine relative Reichlichkeit von Pollen des Eichenmischwaldes. Dies ist früher als warme Zwischenzeit ausgelegt und oft mit der Allerödzeit identifiziert worden. Wird aber die Spärlichkeit der NBP auf örtlichen Kiefern-Birken-Mischwald zurückgeführt, so könnte ein EMW weder als

¹⁾ Die Möglichkeit, die NBP-Gipfel bei ihrer Deutung auf Ufernähe zurückzuführen, ist in jedem Einzelfalle noch gesondert zu besprechen.

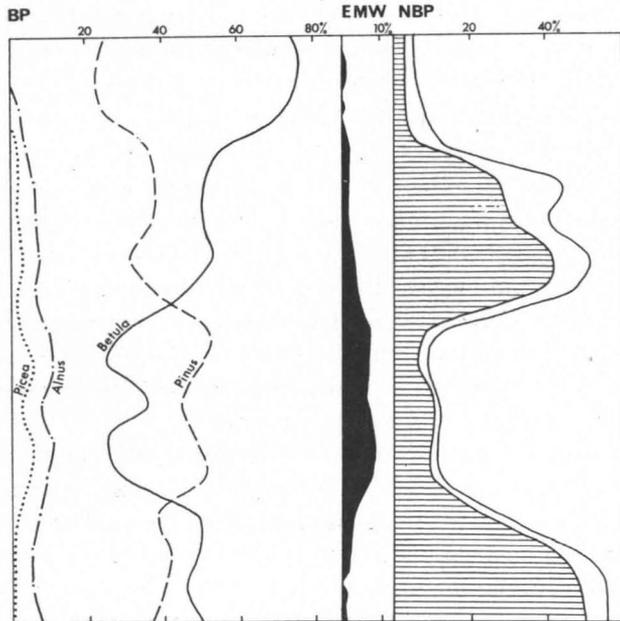


Abb. 4. Schematische Darstellung der Pollenverhältnisse in finnischen spätglazialen minerogenen Sedimenten.

Sekundärpollengehalt noch als Fernflug einen so hohen Prozentsatz erreichen, sondern müsste durch örtliche Reichlichkeit von Bäumen des EMW in Wäldern bedingt sein. Dies ist jedoch nicht möglich, da selbst in Norddeutschland in der Allerödzeit solche Laubbäume nicht aufgetreten sind.

Übrig sind drei Möglichkeiten. Die besagte Zwischenzeit kann eine kältere Zeit als die ihr vorausgegangene und die auf sie folgende vertreten, wodurch auch die örtliche Kräutervegetation dann geschwächt und die Menge der Kümmerbirke (*Betula nana*) vermindert worden sowie etwa früher spärlich dagewesene Fjeldbirke (*Betula tortuosa*) geschwunden wäre. Dabei hätte der Anteil von Sekundärpollen und Fernflugpollen zugenommen und ein Vermehren von Kiefer und Bäumen des EMW bewirkt. Zu demselben Ergebnis könnte ein Ansteigen des Meeresspiegels führen, durch das zahlreiche Inseln überflutet worden wären und der Anteil der Pflanzendecke der Nahumgebung sich somit vermindert hätte. Eine dritte Möglichkeit wäre, dass die Sedimentationsgeschwindigkeit zugenommen hätte, was auf einer Temperatursteigerung und einem ihr entsprechenden schnelleren Abschmelzen des Inlandeises hätte beruhen können. Infolge schneller Sedimentation wäre der Anteil des aufs neue abgesetzten Sekundärpollengehaltes gewachsen und der der Pollenflora naher Umgebung entsprechend vermindert worden.

Da viele Deutungsmöglichkeiten bestehen, können diese ältesten Pollenzonen Südfinnlands vorläufig dem andernorts festgestellten Warm-Kalt-Rhythmus nicht zugeordnet werden.

Die NBP-Verminderung und das Zunehmen der Birkenpollen im oberen Teil des Diagrammes lassen das Eintreffen von Birkenwäldern im Gebiet erkennen. Indem Donner die Lage dieser Grenze (III/IV) mit Ufersystemen, die anhand von Bändertonen datiert sind, vergleicht (1964 S. 34), erhält er als Alter des Niveaus in Südfinnland rd. 9800 J.v.h., was 200—300 J. weniger ist als der in Mitteleuropa durch Radiocarbon-Untersuchungen herausgestellte Wert (Godwin 1960, T. Nilsson 1964). In Südfinnland hat man C^{14} -Bestimmungen für die Zone IV aus Pölläkkälä auf der Karelischen Landenge (Su-8, 9850 \pm 300 J.v.h.), aus Vävarsbacka an der Küste des finnischen Meerbusens bei Porvoo (Su-13, 9600 \pm 350 J.v.h.) und aus Kihniö im nördlichen Satakunta (Su-15, 9850 \pm 320 J.v.h.) (Hyypä, Hoffren, Isola 1962, 1963, Salmi 1962, Hyypä 1963). Die Grenze III/IV muss also älter als 9850 J. sein und somit von ungefähr gleichem Alter wie in Mitteleuropa. Zwischen so weit voneinander entfernt gelegenen Gebieten wäre jedoch ein Altersunterschied zu erwarten gewesen. Die Gleichaltrigkeit scheint auf eine verhältnismässig schnelle Klimaveränderung hinzuweisen. In jener Zeit war das Inlandeis gewiss in schneller Verminderung begriffen. Auf der anderen Seite ist zu berücksichtigen, dass das Inlandeis die Richtung der Isothermen beeinflusst hat, so dass der Temperaturunterschied zwischen Norddeutschland, Dänemark und Südfinnland kaum ebenso gross wie gegenwärtig gewesen sein dürfte (L. Aario 1944 b, S. 697).

Der nächste in den finnischen Pollendiagrammen hervortretende deutliche Anhaltspunkt ist die Grenze zwischen der subpolaren Birkenzone (IV) und der auf sie folgenden Kiefernzzone (V). Da diese Grenze einen Wechsel der in den Wäldern vorherrschenden Holzart bezeichnet, ist auf Grund des Obigen zu erwarten, dass auch er metachron sei. L. Aario hat denn auch eine Verjüngung dieser Grenze in süd-nördlicher Richtung festgestellt (1939, 1943, 1944 a, b), wie später auch Hyypä und Sorsa auf Grund der Radiocarbon-Bestimmungen (Hyypä, Hoffren, Isola 1962, 1963, Sorsa 1964).

Auch sind das Enden der obengenannten »borealen« kiefernreichen Zeit und das spätere, oft zur Zeit der Fichtenverhäufung vor sich gehende Vermehren von Kiefer sogar auch in verhältnismässig nahe beieinander gelegenen Gebieten als ungleichaltrig erwiesen worden (u.a. R. Aario 1965 a, S. 184).

Erlenverhäufung (V/VI), Fichtenverhäufung (VIII/IX) sowie Anfang und Ende des Vorkommens von Bäumen des Eichenmischwaldes (VI/VII, VIII/IX) bezeugen Verschiebungen der Grenzen einzelner Pflanzenarten, und somit sind auch sie von metachronem Gepräge (u.a. Firbas 1954). Die Fichtenverhäufung ist bereits oben besprochen worden. Die Erlenverhäu-

figung habe ich zuvor (1965 a) u.a. durch C^{14} -Bestimmungen in den verschiedenen Teilen des Vor-Päijänne-Beckens als ungleichaltrig festgestellt. Zuvor hatte man auch die Erlenverhäufung mit gleicher Begründung wie die Fichtenverhäufung allgemein als verhältnismässig genau synchron in einem grossen Teil Finnlands angesehen, und sie war in den Untersuchungen über Niveauverschiebungen einer der wichtigsten Anhaltspunkte gewesen (u.a. Sauramo 1954). Da Erle bestimmte Standorte, insbesondere Ufer, bevorzugt und da es sich bei ihr um zwei ökologisch verschiedene Arten (*Alnus incana* und *glutinosa*) handelt, von denen jede in verschiedenen Gegenden zuerst als sich verhäufigende Art aufgetreten ist, wie u.a. Vasari festgestellt hat (1962, S. 110), sind verhältnismässig grosse, sogar ganz örtliche Altersunterschiede möglich. Nach Fromm (1963, S. 52) wechselt der Zeitpunkt der Erlenverhäufung auch in den verschiedenen Teilen Schwedens.

Die gegen Norden zunehmende Verzögerung im Auftreten edler Laubbäume und im Beginn ihrer zusammenhängenden Kurven ist in vielen Untersuchungen nachgewiesen worden. Verfasser hat auch schon im Bereich des Vor-Päijänne-Beckens eine deutliche Ungleichaltrigkeit feststellen können (1965a, S. 185). Das Aufhören des Vorkommens von Eichenmischwald liegt im allgemeinen in den nördlichen Teilen früher als in den südlichen. Auch dieser natürliche Umstand ist in vielen finnischen Diagrammen zu erkennen. Für Westeuropa und die südlichen Teile Skandinaviens ist das durchschnittliche Verjüngen des Ulmenfalles in nord-südlicher Richtung bekannt (Nilsson 1964). In den hainwaldreichen Gegenden sind Bäume des EMW wahrscheinlich früher häufig geworden und am längsten heimisch gewesen. Die Verringerung im Betrage von EMW ist neben der Fichtenverhäufung als Indikator des Einsetzens von Zone IX angesehen worden. Da sie oft miteinander in Widerspruch stehen, ist bald der einen, bald der anderen der Vorrang beigelegt worden, was leicht zur Willkür führt. Wie aus dem Obigen hervorgeht, kann beides nicht als synchron erachtet werden.

Da Finnland in der Grenzgegend des Eichenmischwaldes liegt, sollte sein Frequenzmaximum theoretisch im ganzen Land einen so gut wie synchronen Punkt darstellen. Doch wechselt die Frequenz dieser Holzarten in den verschiedenen Diagrammen verhältnismässig beträchtlich und ist selbst das Maximum unbestimmt und schwach. Dies mag zur Hauptsache daran liegen, dass die edaphischen Faktoren in recht reichlicher Masse auf ihr Vorkommen einwirken. Daher hat dieses Maximum nicht als genauer chronologischer Anhaltspunkt benutzt werden können.

SCHLUSSFOLGERUNGEN

Auf Grund der obigen Betrachtung ist festzustellen, dass alle in den finnischen Pollendiagrammen als Zonengrenzen benutzten Leitniveaus schon

ihrer Natur nach metachron sind und auch in Wirklichkeit als solche nachgewiesen werden können. Der Zeitpunkt der Fichtenverhäufung wird in ost-westlicher Richtung allgemein um rd. 300—400 J./100 km jünger (S. 220, Abb. 3). Über die Metachronie der übrigen in den Pollendiagrammen auftretenden Anhaltspunkte liegen nicht ebenso genaue Angaben vor, aber sogar bedeutende regelmässige Altersunterschiede haben bereits erkannt werden können. Wie gross die Gebiete sind, in denen die einzelnen Zonengrenzen der Pollendiagramme als so gut wie synchron angesehen werden können, richtet sich nach der Verschiebungsgeschwindigkeit der ihnen entsprechenden Grenzen der Vegetationszonen sowie nach der vom Anwendungszweck geforderten Genauigkeit der Synchronie. Bevor die Pollenzonen zuverlässig für Altersbestimmungen benutzt werden können, ist zunächst das Alter der Grenzstellen der Zonen über ganz Finnland für so viele Beobachtungsstellen zu bestimmen, dass dadurch das Zeichnen einer Isochronenkarte ermöglicht wird. Dadurch werden sich auch die Altersdaten der mit der entsprechenden Ziffer Mitteleuropas vermerkten Zonen vergleichen lassen.

Nachdem die Vorstellung von der Synchronie der Pollenzonen aufzugeben ist, besteht Anlass, das Verhältnis der Pollenzonen zu den Klimazeiten und deren Verhältnis zu der absoluten Chronologie genauer festzulegen. Viele Forscher erachten z.B. Präboreal, Boreal, Atlantikum, Subboreal und Subatlantikum als überall gleichaltrig, andere aber binden diese klimahistorischen Perioden an Pollenzonen und halten sie somit für ungleichaltrig. Nach der erstgegebenen Fassung wäre es u.a. möglich, dieselben Perioden auch in denjenigen Räumen zu unterscheiden, in denen die in Mittel- und Nordeuropa allgemein benutzte Pollenzoneneinteilung nicht angewandt werden kann, wie z.B. in der gegenwärtigen Waldgrenzgegend im Norden. Demgegenüber ist man ziemlich allgemein daran gewöhnt, die obengenannten Blytt-Sernanderschen Zeiträume mit der Waldgeschichte und also mit den Pollenzonen zu verbinden, und auf diese Bindung ist nicht leicht zu verzichten. Die Grenzen der auf diese Weise bestimmten, nicht synchronen Zeiträume sind im allgemeinen in den Pollendiagrammen verhältnismässig leicht festzustellen. Bei weitem nicht immer ebenso leicht wird die Gliederung der Diagramme, wenn man an der Gleichaltrigkeit der Grenzen von Klimaperioden in verschiedenen Gebieten festhält.

Die Hauptsache ist es, das gegenseitige Verhältnis von Pollenzonen, Klimaperioden und absoluter Chronologie nunmehr genauer festzulegen und dadurch eine Grundlage für einen einheitlichen Gebrauch dieser Begriffe zu schaffen. Meinerseits wäre ich aus mehr als einem Grunde geneigt, synchronen, von den Pollenzonen unabhängigen Klimaperioden den Vorzug zu geben. Sollten die Blytt-Sernanderschen Perioden nicht dazu geeignet sein, so liessen sich in den Ländern um die Ostsee in nunmehr höherem Masse die Pollendiagramme mit den Ostseestadien verbinden.

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