Geological Survey of Finland

Bulletin 314

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by Seppo I. Lahti

Geologinen tutkimuslaitos
Espoo 1981

Geological Survey of Finland, Bulletin 314

ON THE GRANITIC PEGMATITES OF THE ERÄJÄRVI AREA IN ORIVESI, SOUTHERN FINLAND

by

SEPPO I. LAHTI

with 23 figures and 23 tables in the text and 3 appendices

GEOLOGINEN TUTKIMUSLAITOS
ESPOO 1981

Lahti, Seppo I. 1981. On the granitic pegmatites of the Eräjärvi area in Orivesi, southern Finland. *Geological Survey of Finland, Bulletin 314.* 82 pages, 23 figures, 23 tables and 3 appendices.

The pegmatites of the Eräjärvi area (Orivesi, Finland) are middle Proterozoic granitic dyke rocks. The pegmatite deposits seem to be associated genetically with the late orogenic granites occurring in the research area. Most of the dykes are met with in schists near or at the contacts of granite plutons. The pegmatite dykes and lenses are generally parallel to the schistosity, although perpendicular and diagonal deposits are also common.

The pegmatites are nearly homogenic, zoned or banded rocks. Altogether 61 complex pegmatites have been met with in the area. Of these, 2/3 contain tourmaline, beryl, columbite—tantalite or tapiolite as characteristic accessory minerals; the remaining 1/3 include various Li silicates and especially Li phosphates as well. The number of minerals in the pegmatites is about 80. The author has identified 29 different phosphate minerals including many alteration and replacement products of triphylite and lithiophilite. A new monoclinic phosphate $(Na, K)_{0.92} (Ca, Mn, Fe, Mg)_{1.04} Al_{1.07} P_{0.97} O_{4.11} (F, OH)_{2.90}$ (space group $P2_1$ or $P2_1/m$, $a = 6.83$ Å, $b = 7.14$ Å, $c = 5.45$ Å, $\beta = 109°22'$) named viitaniemiite in allusion to the locality (Viitaniemi), is described in this study. Also described in detail are pink muscovite, bertrandite, eosphorite, beryllonite, wodginite, bismuthian antimony and Li-Fe muscovite; the chemical analyses, principal physical properties and unit cell dimensions are given. In addition, several columbite—tantalites, some K-feldspars and plagioclases were analysed from the pegmatite deposits.

Key words: pegmatite, minerals, phosphates, viitaniemiite, new minerals, silicates, elements, Eräjärvi, Orivesi, Finland

Seppo I. Lahti, Geological Survey of Finland,
SF-02150 Espoo 15

ISBN 951-690-137-9
ISSN 0367-522X

Vammala 1981, Vammalan Kirjapaino Oy

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INTRODUCTION

The Eräjärvi pegmatite area is located in Orivesi, southern Finland about 50 km ENE of Tampere. Observations on the granite pegmatites in the area were reported back in the 1890's in association with field studies (Andersson 1891; Frosterus 1903) connected with work on the geological map sheet of Mikkeli (1 : 400 000; Frosterus 1900). The deposits became better known in the 1910's and 1920's when quartz and later mainly potassium feldspar were extracted.

The first pegmatite deposit of economic importance in the area, the Juurakko pegmatite, was reported to be found by A. Pajala in 1916 (Hirsjärvi 1933; Luoto 1967). Several other pegmatites, some of them sizeable, were later discovered and mined from time to time. The companies mainly concerned were Uiherlan Sälpä Oy and Suomen Mineraali — Paraisten Kalkki Oy. The most important deposit in the area is the Viitaniemi pegmatite which was discovered in 1935 and mined without interruption until the mid-1960's. The Viitaniemi and also the big Juurakko deposit have yielded several tens of thousands of tons of potassium feldspar. The Mattila and Eräpyhä pegmatites have produced a few thousands of tons; the others conciderably less (Volborth 1954c; Virkkunen 1964). Altogether some 30 pegmatite deposits have been mined in the area. In addition to potassium feldspar, they have produced notable amounts of quartz, and some beryl, muscovite and columbite—tantalite. The Viitaniemi deposit has yielded montebrasite and pollucite as well. All the products were separated by hand picking during mining and sold as lump ore.

The Viitaniemi pegmatite is characterized by the exceptional abundance of phosphate minerals containing lithium, manganese, iron and beryllium. The deposit is the type-locality of the Mn-Be phosphate väyrynenite (Volborth 1954a; Mrose and Knorring 1959; Mrose and Appleman 1962), and of viitaniemiite described in this paper. Volborth (1951, 1954a-c, 1956, 1960) has published several papers on the minerals in the Viitaniemi pegmatite. In contrast, there are few studies on the mineralogy of the other pegmatites in the Eräjärvi area. Vesasalo mapped the pegmatite deposits in terms of economic geology in the 1950s. The results were compiled and posthumously published by Virkkunen in 1964. Their research reports include a geological map of the southern part of the predominant dyke swarms. The bedrock of the study area is also partly shown on the Kangasala map sheet of pre-Quaternary rocks (1 : 100 000) drawn up by Matisto (1964). The explanation to that map sheet (Matisto 1976) includes a review of the granite pegmatites in the area.

The present author encountered several rare mineral pegmatites during geological mapping on a scale of 1 : 100 000 conducted in the northern part of the area, within the map sheet 2142 Orivesi. The mineralogy of the pegmatites in the Eräjärvi area was reviewed in my master's thesis (Lahti 1974, see also Lahti 1975). Since then, in the course of detail studies, I have found a number of previously unknown pegmatites that contain rare minerals within the areas of the map sheets 2141 Kangasala and 2142 Orivesi.

The pegmatites in the Eräjärvi area are characterized by various rare accessory minerals containing boron, beryllium, niobium, tantalum, phosphorus, lithium, tin and manganese. In this study pegmatites are called »simple» if they contain common rock-forming minerals, tourmaline and sporadic beryl; the term »complex» is reserved for the varieties that contain columbite—tantalite, cassiterite, lithium phosphates and lithium silicates as well. Most of the complex pegmatites in the area contain only tourmaline, beryl and niobium-tantalum minerals (generally columbite—tantalite) as accessory minerals. The present author calls these deposits beryl—columbite pegmatites (or Be-Nb pegmatites) to distinguish them from the lithium pegmatites (Li pegmatites) that, apart from the above minerals, also contain various Li silicates or Li phosphates, or both. The abundance of rare minerals varies between and within the occurrences to such an extent that the classification is only approximate; it is not possible to make a sharp distinction between the three types.

In regional studies on granite pegmatites the deposits are often classified by origin as magmatic or metamorphic. According to Schneiderhöhn (1961), the latter are usually pegmatites associated with migmatites and resulting in metamorphic differentiation processes, whereas the former include deposits crystallized from pegmatitic magma at higher levels in the Earth's crust. The pegmatites in the Eräjärvi area, which occur mainly as dykes and lenses in various mica-rich schists and granites, often show sharp and straight contacts with the wall rocks. They are magmatic rather than metamorphic pegmatites in origin. Pegmatitic dykes and bodies associated with migmatitic mica schists and gneisses occur south of the appended map (Appendix 3) and in the southeastern corner of the study area. They are, however, outside the complex pegmatite area proper and are not discussed in the present study.

All the known deposits that contain rare minerals and as many as possible of the simple pegmatites are marked on the appended geological map of the area (Appendix 3). The complex pegmatite deposits are numbered from left to right and denoted by overprints. Usually, when a deposit is referred to in the text, the numbers on the map are given in brackets after the name of the deposit. During geological mapping the environments of the Mikkola (20), Seppälänranta (31) and Niemelä (54) pegmatites were submitted to detailed studies, whereas elsewhere only a fraction of the outcrops were visited.

The present study contains a brief description of the geological structure of the Eräjärvi area, the mode of occurrence of the pegmatites and their relation to tectonics as well as of the internal structures of the occurrences; the emphasis, however, is on the mineralogy of the complex pegmatites. Hence, the study area was outlined so that all the occurrences known to contain rare minerals fall within the area of the appended geological map. All the minerals discovered in the complex pegmatites in the area are listed in Tables 2 and 3a—c. Efforts were made to make as complete a list as possible of pegmatites that have been mined; mineralogical observations on the other pegmatites were made on samples from the surface of the rock, usually during two visits to the locality.

LABORATORY METHODS

The mineralogical laboratory studies were conducted at the Geological Survey of Finland in Espoo. The minerals were identified by a Philips X-ray diffractometer and Debye-Scherrer cameras (∅ 57.3 and 114.6 mm). The single crystal investigations were performed on a Buerger precession camera. The intensities of the reflections of the X-ray diffractograms were measured directly from the peak heights on the chart. The triclinicity ($\triangle$) of the potassium feldspars was always determined from the reflections (131) and ($1\bar{3}1$) of the powder pattern by using the equation of Goldsmith and Laves (1954a and b) $\triangle = 12.5 \times (d(131) - d(1\bar{3}1))$.

The refractive indices of the minerals were measured by the immersion method using gelatine-coated microscope slides. The measurements were performed at ambient temperature (23°—24°C) in Na light. The refractive indices of the immersion liquids were invariably measured on an Abbe refractometer. The specific gravity of the heaviest minerals were determined on a Berman balance; all the others were determined by heavy liquids and on a Westphal balance. The symbol D in the tables refers to the calculated density and G to the measured specific gravity.

In addition to the X-ray and optical studies the composition of many rare or otherwise mutually indistinguishable minerals were assayed by T. Hautala at the Geological Survey of Finland with the aid of a Geoscan microanalyser using natural, wet chemically analysed samples as standards. The determinations were made from polished samples by assaying at least three adjacent points within a given part of a mineral grain. The X-ray fluorescence and most of the wet chemical analyses for this study were made in the Chemistry Department of the Geological Survey. The majority of the computations referring to the chemical composition of the minerals were performed by an HP 3000 computer at the Geological Survey using the ROCK program reported by Saltikoff, Honkamo and Kallio (1976). The X-ray crystallographical computations were carried out by the unpublished programs written by Pekka Kallio. The wavelengths of the X-rays used in the crystallographical computations were obtained from the book »International tables for X-ray Crystallography» (Ibers and Hamilton 1974).

REGIONAL GEOLOGY

Lithology

The study area (Appendices 2 and 3) is located in the eastern part of the Tampere schist belt, which is part of the Precambrian Svecokarelidic schist formation. The bedrock is characterized by various bedded or graded-bedded mica schists, phyllites and greywacke schists (Fig. 1) that underwent folding and metamorphism during the Svecokarelidic orogeny 1800 to 1950 million years ago (Kouvo and Tilton 1966; Simonen 1971). The micaceous pelitic schists reveal narrow intercalations of quartz–feldspar schist, sericite quartzite, black schist, biotite–plagioclase gneiss and conglomerate schist. In the northern part of the study area there are also narrow zones of metavolcanics that vary in composition, being in general ultramafic or intermediate metamorphic pyroclastic rocks, lavas

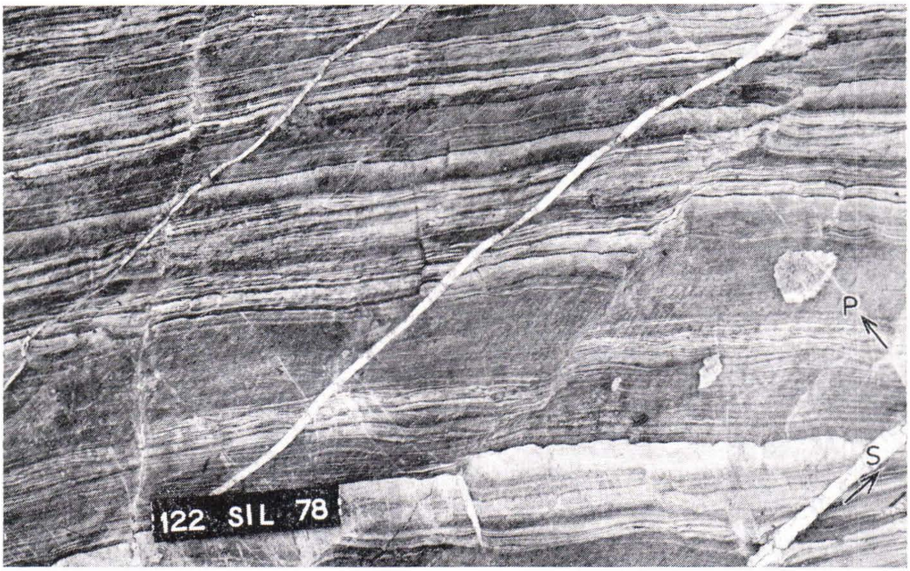


Fig. 1. Bedded mica schist intersected by quartz and quartz-feldspar veins (s, arrow) parallel to schistosity. Note also the very thin quartz veins (p, arrow) perpendicular to schistosity. The outcrop is near the Attilanvuori pegmatite (37). Length of scale tag 12 cm.

and sills. The majority of the metavolcanics are plagioclase porphyrites, although amphibolite, uralite porphyrite and uralite—plagioclase porphyrite are also encountered locally.

The schists in the area were metamorphosed under the conditions of low pressure amphibolite facies (cf. Seitsaari 1954), and in places the mica schists and phyllites show porphyroblasts of andalusite, garnet and staurolite. The mica schists, greywacke schists and phyllites in the northern part of the study area often exhibit distinct primary structures, whereas the micaceous schists in the south and southeast are usually homogenous and often intensely granitized.

A series of plutonic rocks, ranging in composition from gabbros and diorites to granitic rocks, intruded the schists in association with regional metamorphism and folding. In general the plutons are elongated parallel to the schistosity and vary in size. The largest plutons are heterogenous, often containing several differentiates (Matisto 1964, 1976, 1977). In the central part of the Eräjärvi area the plutonic rocks form an extensive complex composed of silicic and ultramafic intrusions. The sizeable ellipsoidal diorite area in the middle of the appended map (Appendix 3) is fairly homogenous and has quartz-rich marginal variants only at its eastern end. In the diorite pluton located in the western part of the study area, just at the left-hand margin of the appended map, the mineral composition varies considerably. Andesine, hornblende and biotite are the major minerals in diorite, but the rock grades locally into a dark hornblende—augite gabbro. The diorite—gabbro pluton is surrounded by an arch-shaped quartz diorite that grades further into microcline granite (the Rappuvuori granite intrusion). The contact between quartz diorite and granite is covered by overburden, and hence the crosscutting and age relations of the rocks have not been very well established.

The elliptic NE-trending Rappuvuori granite intrusion between the diorite—gabbro areas in the western part of the map is of homogenous and medium-grained microcline granite. The main minerals are microcline ($\triangle = 0.9$), plagioclase (An_{16-26}), quartz and biotite. In places the granite shows narrow pegmatite dykes or swarms of dykes. The largest and mineralogically most interesting pegmatite deposits are, however, in or near the contacts of the granite intrusion.

The Kultavuori pegmatite granite area between Eräjärvi and Enonselkä, Längelmävesi, is very heterogenous. The western part of the area is occupied by muscovite pegmatite and pegmatite granite, whereas the middle of the pluton is dominated by a distinctly oriented two-mica pegmatite granite. In some places it contains coarse-grained pegmatite lenses that show gradual contacts with the wall rock and in others pegmatite dykes with sharp contacts. In a few outcrops in the middle of the pluton the rock is medium-grained microcline granite. The part of the Kultavuori intrusion that extends to the east of Eräjärvi is of the same kind of grey microcline granite as the Rappuvuori intrusion.

Mica gneisses and veined gneisses with local coarse pegmatitic veins, dykes and lenses predominate in the southeastern and southern parts of the study area. The contacts of the larger pegmatite granite lenses with the enveloping schists are often indistinct, and the wall rock is intensely migmatized. The pegmatites of the veined gneisses are simple in mineralogy. They may include as accessory minerals schorl, apatite, zircon and sometimes andalusite, cordierite and even sillimanite. These deposits have not been studied in detail in this context, and hence, only the largest pegmatites in the gneisses in the southeast and south are marked on the appended map (Appendix 3).

From their contact and breccia observations Simonen (1952) and Matisto (1976) concluded that the ages of the plutonic rocks in the Tampere schist belt are consistent with their composition, in such a way that the ultramafites are the oldest synorogenic rocks, and that the granites and pegmatites, which often cut even the granites, are younger and late orogenic. Radiometric dating by O. Kouvo and M. Sakko on zircon from the Viitaniemi pegmatite in the study area indicates for the occurrence a U-Pb age of about 1780 million years from the concordia diagram. Likewise, a preliminary lead isotope analysis on uranothorite from the Eräpyhä pegmatite shows a 207 Pb/206 Pb age of 1793 million years. The granites in the study area have not been dated. In general, however, Svecokarelidic late orogenic granites are between 1800 and 1860 million years old (Kouvo and Tilton 1966; Vaasjoki 1977). Thus, agewise, the pegmatite deposits in the Eräjärvi area seem to be typical Svecokarelidic late-orogenic dyke rocks.

The olivine diabase, which in places cuts pegmatites as well, is, however, the youngest igneous rock in the area. The occurrences are part of the extensive swarm of Häme olivine diabase dykes whose age is comparable to that of the rapakivi granites in southern Finland (Laitakari 1969).

Tectonics

The predominant trend of the schists in the Tampere schist belt is nearly east—west. In the Eräjärvi area the igneous rocks subdivide the schists into two parts. Adjacent to the plutons the schists are conformable with the contacts, having, however, undergone folding during the emplacement of the plutons.

The tectonics of the study area are illustrated in the Appendix 2. The dash-dot lines indicate the subareas that were studied in detail for the trends of the pegmatite dykes and tectonic directions. The areas, marked with Roman numerals, are: I) Viitaniemi subarea, II) Seppälä subarea, III) Mikkola subarea, IV) Leväslahti subarea. As many tectonic observations as possible were made in each of them. Appendix 1 shows the projections of the strike and dip of the schistosity within the subareas. Included are the axes of the minor folds and observations of lineations. In the northern part of the map area (Appendix 3), transversal schistosity is fairly common and hence, a projection was drawn that illustrates the strike and dip of bedding in the Leväslahti subarea. All the projections in Appendix 1 were drawn by computer by means of the GELI-2 program (Behrens und Siehl 1975) modified at the Geological Survey of Finland.

The metapelites exhibit well-developed schistosity. In the northern and southwestern parts of the study area the schistosity strikes almost east—west, often with subvertical dips. Near the intrusions the strike of the schistosity conforms with the contacts of the plutons. In the south, close to the plutons, the dips are often slightly southwards; in the north they are correspondingly northwards. On account of the proximity of a diorite intrusion the schistosity in the Viitaniemi (I) and Seppälä (II) subareas striked ENE and in the Mikkola (III) subarea towards NE.

Mica schists, phyllites and greywacke schists often show distinct bedding or graded bedding. The beds are often a few millimetres or centimetres wide. In most parts of the area schistosity and bedding coincide. Transversal schistosity is common only in the north-east, where the schistosity bends more towards NE than does the bedding. Since the angle between the schistosity and bedding planes is often small, the directions do not differ clearly from each other on the projections either (see Appendix 1).

The schists in the area tend to be very rich in mica; hence they were folded plastically and the folds are often narrow and isoclinal. The top of the beds therefore varies over a short distance across the schist zone. Observations on the strike of the bedding marked on the tectonic map were usually obtained from graded-bedded schists or in a few cases from current bedding. The strikes of the minor fold axes are often parallel to schistosity, which developed mainly parallel to the axial plane. Lineation was measured from the stretching of biotite and hornblende along the schistosity plane. The projections indicate that the lineations are often parallel to the strike and plunge of the fold axes.

Topographic maps and aerial photos of the area show that the bedrock was broken up into blocks bordered by valleys and water ways during later deformation of the Earth's crust. The prevailing deformation directions run from NW to SE but lineaments trending NE—SW and E—W are also common. The first mentioned direction, which is the predominant direction of glacial flow in the area, is also an old fracture and fault direction (cf. Laitakari 1969).

SHAPE, SIZE AND REGIONAL DISTRIBUTION OF THE PEGMATITES

Most of the pegmatites in the study area occur in different kinds of schists and in granitic plutonites. Deposits are rare in quartz diorites and are lacking completely from diorite plutons except in their marginal zones. The occurrence of pegmatites in the Eräjärvi area is largely controlled by the proximity of the granite intrusions. All the largest deposits and half of the complex pegmatites are in the contacts of the granite plutons or in the schists close to the contacts. Some of the marked pegmatites also occur within the Kultavuori pegmatite granite area east of Eräjärvi. Pegmatites over 10 m wide are not encountered farther than 1.5 km from the granites and even dykes 5 to 10 m wide are only met with sporadically. Exceptions are the Niemelä pegmatite (54) and some other deposits in the same dyke swarm.

The dykes within the microcline granites are generally two-mica and tourmaline-poor simple pegmatites that often show sharp and linear contacts with the wall rocks. The pegmatite dykes are often narrow, only 0.5 to 2.0 m wide, and up to a few tens of metres long. The Rappuvuori granite in the western part of Eräjärvi is largely free from pegmatites, although in places they occur as dyke swarms. The Kultavuori pegmatite granite in the southern part of the study area has several pegmatite dykes and lenses. Since the deposits are sometimes very coarse-grained and several metres wide, some of them have been mined for potassium feldspar. Beryl and columbite–tantalite have been encountered in five pegmatites. One of these, the Katila pegmatite (6), close to the western contact of the pluton, contains small amounts of Li silicates and Li phosphates as well.

The northern and southern contact of the pegmatite granite is covered by overburden. Only the part of the pluton that extends to the Seppälä pegmatite area (II) east of Eräjärvi and which is composed of grey microcline granite, crops out locally. Pegmatite dykes or lenses are more or less ubiquitous between the schists and granite. In some places along the contact diagonal or perpendicular dykes extend farther into the schists and have obviously acted as feeding channels for the pegmatite melt.

More than half of the Li pegmatites as well as several Be-Nb pegmatites and simple pegmatites occur in the Seppälä area. Most of the complex pegmatites are in the schists beyond the contact with granite; even so, they are close to it and never more than 500 m from the pluton. In the southern contact of the microcline granite tongue there is a marked fracture and probably also a fault zone. The schists SE and S of it grade into migmatized mica gneisses, mica schists and veined gneisses. The pegmatite bodies, veins and dykes encountered in them are often two-mica pegmatites, homogenous or slightly zoned and mineralogically simple. All the known complex pegmatites are located NW and N of the fault line, always in less intensely altered schists and in the plutonites within them.

The eastern contact of the Rappuvuori granite intrusion in the western margin of the study area is also partly exposed and, as in the Seppälä area, contains several successive pegmatite lenses and dykes. Some of the deposits are coarse-grained or very coarse-grained simple pegmatites; some are complex pegmatites. The latter contain small amounts of Be minerals and Nb-Ta minerals. Li silicates have been encountered only in the Leikattu (7) pegmatite and in the Juurakko (5) pegmatite, which is somewhat farther away from the contact of granite. The Mattila pegmatite (8), which is the largest occurrence in the area, is a lens-shaped deposit c. 150 m wide between the schists and granite. In grain size and mineralogy the pegmatite varies considerably and in places it contains large microcline granite inclusions. North of the deposit there are several other pegmatites but, for lack of outcrops, their size and shape are not well established.

The dyke swarm in the Mikkola area (III), the sizeable Juurakko deposit (5) south of the Mattila pegmatite and the pegmatites in the environment are in the schists, yet not far from the Rappuvuori granite. Diamond core drilling (3 drill holes) suggest that the Juurakko pegmatite (5) is a subhorizontal

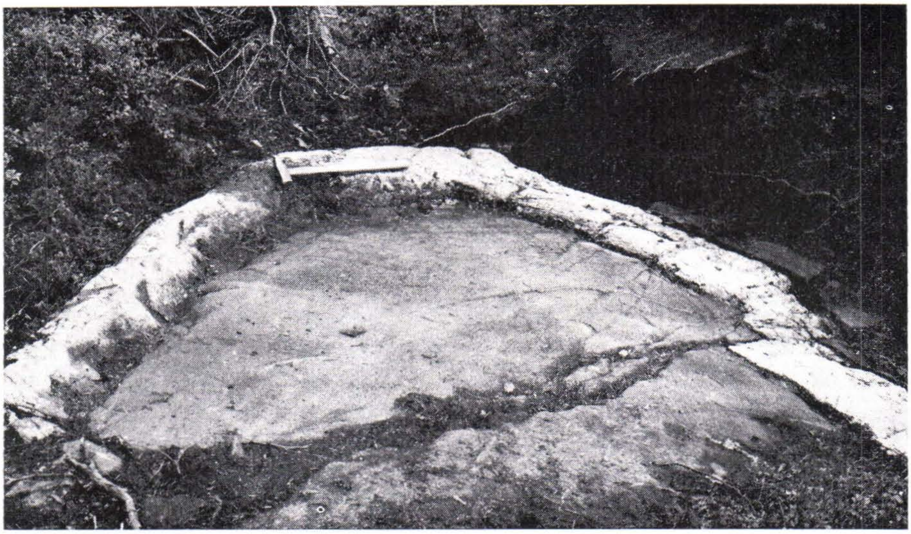


Fig. 2. Phacolite (pegmatite) in mica schist near the Keskimetsä quarry (46). Note the fault on the right. Length of hammer handle 50 cm.

lens. On the ground plan, however, it appears as a forked and ill-defined body (cf. Hirsjärvi 1933, Matisto 1964, Virkkunen 1964).

In the schists farther from the granites the pegmatites often occur as groups of dykes. Exceptionally, however, in the Viitaniemi area (I) around the Niemelä (54) and Rauhala (50) deposits there is a dyke swarm with scores of pegmatites. Most of the deposits in the northern part of the study area are simple pegmatites, although dykes that contain small amounts of beryl and columbite—tantalite are not unusual. Besides the Viitaniemi pegmatite (47), lithium minerals have been encountered in three deposits at the northern end of Vähä-Löytäne and rarely also in the Jussinvuori pegmatite (24) east of Koljonselkä, Längelmävesi.

Being harder than schist, pegmatite has resisted the abrasive action of glacial ice; particularly the larger deposits stand up from their surroundings. Most of the pegmatites in the schists are tabular dykes in shape, although some of the largest deposits are lens-like. Furthermore, a few rather small phacoliths (Fig. 2) have been encountered in the area. The narrow dykes (less than 5 m wide) are usually a score or so of metres, at the most 100 to 200 m long. The wider dykes, however, can often be traced for several hundreds of metres. The Niemelä pegmatite (54), which is one of the largest deposits in the study area, is about 30 m wide at the middle; it is at least 800 m long, but shows several discontinuations.

In general the pegmatite deposits in schists exhibit one particular trend. In some cases, however, the strike of the dykes may alternate between two or more tectonic directions, in which case the trend of the deposit appears very capricious. The pegmatite dykes in schists commonly form *en echelon* patterns. The dykes parallel to schistosity are usually of constant width and often longer than the dykes diagonal and perpendicular to the schistosity, in which the width may vary considerably. In some places the contacts of the lens-like deposits are so sinuous that it is not easy to estimate the size of the peg-

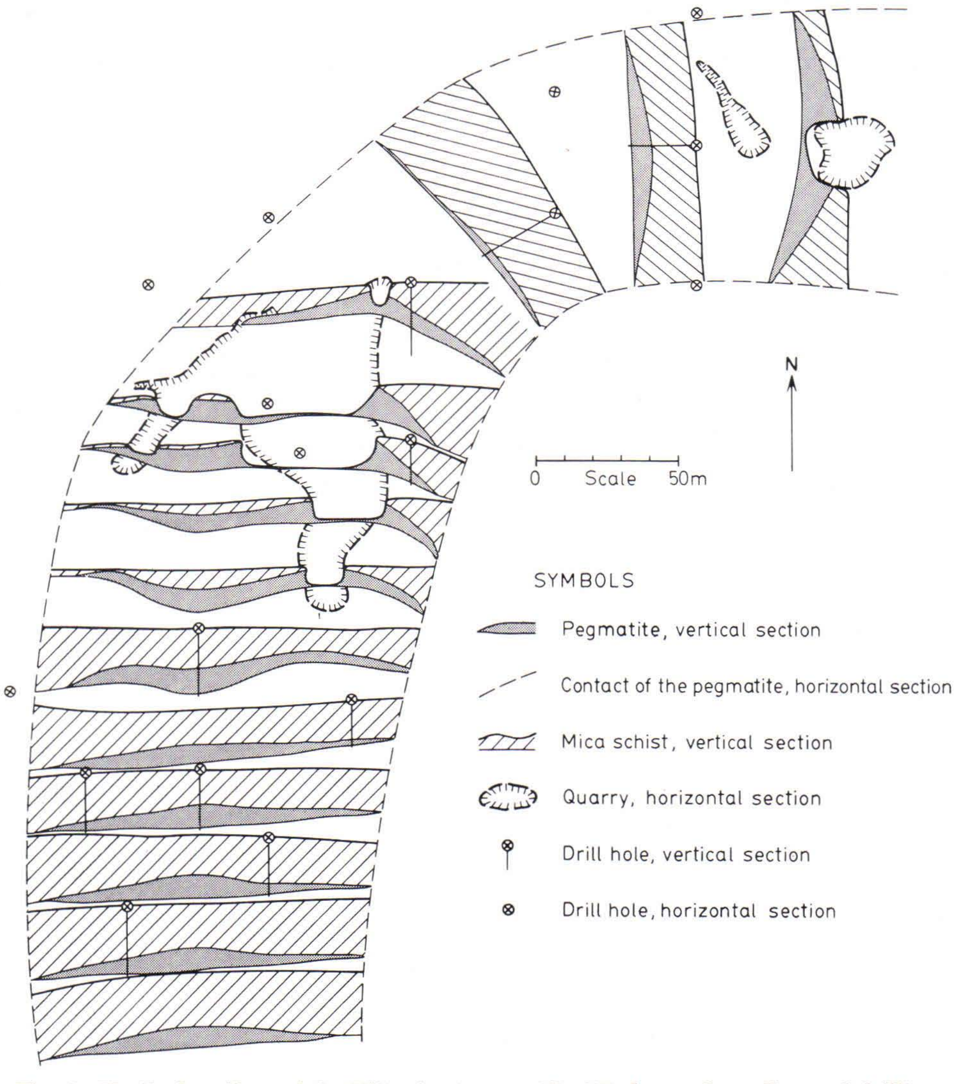


Fig. 3. Vertical sections of the Viitaniemi pegmatite (47) drawn from diamond drillings.

matites. The Viitaniemi (47) and Juurakko (5) deposits at least, as well as possibly some other large lens-like pegmatites, are horizontal occurrences of which only a random section is exposed on the surface.

Fig. 3 illustrates sections across the Viitaniemi pegmatite drawn from diamond core drilling data. The drilling data gathered by Suomen Malmi Oy (18 drill holes) reveal that the pegmatite runs in the schists almost from east to west as a body 100 to 130 m wide and in places 10 m thick. In the western part of the largest quarry the pegmatite turns sharply southwards (Lonka 1961). In the pits the pegmatite emerges towards the surface; between them, however, it descends and thins out. Hence, the contacts of the deposit on the surface plan show many forks and, without drilling, it would be impossible to evaluate the size and shape of the pegmatite.

RELATION OF THE PEGMATITES TO TECTONICS

The correlation between tectonics and dyke trends was studied in detail in four subareas outlined on the tectonic map of the study area (Appendix 2). Appendix 1 shows the trends of the simple and complex pegmatites arranged according to subareas and stereographic projections of the predominant tectonic directions. The subareas were chosen so that each one has the maximum dyke frequency and also so that there is no random variation in the schistosity.

The majority of the pegmatites in the schists zone conform with the schistosity but, as shown by the projections and the geological map, occurrences normal or diagonal to schistosity are encountered fairly frequently (cf. Fig. 1). Besides dykes parallel to the schistosity plane, dykes of this kind intruding shear and tension joints are very common in fold mountains (cf. Whitten 1966, Billings 1972). In the study area the pegmatite dykes trend only in a certain tectonic direction, although in some cases the trend of the deposits may alternate between two or more directions, so much so that in poorly exposed areas there is no way of tracing the dykes. Particularly in the schists in which the bedding and schistosity differ from each other, the pegmatite dykes sometimes follow the bedding and sometimes the schistosity (e.g. the Jussinvuori pegmatite). In rocks in which mica-rich and mica-poor competent layers alternate, the dykes preferably trend along the bedding, even if the rock has transversal schistosity.

The Löytäne pegmatite (57) is a good example of a dyke that follows several tectonic directions. Its predominant trend is diagonal, but in places the dyke is parallel or perpendicular to the schistosity. The large pegmatites that cut the schistosity often have narrow apophyses parallel to the schistosity. Crossing pegmatite occurrences are common, particularly in the Seppälä subarea (II), although they are encountered elsewhere too. The different directions of one and the same main dyke have been marked as separate observations on the projections showing the trends of the pegmatite dykes. Nevertheless, there are several deposits in the study area whose trend has been impossible to determine accurately. Hence the total number of observations of the projections (Appendix 1) differs somewhat from the actual number of dykes.

Volborth (1954a) has suggested that there are two pegmatite generations of different age in the Eräjärvi area and that the Viitaniemi occurrence belongs to the younger generation whereas the narrow simple pegmatites parallel to the schistosity are older. The study by Volborth (1954a) does not, however, include projections or diagrams showing the trends of dykes, and his observations obviously derive from a limited area. The present author found no mutually cross-cutting dykes in the course of his field work; hence, orientation projections were drawn for complex and simple pegmatites to establish possible differences in trend.

The projections in Appendix 1 demonstrate that in the northern part of the study area, in the Mikkola (III) and Leväslahti (IV) subareas, both the simple and complex pegmatites are often parallel to the schistosity. In the Viitaniemi (I) and Seppälä (II) subareas the simple pegmatites are also generally parallel to the schistosity, whereas the complex pegmatites show a larger scatter in dyke trends, and the pegmatite deposits normal and diagonal to schistosity are of almost equal frequency. At least in these subareas (Viitaniemi and Seppälä) the complex pegmatites intruded into different tectonic environments than the simple pegmatites, and some of the dykes may differ slightly in age. Since mutually crosscutting pegmatite dykes have not been encountered with certainty in the area, the order of the relative ages of the occurrences has not been established.

INTERNAL STRUCTURES OF PEGMATITES

The pegmatites interpreted as magmatic crystallized in several steps, and their intraformational structures are due to a number of processes. The pegmatite melt was segregated from the parent magma and intruded along channels opened up by tectonic movements higher up in the Earth's crust. At the same time the melt was differentiated and enriched in certain elements. The pegmatite deposits are homogenous, zoned, or they are characterized by banded or layered structures (Cameron *et al.* 1949, Jahns 1955, Schneiderhöhn 1961, Jahns and Tuttle 1963, Matrosov 1978). The complex pegmatites in particular are often characterized by a distinct zoned structure. The deposits are composed of a border, wall and intermediate zone or zones and a quartz core or cores (Cameron *et al.* 1949, Schneiderhöhn 1961). The zones differ from each other in mineralogy and structure, and their formation is frequently attributed to the fractional crystallization of the pegmatite melt (Schneiderhöhn 1961, Stewart 1978).

Zonality is a typical feature of the pegmatites; banded or layered dykes have seldom been reported. One reason for this may be that these structures tend to favour simple pegmatites and aplite—pegmatite dykes, the very rocks whose mineralogy and internal structure have not attracted much attention. The banded and layered structures that occur in various forms in pegmatites have been attributed either to primary magmatic factors (Orville 1960, Jahns and Tuttle 1963, Windley and Bridgwater 1965, Ashraf 1975) or to metasomatic and metamorphic processes (San Miguel 1969). The mode of formation of the structures obviously varies from area to area and from occurrence to occurrence. Thus, Matrosov (1978), for example, has reported six genetic main types (relict, crystallization, segregation, fluidal, tectonic and injection banding) from Central Asia, Kazahstan, Tuva, East Transbaykal, the Urals and Mongolia.

The complex pegmatites in the Eräjärvi area are often distinctly zoned whereas the simple and complex pegmatites with very low content of rare minerals are only indistinctly zoned or almost homogenous. Furthermore, numerous simple and complex pegmatites and aplite—pegmatite dykes characterized by banded structure are encountered in the area. Both zoned and banded pegmatites contain fracture fillings and replacement bodies formed during crystallization. In the complex pegmatites beryl, Li, Be and Fe-Mn phosphates, Li silicates and Nb-Ta minerals are enriched in the intermediate zone or albite-rich replacement bodies and in fracture fillings, of which some at least represent portions crystallized from supercritical fluids. Under postmagmatic conditions, the crystallization of the minerals may have continued down to very low temperatures. The complex pegmatites in particular contain as fracture and cavity fillings low-temperature secondary minerals that often include minerals very rich in water and fluorine.

Zoned structures

The large complex pegmatites, but occasionally also the simple pegmatites, show a

Fig. 4. The contact of the Löytäne pegmatite dyke (57). Note the zoned structure of the pegmatite: a, wall rock (mica schist); b, border zone; c, wall zone; d, intermediate zone. Length of scale tag 12 cm.

distinct zoned structure. The deposits are composed of a border, wall and intermediate zone and, in the middle of the deposit, quartz cores (Fig. 4). The zonality is generally symmetric, although the quartz cores are not always right in the middle of the dyke. Wholly homogenous pegmatites are rare, whereas all sorts of transitional forms between distinctly zoned and homogenous pegmatites are common. Most of the simple pegmatites and many of the complex pegmatites have only two clear zones: the border zone and the homogenous intermediate zone in the middle of the dyke. The dykes with indistinct zonality sometimes, but by no mean always, contain quartz cores. As a rule the clearer the zoned structure in the deposit, the larger are the quartz cores.

Border zone

The border zone is a narrow strip of pegmatite up against the wall rock. Owing to its small grain size (less than 3 cm), it often differs clearly from the inner wall zone and displays a sharp contact with the wall rock (Fig. 4). The width of the border zone varies from 3 to 30 cm. The narrow (0.5 to 2.0 cm wide) seam composed almost entirely of plagioclase that often occurs against the wall rock, makes the zone look as if it had two parts. In the pegmatite deposits in the schists the border zone is generally composed of plagioclase (oligoclase or albite), quartz and muscovite. There may also be abundant black tourmaline and in places potassium feldspar as well (microcline or orthoclase); its abun-

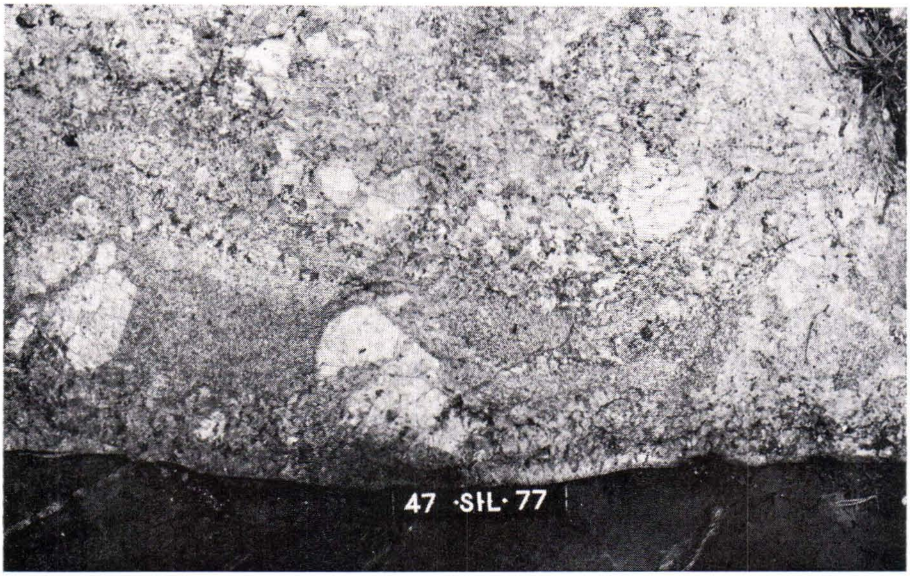


Fig. 5. Microcline megacrysts in the aplitic contact rock of the Tahkonkallio pegmatite dyke. Length of scale tag 12 cm.

dance, however, is always lower than that of plagioclase. Sporadic potassium feldspar megacrysts sometimes occur (Fig. 5) when the contact rock is a fine-grained aplite. In some dykes associated with the schists the border zone is composed of quartz and muscovite alone. This mineral association is typical especially in complex pegmatites.

The border zone of the pegmatites associated with the plutonites has plagioclase, quartz, micas and often also microcline as major minerals; microcline, however, is invariably subordinate to plagioclase. Tourmaline tends to be rare or absent. The predominant mica is usually biotite, but sometimes muscovite and sometimes both. The border zones of the pegmatites in the plutonites and schists contain garnet, apatite, zircon and beryl as accessory minerals. Elongated crystals (tourmaline, beryl, apatite and microcline) often grow perpendicular to the contact, widening in places like cones towards the innermost parts of the deposit.

Wall zone

Towards the middle of the pegmatite the border zone is followed by the wall zone. It is usually medium-grained (grain size from 2.5 to 10 cm) or coarse-grained (grain size between 10 and 30 cm) pegmatite with plagioclase (oligoclase or albite), microcline, muscovite, quartz and occasionally tourmaline as main minerals. In the area of plutonites or in their contact zone the micas include muscovite and biotite or sometimes only biotite. The abundance of plagioclase is the same as or somewhat higher than that of microcline. The accessory minerals include beryl, garnet, apatite, sulphides, arsenides and some thucholite. The contact with the inner intermediate zone is often gradual. In the clearly zoned pegmatites the intermediate zone is always distinctly more coarse-grained. In the large deposits with well-developed zonality the width of the wall zone may attain several metres. In the pegmatites with indistinct zoning the wall zone is very narrow or absent, and the middle of the occurrence is composed of a coherent intermediate zone.

Intermediate zone

The intermediate zone constitutes together with the quartz cores the interior of the pegmatite deposit. The zone is coarse-grained (grain size from 10 to 30 cm) or very coarse-grained (grain size over 30 cm) potassium feldspar predominant pegmatite. The microcline crystals in the middle of the occurrences may be huge, even as much as several metres long. They are interspersed with albite—quartz—muscovite pegmatite which, in the complex pegmatites, often contains beryl, columbite—tantalite and other less common minerals. In the lithium pegmatites, muscovite may be replaced by lepidolite associated with other lithium minerals. The rare minerals may occur in only a certain part of the intermediate zone, in which case the zone may consists of several parts. On the other hand, the albite replacement bodies and fracture fillings may cut the central parts of the deposits, which makes the zoned structure very complicated.

Core

The deposits with poorly developed zonality lack quartz cores whereas some of the large and distinctly zoned pegmatites contain cores that may have a volume of several cubic metres. Several cores generally occur together and are elongated or irregular in shape. Quartz is often white or grey. The banded variety is also rather common, and rose quartz has been encountered in some deposits. In some places the cores were apparently formed in several steps. Thus, in the Viitaniemi pegmatite, for example, brecciated banded quartz is locally embedded in a white quartz of a later generation.

Banded structures

Banded structure is met with in the study area in numerous pegmatite or aplite—pegmatite dykes associated with schists. The banding is due to the rhythmic alternation of the main minerals in the rock or of the aplitic and pegmatitic layers. On the weathered surface of the fine-grained pegmatites and aplites the structure is shown by garnet and tourmaline bands whose dark colour readily distinguishes them from the other minerals in the rock. The banding is distinct particularly in the margins of the pegmatites, which are sometimes composed of fine-grained albite- or oligoclase-rich aplite. The banding often disappears in the interior of the dykes as the pegmatite becomes coarser and richer in potassium feldspar. With some exceptions, the contact between the banded marginal parts and the coarser and homogenous middle part of the dyke is gradual.

In the northern part of the Niemelä pegmatite (54), the margins of the dyke are of banded tourmaline- and garnet-bearing aplite that occasionally shows cross-bedding structures (Fig. 6) due to melt flow. The interior of the dyke is markedly more coarse-grained potassium feldspar-predominant pegmatite in which banding is only sometimes visible. In some parts of the Tahkonkallio pegmatite (about 200 m north of the Mikkola pegmatite dyke, 20), homogenous aplite and pegmatite layers (Fig. 7) alternate close to the contact of the deposit. The layers vary in thickness from a few centimetres to some decimetres. The rock contains sporadic microcline megacrysts surrounded by narrow layers of undulating aplite. In places the interior of the dyke consists of banded pegmatite, in others of homogenous potassium feldspar-predominant pegmatite. Occasionally it shows aplite autoliths that probably derived from the margins of the occurrence during the intrusion of the melt.

Fig. 6. Banded aplite near the contact of the Niemelä pegmatite dyke (54). Note the cross-bedding structures which are due to melt flow. Length of scale tag 12 cm.

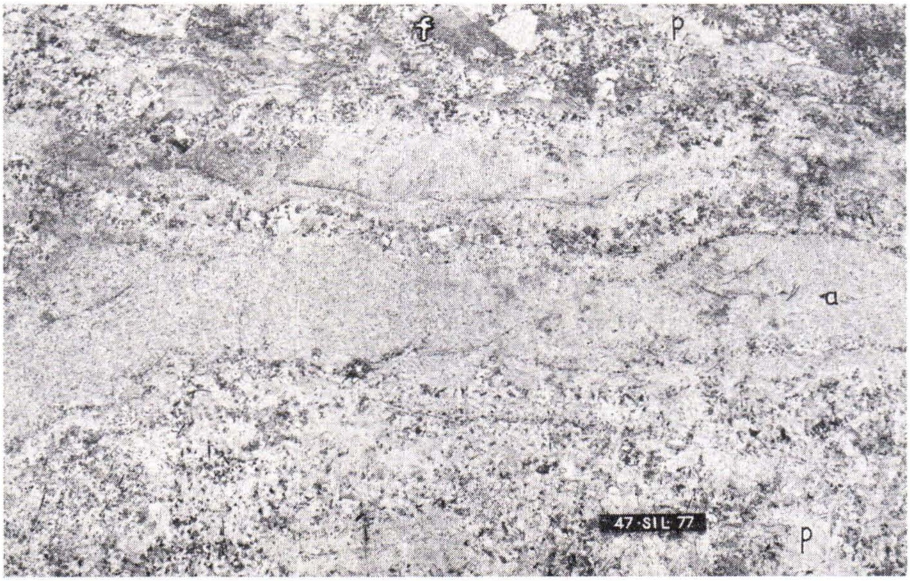


Fig. 7. Alternating layers of pegmatite (p, coarse-grained parts) and aplite (a, fine-grained parts) near the contact of the Tahkonkallio pegmatite dyke. Note the aplite autolith (f) beside the large microcline megacryst in the pegmatitic layer. Length of scale tag 12 cm.

The banding is usually subparallel to the contacts of the pegmatite dykes. Now and then, however, as in the Kanervikko dykes (42, 43), for example, the bands are folded and bend around the lens-shaped and more coarse-grained pegmatite portions (Fig. 8).

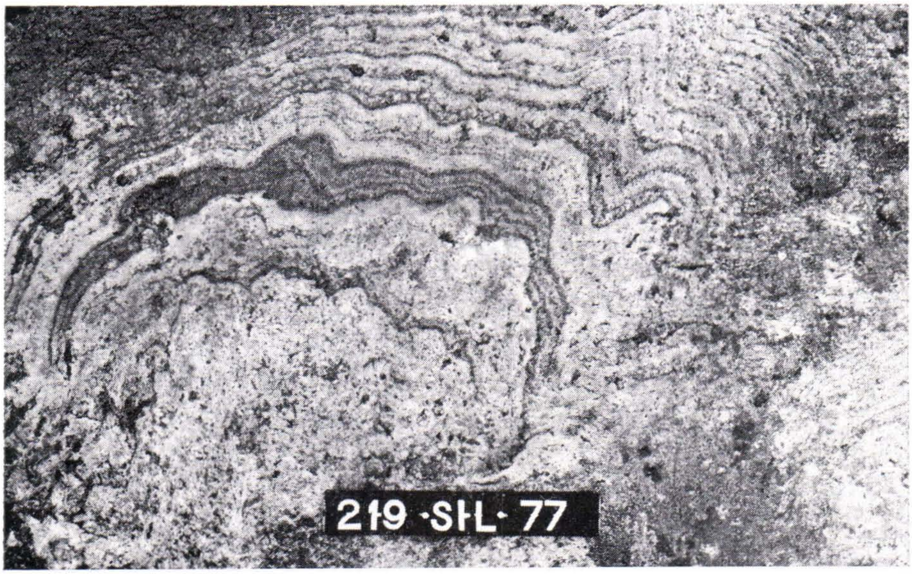


Fig. 8. Folded bands in the garnet- and tourmaline-bearing aplite. The aplite-pegmatite dyke of Kanervikko S (43). Length of scale tag 12 cm.

The banded rock is very fine-grained aplite in which narrow bands rich in garnet or tourmaline, or both, alternate with quartz, albite and muscovite, the main minerals. Similar undulating or folding banding has also been observed in the intraformational dykes with distinct contacts that cut the pegmatites. It is not always possible to conclude whether the banded structure is primary magmatic due to the intrusion of melt, or whether it is due to the postmagmatic fracture fillings that were formed during the final stages of the crystallization of pegmatite.

Replacement bodies and fracture fillings

The pegmatites in the study area show various mineralogically distinct replacement bodies and fracture fillings that were formed in association with the crystallization of the main dyke. Tectonic movements opened up fractures in the pegmatite dykes; these were then intruded by pegmatite melt, supercritical fluids or hydrothermal solutions. As the fractures opened the wall rock of the occurrences often reacted plastically and penetrated as tongues towards the interior of the main dyke.

The narrow and rather coarse-grained intraformational pegmatite dykes that crystallized in the fractures opened in the margins of the occurrences are quite common in the pegmatites. The interior of the main dyke was still not solidified, and the fractures filled with pegmatite melt or with melt and crystals. The fracture-filling dykes often start in the middle of the deposit and usually run perpendicular, sometimes also diagonally, towards the contact of the main dyke, where they come to an end. In some places, however, the intraformational dykes cut the central parts of the main dyke as well. In those cases the melt obviously derived from the lower parts of the feeding channel, and the dykes represent a later emplacement. Several

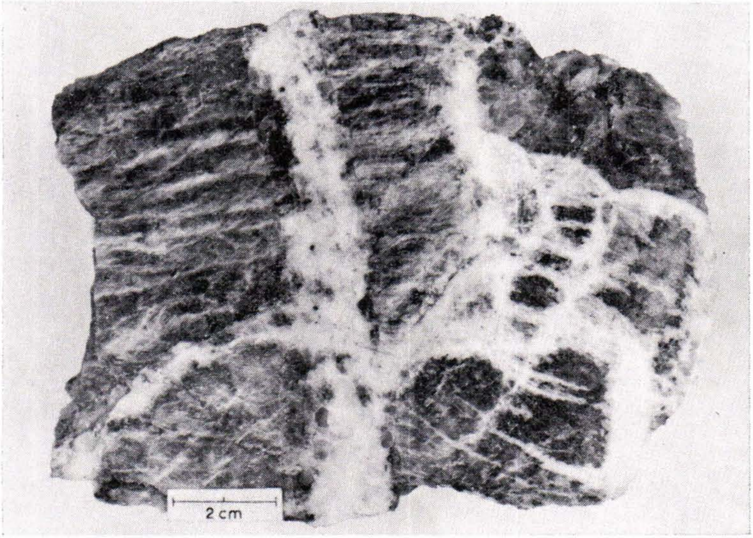


Fig. 9. Part of a large microcline crystal (dark grey) replaced by small sugar albite veins (white). The sample was taken from the wall zone of the Viitaniemi pegmatite (50 cm to the upper contact of the pegmatite).

fracture fillings a few decimetres wide are seen to cut sharply the aplitic marginal parts of the occurrences, e.g. in an outcrop near the northernmost quarry of the Niemelä pegmatite (54). The intraformational dykes are composed of medium-grained pegmatite. Their main minerals are microcline, quartz, muscovite and albite; the accessory minerals are tourmaline, beryl, apatite, garnet and occasionally small amounts of cassiterite and columbite—tantalite, i.e. the same minerals that occur at random elsewhere in the interior of the pegmatite.

The intermediate zones of the complex pegmatites often show homogenous, but sometimes banded or zoned dykes a few centimetres or decimetres wide that were formed as fracture fillings and contain mainly cleavelandite or sugar albite, quartz, muscovite and small amounts of potassium feldspar. The contacts of the fracture fillings are sharp and linear or gradual and the dykes may fork in many directions. The large microcline crystals, particularly in the interior of the deposits, often reveal albite-rich veins that replace potassium feldspar (Fig. 9) and generally come to an end in a large cleavelandite or sugar albite body in the middle of the pegmatite dyke. In the complex pegmatites the replacement bodies and albite fracture fillings are enriched in beryl, columbite—tantalite, lepidolite, elbaite, spodumene, amblygonite—montebrasite, manganiferous apatite and triphylite—lithiophilite. The albite-rich dykes and replacement bodies with lithium minerals and other rare pegmatite minerals were formed during the final stages of magmatic crystallization from melts very rich in volatiles or from postmagmatic fluids. These may then have reacted with minerals that had crystallized earlier and thus given rise to replacement structures and pseudomorphs.

The zoned or banded fracture-filling dykes are rare. The zonality appears in them in such a way that the interior of the dyke is composed of a quartz or quartz—muscovite core whereas the margins are predominantly albite. In the banded dykes, narrow layers rich in garnet or tourmaline, or both, alternate which the main minerals: sugar albite, quartz and muscovite. In the Eräpyhä pegmatite (1), exceptional fracture fillings are encountered that contain abundant zircon as well as spessartine. The bands are sometimes composed of fine-grained cassiterite alone.

Crosscutting narrow quartz or quartz—tourmaline dykes are very common in the pegmatites of the study area. They run across the deposit, either perpendicular or diagonal to it, and were obviously crystallized in the fractures of the pegmatities from postmagmatic fluids and water solutions. The quartz dykes tend to be a few centimetres wide at the most, and they invariably come to an end in the contact. The deposits without quartz cores sometimes show abnormally wide intraformational and crosscutting quartz dykes. In some pegmatites the quartz core seems to have been formed as a fracture filling and, like a dyke, runs with linear contacts through the middle of the deposit.

Calcite, fluorite, löllingite and various sulphides are occasionally encountered as narrow fracture fillings in the pegmatites. They often seem to represent the last hydrothermal crystallization products of the pegmatites, although they may also derive from the wall rocks. The veins with fluorite, sphalerite, löllingite and arsenopyrite obviously crystallized from the differentiation products of the pegmatite melt, for these minerals also occur in the pegmatite in miarolitic cavities as euhedral or subhedral crystals.

CONTACTS OF THE PEGMATITES

The contacts of the pegmatite dykes with the wall rock are generally sharp and linear. Only in the Kultavuori pegmatite granite area and occasionally in the contact zone of the microcline granite plutons may the contacts of the pegmatites with the wall rock be gradual.

The intruding pegmatite melt has often snatched fragments from the wall rock. The xenoliths close to the contacts of the dykes are evidently very local. Seldom having rotated in the melt, they have retained the orientation of the wall rock. In some places thermal metamorphism due to pegmatite melt has thoroughly altered the mineral composition of the fragments and the wall rock; in other places the only alteration is the enrichment of the contact zone in black tourmaline. The plutonites adjacent to the pegmatites show sericitization and saussuritization of feldspars and chloritization of biotite.

In the schists close to the pegmatite deposits there is often a seam very rich in tourmaline. Small dark brown or black acicular crystals may occur up to several metres from the contact in the mica-rich layers in the wall rock. The tourmaline crystals often exhibit zonality and contain abundant quartz inclusions either as stringers parallel to the bedding or as rims around the cores of the crystals. Electron microprobe determinations show that the tourmalines in the contact rocks of the Viitaniemi (47) and Seppälänranta (31) pegmatites contain 4.3 to 6.8 wt% MgO, 7.8 to 10.3 % FeO (total iron) and 0.1 to 0.2 % MnO. Thus, they are rather rich in iron and magnesium and belong to the schorl—dravite series.

When in association with Fe-Mg tourmaline, the wall rock biotite has often disappeared and been replaced by porphyroblasts (∅ 0.5 to 5.0 mm) of a pale-brown iron- and magnesium-rich muscovite or phengite. Electron microprobe determinations show that the micas in the contact rocks of the Viitaniemi (47), Seppälänranta (31), Niemelä (54) and Löytäne (57) pegmatites contain 1.0 to 2.7

Table 1

The chemical composition (wt$^0/_0$) and physical properties of lithium- and iron-bearing muscovite from the mica schist in the contact of the Viitaniemi pegmatite. Wet chemical analysis by R. Saikkonen.

		The content of unit cell on the basis of 48 anions		Physical properties
SiO_2	46.40	Si	12.70	The unit cell:
TiO_2	0.91	Ti	0.19	Monoclinic C2/c
Al_2O_3	26.73	Al	8.62	Polytype $2M_1$
Fe_2O_3	1.26	Fe^{+3}	0.26	a = 5.186 ± 0.005 Å
FeO	5.63	Fe^{+2}	1.29	b = 9.097 ± 0.005 Å
MnO	0.16	Mn	0.04	c = 19.921 ± 0.005 Å
MgO	2.70	Mg	1.10	$\beta = 95°40' \pm 5'$
CaO	0.16	Ca	0.05	V = 935.221 Å^3
Na_2O	0.31	Na	0.16	G = 2.899
K_2O	9.70	K	3.39	D = 2.909 g/cm^3
Li_2O	1.33	Li	1.46	
Rb_2O	0.28	Rb	0.09	Optical properties:
Cs_2O	0.12	Cs	0.01	$\alpha = 1.557 \pm 0.001$
BaO	0.18	Ba	0.02	$\beta = 1.588 \pm 0.001$
SrO	0.01	Sr	0.00	$\gamma = 1.590 \pm 0.001$
H_2O+	2.50	OH	4.56	$\gamma - \alpha = 0.033$
H_2O-	0.13	F	3.13	—2V = 28° ± 1°
F	3.62	O	40.30	—2V calc. = 28.1°
	102.13			X = colourless
$-O = F_2$	1.52			Y = Z = light brown
Total	100.61 *			

* Omitting the results of the spectral analysis by R. Danielsson: Sn 900 ppm, Pb 900 ppm, V 92 ppm, Cr 100 ppm, B 85 ppm

wt$^0/_0$ MgO and 2.0 to 7.2 wt$^0/_0$ FeO (total iron). In thin section the micas show weak pleochroism (X = colourless, Y = Z = pale brown). Universal stage determinations indicate that they have a small axial angle, $-2V = 28°—38°$, as have many Li-Fe-Mg muscovites and phengites reported in the literature (cf. Deer, Howie and Zussman 1967 p. 21, Koval' *et al.* 1972, Syritso *et al.* 1977, Bazarova *et al.* 1977).

The iron-rich mica in the wall rock of the Viitaniemi pegmatite was studied in detail. The contact rock contains quartz, Mg-Fe tourmaline and pale brown muscovite as main minerals, and apatite, orthoclase, zircon, graphite and Fe sulphides as accessories. In places the contact zone is several decimetres wide. Farther from the pegmatite the rock is mica schist or greywacke schist with some muscovite in addition to the main minerals biotite and quartz.

Pale brown muscovite was separated from a ground sample of contact rock by means of heavy liquids and a Franz isodynamic magnetic separator. The mica fraction thus obtained was homogenous, as was demonstarted by refractive liquids, X-ray powder diffraction pattern and semiquantitative electron microprobe determinations on some scales. Thin section examination revealed that the only mica in the sample selected for study was brown muscovite.

The composition of the mica from the contact rock of the Viitaniemi pegmatite was analysed by wet chemical methods. The analytical data together with the physical parameters are listed in Table 1. Based on 48 anions and Z = 4, the formula of the mica is:

$$(K_{0.85}Na_{0.04}Rb_{0.02}Ba_{0.01})_{0.92}(Al_{1.34}Li_{0.37}Fe^{+2}{}_{0.32}Fe^{+3}{}_{0.07}$$
$$Mg_{0.28}Ti_{0.05}Mn_{0.01})_{2.44}(Si_{3.18}Al_{0.82})_{4.00}O_{10.08}((OH)_{1.14}F_{0.78})_{1.92}$$

The mica differs from normal muscovite in that it is exceptionally rich in lithium, iron, magnesium and fluorine. The sum of the cations in the octahedral coordination is considerably more than 2 atoms per formula. Hence the mica is not an ideal phengitic muscovite but a transitional type between dioctahedral and trioctahedral micas.

In addition to chemical data, Table 1 also includes the physical properties determined from the Li- and Fe-rich muscovite in the wall rock of the Viitaniemi pegmatite. 0-, 1- and 2- level a-, b- and c-axis precession photographs were produced for the mica with Zr-filtered Mo radiation. The lattice parameters were determined from the precession films calibrated with silicon. On the basis of systematic extinctions, the space group of the mica is C2/c and the polytype $2M_1$. Furthermore, 0-level a- and b-axis photographs were taken for two scales but neither of them differed in structure nor essentially in lattice parameters from the crystal studied in detail.

The mineral exhibits weak pleochroism with X = colourless and Y = X = pale brown. The optic angle was measured on a universal stage from a thin section made of separated mica fraction. Several determinations showed that the optic angle did not vary between crystals.

The Li-, Fe- and Mg-rich micas encountered in the contact rocks of the pegmatites in the study area were formed through reactions between the wall rock and the constituents segregated from the melt. The magnesium and iron in biotite together with the boron, fluorine and alkalies derived from the pegmatite melt crystallized partly as tourmaline, partly as phengitic and possibly lithium-rich micas.

In addition to muscovite or biotite, or both, the contact rocks of the complex pegmatites have now and then been reported to contain lepidolite or zinnwaldite, although detailed studies, on the micas are frequently lacking (cf. Heinrich 1965). Lithium- and iron-bearing muscovites nearly similar in composition and physical properties to those in the wall rock of the Viitaniemi pegmatite have been reported from apogranites, greisens and pegmatites (Koval' *et al.* 1972, Syritso *et al.* 1977, Bazarova et al. 1977). The isomorphism between dioctahedral and trioctahedral micas is apparently a typical feature of these micas. According to Bazarova *et al.* (1977), in the micas of the muscovite—phengite—protolithionite—zinnwaldite series the sum of the atoms in the octahedral coordination of the mica formula is often 2.4 to 2.6, as it is in the muscovite in the contact rock of the Viitaniemi pegmatite.

MINERAL PARAGENESES OF THE PEGMATITES

In the course of field work more than three hundred pegmatite deposits were encountered in the study area. Sixty-one of these contained rare pegmatite minerals; most of them, 41 deposits in all, are beryl—columbite pegmatites and the rest lithium pegmatites. Altogether some 80 mineral species have been identified from the pegmatites. The list of minerals in the complex pegmatites is given in Table 2 (Be-Nb pegmatites) and Tables 3 a-c (Li pegmatites). Most of the rare minerals occur in the Li pegmatites, which are characterized especially by certain phosphate minerals. Apatite is the only known phosphate in the Be-Nb and simple pegmatites, whereas 29 phosphate minerals have been identified from the lithium pegmatites. Rare or otherwise mutually indistinguishable

Table 2

The minerals of the Be-Nb pegmatites in the Eräjärvi area.

	Eräpyhä (1)	Päimäki (2)	Makkarakorpi (3)	Niittymäki (4)	Mattila (8)	Kaunismäki (9)	Karvaanmaa (10)	Lintuvaara (11)	Maijaanvuori (12)	Kullanpää (13)	Omenajärvi (14)	Haavikko (15)	Vanhantalonsaari (16)	Kultavuori NW (17)	Suonaanmäki (18)	Metsänen (19)	Mikkola (20)	Kultavuori (21)	Paavola (22)	Vähävuori (23)	Retula (25)
Graphite & thucholite			×	×	×	×	×	×				×	×		×		×		×	×	×
Sphalerite	×										×										
Chalcopyrite	×										×										
Pyrrhotite	×										×										
Pyrite	×					×			×		×										
Löllingite					×																
Arsenopyrite	×	×	×	×			×	×			×	×	×	×				×		×	×
Molybdenite	×												×								×
Fluorite			×		×				×				×						×		
Quartz	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Microlite	×								×							×	×				
Cassiterite	×																				
Tapiolite																					
Columbite-Tantalite	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Uraninite	×																				
Calcite					×																
Apatite	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Scorodite				×	×																
Garnet	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Zircon	×	×	×		×																×
Bertrandite			×	×					×				×								
Beryl (yellowish)	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Beryl (pink)																					
Tourmaline (black)	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Muscovite	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Biotite	×	×			×		×			×											
Chlorite	×																				
Kaolinite			×						×												
K-feldspar	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Plagioclase	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×

Continued on page 26

pegmatite minerals were identified by means of X-ray diffraction powder patterns. Several of the rare minerals were also assayed chemically or identified by means of refractive indices. The pegmatites submitted to mining were studied in detail; the others were usually examined during two visits when samples were taken from the surface of the rocks.

The main minerals in the pegmatites are potassium feldspar, quartz, plagioclase, muscovite and/or biotite. Muscovite is the only mica in most of the occurrences. In the area of the plutonites and at their contacts, however, two-mica pegmatites or those with biotite alone are also encountered. The abundance of main minerals varies considerably, particularly within the zoned pegmatites. Black tourmaline and garnet may also be enriched in several zones. The rare pegmatite minerals usually only occur in the intermediate zone, in the albite-rich replacement bodies and in fracture fillings. Apart from black tourmaline and garnet only beryl,

Table 2 (continued)

	Väinävuori (26)	Seppälänniemi (28)	Hevoshaka (29)	Seppälä (34)	Kivistö (36)	Attila (37)	Attilanvuori (38)	Kangas (40)	Iilijärvi (45)	Ruokolahti (48)	Aukeessuo (49)	Rauhala (50)	Leppäjärvi (51)	Voikallionmäki (52)	Mäensyrjä (53)	Niemelä (54)	Rajala (55)	Sammalikko (56)	Löytäne (57)	Selkäluoto (61)
Graphite & thucholite	×	×		×	×					×	×				×	×	×			
Sphalerite																×				
Chalcopyrite																				
Pyrrhotite												×				×				
Pyrite												×								
Löllingite																				
Arsenopyrite	×	×		×	×	×	×			×		×		×					×	×
Molybdenite																				
Fluorite										×		×	×	×		×				×
Quartz	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Microlite																			×	
Cassiterite										×						×				×
Tapiolite		×																		
Columbite—Tantalite	×		×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Uraninite																				
Calcite																×				
Apatite	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Scorodite																×				
Garnet	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Zircon										×		×				×				
Bertrandite										×		×		×		×				
Beryl (yellowish)	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Beryl (pink)																×				
Tourmaline (black)	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Muscovite	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Biotite		×																		
Chlorite																				
Kaolinite										×	×					×	×			
K-feldspar	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Plagioclase	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×

apatite, zircon, thucholite and some sulphides and arsenides are typical accessory minerals in the wall and border zones of the deposits. These minerals, too, are often concentrated in the central parts of the pegmatites.

The beryl—columbite pegmatites are characterized by beryl in particular. The mineral occurs here and there within the pegmatites either as individual crystals or as groups of several crystals. Niobium-tantalum minerals, cassiterite, zircon, thucholite, garnet and apatite are often associated with it.

In the lithium pegmatites the rare minerals may be either homogenously distributed in the middle of the deposit or concentrated in a certain part of the dyke. The various lithium phosphates and lithium silicates frequently occur together. In the pegmatites of the study area lithium is mainly incorporated in phosphates. The most common Li mineral is triphylite—lithiophilite, which has been encountered in almost all the Li pegmatites. In eleven pegmatites it is the sole primary lithium mineral. In addition to triphylite—

Table 3a

The silicates of the Li pegmatites in the Eräjärvi area.

	Juurakko (5)	Katila (6)	Leikattu (7)	Jussinvuori (24)	Lahdensuu (27)	Talas (30)	Seppälänranta (31)	Rinne (32)	Töyräs (33)	Männikkö (35)	Pellonreuna (39)	Maantienvarsi (41)	Kanervikko N (42)	Kanervikko S (43)	Suonlaita (44)	Keskimetsä (46)	Viitaniemi (47)	Hannula (58)	Myllyrinne (59)	Karpalepohja (60)
Garnet	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Zircon	×	×	×				×					×			×	×	×			
Thorite			×														×			
Topaz	×						×										×			
Bertrandite	×	×	×				×					×					×			
Beryl (yellowish)	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Beryl (pink)	×		×				×										×			
Tourmaline (black)	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Tourmaline (pink & light green)	×	×				×	×					×					×			
Spodumene		×				×						×			×					
Muscovite	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Biotite		(×)																		
Lepidolite	×	×	×			×	×					×					×			
Bityite												×								
Montmorillonite			×														×			
Chlorite	×											×							×	
Cookeite			×									×					×			
Kaolinite		×					×					×			×		×			
Pollucite																	×			
K-feldspar	×	×	×	×*	×	×*	×*	×*	×*	×	×	×	×	×*	×	×	×*	×*	×*	×*
Plagioclase	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×

* Orthoclase in addition to microcline

lithiophilite, the deposits enriched markedly in lithium contain amblygonite—montebrasite and various lithium silicates, lepidolite, elbaite and spodumene. In the final stages of pegmatite crystallization or possibly partly under supergene conditions a number of phosphates were formed as alteration products of triphylite and lithiophilite. The secondary phosphates contain iron or manganese, or both, and often also water of crystallization or (OH) groups, or both. Altogether 15 such secondary phosphates have been identified from the pegmatites in the area (Fig. 10 and 15). Usually, however, they occur in the pegmatites in very small amounts.

The Viitaniemi pegmatite is exceptionally rich in amblygonite—montebrasite. Blue-green Mn-rich apatite, iron-, manganese- and beryllium-bearing phosphates are often encountered between the crystal plates of amblygonite—montebrasite as small nodules or as larger aggregates. Since the Viitaniemi pegmatite also contains abundant secondary Fe-Mn phosphates (Volborth 1954a), and since little attention has previously been paid to these, in this study secondary phosphate minerals were identified in greater detail. All the phosphates encountered in the deposit are listed in Fig. 10. The primary phosphates in the deposit are given from left to right in the

Table 3b

The phosphates of the Li pegmatites in the Eräjärvi area.

	Juurakko (5)	Katila (6)	Leikattu (7)	Jussinvuori (24)	Lahdensuu (27)	Talas (30)	Seppälänranta (31)	Rinne (32)	Töyräs (33)	Männikkö (35)	Pellonreuna (39)	Maantienvarsi (41)	Kanervikko N (42)	Kanervikko S (43)	Suonlaita (44)	Keskimetsä (46)	Viitaniemi (47)	Hannula (58)	Myllyrinne (59)	Karpalepohja (60)
Beryllonite																	×			
Hurlbutite																	×			
Triphylite—Lithiophilite		×		×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Ferrisicklerite—Sicklerite		×		×	×	×	×	×	×			×	×	×	×	×	×	×	×	
Heterosite—Purpurite		×			×	×	×	×	×			×	×	×	×	×	×	×	×	
Beusite				×																
Alluaudite		×			×	×	×	×	×		×	×	×	×	×	×	×	×	×	×
Herderite																	×			
Väyrynenite				×													×		×	
Amblygonite—Montebrasite		×				×	×									×	×			
Triplite																	×			
Rockbridgeite—Frondelite												×					×			
Viitaniemiite																	×			
Crandallite																	×			
Apatite	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	
Hureaulite		×				×	×					×			×		×		×	
Reddingite		×															×			
Phosphosiderite																	×			
Strengite																	×			
Switzerite		×					×					×			×	×	×			
Vivianite		×		×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Metavivianite		×															×			
Fairfieldite—Messelite		×				×	×					×			×	×	×		×	
Moraesite																	×			
Strunzite																	×			
Laueite																	×			
Eosphorite																	×			
Jahnsite		×					×					×			×	×	×			
Morinite																	×			

upper part of the figure. The phosphates that were formed at lower temperatures are below them. Common mineral associations are joined by a line. The minerals in boxes occur together or are otherwise associated with similar parageneses, although their mutual order of crystallization has not been established. Of the alteration products of lithiophilite given here reddingite, strunzite, laueite, phosphosiderite and strengite have so far been identified in the study area only in the Viitaniemi deposit. Nevertheless, they may occur in other pegmatites as well, because the alteration products are frequently alike and obviously formed under similar conditions.

The chemical formulae of the phosphate minerals in Fig. 10 are mainly from the »1975 Glossary of Mineral Species» (Fleischer 1975). In addition to viitaniemiite, crandallite, jahnsite, metavivianite, switzerite, laueite, strunzite, reddingite and phosphosiderite have not been reported before from Finland. Since the secondary phosphates often occur in very low abundances their identification is a tedious task. Hence the list of phosphates

Table 3c

The elements, sulphides, sulphosalts, arsenides, halogenides, oxides, carbonates and arsenates of the Li pegmatites in the Eräjärvi area.

	Juurakko (5)	Katila (6)	Leikattu (7)	Jussinvuori (24)	Lahdensuu (27)	Talas (30)	Seppälänranta (31)	Rimne (32)	Töyräs (33)	Männikkö (35)	Pellonreuna (39)	Maantienvarsi (41)	Kanervikko N (42)	Kanervikko S (43)	Suonlaita (44)	Keskimetsä (46)	Viitaniemi (47)	Hannula (58)	Myllyrinne (59)	Karpalepohja (60)
Antimony																	×			
Bismuth																	×			
Bi-antimony																	×			
Graphite & thucholite	×	×		×	×	×	×		×	×	×	×			×	×	×	×	×	×
Sphalerite		×	×				×									×	×			
Chalcopyrite	×																×			
Germanite																	×			
Tetrahedrite																	×			
Pyrrhotite																	×			
Herzenbergite																	×			
Stibnite																	×			
Kermesite																	×			
Pyrite	×	×	×												×	×	×		×	
Löllingite	×	×													×	×	×			
Arsenopyrite		×	×	×		×	×	×		×	×	×			×	×	×	×	×	×
Molybdenite	×																×			
Fluorite	×	×	×			×	×								×	×	×		×	
Valentinite																	×			
Senarmontite																	×			
Stibiconite																	×			
Quartz	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×	×
Cassiterite	×	×	×			×	×					×			×	×	×	×	×	
Wad							×								×		×			
Tapiolite							×				×						×			
Wodginite																	×			
Columbite—Tantalite	×	×	×	×	×	×	×	×	×	×		×	×	×	×	×	×	×	×	×
Limonite	×	×	×				×								×	×	×			
Calcite			×				×					×								
Siderite		×																		
Malachite																	×	×		
Scorodite	×	×	×				×								×		×			

in Table 3b was made as complete as possible for the Viitaniemi pegmatite, from which it was easiest to obtain samples.

The main lithium silicates in pegmatites are lepidolite, elbaite and spodumene. In the pegmatites of the study area, however, they tend to be rare. The exception is the Viitaniemi pegmatite, which contains abundant green tourmaline and lepidolite. Niobium-tantalum minerals, cassiterite, beryl, Mn-rich apatite, Mn-rich garnet, zircon and in some pegmatites topaz are often associated with the lithium minerals.

Most of the complex pegmatites in the study area contain the forementioned characteristic as well as other less common minerals scantily or very scantily. The lithium minerals are enriched particularly in the Viitaniemi pegmatite, although abundant triphylite—lithiophilite is also encountered in some pegmatites in the Seppälä subarea (II). Beryl and niobium-tantalum minerals

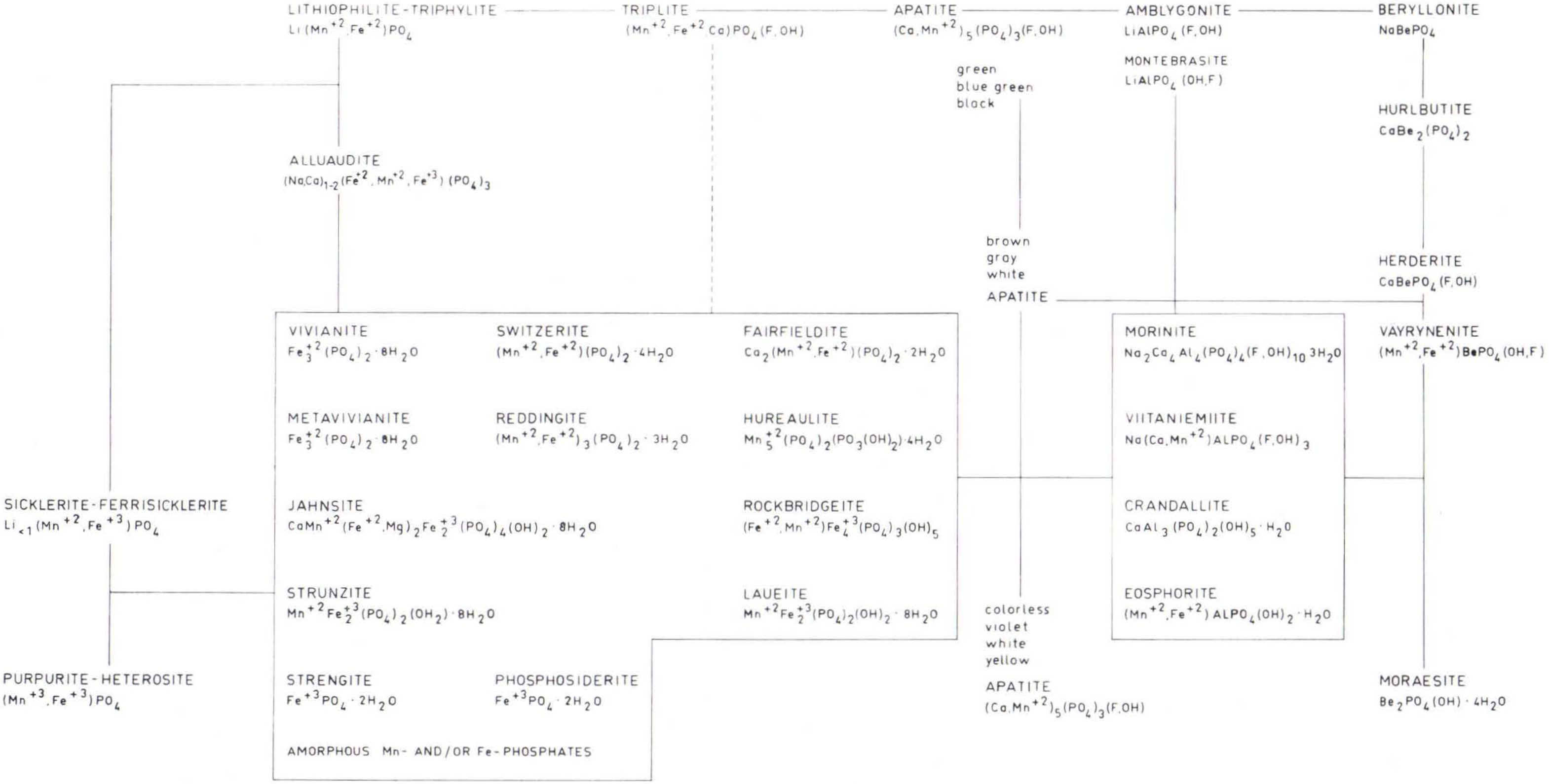


Fig. 10. A paragenetic tree of the phosphate minerals in the Viitaniemi pegmatite.

are most abundant in the complex pegmatites near or at the contacts of the granites and in some dykes in the Viitaniemi area. In the deposits farther from the granites, however, especially columbite—tantalite is rare, occurring only as tiny crystals.

MINERAL DESCRIPTIONS

Silicates

Potassium feldspar

Potassium feldspar which is enriched particularly in the intermediate zone of the pegmatites, decreases in amount towards the contacts. The better developed is the zonality in the occurrence, the larger are the potassium feldspar crystals in the intermediate zone. In the largest zoned dykes the microcline crystals attain several metres in length.

In the course of field work potassium feldspar samples were collected for detail studies from 70 deposits. According to the classification of the present author, 18 of them are Li pegmatites, 32 Be-Nb pegmatites and 20 simple pegmatites. The samples were taken from as close to the contact as possible, but sometimes also from the different zones of the dykes. Most of the feldspars in the samples studied were microcline ($\triangle$ usually 0.50 to 1.00), although orthoclase was also encountered in the border zones of ten pegmatites. Triclinicity of feldspars is lowest close to the contacts and highest in the interior of the deposits. Only one exception was found. In the Ruokolahti pegmatite (48) monoclinic potassium feldspar (adularia) crystals were encountered associated with euhedral bertrandite and fluorite crystals in a small cavity in the middle of the dyke. In addition, the microcline in the pegmatites close to the contacts of the olivine diabase dykes (1.5 to 2.0 m wide) that intersect the Seppälänranta (31) and Viitaniemi (47) pegmatites, has turned into orthoclase (up to 5 cm from diabase) evidently owing to the thermal effect of diabase melt and subsequent rapid cooling (cf. Haapala 1977, p. 76). Except for the Ruokolahti pegmatite, the deposits with orthoclase are all Li pegmatites whereas the simple pegmatites and Be-Nb pegmatites always have either intermediate or maximum microcline.

The orthoclase in the border zones of the pegmatites is grey or white in colour and either crypto- or microperthitic. The crystals are small, measuring no more than a few centimetres. 0-level precession photographs taken perpendicular to the b- and c-axis for the grey cryptoperthitic orthoclase collected from the contact of the Viitaniemi pegmatite showed up the albite phase as very weak spots indicating twinning according to the albite law.

The perthite structures grow coarser as the triclinicity increases towards the interior of the dykes. The microcline in the pegmatites of the area is usually grey or white. Cross-hatched twinning is very common, and string, vein and patch perthites predominate. The structures are often visible to the naked eye, and the big crystals in the intermediate zone may show very coarse macroalbite veins up to 0.5 to 2.0 mm wide. The large macroperthitic microcline crystals are occasionally enveloped by a rim of white microperthitic microcline. A similar pure white microcline whose triclinicity is unity or close to unity often occurs in the replacement bodies and

Table 4

Partial chemical analyses (oxides: wt%; albite = Ab, orthoclase = Or and anorthite = An: mol%), optic angle (—2V, U-stage measurements) and triclinicity (Δ) of potassium feldspars from the Viitaniemi, Eräpyhä and Pajala pegmatites. X-ray fluorescence analyses by V. Hoffrén.

	VIITANIEMI			ERÄPYHÄ			PAJALA	
	bz	wz	rb	bz	wz	iz	wz	iz
Na_2O	1.23	1.14	0.62	1.87	1.65	2.23	1.83	2.40
K_2O	14.58	14.26	14.90	14.02	13.53	13.36	13.84	13.02
CaO	0.08	0.08	0.16	0.09	0.08	0.09	0.09	0.09
Rb_2O	0.21	0.23	0.27	0.06	0.09	0.11	0.07	0.07
Cs_2O	0.02	0.05	0.10	0.00	0.00	0.00	0.00	0.00
Fe_2O_3	0.07	0.05	0.05	0.06	0.05	0.06	0.06	0.05
Ab	11.32	10.79	5.90	16.78	15.57	20.14	16.66	20.14
Or	88.28	88.79	93.26	82.77	84.01	79.41	82.89	79.41
An	0.41	0.42	0.84	0.45	0.42	0.45	0.45	0.45
—2V°	60	68	82	75	79	84	79	80
Δ	0.00	0.57	0.96	0.84	0.89	0.97	0.81	0.90

bz = border zone
wz = wall zone
iz = intermediate zone
rb = replacement body

in fracture fillings, and occasionally as euhedral crystals in the cavities.

Compiled in Table 4 are the triclinicity, optic angle and partial analyses of potassium feldspar samples taken from different parts of the Viitaniemi (47), Eräpyhä (1) and Pajala pegmatites. The Viitaniemi pegmatite represents Li pegmatites, the Eräpyhä pegmatite Be-Nb pegmatites and the eight metres-wide Pajala pegmatite, north of the Haavikko pegmatite (15), simple pegmatites. They all show well-developed zonality, and the deposits are rather large (8 to 20 m wide). The potassium feldspars from different parts of a deposit vary little in sodium and potassium contents. An exception is the white microperthitic microcline from the replacement body of the Viitaniemi pegmatite. The sodium content of the mineral appears to be conspicuously lower than that of the orthoclase from the contact zone and that of the microcline from the wall zone. Between occurrences, however, the potassium feldspars show considerably larger variations in composition. The sodium content in the potassium feldspars of the Viitaniemi pegmatite is markedly lower than that in the Eräpyhä and Pajala pegmatites.

Rubidium and cesium are generally enriched in the complex pegmatites in the feldspars and micas formed during the final stages of crystallization. In the study area the Rb contents of K-feldspars are fairly high, particularly in the intermediate zone of the Li pegmatites, replacement bodies and fracture fillings (cf. Table 4). Volborth (1956) reports that the microcline in the intermediate zone of the Viitaniemi pegmatite contains up to 0.96 wt% Rb_2O (= 0.88 % Rb). According to Lappalainen and Neuvonen (1968), the microcline from the Juurakko Li pegmatite contains 0.42 wt% Rb_2O. The potassium feldspars in the simple pegmatites of the Eräjärvi area (Ahonen, Männistö and Kultavuori) studied by Virkkunen (1964) assay only 0.07 to 0.08 wt% Rb_2O. In the Mattila (8) and Karvaanmaa (10) pegmatites, which the present author classifies as Be-Nb

Table 5

Partial chemical analyses (oxides: wt%; albite = Ab, orthoclase = Or and anorthite = An: mol%) of plagioclases from the Viitaniemi, Eräpyhä and Pajala pegmatite deposits. Electron microprobe analyses by T. Hautala (wet chemically analysed plagioclases as standards).

	VIITANIEMI			ERÄPYHÄ				PAJALA		
	bz	wz	rb	bz	wz	iz	ff	bz	wz	iz
Na_2O	10.5	11.7	11.7	10.1	11.4	11.6	11.6	0.9	10.2	10.1
K_2O	0.2	0.2	0.1	0.1	0.1	0.1	0.1	0.3	0.2	0.1
CaO	2.2	0.0	0.0	3.1	0.8	0.5	0.3	3.1	2.5	2.3
Ab	88.6	98.9	99.4	85.0	95.7	97.1	98.0	83.8	87.0	87.8
An	10.3	0.0	0.0	14.4	3.7	2.3	1.4	14.5	11.8	11.0
Or	1.1	1.1	0.6	0.6	0.6	0.6	0.6	1.7	1.1	1.1

bz = border zone
wz = wall zone
iz = intermediate zone
rb = replacement body
ff = fracture filling

pegmatites, the corresponding figure is somewhat higher, 0.13 wt% Rb_2O. The above mentioned Rb analyses are not directly comparable with the results in Table 4, because recent studies show that the Rb content of the international G-1 standard sample is much lower than previously supposed (cf. Ahrens 1951 and Fleischer 1965).

Oligoclase and albite

Oligoclase usually occurs in the margins of the pegmatites. The anorthite content of the mineral decreases towards the middle parts of the dykes, which often consist of albite. Oligoclase and albite are usually grey or white in colour. The small euhedral albite crystals that are occasionally met with in the miarolitic cavities are often colourless and with abundant crystal faces.

Oligoclase is always granular, whereas albite may be granular, platy (cleavelandite) or sugary (sugar albite). Cleavelandite and sugar albite often characterize the intermediate zone of the complex pegmatites, the replacement bodies and the fracture fillings.

Table 5 gives partial analyses of albite and oligoclase specimens from different parts of the Viitaniemi (47), Eräpyhä (1) and Pajala pegmatites. Analytical data show that the anorthite content of the mineral varies not only within the same occurrence but also between corresponding zones in different pegmatites. Hence, in the Pajala pegmatite oligoclase is still encountered in the middle of the dyke, whereas in the Viitaniemi pegmatite albite occurs almost from the contact inwards.

The structural state of oligoclase and albite was studied in some samples taken from the contact zones and interiors of the Viitaniemi, Eräpyhä and Pajala pegmatites by applying the (131) peak method ($\triangle = 2\Theta\ (131)—(1\bar{3}1)$; Bambauer *et al.* 1967). X-ray diffractograms show that the minerals studied are low-temperature modifications ($\triangle = 1.09$ to 1.38).

Muscovite

Muscovite, the predominant mica in the pegmatites in the study area, is common throughout the deposits. The mica sheets are

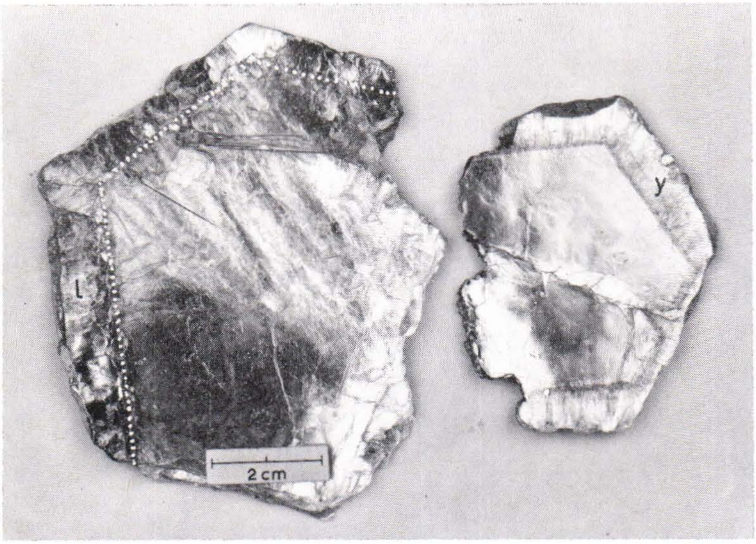


Fig. 11. Zoned mica sheets. Left: Lepidolite rim (1) around yellow muscovite from the Viitaniemi pegmatite (47). Right: Yellow muscovite rim (y) around brownish muscovite sheet from the Juurakko pegmatite (5).

often 2 to 10 cm in diameter in the core of the dykes but considerably more in some of the largest deposits. In the two-mica dykes muscovite seldom appears before the intermediate zone, where it commonly occurs as intergrowths with biotite. As a rule muscovite is pale yellow or pale brown, but exceptionally it may be also pale green, yellow or pink. Occasionally the muscovite sheets are zoned, and in the lithium pegmatites lepidolite often forms a rim around the muscovite sheets (Fig. 11).

Pink muscovite is encountered here and there in the northernmost pit of the Niemelä pegmatite (54) and in the Rauhala pegmatite quarry (50). The mica occurs as a fine-scaled mass (scales ∅ 1 to 5 mm) in cleavelandite fracture fillings. Besides albite and quartz, they also contain small amounts of blue manganiferous apatite, cassiterite and columbite—tantalite. On account of its unusual colour, the pink mica from the Niemelä pegmatite was analysed and studied in detail. The muscovite was separated as a pure fraction with the aid of heavy liquids. The fraction was assayed by wet chemical methods and the results are compiled to the Table 6 together with some other analyses of rose or pink muscovites. The chemical formula of the mica, based on 48 anions and Z = 4, is:

$$(K_{0.81}Na_{0.07}Rb_{0.01})_{0.89}(Al_{1.98}Fe^{+3}{}_{0.01},Mn_{0.01}$$
$$Li_{0.02})_{2.02}(Si_{3.28}Al_{0.72})_{4.00}O_{10.19}((OH)_{1.74}F_{0.07})_{1.81}$$

The mica is close to the ideal muscovite in chemical composition although there is a slight deficiency in alkalies and water + fluorine. The muscovite contains only small amounts of lithium although it looks much like lepidolite.

The space group and polytype of the pink muscovite was determined by a precession camera. 0-, 1-, 2- and 3-level a-, b- and c-axis photographs were prepared with the aid of Zr-filtered Mo radiation. X-ray diffraction studies showed that the mica is monoclinic with space group C2/c and polytype $2M_1$ 0-level a- and c-axis photographs were also prepared for three other crystals to establish the possible existence of other polytypes. All the crystals, however, showed the same extinction rules, and there was no indication of any »extraordinary» reflections in the powder

Table 6

The chemical composition (wt%) of some rose or pink muscovite specimens.

	1	2	3	4	5	6	
SiO_2	45.00	45.10	46.01	47.02	49.90	Si	13.13
TiO_2	0.02	0.56	0.00	—	0.04	Ti	0.00
Al_2O_3	37.41	34.60	35.64	36.83	34.78	Al	10.78
Fe_2O_3	0.12	2.81	0.13	0.51	0.11	Fe^{+3}	0.02
FeO	0.05	0.1	0.00		0.03	Fe^{+2}	0.01
MnO	0.05	0.55*	0.09	1.05	0.17	Mn	0.04
MgO	0.06	0.64	0.04	0.26	0.03	Mg	0.01
CaO	0.03	0.1	1.12	—	0.03	Ca	0.01
Na_2O	0.63	0.21	1.88	—	0.58	Na	0.30
K_2O	10.50	10.4	8.19	9.80	9.66	K	3.24
Li_2O	0.02	—	0.69	0.30	0.06	Li	0.06
Rb_2O	—	0.01	1.20	—	0.29	Rb	0.05
Cs_2O	—	—	0.20	—	0.05	Cs	0.01
BaO	—	0.09	—	—	0.0	Ba	0.00
SrO	—	—	—	—	0.0	Sr	0.00
H_2O+	4.48	4.85	4.65	3.90	3.96	OH	6.95
H_2O-	0.68	—	0.08	—	0.00	F	0.27
F	0.78	—	0.54	0.52	0.33	O	40.77
	99.93	99.94	100.46	100.19	100.02		
$-O = F_2$	0.33	—	0.23	0.22	0.14		
Total	99.60	99.94	100.23	99.97	99.88		

* All manganese recalculated as Mn_2O_3

— Not determined

1. Rose muscovite, Birkeland 2, Iveland, Norway (See Heinrich and Levinson 1953).
2. Pink muscovite, Archer's Post, Kenya (Richardson 1975).
3. Rose muscovite, Varuträsk, Sweden (Berggren 1941).
4. Rose muscovite, Goshen, Mass., USA. (See Heinrich and Levinson 1953).
5. Pink muscovite, the Niemelä pegmatite. Wet chemical analysis by P. Ojanperä.
6. The unit cell content of the Niemelä muscovite on the basis of 48 anions.

pattern either. The lattice parameters listed in Table 7 were calculated from the powder pattern on the basis of nine indexed reflections. The corresponding d-values measured from the powder pattern are similar within the limits of measuring accuracy, and the heights of the peaks also correspond to those reported for normal muscovite (cf. Borg and Smith 1969).

Red or pink muscovites are rare in nature. Most descriptions are from complex pegmatites, in which the mica occurs in replacement bodies formed during the final stages of crystallization, and from hydrothermal manganiferous dykes (Heinrich and Levinson 1953; Askvik 1972; Richardson 1975, 1976; Annersten and Hålenius 1976). The pink muscovite usually contains some iron, manganese, lithium and rubidium (see Table 6), otherwise its physical properties and chemical compo-

Table 7

Optical properties and the unit cell of the pink muscovite from the Niemelä pegmatite.

The unit cell	Optical properties
Monoclinic	$\alpha = 1.559 \pm 0.001$
Space group C2/c	$\beta = 1.590 \pm 0.001$
Polytype $2M_1$	$\gamma = 1.596 \pm 0.001$
a = 5.176 ± 0.004 Å	$\gamma - \alpha = 0.037$
b = 8.974 ± 0.004 Å	$-2V = 48° \pm 1°$
c = 20.108 ± 0.006 Å	$-2V$ calc. = 46.8°
$\beta = 95°49'$	
V = 929.322 Å^3	
G = 2.863	
D = 2.868 g/cm^3	

sition are close to those of the ideal muscovite (Heinrich and Levinson 1953). Chemical analyses show that such micas contain very little ferrous iron but always a certain amount of ferric iron and manganese, occasionally even several percent. Hence Mn^{+3} and Fe^{+3} ions are widely considered to be chromophors of the mineral (Deer, Howie and Zussman 1967, p. 21; Richardson 1975; Annersten and Hålenius 1976). The muscovite of the Niemelä pegmatite is nearly similar in chemical composition and physical properties to the pink muscovite reported in the literature. The pink colour of the mica is also probably due to the small amounts of iron or manganese or both, in the mineral.

Biotite

Biotite usually only occurs in dykes associated with granites and in their contact zones. The mica is brown in colour and occurs either as separate sheets or as intergrowths with muscovite. The biotite-rich pegmatites usually contain only small amounts of or are entirely free from tourmaline and garnet. The Li pegmatites in the study area contain no biotite at all. Some Be-Nb pegmatites have biotite as their main mineral together with feldspars and quartz in the border, wall and occasionally also in the intermediate zone. Biotite is totally lacking from the albite rich replacement bodies and fracture fillings.

Tourmaline

Tourmaline, which is often schorl (black tourmaline) in the pegmatites of the area, is a typical mineral, particularly in the dykes in schists. The schorl crystals are often 0.5 to 3.0 cm in diameter, although considerably larger crystals (∅ up to 12 cm) have been encountered in some pegmatites. The crystals often show distinct prism faces whereas pyramid faces are rare. In the interior of the deposits the crystals tend to be long but occasionally bent and broken; in the contacts they are short and locally cone-shaped. In some pegmatites tourmaline crystals are intensely altered and the pseudomorphs consist of fine-scaled yellow muscovite.

In addition to schorl, the cores of the deposits may contain green, red or blue tourmaline. Dark green tourmaline occurs together with various minerals and occasionally even close to the margins of the deposits. In contrast, red tourmaline is usually only met with in the interior of the pegmatites in association with lepidolite, montebrasite and spodumene. Blue tourmaline is sometimes found as small clusters of crystals in the margins of triphylite—lithiophilite aggregates. The red, blue and green tourmaline crystals are often cloudy; transparent green gem-quality tourmaline has only been encountered in the Viitaniemi pegmatite. The tourmaline crystals in the Seppälänranta (31), Talas (30) and Katila (6) pegmatites occasionally show transversal or longitudinal zonality, or both. The core of the crystals is then dark green and the margin red. The outermost rim may sometimes still be pale green. The colour in the crystals with longitudinal zoning changes in the corresponding way as in the crystals with transversal zoning.

Electron microprobe determinations on the tourmaline from the Seppälänranta pegmatite (31) show that the red and pale green variants are elbaite. They contain only very small amounts of iron (as well as manganese, calcium and magnesium). The blue and dark green tourmaline is often considerably richer in iron and closer to schorl in composition.

Garnet

Garnet is frequently reddish brown or pink and usually occurs as small (∅ 0.4 to 5.0 mm) subhedral or euhedral crystals. The garnet crystals in the interior of the complex pegmatites are commonly large, sometimes up to several centimetres in diameter. Some semiquantitative analyses, cell parameters and refractive indices (Lahti 1974) suggest that the garnets are either almandine or spessartine. The composition of the garnets varies between and within the pegmatites. The garnets richest in manganese occur in the interiors, which were the last to crystallize, whereas almandine favours the margins of the deposits.

Beryl

Beryl occurs in the pegmatites of the area as yellowish, green-white or pure white, sometimes pink crystals. They often show distinct prism faces, whereas basal pinacoids and pyramid faces are rare. The beryl crystals are up to 15 cm in diameter in the interior of the larger pegmatites but markedly smaller close to the contacts and in narrow dykes. In the Eräpyhä pegmatite (1) the beryl crystals may be abnormally large, even as much as 40 cm in diameter.

Pink beryl, which is often rich in Cs, usually occurs in association with Li minerals or as a beryl variant in the parts of deposits that were the last to crystallize. In places the crystals are zoned, yellow beryl being surrounded by a zone of pink beryl. Usually, however, the pink beryl occurs as individual and short euhedral to anhedral crystals in cleavelandite together with lepidolite, amblygonite-montebrasite, tantalite and topaz. X-ray fluorescence analyses show that the yellow beryl in the Viitaniemi pegmatite contains 0.18 wt% but the pink beryl 0.66 wt% Cs (cf. Volborth 1956, p. 281).

Bertrandite

Due to the influence of supercritical fluids or hydrothermal solutions, or both, the beryl crystals have occasionally been altered either partly or completely into other minerals. The crystals have usually preserved their shapes, and the beryl is replaced by bertrandite, various aluminium-rich minerals (such as micas, feldspars and clay minerals), fluorite, quartz and apatite (Fig. 12). The Viitaniemi pegmatite often has beryllium-bearing phosphates

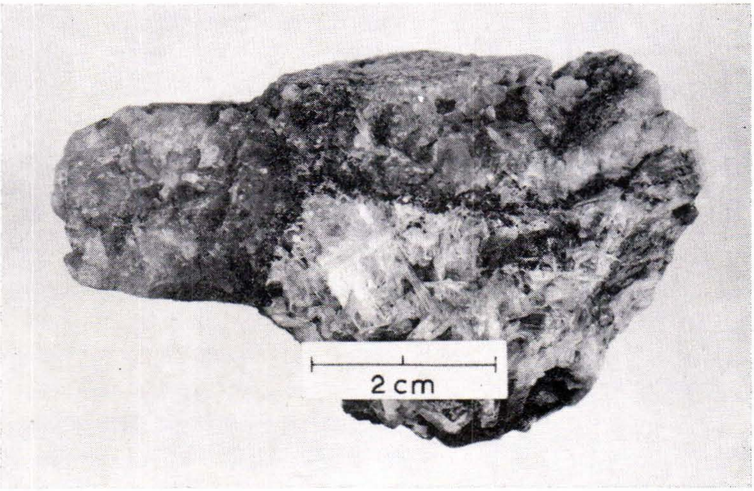


Fig. 12. Bertrandite crystal plates in the pseudomorph after beryl (lower part). Upper part of the sample consists of quartz. The Juurakko pegmatite (5).

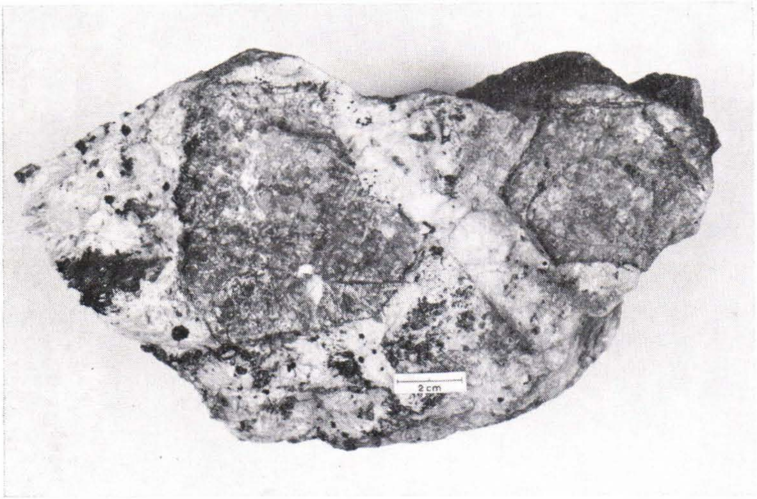


Fig. 13. Two pseudomorphs after beryl filled with fine-grained väyrynenite, herderite, apatite and muscovite. The white mineral around the pseudomorphs is cleavelandite. The black roundish grains are thucholite. The Viitaniemi pegmatite (47).

instead of bertrandite (Fig. 13). The pseudomorphs are frequently so fine-grained that the minerals can seldom be identified with the naked eye. Thin sections show that bertrandite occurs either as individual small tabular crystals or as spherulites in association with other alteration products. The alteration of beryl has not always reached completion, and bertrandite may occur as a rim around the beryl crystals or in fractures. In some places there are small cavities in the pseudomorphs where bertrandite occurs as euhedral crystals, often together with fluorite that is sometimes an abundant constituent in the alteration products.

Bertrandite is exceptionally abundant in the Juurakko pegmatite (5), where it often occurs as tabular crystals flattened along {001} and measuring several centimetres, and as crystal aggregates (Fig. 12). The mineral is invariably colourless, although the crystal plates often show a thin coating of brown limonite. In some miarolitic cavities bertrandite occurs as euhedral or subhedral transparent crystals. The largest euhedral bertrandite crystal encountered in the Juurakko occurrence measures 11 mm; it shows faces {001}, {010}, {100} and {110}, and cleavages along {110}, {001} and {010}.

The Juurakko bertrandite was analysed by wet chemical methods. Table 8 gives the chemical analysis of the mineral compared with some bertrandite analyses reported in

Table 8

The chemical composition (wt%) of bertrandite.

	1	2	3	4	
SiO_2	49.63	55.37	50.80	Si	8.05
Al_2O_3	—	0.86	0.04	Al	0.01
Fe_2O_3	—	0.40	0.08	Fe^{+3}	0.01
FeO	—	—	0.10	Fe^{+2}	0.01
BeO	41.60	35.07	41.44	Be	15.78
MgO	—	0.29	0.00	Mg	0.00
CaO	—	1.34	0.00	Ca	0.00
Na_2O	—	—	0.06	Na	0.02
K_2O	—	—	0.03	K	0.01
H_2O+	8.65	7.14	7.67	OH	8.11
H_2O-	—	—	0.01	O	27.89
Total	99.88	100.47	100.23		

— Not determined

1. Bertrandite, Kolsva, Västmanland, Sweden (Mårtensson 1960).
2. Bertrandite, Zabaikalja, Sovjet Union (Grigor'ev and Dolomanova 1955).
3. Bertrandite, Juurakko. Wet chemical analysis by P. Ojanperä.
4. The unit cell content of the Juurakko bertrandite on the basis of 36 anions.

Table 9

X-ray powder diffraction data of bertrandite:

1. Bertrandite, Juurakko. Diffractometer, scanning speed 1/4°/min., Ni-filtered Cu radiation, silicon as an internal standard. The reflections marked with asterisks were used in calculating unit cell dimensions. b = broad peak.
2. Bertrandite, Mica Creek, Queensland. Camera, diameter 114.6 mm (Vernon and Williams 1960, see also JCPDS-card 17-515).

hkl	1			2	
	d meas.	d calc.	I	d	I
110 *	7.566	7.565	16	7.56	10
130 *	4.392	4.392	44	4.85	< 2
200 *	4.356	4.356	100	4.38	100
021, 111	3.916	3.920, 3.911	40	3.94	40
040 *	3.814	3.814	23	3.80	7
131 *	3.166	3.166	70	3.19	90
221, 041 b	2.914	2.914, 2.928	< 5	2.93	10
150	2.880	2.880	7	2.88	10
240	2.870	2.869	8	2.54	80
310	2.853	2.853	29	2.42	5
060 *	2.543	2.543	31	2.28	60
330 *	2.522	2.522	68	2.22	60
311 *	2.420	2.420	8	2.18	2
002 *	2.285	2.285	42	2.10	3
061 *	2.222	2.222	46	2.02	5
331, 260	2.207	2.208, 2.196	17	1.983	20
400 *	2.178	2.178	11	1.923	5
350	2.104	2.104	8	1.787	5
132 b	2.027	2.027	5	1.698	15
261 *	1.979	1.979	23	1.650	10
042	1.960	1.960	9	1.628	< 2
351, 080 b	1.910	1.911, 1.907	8	1.579	3
312, 242	1.785	1.783, 1.787	< 5	1.555	30
062 b	1.700	1.699	5	1.491	< 2
332 b	1.694	1.693	6	1.465	30
190 b	1.665	1.664	5	1.440	10
530 *	1.648	1.654	11	1.363	2
371 b	1.629	1.629	6	1.338	3
402 b	1.576	1.576	5	1.305	40
191	1.563	1.563	7	1.251	10
461	1.555	1.555	7	1.233	5
531	1.552	1.551	7	1.220	20 b
352 b	1.549	1.547	5	1.167	10 b

the literature. The chemical formula of the Juurakko bertrandite (when Z = 4) is:

$Be_{3.95}Na_{0.01}Si_{2.01}O_{6.97}(OH)_{2.03}$

The mineral is thus very close to the ideal composition of bertrandite: $Be_4Si_2O_7(OH)_2$.

On the basis of b- and c-axis zero- and first-level precession photographs the mineral is orthorhombic with space group $Cmc2_1$. The unit cell parameters were computed from the indexed powder pattern (cf. Borg and Smith 1969) on the basis of the 12 sharpest reflections. The measured and calculated d-values arrived at are presented in Table 9, together with the corresponding values for bertrandite of Mica Creek (Vernon and Williams 1960). The cell dimensions obtained were:

$a = 8.712 \pm 0.005$ Å
$b = 15.255 \pm 0.005$ Å
$c = 4.569 \pm 0.003$ Å
$V = 607.239$ Å^3

Table 10

Optical properties of bertrandite.

	1	2	3	4
α	1.587 ± 0.001	1.589 ± 0.003	1.589	1.589
β	1.606 ± 0.001	1.603 ± 0.002	1.604	1.604
γ	1.615 ± 0.001	1.613 ± 0.002	1.612	1.612
$\gamma - \alpha$	0.028	0.024 (0.025*)	0.023	0.023
—2V° meas.	70 ± 1	—	72	72
—2V° calc.**	68.4	79.8	71.7	71.7

* Using Berek compensator
** Calculated by the present author
— Not determined
1. Bertrandite, Juurakko.
2. Bertrandite, Mica Creek, Queensland (Vernon and Williams 1960).
3. Bertrandite, Siberia (Getmanskaja 1966).
4. Bertrandite, Ilimaussaq alkaline intrusion, South Greenland (Andersen 1967).

The unit cell dimensions deviate somewhat from those ($a = 8.73$ Å, $b = 15.31$ Å, $c = 4.56$ Å, $V = 609.47$ Å^3) reported by Solov'eva and Belov (1961). The optical properties of the mineral are given in Table 10. The optic angle and the refractive indices do not differ significantly from those reported in the literature. The specific gravity of the mineral obtained with heavy liquids is 2.595. The calculated density for the mineral is 2.608 g/cm^3, a value that is very close to the measured specific gravity.

Bertrandite often has been described from greisens and pegmatites. The mineral occurs in pseudomorphs after beryl or as euhedral crystals in miarolitic cavities (Vlasov 1966 p. 92; Černý 1968; Haapala and Ojanperä 1972; Henderson 1975). The mineral composition of the pseudomorphs varies greatly, depending on the composition and pH of the postmagmatic decomposing solutions (Govorov and Stunzhas 1963; Bukin 1967; Černý 1968). The pseudomorphs in the study area are very similar in mineral composition. The aluminium of beryl is incorporated in various micas, clay minerals and feldspars. Beryllium has entered into bertrandite, bityite (the Maantienvarsi pegmatite, 41) and, in the phosphorus-rich Viitaniemi pegmatite (47), into a variety of beryllium-bearing phosphates. The abundant fluorite and other fluorine-bearing minerals, which are often encountered in association with the Be minerals, suggest that the concentration of fluorine in the fluids and hydrothermal solutions that led to the decomposition of the beryl crystals must have been appreciable.

Lepidolite

Lepidolite is met with in several Li pegmatites in the study area as either a fine-scaled mass or as larger individual sheets. It is a reddish purple mineral except in the Viitaniemi pegmatite where it is reddish brown. Electron microprobe analyses show that the Viitaniemi lepidolite is rather rich in iron and manganese, containing 2.0 wt% FeO (total iron) and 1.5 wt% MnO. The large mica sheets in the lithium pegmatites are frequently zoned lepidolite forming a narrow rim around muscovite (Fig. 11). Powder patterns show that lepidolite from different quarries includes several polytypes; without single-crystal studies, however, they cannot be identified.

Spodumene

Spodumene has been encountered in no more than four pegmatites, and even then only rarely. It is a grey or white mineral and occurs in the interior of the pegmatites as either fine-grained aggregates or as individual anhedral crystals several centimetres long. In the Maantienvarsi pegmatite (41) spodumene occurs, exceptionally, intergrown with quartz in the middle of the deposit. In places the spodumene laths are intensely altered, and clay minerals, muscovite and cookeite have been identified from the pseudomorphs.

Topaz

Topaz occurs in the Viitaniemi (47), Seppälänranta (31) and Juurakko (15) pegmatites in association with Li minerals. It is rare in the latter two deposits but very common in the Viitaniemi pegmatite. The topaz crystals are generally subhedral or euhedral and bluish-green, green or grey in colour. Topaz is frequently clouded, although occasionally transparent bluish-green euhedral crystals have been found in the miarolitic cavities of the Viitaniemi pegmatite and in association with some low-temperature minerals (clay minerals, eosphorite, väyrynenite, herderite etc.). The large topaz crystals are often zoned, the core being bluish-green, the outer margin grey and the outermost surface occasionally transparent.

Zircon

Zircon occurs as subrounded brown crystals throughout the pegmatites. It is often enriched in the interior of the deposits, where the crystals are subhedral or euhedral and up to 10 mm in diameter. When in association with thucholite, cassiterite, tantalite and lepidolite, euhedral zircon crystals with abundant crystal faces are common in the Viitaniemi pegmatite. Thin section examination shows that the largest crystals are very heterogenous and contain dark spots or alternating light and dark zones. Zircon is usually metamict and can be identified by X-ray diffractometer only after heating (cf. Lima de Faria 1964).

Thorite

Thorite occurs in the Viitaniemi (47), Eräpyhä (1) and Leikattu (7) pegmatites as epoxy-like anhedral crystals, dark brown or black in colour, and often in association with columbite—tantalite, cassiterite and zircon. Thorite is common in the Leikattu pegmatite, where it is encountered as unusually large aggregates up to several centimetres in diameter; it is rare in the other deposits. The samples studied were metamict, and the minerals were identified by X-ray diffractometer after heat treatment.

Pollucite

Pollucite has only been discovered in the Viitaniemi pegmatite (47), where it has formed a large core, several tons in weight, in the eastern part of the pegmatite. It is a white mineral or, when altered, pale brown. The chemical composition of the Viitaniemi pollucite has been reported by Erämetsä and Sihvonen (1973, see also Mikkonen 1963). According to them the fresh mineral contains about 18 mol% analcime as an isomorphous mixture ($Na_2O = 1.64$ wt% and $Cs_2O = 33.85$ wt%).

Chlorite and clay minerals

Chlorite group minerals and clay minerals are usually alteration products of various

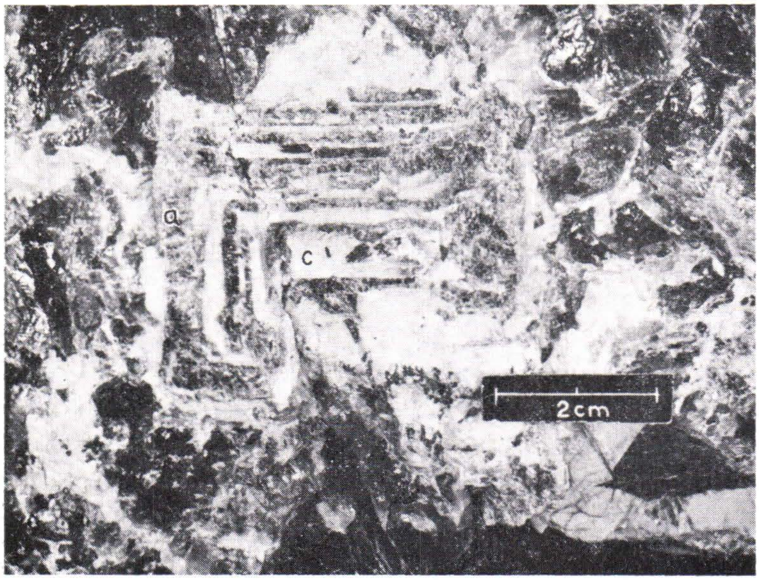


Fig. 14. Cleavelandite plates (c, white) parallel to prism and basal planes of a fluorapatite crystal (a, grey). The c-axis of apatite is horizontal in the figure. The black mineral around apatite is quartz with white cleavelandite. The Niemelä pegmatite (54).

aluminous minerals. Cookeite is the only mineral of the chlorite group to have been encountered sporadically among the alteration products of spodumene. In the Viitaniemi pegmatite, however, cookeite also occurs as a fine-scaled, pale green mass in cavities together with other low-temperature minerals. Black, iron-rich chlorite often forms thin reaction rims around sulphides. In the interior of the Myllyrinne pegmatite (59), dark green chlorite is met with as narrow fracture fillings that often contain small väyrynenite crystals as well. The clay minerals, kaolinite and montmorillonite, which are the usual alteration products of feldspars, montebrasite and beryl, occur in association with the latter or in miarolitic cavities.

Bityite

Bityite has only been encountered in the Maantienvarsi pegmatite (41). The mineral occurs as small yellowish-white scales in pseudomorphs after beryl together with bertrandite and fluorite. Bityite has been identified on the basis of the diffractograms and partial chemical analyses ($Li_2O = 2.32$ and $BeO = 7.21$ wt%).

Apatite-group phosphates

Apatite, the predominant phosphate mineral in the pegmatites of the study area, is ubiquitous in the dykes. Its abundance is highest in the interior of the deposits, where it occurs as subhedral or anhedral crystals (∅ usually from 1 to 5 cm) or as crystal aggregates. The complex pegmatites in particular are occasionally rich in apatite, and the crystal aggregates may weigh as much as several kilograms. In some occurrences the apatite crystals contain cleavelandite and columbite—tantalite laths as inclusions that are sometimes oriented along the prismatic, basal and pyramidal planes (Fig. 14).

Apatite is usually pale green. Its colour varies considerably, however, particularly in miarolitic cavities and with the manganiferous minerals (see Fig. 15). The bluish green apatite analysed by Volborth (1954a) from the

Viitaniemi pegmatite is manganiferous fluorapatite with a fairly high (3.01 wt%) fluorine content. As shown by the refractive indices determined by the present author, the common green apatite is also fluorapatite. The apatite in the interior of the lithium pegmatites that contain manganese-bearing minerals is generally bluish-green or blue, sometimes black, grey, yellow or brown in colour and often shows a marked manganese content. In places the crystals are zoned, the core being bluish-green and manganiferous, and the margins pale blue or grey and poor in manganese. Electron microprobe determinations show that the bluish green apatite from the Viitaniemi (47) and Niemelä (54) pegmatites contains up to 9.0 wt% MnO and 0.1 wt% FeO. The corresponding figures for the apatite from the Seppälänranta pegmatite (31) are 9.8 wt% MnO and 2.1 wt% FeO.

In some pegmatite cavities apatite occurs as euhedral crystals rich in faces. In the Viitaniemi pegmatite with its abundant miarolitic cavities, apatite occurs as colourless or purple and commonly zoned crystals together with fluorite, kaoline and quartz (apatite IV according to Volborth 1954a).

Iron- and manganese-bearing phosphates

Altogether 29 phosphate minerals have been encountered in the pegmatites in the study area, of which 21 contain iron or manganese, or both, as a main component (Fig. 15). The primary Fe-Mn phosphates are lithiophilite, triphylite, triplite (Viitaniemi) and beusite (Jussinvuori). The apatite associated with the above mentioned minerals may also contain appreciable amounts of manganese and some iron. In terms of species, the largest phosphate mineral group is that of secondary phosphates, which are alteration and replacement products of lithiophilite and triphylite. They occur, albeit in rather small amounts, in the margins and cleavages of lithiophilite—triphylite aggregates (Fig. 16). In some places close to the surface, however, triphylite and lithiophilite are completely altered and their zoned pseudomorphs are composed of various secondary phosphates. Small crystal cavities are fairly common in the alteration products and apart from Fe-Mn phosphate minerals, they may also contain apatite, calcite, siderite and amorphous iron- and manganese-bearing minerals (e.g. limonite and wad).

Several secondary phosphates contain variable amounts of (OH) groups or molecular water or both. In most of them iron is ferrous, and ferric iron occurs only in the phosphates that were the last to crystallize. The valence of manganese is +3 only in heterosite and purpurite; in the other minerals it is +2. The yellow and brown amorphous oxides, hydroxides and phosphates that formed as final alteration products in cavities and around the pseudomorphs also contain ferric iron and trivalent manganese.

In the phosphates that contain water or hydroxyl groups, or both, iron and manganese tend to form minerals of their own (cf. Fig. 15). The acidity (pH) of the hydrothermal solutions that decomposed triphylite and lithiophilite, the difference between iron and manganese in redox potential and element concentrations in solutions were obviously the primary factors controlling the crystallization of the secondary phosphates. Many of the Fe-Mn phosphates that contain molecular water can crystallize under the conditions prevailing on the surface (Nriagu and Dell 1974; Warry and Kramer 1976). Some of the secondary phosphates in the research area may also be of supergene origin (cf. Moore 1973).

Triphylite—lithiophilite together with their replacement and alteration products are described only briefly. The eosphorite from the Viitaniemi pegmatite is the only Fe-Mn phosphate described in detail. The mineral is

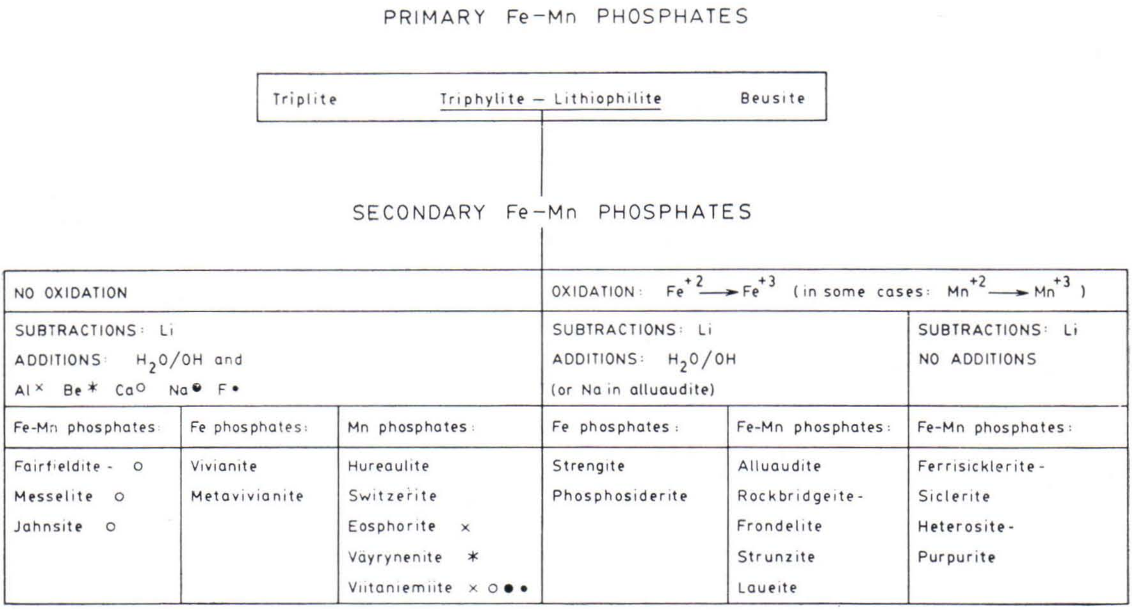


Fig. 15. The primary and secondary Fe-Mn phosphates of the pegmatites in the Eräjärvi area. The scheme shows the alteration of triphylite-lithiophilite into secondary iron, manganese and iron-manganese phosphates under different conditions. Viitaniemiite has not been found in close association with triphylite-lithiophilite.

often paragenetically associated with amblygonite–montebrasite and, like them, contains aluminium in its formula.

Triphylite–lithiophilite

Triphylite–lithiophilite often occurs in the pegmatites as subrounded nodules or crystal aggregates (Fig. 16) up to 30 cm in diameter. In the Seppälänranta (31) pegmatite the mineral is sometimes encountered as subhedral crystals in cleavelandite as well. Triphylite and lithiophilite are grey or in the Viitaniemi pegmatite sometimes pale pinkish brown. The small amounts of alteration products often impart bluish or greenish hues to the mineral.

X-ray fluorescence and electron microprobe analyses of several samples demonstrate that the iron and manganese content of the mineral varies markedly between occurrences. The Katila (16), Maantienvarsi (41), Keskimetsä (46), Suonlaita (44) and Viitaniemi (47) pegmatites contain lithiophilite, whereas the Seppälänranta (31), Talas (30), Lahdensuu (30), Männikkö (35), Pellonreuna (39) and Karpalepohja (60) pegmatites contain triphylite. The lithiophilite richest in manganese (80 mol⁰/₀ $LiMnPO_4$) was encountered in the Katila pegmatite and the triphylite richest in iron (77 mol⁰/₀ $LiFePO_4$) in the Talas pegmatite dyke.

According to Volborth (1954a), the Viitaniemi pegmatite has three lithiophilite generations, which differ from each other in colour, mode of occurrence and composition. The present author also discovered triphylite (53 mol⁰/₀ $LiFePO_4$, microprobe determination) as a greenish grey rim around dark brown triplite aggregates and as narrow lamellae in triplite.

The alluaudite group

The alluaudite group of minerals (see Moore 1971) usually occur in the margins of

Fig. 16. Lithiophilite aggregate (l, lithiophilite, white or grey) partly altered into sicklerite and purpurite (p, dark rim). Alteration has proceeded along cleavage planes. Polished sample. The Keskimetsä pegmatite (46).

the triphylite—lithiophilite aggregates as a green and fine-grained mass that replaces the host mineral along the cleavages up to the interior of the aggregates. Alluaudite was usually formed before crystallization of the water-bearing secondary phosphates, which often constitute the outermost zone in the alluaudite-bearing triphylite—lithiophilite aggregates. In some cases the crystallization of alluaudite has obviously continued to very low temperatures, because it also occurs in cavities together with H_2O- and OH-bearing phosphates.

Ferrisicklerite—sicklerite and heterosite—purpurite

Ferrisicklerite—sicklerite and heterosite—purpurite always occur on or close to the surface of outcrops as alteration products of triphylite—lithiophilite. Ferrisicklerite—sicklerite occurs as a narrow rim between heterosite—purpurite and triphylite—lithiophilite. The alteration products often form around the aggregates a narrow purple or brown shell that is usually no more than a few millimetres or in some cases a few centimetres thick (Fig. 16). According to Volborth (1954a), purpurite and sicklerite have not been encountered in the Viitaniemi pegmatite deeper than 2 to 4 m. In the other deposits the minerals usually occur only near the surface. Thus, most ferrisicklerite—sicklerite and purpurite—heterosite were evidently formed as supergene alteration products of triphylite—lithiophilite.

Vivianite, hureaulite, fairfieldite—messelite and jahnsite

Vivianite, hureaulite, fairfieldite—messelite and jahnsite often occur together. The three first mentioned secondary phosphates are very common. Vivianite, in particular, is frequently a predominant mineral in pseudomorphs after triphylite—lithiophilite and in cavities filled with alteration products. Vivianite is blue or green and, like the others, occurs as either euhedral crystals (1 to 3 mm long) or as a fine-grained mass. Hureaulite and jahnsite occur in places between the vivianite crystals as tiny reddish brown, yellowish brown or dark brown crystals that cannot be distinguished by colour.

Jahnsite is considerably more rare than hureaulite, which, together with vivianite, is often one of the main minerals in cavities. Jahnsite was identified on the basis of a powder pattern. The d-values in samples from different occurrences were almost identical to those reported by Moore and Ito (1978) for jahnsite in the Fletcher pegmatite (U.S.A.). Fairfieldite—messelite occurs in cavities as colourless or white crystals (1 to 3 mm long) and sometimes as a fine-grained white mass in association with other Fe-Mn phosphates. In places it is found together with small crystals of white or colourless apatite.

Switzerite

Switzerite, which has been encountered in numerous lithium pegmatites in the area, but always in small abundances, is a reddish brown mineral that resembles mica under the microscope. It occurs on cleavage surfaces in lithiophilite, in the margins of aggregates and in cavities. It often forms narrow seams between lithiophilite and the other secondary phosphates (vivianite, fairfieldite, hureaulite, metavivianite).

Rockbridgeite—frondelite

Rockbridgeite—frondelite was identified from the Viitaniemi (47) and Maantienvarsi (41) pegmatites. In both deposits the mineral appears as a green, fine-grained rim associated with hureaulite in the margin of lithiophilite nodules.

Metavivianite and reddingite

Metavivianite and reddingite were identified from the Viitaniemi (47) and Katila (6) pegmatites. Metavivianite occurs as rather small, green subhedral crystals together with vivianite, fairfieldite, hureaulite and siderite. Tiny reddish brown reddingite crystals are associated with metavivianite in both occurrences.

Strunzite and laueite

Strunzite and laueite have only been met with in two samples taken from the Viitaniemi (47) pegmatite. Strunzite occurs as small, yellow acicular crystals and crystal aggregates in the cavities of lithiophilite aggregates. Associated with it are reddish brown laueite and hureaulite as separate crystals (1 to 2 mm long) and vivianite as somewhat larger crystals.

Phosphosiderite and strengite

Phosphosiderite and strengite have only been identified from the Viitaniemi (47) pegmatite. Both occur together with yellow amorphous secondary phosphates as yellowish alteration products around the lithiophilite aggregate.

Beusite

Beusite has been encountered only in the Jussinvuori pegmatite (24). The beusite grains are small, about 0.5 cm in diameter. The

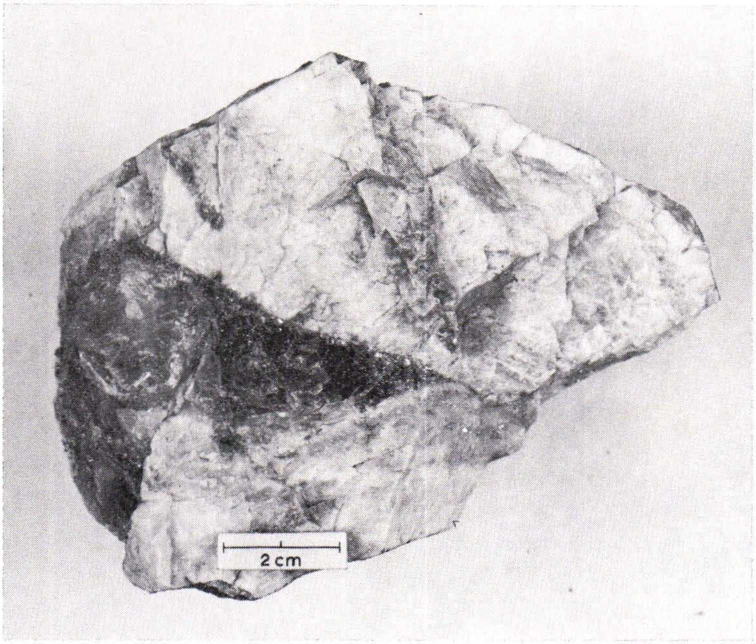


Fig. 17. Eosphorite (black) as a cavity filling between montebrasite crystal plates (white). The Viitaniemi pegmatite (47).

mineral is intergrown with triphylite—lithiophilite (Fe:Mn = 1.0—1.1), which occurs as thin lamellae (thickness 0.01—0.1 mm) in beusite (Fe:Mn = 0.6, microprobe determination).

Triplite

Triplite has been encountered in the study area only in the Viitaniemi pegmatite, where it occurs as nodules and aggregates of variable size (∅ up to 20 cm). It is either dark brown (triplite I) or reddish brown (triplite II, Volborth 1954a). The reddish brown triplite often occurs as inclusions in lithiophilite, whereas the dark brown triplite is associated with triphylite. Electron microprobe analyses show that the reddish brown triplite at Viitaniemi is markedly poorer in iron and richer in manganese (Fe:Mn = 0.14) than the dark brown triplite (Fe:Mn = 0.37). Volborth (1954a) has reported manganese oxide (wad), vivianite and strengite as alteration products of triplite. Usually, however, triplite is very resistant and the secondary phosphates associated with it probably derive from the altered triphylite—lithiophilite lamellae and inclusions in the mineral. Hence, in Fig. 10, which shows the phosphate minerals in the Viitaniemi pegmatite, triplite has been joined with a dashed line to the box of the secondary Fe-Mn phosphates.

Eosphorite

Eosphorite, the Mn-member of the childrenite—eosphorite series $(Fe,Mn)AlPO_4(OH)_2 \cdot H_2O$, has been encountered in the study area only in the Viitaniemi pegmatite (47). It is rather common and usually occurs as nodules or as irregular cavity (Fig. 17) and fracture fillings. The mineral often seems to have formed through the replacement of montebrasite by manganese-rich hydrothermal solutions.

Eosphorite has never been encountered to-

Table 11

Comparison between the physical properties and the unit cell dimensions of eosphorite from various localities.

	1	2	3	4
α	1.629	1.629 ± 0.001	1.638	1.644
β	1.650	1.657 ± 0.001	1.660	1.662
γ	1.658	1.663 ± 0.001	1.667	1.671
$\gamma - \alpha$	0.029	0.034	0.029	0.027
—2V° meas.	45—50	48 ± 1	50	∼ 50
—2V° calc.	62.7 *	49.7 *	58.2 *	69.9 *
a (Å)	—	10.452 ± 0.003	10.45	10.41
b (Å)	—	13.510 ± 0.003	13.49	13.42
c (Å)	—	6.936 ± 0.003	6.93	6.92
V ($Å^3$)	—	979.408	976.93 *	966.74 *
G	3.10	3.052	3.06—3.08	3.11—3.15
FeO wt%	1.38	2.18 (tot.)	10.10	14.86
MnO w%	29.94	24.30	—	15.17

* Calculated by the present author
— Not determined
1. Eosphorite, Buckfield, Maine (see Hurlbut 1950).
2. Eosphorite, Viitaniemi.
3. Eosphorite, Newry, Maine (see Hurlbut 1950).
4. Childro—eosphorite, Hagendorf, Germany (Strunz and Fischer 1957).

gether with triphylite—lithiophilite or triplite and it is very rare in the pseudomorphs after these minerals. The only iron- and manganese-bearing phosphates that are closely associated with it are viitaniemiite and väyrynenite, both of which crystallized before eosphorite. The Viitaniemi eosphorite sometimes shows subhedral topaz and väyrynenite crystals as inclusions. Other low-temperature minerals that are also often associated with eosphorite are purple or colourless fluorapatite, fine-scaled green micas, fluorite and kaolinite.

Pure eosphorite is pink or orange in colour, although small inclusions in the mineral often tinge it brown or grey. It shows one rather distinct cleavage, which, according to precession photographs, is parallel to the {100} plane. Measurements on universal stage show that the axial plane of the Viitaniemi eosphorite is parallel to or very close to the cleavage. It was not possible to determine accurately the position of the axial plane, because the cleavages in the mineral are always somewhat uneven and no euhedral, not even subhedral, eosphorite crystals were found for studies.

According to Hurlbut (1950), and Strunz and Fischer (1957), there is a discrepancy between the optical and X-ray diffraction data on eosphorite. X-ray diffraction studies (Hurlbut 1950; Barnes and Shore 1951) show that eosphorite is orthorhombic with the space group Bbam. Optical determinations, however, suggest that eosphorite is monoclinic. The optic axial plane of the mineral deviates slightly from the {100} plane. Hence Z∧c varies between 3° and 8°, depending on the Fe/Mn ratio of the mineral. Hanson (1960) analysed the crystal structure of the mineral in detail and came to the conclusion that the space group is Bbam. Referring to the optical studies, however, he points out that the orthorhombic structure shown by the X-ray

Table 12

X-ray powder diffraction data of eosphorite:

1. Eosphorite, Viitaniemi. Diffractometer, scanning speed 1/4°/min., Mn-filtered Fe radiation, silicon as an internal standard. The reflections marked with asterisks were used in calculating unit cell dimensions.
2. Eosphorite, Newry, Maine. Camera, diameter 114.6 mm, Ni-filtered Cu radiation. The FeO content of the mineral is 10.1 wt%. Data from JCPDS card 17—131. b = broad peak.

hkl	1			2	
	d meas.	d calc.	I	d	I
020 *	6.758	6.755	17	6.75	40
111 *	5.313	5.314	13	5.23	50
200 *	5.226	5.226	60	4.87	10
210	4.874	4.874	10	4.39	50
121 *	4.391	4.392	44	4.15	30
220 *	4.133	4.133	35	3.55	50
131 *	3.552	3.552	34	3.41	50
002	3.466	3.468	7	2.826	100
230 *	3.411	3.412	27	2.610	5
040	3.378	3.378	10	2.422	60 b
022 *	3.085	3.085	21	2.253	10
141	2.916	2.916	10	2.077	30
212, 321	2.826	2.826, 2.827	100	1.989	30
400 *	2.613	2.613	19	1.885	30
410, 331	2.563	2.565, 2.561	< 10	1.811	10
151	2.448	2.448	10	1.778	10
232	2.431	2.432	20	1.734	20
250 *	2.400	2.400	25	1.687	5
060	2.255	2.252	12	1.535	50 b
402 *	2.087	2.087	16	1.469	5
260, 440	2.068	2.068, 2.067	11	1.412	10
422 *	1.994	1.994	10		
143	1.877	1.877	12		
262	1.776	1.776	< 10		
004	1.736	1.734	10		

diffraction data may be an average structure and that the mineral is really monoclinic (with $\beta = 90°$).

0-, 1-, 2- and 3- level a- and b-axis precession photographs were prepared for the eosphorite of the Viitaniemi pegmatite. Further three 0-level a-axis photographs were taken with exposure times of 20, 50 and 100 hours. In spite of overexposure examination of the precession photographs revealed no deviation from orthorhombic symmetry, and the systematically absent reflections showed that the space group of the mineral is Bbam (or Bba2). An X-ray diffractogram was obtained for the mineral with silicon as internal standard. The cell parameters in Table 11 were calculated from the powder patterns; the observed and calculated d-values of the mineral are given in Table 12 together with the corresponding values for eosphorite from Newry, Maine (see JCPDS card 17—131).

The optical properties of the Viitaniemi eosphorite are given in Table 11 together with the corresponding properties for some other eosphorite specimens described in the literature. The chemical composition was establisbed by wet chemical methods; the analytical data are shown in Table 13 with some other eosphorite analyses. The chemical formula of the mineral, calculated on the basis of unit cell volume and density (when Z = 8), is:

Table 13

Chemical analyses (wt%) of eosphorite.

	1	2	3	4	5	
SiO_2	—	0.16	—	—	Si	0.05
Al_2O_3	22.37	24.50	20.51	19.13	Al	8.65
Fe_2O_3	—	0.22	—	0.02	Fe^{+3}	0.05
FeO	1.38	1.98	3.74	14.86	Fe^{+2}	0.50
MnO	29.94	24.30	27.65	15.17	Mn	6.17
MgO	—	0.00	—	0.12	Mg	0.00
CaO	—	0.06	0.84	3.58	Ca	0.02
Na_2O	—	0.15	—	—	Na	0.04
Li_2O	—	0.96	—	—	Li	1.16
K_2O	—	0.02	—	—	K	0.00
P_2O_5	29.89	32.90	30.38	30.29	P	7.80
F	trace	0.41	—	—	F	0.39
H_2O+	15.34	14.30	15.59	14.94	OH	15.61
H_2O-	—	0.02	—	—	H_2O	6.50
Unsol.	0.90	0.00	1.82	1.34	O	33.44
	99.82	99.98	100.53	99.45		
$-O = F_2$	—	0.17	—	—		
Total	99.82	99.81 *	100.53	99.45		

* Omitting the spectrochemical analysis by R. Danielsson: Ti 3600 ppm, Cu < 10 ppm, Cr < 30 ppm.
— Not determined

1. Eosphorite, Buckfield, Maine (see Hurlbut 1950).
2. Eosphorite, Viitaniemi. Wet chemical analysis by R. Saikkonen.
3. Eosphorite, Newry, Maine (see Hurlbut 1950).
4. Childro—eosphorite, Hagendorf-Süd (Strunz and Fischer 1957).
5. The unit cell content of the Viitaniemi eosphorite on the basis of specific gravity and unit cell volume (see column 2).

$$(Mn_{0.77}Li_{0.14}Fe^{+2}_{0.06}Fe^{+3}_{0.01}Na_{0.01})_{0.99}(Al_{1.08}Si_{0.01})_{1.09}P_{0.97}O_{4.18}((OH)_{1.95}F_{0.05})_{2.00} \cdot 0.81H_2O$$

In the calculation of the formula the water shown by the analysis was allocated so that first the sites of (OH,F) were filled in, and then the rest of the water was converted into molecular water. The amount of molecular water shows a distinct deficiency. Otherwise the formula of the mineral is close to the ideal formula of eosphorite.

The iron content in the eosphorite studied is very low and the lithium content abnormally high. In their studies on childrenite—eosphorite, Hurlbut (1950), and Strunz and Fischer (1957) mention only one deposit, the Buckfield pegmatite (Maine, U.S.A.), in which the iron content of the mineral is somewhat lower (see Table 13). The physical properties of the childrenite—eosphorite depend greatly on the Fe/Mn-ratio of the mineral as showed by Hurlbut (1950). The optical properties, specific gravity and unit cell dimensions of the Viitaniemi eosphorite do not differ significantly from those reported earlier in the literature (see Table 11 and Hurlbut 1950).

Aluminium—bearing phosphates

Amblygonite—montebrasite, the predominant aluminous phosphate, has been encountered in five pegmatites in the study area. The mineral abounds in the Viitaniemi deposit, where it is associated with some other aluminium-bearing phosphates. Of them, eosphorite is rather abundant, but the others, i.e. morinite, viitaniemiite and crandallite, are rare.

Amblygonite—montebrasite

Amblygonite—montebrasite occurs in pegmatites as oblong nodules, tabular and anhedral crystals or crystal aggregates (∅ often 10 to 30 cm). In the Viitaniemi (47) and Seppälänranta (31) pegmatites, however, a few subhedral montebrasite crystals have been encountered in cleavelandite associated with other Li minerals. As a rule the mineral is white, but in the Viitaniemi pegmatite it may be yellow (F-rich types) or pinkish white. It usually shows polysynthetic twinning in thin section. The twin lamellae are in fact often so wide that they can even be recognised with the naked eye. With the aid of optical determinations, Volborth (1951) has shown that Viitaniemi montebrasite has ($\bar{1}\bar{1}1$) as the twin plane.

Some amblygonite—montebrasite samples were taken from various pegmatites in the study area and were assayed for fluorine by an X-ray powder diffraction data using the (131) peak method described by Moss *et al.* (1969). Except for the Viitaniemi pegmatite, the samples studied were montebrasite with a fluorine content of between 3 and 6 wt%. The fluorine content in the amblygonite—montebrasite from the Viitaniemi deposit is usually more or less constant. Volborth (1951) has described from the same deposit montebrasite that, according to wet chemical analysis, contained 3.55 wt% F. The present author, however, has noticed that the crystal plates are often zoned, the core being yellow or whitish yellow amblygonite with a fluorine content of up to 7.9 wt%, and the margins white montebrasite. The intensely altered montebrasite crystals are invariably very poor in fluorine. Hence the montebrasite that has partly altered into kaolin may contain only 0.9 wt% F. The variation in fluorine content causes variation in the optic angle of the mineral as well (Dubois *et al.* 1972). Volborth (1951) has reported a large variation in 2V in the Viitaniemi amblygonite—montebrasite, although he attributes it to fracturing and deformation in the crystal plates after crystallization.

Morinite

Morinite has been encountered in the study area only in the Viitaniemi pegmatite. It is a rare mineral that occurs as fine-grained red nodules invariably in association with amlybonite-montebrasite. Volborth (1954a, see also Fischer and Volborth 1960) has described in detail the mode of occurrence of the Viitaniemi morinite. For the present study several red phosphate nodules were identified by X-ray diffractometer. Most of them turned out to be fine-grained väyrynenite or eosphorite; only three samples contained morinite. According to Volborth (1954a), the morinite nodules include Ca-rich hureaulite, väyrynenite and ježekite. The only minerals that could be identified from the morinite samples found by the present author were viitaniemiite and eosphorite, which occurs as a rim around the morinite nodules against montebrasite.

Viitaniemiite

Viitaniemiite is a new mineral species that was found during the identification of the minerals in a montebrasite sample taken from the Viitaniemi pegmatite by Erkki Halme. The Commission on New Minerals and Mineral Names of I.M.A. approved the name of the mineral and accepted it as a new mineral species by vote on 20. 12. 1977. The name of the mineral alludes to the deposit where it

Fig. 18. An apatite- (a, dark grey) and crandallite- (c, white) bearing cavity filling in viitaniemiite (v, dark and white crystal plates). Note the large eosphorite (e, grey) grains at the right. Nicols crossed. The Viitaniemi pegmatite (47).

was found. The type sample is stored in the Mineralogical Museum at the Geological Survey of Finland.

To date viitaniemiite has only been identified for sure from two mineral samples. In the sample studied by the present author the mineral occurs in eosphorite as an inclusion weighing about four grams. The eosphorite aggregate is surrounded by montebrasite crystal plates. With the aid of the description of the present author, Professor O. von Knorring has identified viitaniemiite from a sample he had taken from the type locality. In his specimen viitaniemiite occurs as a grey rim about 5 mm wide around a small morinite grain.

The montebrasite in the viitaniemiite sample studied in detail by the present author is intensely altered (kaolinized and apatitized), and its fluorine content is merely 0.9 wt% as measured from the (131) reflection of the X-ray powder diffraction data (cf. Moss *et al.* 1969). According to partial electron microanalyses, the eosphorite around the viitaniemiite contains approximately 10 mol% childrenite, the Fe-member of the childrenite—eosphorite series. Viitaniemiite occurs in eosphorite as an elongated inclusion 5 to 10 mm wide. The thin section shows that fluorapatite ($\varepsilon = 1.627$, $\omega = 1.631$) occurs as small cavity fillings between laths of viitaniemiite. There is always a narrow seam (1 to 3 mm wide) of crandallite (Fig. 18) between viitaniemiite and apatite. Fluorapatite and crandallite were identified by X-ray powder diffraction pattern during separation. Both minerals were also submitted to semiquantitative analyses by an electron microprobe. The analytical data show that crandallite does not contain impurities, whereas apatite has some manganese (c. 0.5 wt% MnO) and sodium (c. 1.0 wt% Na_2O).

Viitaniemiite is grey or white in colour and exhibits a glassy lustre. In thin section

Fig. 19. Fan-shaped viitaniemiite crystal plates partly altered into other minerals (dark parts). Nicols crossed. The Viitaniemi pegmatite (47).

it occurs as colourless, sometimes fan-shaped, tabular crystals 0.02 to 0.20 mm wide (Fig. 19). The fine-grained alteration products (probably apatite, kaolinite and crandallite) in the margins of the tabular crystals render the mineral cloudy. The hardness of viitaniemiite (Mohs) is 5 and the streak is white. The well-developed $(10\bar{1})$ cleavage, which was determined from precession photographs, is parallel to the crystal plates.

The physical properties of viitaniemiite are given in Table 14. The mineral is optically biaxial and negative. The optic angle was determined from a thin section on a universal stage. These determinations indicated that the optic axial plane (= O.A.P.) parallels the (010) plane and Y//b (Fig. 20). The specific gravity of viitaniemiite, which was determined with heavy liquids, is 3.245. The mineral is not fluorescent when irradiated with long or short wave UV radiation.

Viitaniemiite was initially analysed semi-quantitatively by an electron microprobe and then by wet chemical methods. The mineral was separated as a pure fraction by heavy liquids, and the purity of the fraction was

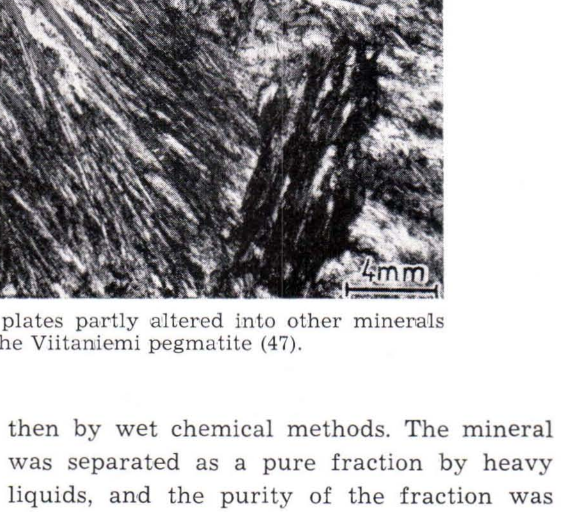


Fig. 20. Optic orientation of viitaniemiite. 1 = X, 2 = Y, 3 = Z, 4 = Optic axis, 5 = Normal to cleavage $(10\bar{1})$. Stereographic projection, lower hemisphere.

Table 14

The chemical composition (wt%) and physical properties of viitaniemiite from the Viitaniemi pegmatite. Chemical analyses by P. Ojanperä.

		Unit cell content on the basis of 14 anions		Physical properties
Al_2O_3	22.4	Al	2.15	The unit cell:
FeO	0.70	Fe^{+2}	0.05	Monoclinic
MnO	10.5	Mn	0.72	space group $P2_1$ or $P2_1/m$
MgO	0.38	Mg	0.05	$a = 6.832 \pm 0.003$ Å
CaO	14.7	Ca	1.28	$b = 7.143 \pm 0.003$ Å
Na_2O	11.6	Na	1.83	$c = 5.447 \pm 0.003$ Å
K_2O	0.27	K	0.03	$\beta = 109°22' \pm 5'$
Li_2O	0.00	Li	0.00	$V = 250.795$ Å^3
P_2O_5	28.3	P	1.95	$G = 3.245$
F	12.3	F	3.13	$D = 3.240$ g/cm^3
H_2O+	4.9	OH	2.66	
H_2O-	0.03	O	8.21	Optical properties:
	105.98			$\alpha = 1.557 \pm 0.001$
$—O = F_2$	5.14			$\beta = 1.565 \pm 0.001$
				$\gamma—\alpha = 0.014$
Total	100.84			$—2V = 81° \pm 1°$
				$—2V$ calc. $= 81.4°$
				O.A.P.//(010)
				b//Y

tested between separations with the aid of refractive liquids and powder pattern. The powdered mineral dissolved readily in nitric acid and sulphuric acid. Alkalies and alkaline earths, iron and aluminium were assayed from diluted solutions by AAS; manganese and phosphorus colorimetrically; fluorine on a fluorine ion-active electrode and water by a Penfield tube. The analyses were made by Pentti Ojanperä at the Geological Survey of Finland. The composition and physical properties of viitaniemiite are listed in Table 14. The formula of the mineral, calculated from the analytical data and assuming Z = 2, is:

$$(Na_{0.91}K_{0.01})_{0.92}(Ca_{0.64}Mn^{+2}_{0.36}Fe^{+2}_{0.02}Mg_{0.02})_{1.04}$$
$$Al_{1.07}P_{0.97}O_{4.11}(F_{1.57}(OH)_{1.33})_{2.90}$$

The idealized formula of viitaniemiite is: $Na(Ca,Mn)Al(PO_4)\ (F,OH)_3$, where $Ca > Mn$ and $F > OH$.

Viitaniemiite was studied by X-ray diffractometry using single crystal and powder methods. 0-, 1-, 2- and 3-level precession photographs of the a-, b- and c-axis were taken of the mineral using Zr-filtered Mo radiation. The parameters of the unit cell were calculated from the 11 reflections of an indexed X-ray powder data calibrated with quartz. The observed and calculated d-values of the mineral are given in Table 15 and the cell parameters in Table 14. The X-ray diffraction studies show that viitaniemiite is a monoclinic mineral. In the precession photographs $k = 2n$ in 0k0 reflections, and thus the space group of the mineral is either $P2_1$ or $P2_1/m$.

An X-ray powder pattern almost identical to that of viitaniemiite has been reported on JCPDS card 13-587 (see Table 15). The unknown mineral referred to was earlier named lacroixite originally described by Slavik (1914, 1915) from Greifenstein (Saxony, East Germany). Since the crystal structure, composition and physical parameters of lacroixite were only partly known, Mrose

Table 15

Comparison between the X-ray powder diffraction data of:

1. Viitaniemiite, the Viitaniemi pegmatite. Diffractometer, Ni-filtered Cu radiation, scanning speed 1/4°/min., quartz as an internal standard. The reflections marked with asterisks were used in calculating unit cell dimensions.
2. Unnamed mineral, Greifenstein, Saxony. See JCPDS card 13-587.
3. Lacroixite, Greifenstein, Saxony. Strongest lines according to Mrose 1971.

	1			2		3		
hkl	d meas.	d calc.	I	d	I	hkl	d	I
001	5.142	5.138	11	5.20	20			
$10\bar{1}$ *	4.885	4.885	33	4.91	40	011	4.739	25
110 *	4.785	4.785	21	4.82		$\bar{1}11$	4.633	25
$11\bar{1}$ *	4.032	4.032	14	4.05	8			
020	3.572	3.572	9	3.59	10			
101 *	3.493	3.493	16	3.53				
200 *	3.223	3.223	46	3.24	50			
111 *	3.139	3.138	21	3.18	20	$\bar{2}11$	3.159	71
$21\bar{1}$	2.966	2.966	24	2.962	20			
210 *	2.937	2.938	56	2.906	100	002	2.900	100
$12\bar{1}$ *	2.883	2.883	100	2.737	6			
002	2.569	2.569	35					
121 *	2.497	2.497	18			220	2.470	85
220	2.393	2.393	15					
$21\bar{2}$	2.311	2.311	16					
$30\bar{1}$	2.268	2.268	17					
031 *	2.160	2.160	40	2.107	2	131	2.168	35
				2.069	2			
				2.017	4			
230	1.915	1.915	28	1.862	60			
$32\bar{1}$		1.915		1.855	4			
				1.802				
040 *	1.786	1.786	27	1.762	2			
$32\bar{2}$	1.750	1.751	11	1.736	2			
				1.690	4			
				1.650	4			
$12\bar{3}$	1.615	1.615	13	1.631	25			
				1.607	2			
				1.585	40			
				1.573				
				1.551	4			
$42\bar{2}$	1.483	1.483	15	1.489	45			
$32\bar{3}$		1.482		+ several other lines				

(1971) studied samples collected from the type locality and redetermined it as a monoclinic fluorine-bearing Na-Al phosphate (space group A2/a, a = 6.890 Å, b = 8.220 Å, c = 6.425 Å and $\beta = 115°30'$). Mrose (1971) mentions another monoclinic mineral (space group $P2_1$ or $P2_1/m$, a = 6.840 Å, b = 7.185 Å, c = 5.479 Å and $\beta = 109°10'$) that she encountered in lacroixite samples, but she did not establish its composition and physical properties. She is of the opinion that the powder pattern on JCPDS card 13-587 and the optical properties reported ($\alpha = 1.543$, $\beta = 1.554$, $\gamma = 1.565$, $2V \approx 90°$) were probably measured from an unknown mineral. The original analysis by Slavik (1914, see also Palache *et al.* 1957) suggests that lacroixite is a fluorine-bearing Na-Ca-Mn-Al phosphate of obscure formula. Spectrographic analyses show that the mineral studied by Slavik (1914) contained (wt%): 14.92 % Na_2O, 19.46 % CaO, 8.43 % MnO, 18.87 % Al_2O_3, 28.83 % P_2O_5, 6.53 % F, 0,95 % SiO_2, loss in ignition 5.46 %, total 100.7 %. Since the composition includes appreciable calcium and manganese, Mrose (1971) maintains that the sample analysed was a mixture of minerals that, in addition to lacroixite, presumably contained the forementioned unknown mineral as well. This is consistent with the fact that lacroixite analysed by Štěrba-Böhm (Slavik 1915) does not contain manganese at all.

Viitaniemiite has the same space group as and practically identical lattice parameters to the unnamed mineral mentioned by Mrose (1971). Likewise the analysis reported by Slavik (1914) was made on a material of much the same composition as viitaniemiite. In viitaniemiite, however, the fluorine content is higher and the sodium and calcium contents lower than in the unnamed mineral. As pointed out by Mrose (1971), the axial ratios of the crystals described by Slavik 1914, 1915) are closer to those of lacroixite than to those of a mineral isotype with viitaniemiite. The refractive indices and density of lacroixite (see Palache *et al.* 1957) reported before the redeterminations differ significantly from those of viitaniemiite. Mrose (1971) has not studied the composition and physical parameters of lacroixite and the associated unknown mineral; hence, without detailed studies, it is impossible to know whether a viitaniemiite type of phosphate occurs in association with the Greifenstein lacroixite and whether the data on JCPDS card 13-587 all refer to the same mineral.

Beryllium-bearing phosphates

Beryllium-bearing phosphates occur in the study area in the Viitaniemi (47), Myllyrinne (59) and Jussinvuori (24) pegmatites. In the last two deposits, however, only very small amounts of a Mn-Be phosphate, väyrynenite, have been met with, whereas in the Viitaniemi pegmatite the abundance of beryllium-bearing phosphates is exceptionally large (see Fig. 10). Their mode of occurrence has been reported by Volborth (1954a), and in the context of the present study only beryllonite has been investigated in detail.

Since the studies by Volborth (1954a), moraesite has been identified from the pegmatite as a Be phosphate new to the deposit (Mrose and von Knorring 1959). The present author has found it as white cavity fillings in väyrynenite. The beryllium-bearing phosphates are often associated with amblygonite—montebrasite as nodules and aggregates. Väyrynenite and herderite may also occur in cavities and pseudomorphs with other low-temperature minerals such as kaolinite, eosphorite, fine-grained green mica (cookeite, muscovite and Li muscovite) and apatite (Fig. 13).

Beryllonite, the first of the beryllium-bearing phosphates to crystallize, often occurs in hurlbutite aggregates as inclusions of variable size (Fig. 21). Hurlbutite is abundant in the pegmatite, and in the core of the pegmatite it occurs as fine-grained brown aggregates weighing up to several kilograms. Their margins frequently contain herderite as white or grey rims and red väyrynenite that formed when manganiferous solutions replaced the above-mentioned minerals (Volborth 1954a).

In the Myllyrinne pegmatite väyrynenite occurs as small separate grains (Ø 1 to 5 mm) enveloped by chlorite or as inclusions in the narrow chlorite veins that cut the deposit. The biggest väyrynenite nodule found in the deposit was 20 mm in diameter, and, besides väyrynenite, it contained dark green or almost black chlorite and white apatite. In the Jussinvuori pegmatite (24) only one väyrynenite grain, a few mm in diameter, has been met with in albite—quartz—

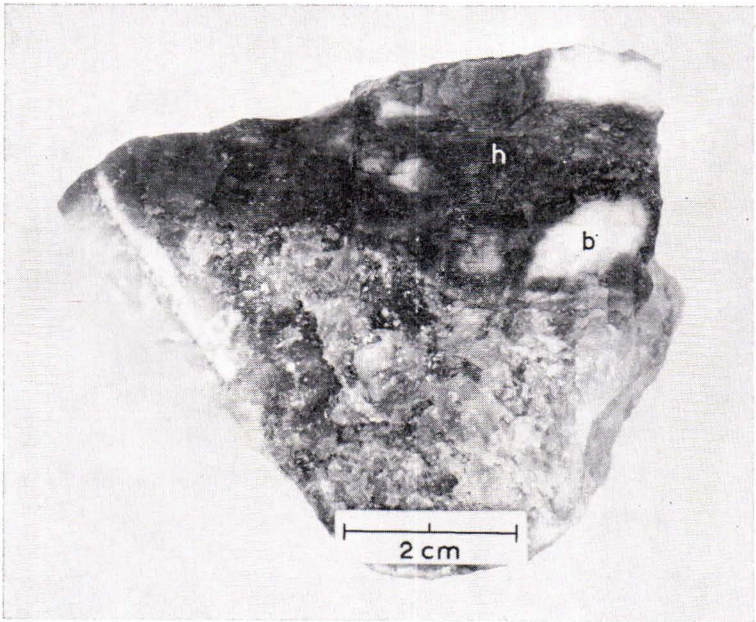


Fig. 21. Beryllium phosphate aggregate: beryllonite (b, white inclusions) which has altered to hurlbutite (h, dark grey). Lower part of sample consists of quartz and fine-grained muscovite. The Viitaniemi pegmatite (47).

muscovite pegmatite in the core of the dyke. As in the Myllyrinne pegmatite, the väyrynenite grain was enveloped by a narrow chlorite rim.

Beryllonite

Beryllonite is a very rare phosphate mineral. As far as the present author knows, it has only been encountered in a few pegmatites in the U.S.A. (Mrose 1952; Moore 1973), Brazil (Cassedane and Cassedane 1978), and in the Altai, USSR (Vlasov 1966, p. 152). The mineral, which occurs in pegmatites containing Li phosphates, Fe-Mn phosphates and beryl, is invariably associated with replacement bodies, fracture fillings and miarolitic cavities formed during the final stages of crystallization.

Beryllonite is rather common in the Viitaniemi pegmatite. It is encountered, 1) as inclusions in aggregates (Fig. 21) and nodules containing other beryllium phosphates, mainly hurlbutite (beryllonite I, according to the terminology of Volborth 1954a); 2) occasionally in association with cleavelandite, purple apatite, topaz, green tourmaline, lepidolite, eosphorite, greenish fine-grained mica or kaolinite as individual crystal aggregates (∅ 1 to 5 cm) that do not contain hurlbutite at all and often only very small amounts of other Be-bearing phosphates (beryllonite II). Beryllonite II represents the later generation, and hence the mineral may be associated with minerals that crystallized at very low temperatures. In the Viitaniemi pegmatite, however, beryllonite never occurs in miarolitic cavities, and it has not been encountered as euhedral or even subhedral crystals as it has in number of other deposits.

Beryllonite is a white mineral. Its cleavage is distinct along $\{010\}$ but poorly developed along $\{101\}$. Thin sections cut parallel to $\{010\}$ always show polysynthetic twinning according to two-twin laws. Optical studies by Volborth (1954a) demonstrated that the twin planes are $\{101\}$ and $\{100\}$.

Table 16

Unit cell dimensions of beryllonite according to various authors.

	1	2	3	4
a (Å)	8.13	8.16	8.178	8.136 ± 0.003
b (Å)	7.76	7.79	7.818	7.797 ± 0.003
c (Å)	14.17	14.08	14.114	14.192 ± 0.005
β°	90	90	90	90
V (Å^3)	893.96	895.01	902.38	900.33

1. Beryllonite, (locality unknown) (Gossner and Besslein 1934).
2. Beryllonite, (locality unknown) (Wehrenberg 1954).
3. Beryllonite, Stoneham, Maine (Giuseppetti and Tadini 1973).
4. Beryllonite, Viitaniemi.

0-, 1-, 2- and 3-level precession photographs of each crystal axis were prepared for beryllonite with the aid of Zr-filtered Mo radiation. The mineral is monoclinic, $\beta = 90°00'$ and the space group, as suggested by systematic extinctions, is $P2_1/n$. The cell parameters given in Table 16 were calculated from the powder pattern obtained with an X-ray diffractometer using Ni-filtered Cu radiation and quartz as internal standard. The observed and calculated d-values and the intensities of the reflections are listed in Table 17.

Since no euhedral crystals of beryllonite have ever been found in the deposit, the cleavage planes of the mineral were established from precession photographs. X-ray diffraction data show that the axes selected by Volborth (1954a) and applied in the present study were identical. Since the mineral invariably exhibits polysynthetic twinning and the individual crystals are very tiny (see Volborth 1954a, Fig. 18), it was impossible to obtain a single crystal for the precession camera. Thus, the precession photographs always show »extra» spots due to unoriented crystals twinned according to the $\{101\}$ law.

The crystallography of the beryllonite from the Viitaniemi pegmatite has not previously been studied by X-ray diffractometry, and the reported cell parameters obviously refer to samples from the U.S.A. As the compilation in Table 16 shows, the cell parameters given in different studies vary somewhat. This indicates that different methods were applied and that the composition may vary. The cell parameters of the Viitaniemi beryllonite, however, are not significantly different from those of the others, although its d-values differ somewhat from the d-values of the beryllonite studied by Mrose (1952; see Table 17). Partly provided with indices, these are also given on JCPDS card 6-0443. The card gives erroneously $P2_1/c$ as the space group of beryllonite (it should be $P2_1/n$, see Wehrenberg 1954, Giuseppetti and Tadini 1973). Some of the indices are also incorrect.

The refractive indices of beryllonite, determined by the immersion method, are listed in Table 18. The optic angle was measured on a universal stage in a thin section cut parallel to $\{010\}$ cleavage. The sections normal to cleavage show that Z and Y are parallel to $\{010\}$, whereas X is normal to it (or that X//b).

The optic data of beryllonite are compiled in Table 18. Unfortunately, however, the optic properties, chemical composition and cell parameters have not previously been determined from the same sample and, except for the data given in the studies by Volborth

Table 17

X-ray powder diffraction data of beryllonite:

1. Beryllonite, Viitaniemi. Diffractometer, scanning speed 1/4°/min., Ni-filtered Cu radiation, quartz as an internal standard. The reflections marked with asterisks were used in calculating unit cell dimensions. b = broad peak.
2. Beryllonite, Stoneham, Maine (Mrose 1952, see also JCPDS-card 6-443).

hkl	1			2	
	d meas.	d calc.	I	d	I
110	5.631	5.629	4	4.44	40
112 *	4.410	4.410	27	4.11	20
103 b	4.084	4.090	15	3.92	60
200 b		4.068		3.65	90
020 *	3.899	3.899	54	3.52	20
113 *	3.622	3.621	88	3.40	5
211 *	3.495	3.495	22	3.25	10
022 b	3.420	3.417	3	3.04	10
212 *	3.215	3.215	7	2.91	5
122	3.150	3.150	6	2.84	100
114 *	3.001	3.002	10	2.70	30
213	2.868	2.868	8	2.64	10
123 b	2.82	2.822	100	2.56	5
220 b		2.815		2.44	20
204 b	2.673	2.674	16	2.37	60
015 b		2.667		2.28	70
006	2.365	2.365	10	2.22	10
303 *	2.353	2.353	26	2.19	10
016	2.266	2.264	19	2.11	30
313 *	2.252	2.252	38	2.05	20
321	2.197	2.199	7	1.965	40
133		2.194		1.910	40
231	2.164	2.165	8	1.821	20
034	2.097	2.097	8	1.755	20
400 b	2.032	2.034	4	1.719	5
134 b		2.030		1.675	10
323	2.014	2.014	5	1.593 b	20
225	1.999	1.999	3	1.526	30
040 *	1.949	1.949	17	1.508	20
333 b	1.744	1.744	4	1.443 b	40

(1954a), the refractive index data are very old. The values of β and γ measured on Viitaniemi beryllonite by the present author are the same as those of Newry and Stoneham beryllonite within the limits of measuring accuracy; the value of α, however, differs significantly from the others. Because the mineral exhibits one distinct cleavage parallel to {010} and X//b, it is more difficult to determine α than the other refractive indices with immersion liquids, because the crystal fragments normally prefer to lie with their cleavage surface on the glass slide. Therefore the present author measured the indices using gelatine-coated slides, and since the observed and calculated optic angles are exactly the same, the value of α must be considered correct. The optic angles of the Stoneham and Viitaniemi beryllonite measured by different authors are equal, i.e. —2V = 67° ± 1° (see Table 18), whereas the calculated —2V varies markedly.

Table 19 is a compilation of beryllonite analyses reported in the literature and the wet chemical data of the Viitaniemi beryllonite analysed by O. von Knorring. As shown by the Table, the chemical compositions of the beryllonites from the Viitaniemi, Newry

Table 18

Comparison between the optical properties of beryllonite from various localities. The refractive indices were measured in Na light.

	1	2	3	4	5
α	1.552	1.552	—	1.553 ± 0.001	1.548 ± 0.001
β	1.558	1.558	1.558	1.560 ± 0.001	1.557 ± 0.001
γ	1.561	1.562	1.562	1.562 ± 0.001	1.561 ± 0.001
γ—α	0.009	0.010	—	0.009	0.013
—2V° meas.	67.6	—	—	66—67	67
—2V° calc.	70.3 *	78.2 *	—	56.0 *	67.1 *

* Calculated by the present author
— Not determined
1. Beryllonite, Stoneham, Maine (Dana and Wells 1889).
2. Beryllonite, Stoneham, Maine (Gaubert 1907).
3. Beryllonite, Newry, Maine (Palache and Shannon 1928).
4. Beryllonite, Viitaniemi (Volborth 1954a).
5. Beryllonite, Viitaniemi.

Table 19

The chemical composition (wt%) of beryllonite.

	1	2	3	4	5	6	
Na_2O	23.64	23.28	20.87	22.34	23.10	Na	11.25
Li_2O	—	0.07	0.00	—	—	K	0.02
K_2O	—	0.92	0.36	0.05	0.05	Be	12.51
CaO	—	0.40	3.85	1.94	0.00	P	11.94
BeO	19.84	19.12	18.24	20.04	20.72	O	48.00
Al_2O_3	—	0.21	0.09	—	—		
Fe_2O_3	—	0.07		—	—		
P_2O_5	55.86	55.40	53.09	54.30	56.14		
SiO_2	—	—	1.77	—	—		
H_2O+	0.08 *	0.52 *	0.64	0.76	—		
H_2O-	—	—	0.16	0.16	—		
Insol.	0.00	0.16	0.75	0.00	0.00		
Total	99.42	100.15	99.82	99.59	100.01		

* Ignition loss
— Not determined

1. Beryllonite, Stoneham, Maine (Dana and Wells 1889).
2. Beryllonite, Newry, Maine (Palache and Shannon 1928).
3. Beryllonite, Viitaniemi (Volborth 1954a).
4. Beryllonite, Viitaniemi. Chemical analysis by O. von Knorring from the specimen studied by the author.
5. The chemical analysis (see column 4) after omitting the content of CaO and H_2O and recalculating the sum to 100.
6. The unit cell content of beryllonite on the basis of 48 anions calculated from the analysis in column 5.

and Stoneham pegmatites differ little from each other. The CaO content, however, seems to be anomalously high in the wet chemical data of the Viitaniemi beryllonite, as is the SiO_2 content in the sample of Volborth (1954a). The beryllonite sample studied by the present author was assayed for calcium an silicon on an electron microprobe to establish whether sodium and phosphorus are replaced isomorphically by calcium and silicium or whether they are due to impurities. Electron microanalyses demonstrated that the mineral contains only 0.1 to 0.2 wt% CaO and even less silicon. Thus it seems that the separated beryllonite fraction of the present author was not pure enough for wet chemical analyses. The author suspects that the relatively high SiO_2 content in the analysis of Volborth (1954a) is due to quartz inclusions and the abnormally high CaO content to hurlbutite and herderite. When calcium and water are omitted from the analytical data (see Table 19, columns 4, 5 and 6) and when Z = 12, the formula of the beryllonite analysed becomes:

$Na_{0.94}Be_{1.04}P_{1.00}O_4$,

which is very close to the ideal composition of beryllonite ($NaBePO_4$).

According to Volborth (1954a), the specific gravity of the Viitaniemi beryllonite is 2.777; on the evidence of heavy liquid determinations by the present author, it is slightly higher, i.e. 2.788. On the basis of chemical data (Table 19, columns 4 to 6) and the unit cell volume, the calculated density for the mineral is 2.785 g/cm³, a value that is very close to the measured specific gravity.

Oxides and hydroxides

Quartz

Quartz was particularly abundant in the last volatile-rich residues of pegmatite melt and crystallized in the middle of dykes as rather big monomineral cores or as intraformational crosscutting veins. It is grey, white, brown, or entirely colourless. Banded quartz is also often encountered, primarily in the cores of pegmatites. Rose quartz has been met with in the Mattila (8) and Viitaniemi (47) pegmatites and in some veins within the Kultavuori pegmatite granite. In the pegmatite cavities quartz frequently occurs as euhedral crystals together with albite and microcline. In the Juurakko pegmatite (5) miarolitic cavities are common, and their quartz crystals may be even several centimetres in diameter.

Columbite—tantalite

Columbite—tantalite is encountered in almost every complex pegmatite in the study area. It is very common in most of the Li and Be-Nb pegmatites, although the crystals tend to be rather small, from 0.5 to 5.0 cm in diameter. The mineral is black or dark brown and occurs as tabular crystals or crystal aggregates associated with beryl, cassiterite and Li minerals. The crystals occasionally show some simple faces; euhedral crystals with abundant faces, however, are very rare.

As a part of the identification of pegmatite minerals, an X-ray powder diffraction pattern was recorded for 1 to 6 columbite—tantalite crystals in each deposit, excluding the Viitaniemi (47), Seppälänranta (31) and Niemelä (54) pegmatites in which the number of crystals studied was greater. The powder patterns show that the mineral has an ordered or a disordered structure. The crystals were considered as ordered if their powder pattern exhibited either the (020) reflection (d = 7.1 Å) alone or the (020) and (110) reflections (d = 7.1 and 5.3 Å) that are lacking from disordered crystals (cf. Nickel *et al.* 1963b; Zelt 1975). Altogether some 180 columbite—tantalite crystals were studied for their powder pattern; most of them, or c. 80 % showed disordered structures. A further detailed single crystal study was made on a

Table 20

The chemical composition (wt%) of some columbite—tantalites from the pegmatites of the Eräjärvi area. Electron microprobe analyses by T. Hautala (except no. 9) with Mn-tantalite as a standard. X-ray powder patterns show that samples 1, 5, 6 and 8 are ordered, the others disordered types.

	1	2	3	4	5	6	7
TiO_2	0.1	0.7	0.2	0.6	0.6	0.4	0.5
FeO	0.8	3.6	0.8	11.8	11.5	12.3	12.6
MnO	13.2	11.7	13.1	3.5	5.0	5.4	5.3
MgO	0.0	—	—	—	—	—	—
CaO	0.0	—	—	—	0.2	—	—
SnO_2	0.3	0.6	0.4	—	—	—	—
Nb_2O_5	13.7	12.3	16.4	14.7	24.8	27.6	26.3
Ta_2O_5	74.9	73.0	69.8	65.4	59.5	57.9	56.8
Total	103.0	101.9	100.7	96.0	101.6	103.6	101.5

	8	9	10	11	12	13	14
TiO_2	0.4	1.08	2.2	—	0.7	0.6	0.5
FeO	12.4	1.73	9.1	5.3	10.6	19.5	15.9
MnO	5.5	14.50	4.8	13.5	8.9	3.0	6.1
MgO	—	0.04	—	—	—	—	—
CaO	—	0.22	0.2	—	—	—	—
SnO_2	—	0.33	—	0.2	0.1	0.1	1.0
Nb_2O_5	29.9	37.65	48.3	45.4	49.5	56.4	58.7
Ta_2O_5	55.8	43.45	36.8	33.4	28.8	17.5	16.6
Total	104.0	99.0	101.4	97.8	98.6	97.1	98.8

— Not determined

1. Mn-tantalite, lamellae in wodginite, Viitaniemi (47).
2. Mn-tantalite, Viitaniemi (47).
3. Mn-tantalite, Viitaniemi (47).
4. Tantalite, Seppälanranta (31).
5. Tantalite, Viitaniemi (47).
6. Tantalite, Seppälänranta (31).
7. Tantalite, Seppälänranta (31).
8. Tantalite, Seppälänranta (31).
9. Mn-tantalite, Mattila (8). Wet chemical analysis by P. Ojanperä (not included: $SiO_2 = 0.18$, $Al_2O_3 = 0.17$, $WO_3 = 0.60$, $U_3O_8 = 0.02$, $NiO = 0.00$, $H_2O^+ = 0.28$).
10. Columbite, Eräpyhä (1).
11. Mn-columbite, Leikattu (7).
12. Columbite, Löytäne (57).
13. Columbite, Seppälä (34).
14. Columbite, Niemelä (54).

Mn-columbite from the Leikattu pegmatite (7). The chemical data are listed in Table 20, column 11. The chemical formula of the mineral calculated on the basis of 24 oxygens and when Z = 4, is:

$(Mn_{0.8}Fe_{0.3})(Nb_{1.4}Ta_{0.6})O_6$

0-, 1- and 2-level a- and b-axis photographs were produced for the mineral by Zr-filtered Mo radiation. The powder patterns revealed that the mineral is disordered, and the single crystal study demonstrated that its unit cell is the subcell of the ordered columbite. Precession photographs calibrated by silicon give the following cell parameters for the mineral (space group Pcan):

a = 5.747 ± 0.005 Å
b = 4.748 ± 0.005 Å
c = 5.151 ± 0.005 Å

The powdered mineral was annealed at 1000°C for 1 h, and a new powder pattern was recorded by X-ray diffractometer. The mineral attained an ordered structure during heating, and the b axis of the unit cell tripled in length (unit cell parameters calculated from the powder pattern calibrated by silicon: a = 5.745 Å, b = 14.354 Å and c = 5.150 Å). In the terminology of Nickel *et al.* (1963b), the disordered columbite—tantalites that assume an ordered structure after heating are pseudo-ixiolites; there is no such change in structure in ixiolite proper. Since ixiolite and pseudo-ixiolite have similar X-ray powder diffraction patterns, some of the columbite—tantalite samples studied may be real ixiolites. Thus, the present author annealed ten columbite—tantalite samples besides the Mn-columbite from the Leikattu pegmatite. According to the powder patterns, the minerals all had disordered structures, but each of them turned out to be pseudo-ixiolite and, after heating, showed a powder pattern of ordered columbite.

Some semiquantitative X-ray fluorescence analyses (Lahti 1975) and electron microanalyses (Table 20) show that the composition of the mineral varies largely between and within deposits. Both Fe- and Mn-columbites and Fe- and Mn-tantalites have been encountered in the pegmatites. The samples from the Be-Nb pegmatites are often columbite, whereas tantalite predominates in Li pegmatites. All the columbites listed in the Table have disordered structures, whereas the tantalites were either ordered or disordered. The ordering may depend partly on the chemical composition of the mineral as demonstrated earlier for example by Zelt (1975). The Table, however, shows that tantalites of different structure may be very similar, at least in the pegmatites of the study area. No distinct geologic correlation in the mode of occurrence between the columbite—tantalites of different structure was established. Within a given dyke both disordered and ordered columbite—tantalites occur. Also in Li pegmatites and Be-Nb pegmatites the minerals showed the same ratio of structural types (20 % ordered and 80 % disordered).

Cassiterite

Cassiterite is a common mineral in the area, particularly in the Li pegmatites. It is, however, rare in the Be-Nb pegmatites. The crystals are dark brown or occasionally black, anhedral or subhedral and usually rather small (∅ 3 to 30 mm). In some pegmatites (Seppälänranta 31, Viitaniemi 47, Juurakko 5, Niemelä 54), cassiterite occurs as crystal aggregates several centimetres in diameter in the albite-rich fracture fillings and replacement bodies often associated with niobium-tantalum minerals, beryl and lithium minerals. It often has tiny inclusions of an unidentified tin sulphide (probably herzenbergite). Microanalyser determinations show that cassiterite is frequently rich in tantalum and may contain up to 10 wt% Ta_2O_5 (a sample from the Viitaniemi pegmatite).

Microlite

Microlite has been identified from the Eräpyhä (1), Metsänen (18), Mikkola (19), Löytäne (57) and Maijaanvuori (12) pegmatites. In the latter the mineral is fairly common and occurs as black or yellow anhedral—subhedral crystals (∅ 2 to 10 mm) in the interior of the deposit. It is rare in the other

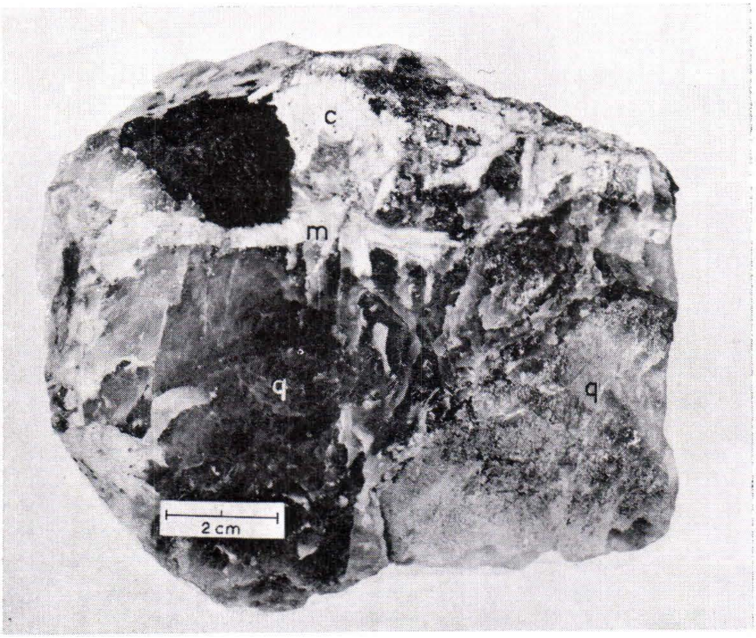


Fig. 22. Large wodginite crystal (black) associated with quartz (q, grey), cleavelandite (c, white) and montebrasite (m, white plates). The Viitaniemi pegmatite (47).

pegmatites, where it is usually intergrown with columbite—tantalite. As a rule microlite is metamict and can only be identified on the powder pattern after heating. To check the identification, some samples were assayed semiquantitatively by an electron microprobe.

Tapiolite

Tapiolite has been encountered sporadically in the Seppälänranta (31), Viitaniemi (47), Pellonreuna (39) and Seppälänniemi (28) pegmatites. It is a dark brown mineral and occurs as fairly small anhedral crystals (∅ 2 to 30 mm) in the middle of the deposit. Debye-Scherrer photographs show that the pegmatites contain both disordered and ordered tapiolites. The composition and physical properties of the mineral have, however, not been established.

Wodginite

Wodginite has only been identified from the Viitaniemi pegmatite, where it occurs as dark brown crystal aggregates (∅ 10 to 30 mm), often in the pegmatite portions that contain montebrasite and other Li minerals (Fig. 22). Up to one half of the aggregates may be composed of manganotantalite, which sometimes occurs as discrete tabular crystals in wodginite. The chemical composition of the ordered tantalite that occurs in the wodginite sample studied is given in Table 20, column 1. In addition to various silicates and tantalite, the mineral has tiny inclusions of an unidentified tin sulphide (obviously herzenbergite). In reflected light on a polished section wodginite is a brown and distinctly anisotropic but only slightly pleochroic mineral. It has abundant small cracks and very strong reddish brown internal reflections. The polished section examined shows some narrow twin lamellae.

The Viitaniemi wodginite was analysed on an electron microprobe, and crystal fragments were taken from the same sample for single crystal study and specific gravity determination. 0-, 1- and 2-level b- and c-axis precession photographs

Table 21

The chemical composition (wt%) and physical properties of wodginite from Viitaniemi. Microprobe analyses by T. Hautala (Mn-tantalite and cassiterite as standards).

		The content of unit cell on the basis of 32 oxygens		Physical properties
				The unit cell:
TiO_2	0.3	Ti	0.1	Monoclinic
FeO	1.2	Fe	0.4	Space group C2/c or Cc
MnO	8.8	Mn	3.2	a = 9.517 ± 0.005 Å
MgO	0.1	Mg	0.1	b = 11.486 ± 0.005 Å
CaO	1.2	Ca	0.6	c = 5.115 ± 0.005 Å
SnO_2	11.8	Sn	2.0	$\beta = 90°55'$
Nb_2O_5	5.5	Nb	1.1	V = 559.061 $Å^3$
Ta_2O_5	70.6	Ta	8.3	G = 7.50
Total	99.5	O	32.0	D = 7.56 g/cm^3

were taken of the mineral with Zr-filtered Mo radiation. Wodginite is monoclinic and, as shown by the systematic extinctions, its space group is C2/c or Cc. The cell parameters, measured from the precession photographs calibrated by silicon, are given in Table 21. The specific gravity of wodginite was determined on a Berman balance. The reported value is an average obtained from determinations on three fragments of various size.

The chemical composition of the Viitaniemi wodginite and the atoms in the unit cell are given in Table 21. The microanalyser data suggest the following formula for the mineral (Z = 16):

$$(Ta_{0.52}Mn_{0.20}Sn_{0.13}Nb_{0.07}Ca_{0.04}Fe_{0.03}Ti_{0.01})_{1.00}O_{2.00}$$

From the wodginite studied a fragment was taken and submitted to wet chemical total iron and ferrous iron determinations to establish the oxidation state of the iron in the mineral. According to the assay by P. Ojanperä, the sample contained 0.79 wt% FeO and 0.14 Fe_2O_3.

Wodginite was described as a new mineral by Nickel *et al.* (1963a). Since then the mineral has been identified from several complex pegmatites (cf. Bourguignon & Mélon 1965; Luna 1965; Khvostova *et al.* 1966; Vorma and Siivola 1967; von Knorring *et al.* 1969a and b; von Knorring 1970a; Burke *et al.* 1970; Grice *et al.* 1972). Like the Viitaniemi pegmatite, the deposits that contain wodginite often also have beryl, columbite—tantalite and possibly some other Nb-Ta minerals and various Li minerals. The mineral is rare and tends to occur as fairly small crystals. As shown by the analytical data, the wodginite from the Viitaniemi pegmatite, like the associated tantalite, is exceptionally poor in iron and niobium and rich in manganese and tantalum. Its calcium content is rather high; otherwise it is very similar in chemical composition and cell parameters to the wodginite from Wodgina, the type locality in Australia (Nickel *et al.* 1963a; Graham and Thornber 1974). Hence the X-ray powder diffraction pattern of the sample studied was not reproduced in the present study.

Uraninite

Uraninite, which has been encountered as small anhedral crystals in the Eräpyhä pegmatite, occurs in albite fracture fillings associated with cassiterite, zircon and spessartine. It is metamict and was identified from the powder pattern of an annealed sample (1 h at 1000°C).

Senarmontite, valentinite and stibiconite

Senarmontite, valentinite and stibiconite have been encountered in small amounts in the Viitaniemi pegmatite as alteration products of native antimony and bismuthian antimony. Volborth (1960) has described in detail the mode of occurrence of the minerals in his paper on bismuthian antimony and other ore minerals in pegmatite.

Limonite and wad

Limonite and wad (amorphous Mn oxides) sometimes occur as a brown or black mass in cavities. Both of these minerals are typical very low-temperature alteration products of iron- and manganese-bearing minerals.

Elements

Graphite

Graphite, which is encountered sporadically in several pegmatites in the area, usually occurs as very narrow fracture fillings in feldspars and other pegmatite minerals. Abnormally abundant graphite is met with in the Juurakko (5) pegmatite, in which it occurs together with fine-scaled muscovite as lenses up to two centimetres thick in cleavelandite-rich fracture fillings and replacement bodies.

In addition to crystalline graphite, the pegmatites in the area often contain small, black carbon spherules known as thucholite (cf. Ramdohr 1969). They are no more than 1 to 5 mm in diameter although in the cores of the complex pegmatites they are markedly larger (∅ up to 20 mm). The thucholite spherules often have small yellow inclusions that, according to microanalyser determinations, contain predominantly uranium. The carbonaceous matter and uranium mineral inclusions (U oxide?) are amorphous, and only after being annealed (1 h in air at 1000°C) do the inclusions show the X-ray diffraction powder pattern of uraninite.

Native antimony, bismuthian antimony and bismuth

Native antimony, bismuthian antimony and bismuth have only been encountered in the Viitaniemi pegmatite. The bismuthian antimony from this deposit was described by Volborth (1960), who encountered native bismuth as well, as tiny inclusions in a löllingite grain. When studying the minerals in the Viitaniemi pegmatite, the present author had microanalyser determinations made on several of the antimony samples collected from the deposit. Only one of them, however, turned out to contain abnormally high amounts of bismuth (10.5 wt%, see Table 22); all the others were pure native antimony.

Native antimony occurs randomly in the central parts of the pegmatite. The mineral has been encountered as aggregates (∅ 1 to 5 cm) in cleavelandite and as fracture fillings in association with antimony oxides or sulphides, or both (Fig. 23). The antimony minerals have crystallized at very low temperatures. Kaolinite, purple apatite, herderite, väyrynenite and fine-grained greenish mica are often associated with them. Bismuthian antimony occurs in a similar manner as a fracture filling (1 to 2 cm wide) in a pegmatite sample that contains montebrasite. Both bismuthian and native antimony are intensely altered. The samples often contain abundant antimony oxides and sulphides. The cores of the antimony aggregates and veins have inclusions that are predominantly senarmontite. The margins are composed of valentinite, stibnite, kermesite and tetrahedrite. Outermost, against the other pegmatite min-

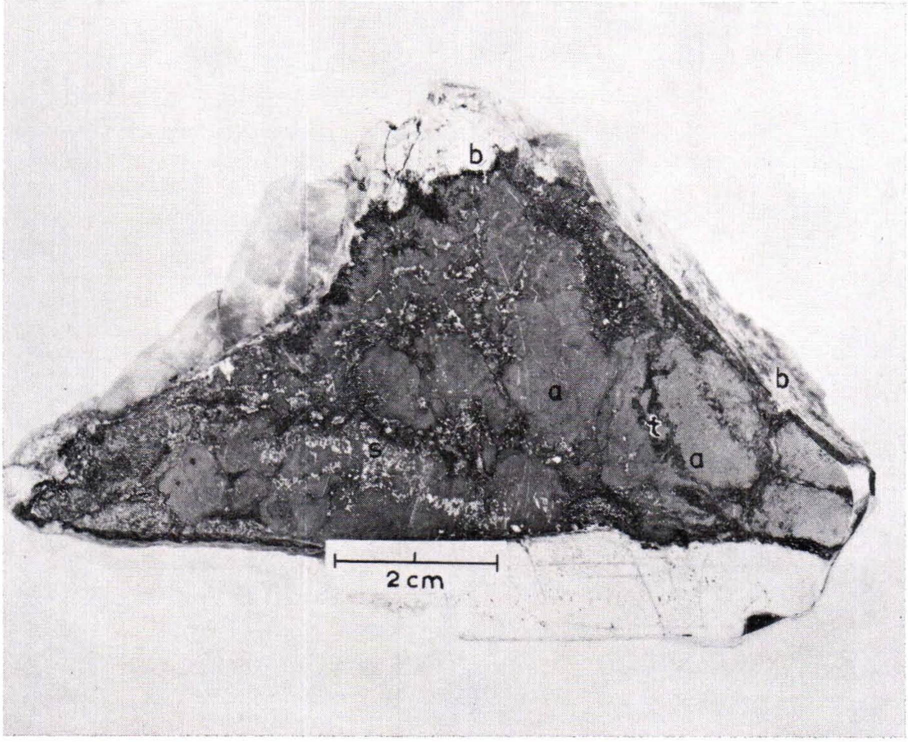


Fig. 23. Antimony mineral aggregate from the Viitaniemi pegmatite (47). The following minerals are present in the sample: a, native antimony, main part; s, senarmontite, white inclusions; t, stibnite, dark ribbons; b, stibiconite, white shell around aggregate. Polished sample.

erals, there is yellow aphanitic stibiconite that, as an oxidation product of the other antimony minerals, was the last to crystallize.

The native and bismuthian antimony are silver-white and invariably very fine-grained. Antimony and bismuthian antimony are always so full of inclusions and alteration products that it has not been possible to measure accurately the reflectance of the minerals. The Vickers hardness number of the bismuthian antimony is from 71.5 to 87.6 (weight 20—50 g) and that of the pure native antimony slightly higher, i.e. from 75.7 to 95.8 (weight 20—50 g).

The analytical data of an impure bismuthian antimony reported by Volborth (1960) and the microanalyser data by T. Hautala of the sample found by the present author are listed in Table 22. The microanalyser determinations indicate that both antimony and bismuth are homogenously distributed in the mineral and that its composition is also homogenous. The electron microprobe analyses suggest the following chemical formula for bismuth-bearing antimony (when $Z = 6$):

$Sb_{0.93}Bi_{0.06}As_{0.01}$

The sample of Volborth (1960) contained somewhat more bismuth than did the sample studied by the present author; hence, either the Bi/Sb ratio varies in the mineral or the material analysed by Volborth (1960) contained bismuth minerals as impurities. The only inclusions in the bismuthian antimony

Table 22

The chemical composition (wt%) of bismuthian antimony and some physical properties of antimony and bismuthian antimony from the Viitaniemi pegmatite.

	1	2
Sb	72.57	88.3
Bi	12.84	10.5
As	—	0.4
S	0.01	0.0
SiO_2	10.48	0.0
H_2O+	0.58	—
H_2O-	0.20	—
Unsol.	0.10	—
Total	96.78	99.2

— Not determined

1. Quartz- and senarmontite-bearing bismuthian antimony. Chemical analysis by H. Vincent (Volborth 1960).
2. Bismuthian antimony. Microprobe analysis by T. Hautala.

Physical properties

The unit cell:
Trigonal
Space group $R\bar{3}m$
Native bismuthian antimony:
$a_{hex.} = 4.326 \pm 0.005$ Å
$c_{hex.} = 11.330 \pm 0.005$ Å
V = 183.589 Å^3
G = 6.76
D = 6.89 g/cm^3
Native antimony:
$a_{hex.} = 4.309 \pm 0.005$ Å
$c_{hex.} = 11.271 \pm 0.005$ Å
V = 181.201 Å^3
G = 6.67
D = 6.69 g/cm^3

studied by the present author were senarmontite, whose Bi/Sb ratio is about the same as that in the host mineral, and small amounts of stibnite, kermesite and tetrahedrite. Microanalyser determination reveal that native antimony contains neither bismuth or sulphur. The arsenic content is also distinctly lower in native antimony (As < 0.4 wt%) than in the bismuthian variety.

The powder patterns of native and bismuthian antimony were recorded by X-ray diffractometer using Ni-filtered Cu radiation and quartz as an internal standard. Both minerals were so fine-grained that it was not possible to perform single-crystal studies on them. The powder pattern of the Viitaniemi antimony and bismuthian antimony, which were indexed on the basis of the powder pattern of antimony reported in the literature (Swanson *et al.* 1954; Berry and Thompson 1962), is given in Table 23; the cell parameters computed from them are in Table 22.

Table 23

X-ray powder diffraction data of antimony and bismuthian antimony from the Viitaniemi pegmatite. Diffractometer, scanning speed 1/4°/min., Ni-filtered Cu radiation, quartz as internal standard. The reflections marked with asterisks were used in calculating unit cell dimensions. b = broad peak.

	Bismuthian antimony			Antimony		
hkl	d meas.	d calc.	I	d meas.	d calc.	I
003	3.776	3.777	7	3.757	3.757	6
102 *	3.125	3.125	100	3.112	3.111	100
014 *	2.259	2.259	30	2.248	2.249	23
110 *	2.163	2.163	25	2.154	2.154	23
105 *	1.939	1.939	6	1.929	1.929	4
006, 113 b	1.884	1.888, 1.877	7	1.874	1.879, 1.869	4
202 *	1.778	1.778	12	1.771	1.771	10
204 *	1.562	1.562	9	1.556	1.556	7
017 *	1.486	1.486	3	1.478	1.478	3
116 *	1.422	1.422	8	1.416	1.416	6
212 *	1.374	1.374	9	1.368	1.368	7

Within the limits of measuring errors, the cell parameters of the Viitaniemi native antimony are identical to those reported by Swanson *et al.* (1954; see also Skinner 1965) for synthetic antimony. The bismuthian antimony has conspicuously longer a- and c-axes than corresponding axes of native antimony, and hence the d-values of bismuthian antimony also differ somewhat from the d-values of the native antimony (Table 22).

Native antimony, which is relatively rare in pegmatites (Ödman 1941; Wretblad 1941; Lawrence 1963; Černý and Harris 1973; Černý and Harris 1978), is encountered mainly in hydrothermal ore deposits (Ramdohr 1969). The bismuth content in antimony tends to be rather low. To the knowledge of the present author the only description of bismuthian antimony similar in composition to that in the Viitaniemi pegmatite is from the Tanco pegmatite in Canada (Cerný and Harris 1978). There bismuthian antimony ($Sb_{0.89}Bi_{0.08}As_{0.03}$) occurs together with various sulphides, whereas antimony oxides, which are typical in the Viitaniemi pegmatite, are altogether lacking.

Sulphides, arsenides and arsenates

Many pegmatites in the study area contain sporadic arsenopyrite, pyrite, pyrrhotite and in places also sphalerite, löllingite, chalcopyrite and molybdenite. The minerals occur as small grain aggregates or narrow fissure veins in the interior of the pegmatites or sometimes as subhedral-euhedral crystals in cavities.

Arsenopyrite is very common and occurs as narrow veins a few millimetres wide. Pyrrhotite, which is encountered in the Omenajärvi pegmatite (14) exceptionally as veins up to 20 cm wide, often contains pyrite as subhedral-euhedral crystals. Some parts of the Mattila pegmatite (8) contain löllingite as monomineralic veins several centimetres wide, and the Juurakko pegmatite (5) has veins that are composed predominantly of löllingite but contain abundant sphalerite, pyrite and some chalcopyrite as well. Scorodite is an alteration product of löllingite and arsenopyrite. It occurs as a greenish-yellow mass in the fractures of these minerals.

Sphalerite, which occurs accessorily, particularly in the Li pegmatites, may be encountered as individual small grains in the middle of the deposits or associated with the other sulphides. As a rule the mineral is deep brown and rich in iron, although, in the cavities of the Viitaniemi pegmatite (47), yellow, iron-poor sphalerite has been met with. According to microanalyser determinations, it contains only 0.6 wt% Fe.

Unlike the other deposits, the Viitaniemi pegmatite contains small amounts of antimony sulphides. Stibnite, tetrahedrite and kermesite have been identified from the deposit by X-ray diffractometry. The minerals occur as fracture fillings in the kaolinized portions of the pegmatite or as inclusions in native antimony. Antimony oxides, valentinite, senarmontite and stibiconite are also often associated with the sulphides. In addition the present author encountered cadmium-rich sphalerite (or hawleyite) and germanite as small grains in an aggregate predominantly composed of valentinite. Herzenbergite which has been identified from the Viitaniemi pegmatite, replaces cassiterite in some of the samples studied.

Halogenides and carbonates

Fluorite, which is one of the most common minerals of this group, occurs as narrow (1 to 3 mm wide) monomineralic veins in numerous complex pegmatites and also in cavities as small purple or colourless crystals. Like fluorite, calcite is met with as narrow veins in some pegmatites. It has also been encountered in association with secondary Fe-Mn phosphates in the Seppälänranta pegmatite (31). Siderite has only been identified from the Katila pegmatite (16), where it occurs as tiny crystals in pseudomorphs after lithiophilite together with metavivianite, fairfieldite and hureaulite. Malachite has been encountered as an alteration product of chalcopyrite in some pegmatites.

COMPARISON WITH OTHER MINERALOGICALLY SIMILAR PEGMATITES

Phosphate-rich lithium pegmatites similar in mineralogy and structure to those in the Eräjärvi area have been described from the U.S.A. from the Black Hills, New England, South California and the southeastern Appalachians (Schneiderhöhn 1961; Roberts and Rapp 1965; Moore 1973). They are also encountered in Brazil (Lindberg *et al.* 1953; Rodrigues da Silva 1975), Africa (von Knorring 1970a and b; Fransolet 1974—1975), Portugal (de Jesus 1933; Fransolet 1974—1975), Germany (Strunz 1974) and in Sweden (Quensel 1955).

In Finland, complex pegmatites enriched in a variety of lithium minerals, niobium-tantalum minerals and beryl are encountered in the Kemiö area (Pehrman 1945), the Somero—Tammela area (Mäkinen 1913; Vesasalo 1959; Aurola 1963), South and Central Bothnia (Neuvonen 1960; Haapala 1966; Nieminen 1978) and the Kitee—Tohmajärvi area (Kallio and Alviola 1975; Nykänen 1975). The latter excluded, all the Finnish lithium pegmatite areas are located in South Finland within the Svecofennides. Deposits with beryl and niobium-tantalum minerals have also been encountered among pegmatites in the Sotkamo—Kajaani area (Alviola 1977). The Hunnakko pegmatite in South Bothnia, the Lemnäs pegmatite in the Kemiö area and some pegmatites in the Kitee–Tohmajärvi area are rich in phosphates and are mineralogically similar to the Eräjärvi pegmatites. The spodumene and petalite pegmatites which are typical e.g. in the Central Bothnia and Somero—Tammela area (Mikkola and Wiik 1947; Vesasalo 1959), are wholly lacking from the Eräjärvi area.

Lithium pegmatites usually contain the same lithium silicates as do those in the Eräjärvi area, i.e. lepidolite, elbaite and spodumene. Petalite is also rather common, particularly in pegmatites rich in spodumene, and eucryptite has occasionally been encountered in some Li pegmatites although, to date, neither of them has been identified from the Eräjärvi area. The phosphate-rich lithium pegmatites may contain sparse or no lithium silicates. Apatite, amblygonite–montebrasite, triplite and triphylite–lithiophilite are very typical and common primary phosphates in the lithium pegmatites. Various secondary phosphates have often formed at lower temperatures as alteration and replacement products of triphylite–lithiophilite. Fairly common examples are sicklerite–ferrisicklerite, heterosite–purpurite, alluaudite, vivianite, hureaulite, rockbridgeite–frondelite and fairfieldite–messelite. In the Eräjärvi pegmatites the water-bearing phosphate minerals that contain iron or manganese, or both, are exceptionally rich in species although they occur only accessorily. Some of the phosphates are rare (e.g. metavivianite, switzerite, jahnsite, laueite and strunzite), and they are known only from a few deposits in the whole world (cf. Moore 1973).

The Viitaniemi deposit differs from the other pegmatites in the study area in that it contains some rare aluminium-rich phosphates (crandallite, morinite, viitaniemiite), very abundant beryllium-bearing phosphates and various antimony minerals. As a rule, beryllium-bearing phosphates are very rare in pegmatites. The present author has found no records of any deposits containing Be-bearing phosphates in such abundance as in the Viitaniemi pegmatite. Similarly, the antimony minerals that occur sporadically in the Viitaniemi pegmatite as aggregates and cavity fillings are generally rare. Topaz, which occurs in some veins in the study area, is typical in complex pegmatites; likewise pollucite, which so far has only been identified from the Viitaniemi pegmatite.

In addition to the pegmatites containing lithium minerals there are also many areas with deposits containing only beryl, niobium-tantalum minerals or tourmaline and yet others with simple pegmatites composed of ordinary rock-forming minerals. The pegmatites often contain the same niobium-tantalum minerals as do the pegmatites in the Eräjärvi area, i.e. columbite—tantalite, tapiolite, occasionally microlite and wodginite. Besides niobium-tantalum minerals, cassiterite is a very typical constituent of such pegmatites.

DISCUSSION

Genesis of the pegmatites

Having crystallized from the volatile-rich residual melts of igneous rocks in the same area, pegmatites with rare minerals are generally considered to be magmatic in origin (Schneiderhöhn 1961, p. 629—631; Vlasov 1968, p. 40—43; Černý and Trueman 1978). As the outcome of differentiation and crystallization in the intrusives, the residual melts became enriched in certain elements that vary from area to area. In many pegmatite areas complex pegmatites are associated with granitic rocks or occur in the immediate vicinity of granitic intrusions. Numerous examples are given in Schneiderhöhn's (1961) »Die Erzlagerstätten der Erde» (Part II, Die Pegmatite) that deals with granite pegmatites in the whole world. In U.S.A. the complex pegmatites of the well-known Black Hills area, which contain Li silicates, Li phosphates, Nb-Ta minerals, beryl, etc., are associated with pegmatite granite (Schneiderhöhn 1961; Černý and Trueman 1978). In Germany the phosphate-rich Hagendorf—Süd, Hagendorf—Nord and Pleystein pegmatites occur in the hanging wall of a large granite intrusion (Strunz 1974). Likewise the Varuträsk pegmatite (Sweden), which resembles the Viitaniemi pegmatite in size, shape, internal structure and mineralogy, is genetically related to the Skellefteå granite (Quensel 1955). In Finland, the phosphate-rich lithium pegmatites in the Kitee—Tohmajärvi area are associated with the late-orogenic Kitee pegmatite granite (Nykänen 1975; Kallio and Alviola 1975).

Similarly, the Eräjärvi pegmatites seem to be related to the microcline granite and pegmatite granite intrusions in that area. The pegmatite dykes and lenses are most abundant in the schists adjacent to the granite plutons. The size of the deposits and the abundance of rare minerals in the pegmatites decrease markedly as the distance from the granites increases.

Pegmatites usually intersect granites with sharp contacts. In the margins of plutons, however, the granite may grade into pegmatite granite and pegmatite. Granites and pegmatites are obviously genetically related to each other. At least the pegmatites adjacent to and within a pluton can be considered to have crystallized from volatile-rich residual granitic solutions. As in numerous areas of zoned pegmatites, deposits with lithium minerals are not in direct contact with granite intrusions but are located in the contact zone or in enveloping rocks (cf. Heinrich 1953; Stewart 1978).

The Kultavuori pegmatite granite in the southern part of the study area obviously represents the roof of a more extensive granite pluton. Pegmatitic parts formed during the crystallization and differentiation of the intrusion from the residual melt richest in volatiles. The pegmatite granite and pegmatite apparently form only a narrow layer in the upper part of the granite pluton. Here and there in the middle of the Kultavuori area medium-grained microcline granite crops out and in some other places the enveloping pegmatite granite contains large fragments of overlying schists. The part of the Kultavuori intrusion that extends to the east of Eräjärvi and the Rappuvuori granite pluton in the western margin of the study area probably represent a deeper section of the late-orogenic granite intrusions. The possibly pegmatitic roofs of these plutons might have been denuded by erosion, and so nowadays the pegmatite deposits only occur in the contact zones of the granites. The interior of the granite plutons is fairly homogenous, having only narrow pegmatite dykes of simple mineralogy as the final and relatively volatile-poor differentiation products of the granite melt.

The bedrock in southern Finland represents a certain section across the roots of a range of Precambrian fold mountains. The metamorphic grade of the rocks suggests that the uplift of the crust in southern Finland has been some 10 to 15 km (Zwart 1967; Hietanen 1975; Vorma 1976; Korsman 1977). The pegmatites in the study area are Svecokarelian late-orogenic dyke rocks that were obviously crystallized mainly at that depth.

The pegmatite melt was segregated from the parent magma and rose higher in the Earth's crust along the fracture zones. In places the contact metamorphic alterations in the rocks enveloping the pegmatites are slight; in others, however, the wall rock is completely recrystallized. Adjacent to the contacts of the pegmatite deposits the andalusite porphyroblasts are altered to muscovite. Occasionally biotite has disappeared from the wall rock and been replaced by magnesium-rich tourmaline and phengitic muscovite or phengite. According to Winkler (1976; see also Velde 1965, 1967), phengite characterizes rocks of low-grade contact and regional metamorphism. Thus the contact metamorphism in the wall rock caused by pegmatite melt took place either under similar or lower pressure-temperature conditions than those prevailing during regional metamorphism, and the intruding melt could not have been superheated.

Experimental studies show that in the orthoclase—albite—quartz—water system the solidus temperature at 4 kb pressure, corresponding to a depth of 15 km in the crust, is 655° to 685°C (Tuttle and Bowen 1958; Luth *et al.* 1964). If plagioclase is anorthite-rich, the solidus temperature rises parallel to the rise in calcium content (Winkler 1976). On the other hand, the elements such as lithium, phosphorus, boron and fluorine that are typical of the pegmatites in the study area, considerably lower the solidus temperature of the granitic melt. According to Wyllie and Tuttle (1961, 1964), if 5 wt% phosphorus pentoxide (P_2O_5) is added to the system granite-water the melting point of the granite at a water vapour pressure of 2.75 kb is lowered by 20°C; with 5 wt% hydrofluoric acid (HF) it is lowered by 70°C and with 5 wt% lithium oxide (Li_2O) by as much as 90°C. Chorlton and Martin (1978) have demonstrated that the addition of 5 wt% boron trioxide (B_2O_3) lowers the melting point of granite by 125°C at a water vapour pressure of 1 kb, i.e. from 720° to 596°C.

Hence the above mentioned elements markedly affect the solidus temperature of the granitic melt. The element concentrations used in the experiments were, however, hypothetical and far too high to be applied to the crystallization temperatures of natural granite pegmatites. In the complex pegmatites with abnormally high lithium, the content hardly exceeds 1.5 to 2.0 wt% Li_2O (Stewart 1978). Nevertheless, experimental studies show that even so much lower lithium abundance notably affects the solidus temperature of granite. According to Stewart (1978), in the (albite, orthoclase)–quartz–eucryptite–water system the eutectic melt at 2 kb pressure likewise contains about 2 wt% lithium (Li_2O). In other respects, too, the total composition of the melt is very close to that of the spodumene and petalite pegmatites richest in Li. The eutectic temperature of the system albite–quartz–eucryptite–water at 2 kb water vapour pressure is 640°C, and that of the system orthoclase–quartz–eucryptite–water 600° to 625°C. When both alkali feldspars are present the temperature drops by at least 35°C (Munoz 1971; Stewart 1978). Hence, depending on the total composition, the granitic water-saturated melt enriched solely in lithium may, under these conditions, crystallize at temperatures ranging from 640° to 560°C.

It is obvious that the crystallization temperature of the pegmatites in the study area was mainly affected by the water content in the pegmatite magma but also to some extent by the abundances of phosphorus, lithium, boron and fluorine. The intruding pegmatite melt was not yet water-saturated and the volatiles were enriched in the intrusion channel at the same time as the pegmatites were crystallized. Water-saturated granitic melt whose temperature is close to the solidus temperature cannot rise in the crust, because the drop in pressure causes the rising melt to crystallize immediately (Cann 1970).

In simple pegmatites and Be-Nb pegmatites boron may have had a marked effect on the crystallization temperature because particularly the dykes in the schists are occasionally very rich in tourmaline. Besides tourmaline, the lithium pegmatites are enriched in various phosphorus, lithium and fluorine minerals. With regard to the abundances of these minerals, the crystallization temperature of the pegmatites may have fallen to 600°C or even somewhat lower. This is corroborated by the crystallization temperature values obtained by the present author from the border zones of the Viitaniemi, Eräpyhä and Pajala pegmatites with the aid of microcline–low-albite–feldspar thermometry. By applying the formula of Whitney and Stormer (1977) to the analytical data in Tables 4 and 5 and assuming that the deposits crystallized at 4 kb pressure (at a depth of about 15 km) we obtain 503°C as the crystallization temperature of the border zone of the Viitaniemi pegmatite, 580°C for that of the Eräpyhä pegmatite and 585°C for the Pajala pegmatite. If, however, the crystallization took place at 2.5 kb pressure (at a depth of about 10 km), the temperatures would have been lower by 15° to 20°C.

The solidus temperature of the complex pegmatites was obviously markedly below that of the simple pegmatites. The melt rich in water, fluorine, boron, lithium and phosphorus was able to intrude farther from its parent magma than was the normal granitic melt. Hence, Li pegmatites often occur in schists around granites. Furthermore, the trends of the dykes in both simple and complex pegmatites in some subareas clearly differ from each other. The difference is most distinct in the Seppälä (II) area, where most of the deposits are Li pegmatites.

Crystallization history and origin of the internal structures

The internal structures of the pegmatites in the study area indicate that crystallization was often a very complicated event. The pegmatite melt started to crystallize in the margins of the deposits and proceeded inwards. The melt evidently started to fractionate during intrusion, and the margins of the feeding channels may have partly crystallized at a different depth than the innermost zones. Hence, the structural features and mineral composition of the pegmatites also depend on the section of dyke exposed (cf. Uebel 1977).

The zoned structures of the pegmatites in the Eräjärvi area are similar to those in numerous other complex pegmatites reported in the literature. The zoning is chiefly due to fractional crystallization (cf. Jahns and Burnham 1969; Luth and Tuttle 1969; Stewart 1978). The margins of the deposits crystallized fairly rapidly, because the border zone is always fine-grained and in sharp contact with the wall zone. The pegmatite melt started to crystallize with plagioclase, quartz and often with micas as well. Potassium feldspar is often not a main mineral until in the wall zone, where, however, its abundance is lower than that of plagioclase. On account of fractional crystallization, the plagioclase in the margins of the deposits is invariably richest in anorthite, and the percentage of the albite component increases towards the portions of pegmatite that were the last to crystallize, i.e. the intermediate zone, the replacement bodies and the fracture fillings.

Banded or layered tourmaline- and/or garnet-bearing aplite and aplite—pegmatite dykes similar to those in the Eräjärvi area have been described by Staatz and Trites (1955) from the Quartz Creek district in Colorado; by Kupfer (1960) and Orville (1960) from the Black Hills area in South Dakota and by Windley and Bridgwater (1965) from the Kînâlik area in Greenland. Besides in the Eräjärvi area, garnet-banded pegmatites have been encountered in Finland in the Peräseinäjoki—Alavus area (Haapala 1966) and in the Kajaani—Sotkamo area (Alviola 1977). In the deposits reported in the literature the banded and layered structures parallel to the contacts of the dykes are often interpreted as primary magmatic structures. The formation of the structures was affected by the crystallization rate and composition of the melt, by alternate over- and under-saturation of the main components and, particularly in the gently dipping deposits, by gravitational differentiation (Jahns and Tuttle 1963).

The banded and layered structures in the Eräjärvi pegmatites seem to be magmatic in origin although in some places they may be due to autometasomatic replacement processes caused by postmagmatic fluids (cf. p. 20). Tectonic movements during crystallization and later deformations gave rise to some secondary parallellism in the minerals of a few of the pegmatites. The joints along which the melt flowed may have opened up gradually, an event that would have caused successive injections. The occasional large variation in the composition of the zones, layers and bands indicates that the composition and water content of the magma pulses varied from time to time. Flowage differentiation also forced the crystal particles to move in the melt and may thus have been one of the chief factors producing banding in the rather homogenous garnet- and tourmaline-bearing aplites. The cross-bedding structures in the margins of the Niemelä aplite—pegmatite dyke as well as the aplite autoliths indicate that the melt was flowing at a higher rate than normal.

The grain size of the potassium feldspar increases together with the coarseness of the perthite structures from the contact of the pegmatite dykes towards the innermost zone.

As in many of the complex intrusives reported in the literature, the Si/Al ordering in potassium feldspar and the exsolution of the perthite phase are two parallel processes (cf. Parsons and Boyd 1971; Vorma 1971; Haapala 1977; Parsons 1978). The first fractions that crystallized from the melt, the margins of the dykes, show the lowest Si/Al ordering in potassium feldspar, but the ordering increases towards the fractions that were the last to crystallize, in the innermost part of the dykes. In the study area orthoclase occurs in the border zone of many lithium pegmatites. On the other hand, in simple pegmatites and Be-Nb pegmatites, microcline shows cross-hatching in the border zone, which probably means that in these pegmatites, too, the potassium feldspar primarily cystallized as orthoclase but converted to microcline below the transition temperature (cf. Laves 1950; Stewart and Wright 1974). The rapid crystallization of the dyke margins at high temperature and the sharp increase in volatile content during fractionation obviously had a decisive influence on the Si/Al ordering (cf. Chaudry 1971). Since, the analytical data indicate that the composition of feldspar (see Table 4) varies considerably, the total composition of the mineral may also have caused orthoclase to survive as a metastable phase and the $\triangle$-values to vary within the dykes.

The zoning in the simple pegmatites is often poorly developed and the replacement bodies, fracture fillings, quartz cores and miarolitic cavities may be entirely lacking. The pegmatite magma was evidently poor in volatiles. In contrast, the complex pegmatite melt was markedly richer in volatiles and its crystallization continued down to quite low temperatures. The replacement bodies and fracture fillings containing lithium silicates, lithium phosphates, niobium-tantalum minerals, cassiterite and other rarer pegmatite minerals presumably crystallized from the volatite-rich residual melts and postmagmatic supercritical fluids that segregated from the melt, invaded the fractures in pegmatites and replaced previously crystallized minerals. The melt did not become saturated with the volatiles until the intermediate zones were crystallizing, because replacement bodies, fracture fillings, miarolitic cavities and quartz cores occur mainly in the innermost parts of the pegmatites.

The large microcline crystals adjacent to the quartz cores of the pegmatites evidently grew owing to fast diffusion from the fluid phase. The core of the crystals is often grey, macroperthitic cross-hatched microcline enveloped by a rim of pure white microperthitic microcline similar to that which occurs in the replacement bodies and fracture fillings. The white microcline does not show cross-hatched twinning, and it presumably crystallized from fluids below the orthoclase—microcline transition temperature (cf. Laves 1950; Stewart and Wright 1974).

The cleavelandite and sugar albite fracture fillings do not usually cut the quartz cores of the pegmatites; hence the cores, too, probably crystallized under postmagmatic conditions. The crystallization of the quartz cores and the intraformational quartz veins intersecting the deposits was a very complicated event that varied from one deposit to the other. In some places the quartz veins are younger than the cores; in others they seem to have crystallized at the same time.

Crystallization conditions varied greatly, particularly in parts of the pegmatites that were the last to crystallize. The crystals are often zoned, and especially the phosphate minerals show several phases crystallized at different temperatures and replacing one another. The pseudomorphs after triphylite—lithiophilite and also after beryl and tourmaline are common in pegmatites. The supercritical fluids sometimes contained abundant fluorine in addition to water, as is shown by the enrichment in fluorine in many of the minerals in replacement bodies, fracture fillings, pseudomorphs and cavities. Many secondary phosphates are alteration products that contain molecular water and probably crystallized from hydrothermal solutions at very low temperatures ($< 250°C$); some perhaps even under supergene conditions.

ACKNOWLEDGEMENTS

This study is based on material that I collected in the course of regional geological research work for the Geological Survey of Finland. I carried out the detailed pegmatite studies in the Eräjärvi area, however, in 1977—1978 with the financial aid of the Finnish Natural Research Council (Suomen Luonnonvarain Tutkimussäätiö) and the Cultural Foundation of Finland (Suomen Kulttuurirahasto).

Dr. Ilkka Laitakari has helped me in many ways during the work, not least with our fruitful discussions on regional geology. Prof. Th. G. Sahama, Prof. Maunu Härme and Dr. Ilmari Haapala have shown continuous interest in my studies and have given me much invaluable advice. My special thanks are due to Prof. Atso Vorma and Mr. Pekka Kallio for their kind suggestions and help during the laboratory studies. For the wet chemical analyses I owe much to Prof. O. von Knorring (at the University of Leeds), to the late Mr. Pentti Ojanperä and to Mr. Risto Saikkonen, for the X-ray fluorescence analyses to M. Väinö Hoffrén, for the electron microprobe analyses to Mrs. Tuula Hautala and for the spectrographical analyses to Miss Ringa Danielsson and Mr. Ari Puisto.

Dr. Olavi Kouvo and Mr. Matti Sakko performed the radiometric age determinations. Mrs. Mirja Saarinen and Miss Raija Mamia helped with the X-ray identification and mineral separations. Mr. Erkki Halme took many of the photographs. Mrs. Helga Leppänen typed the text and Mrs. Elsa Järvimäki, Mrs. Ritva Forsman and Miss Liisa Siren copied the maps and drawings. I am deeply grateful to all the persons mentioned and to anyone unnamed who in one way or another has furthered my research.

The manuscript was translated into English by Mrs. Gillian Häkli. Prof. Jaakko Siivola and Dr. Martti Lehtinen read the manuscript and offered constructive criticism.

Accepted for publication 30th May, 1980

REFERENCES

Ahrens, L. H., 1951. A cooperative investigation of precision and accurracy in chemical, spectrochemical, and modal analysis of silicate rocks. Part 4. Spectrochemical analysis of some of the rarer elements in the granite and diabase samples. U.S. Geol. Surv., Bull. 980, 53—57.

Alviola, Reijo, 1977. Kajaanin alueen graniittipegmatiitit. Manuscript, Department of Geology and Mineralogy, University of Helsinki. 82 p.

Andersen, Steen, 1967. On beryllite and bertrandite from the Ilimaussaq alkaline intrusion, South Greenland. Medd. Groenland 181 (4), 11—28.

Andersson. Aug. A., 1891. Dagbok [Field diary]. Geological Survey of Finland.

Annersten, Hans & Hålenius, Ulf, 1976. Ion distribution in pink muscovite, a discussion. Am. Mineral. 61, 1045—1050.

Ashraf, Mohammad, 1975. Composite albitite–aplite and pegmatite from Mansehra and Batgram area, Hazara district, Pakistan. Fortschr. Mineral. 52, Spec. Issue: IMA-papers 9th meeting, 329—344.

Askvik, Helge, 1972. Red muscovite from a metasedimentary gneiss, Sogn, West Norway. Norsk Geol. Unders. 273, 7—11.

Aurola, Erkki, 1963. On the pegmatites in the Torro area, south-western Finland. Bull. Comm. géol. Finlande 206. 32 p.

Bambauer, H. U.; Corlett, M.; Eberhard, E. & Viswanathan, K., 1967. Diagrams for the determination of plagioclases using X-ray powder methods. Part III of laboratory investigations on plagioclases. Schweiz. Mineral. Petr. Mitt. 47, 333—349.

Barnes, W. H. & Shore, Violet C., 1951. The childrenite–eosphorite problem. Am. Mineral. 36, 509—511.

Bazarova, S. B.; Koval, P. V.; Kovalenko, V. I. & Kashayev, A. A., 1977. Relationships among parameter b optical properties of biotite, muscovite and lepidolite and the composition of mica types. Dokl. Acad. Sci. USSR, Earth Sci. Sect. 226, 107—110.

Behrens, Manfred & Siehl, Agemar, 1975. GELI 2 — ein Rechenprogramm zur Gefüge — und Formanalyse. Geol. Rundsch. 64, 301—324.

Berggren, Thelma, 1941. Minerals of the Varuträsk Pegmatite. XXV. Some new analyses of lithiumbearing mica minerals. Geol. Fören. i Stockholm Förh. 63, 262—278.

Berry, L. G. & Thompson, R. M., 1962. X-ray powder data for ore minerals: The Peacock Atlas. Geol. Soc. Am., Mem. 85. 281 p.

Billings, Marland P., 1972. Structural geology. 3th edition, Prentice-Hall, Englewood Cliff, N. J. 606 p.

Borg, I. Y. & Smith, D. K., 1969. Calculated X-ray powder patterns for silicate minerals. Geol. Soc. Am., Mem. 122. 896 p.

Bourguignon, P. & Mélon, J., 1965. Wodginite du Rwanda. Ann. Soc. Géol. Belg. 88, 291—300.

Bukin, G. V., 1967. Uslovija kristallizatsii fenakit–bertrandit–kvartsevoj assotsiatsii (ehsperimental'nye dannye). Dokl. Akad. Nauk SSSR, Geol. 176 (3), 664—667 (in Russian).

Burke, E. A. J.; Kieft, C.; Felius, R. O. & Adusumilli, S. Maria, 1970. Wodginite from northeastern Brazil. Geol. Mijnbouw 49, 235—240.

Cameron, E. N.; Jahns, R. H.; McNair, A. H. & Page, L. R., 1949. Internal structure of granitic pegmatites. Econ. Geol., Monogr. 2. 115 p.

Cann, J. R., 1970. Upward movement of granitic magma. Geol. Mag. 107, 335—340.

Cassedane, J. P. & Cassedane, J. O., 1978. Ludlamite crystals from Galilea. Mineralogical notes. Mineral. Rec. 9, 41.

Černý, Petr, 1968. Beryllumwandlungen in Pegmatiten — Verlauf und Produkte. Neues Jahrb. Mineral., Abh. 108, 166—180.

Černý, P. & Harris, D. C., 1973. Allemontite and its alteration products from the Odd West pegmatite, southeastern Manitoba. Can. Mineral. 11, 978—984.

Černý, P. & Harris, D. C., 1978. The Tanco pegmatite at Bernic Lake, Manitoba. XI. Native elements, alloys, sulfides and sulfosalts. Can. Mineral. 16, 625—640.

Černý, P. & Trueman, D. L., 1978. Distribution and petrogenesis of lithium pegmatites in the western Superior province of the Canadian Shield. Energy 3, 365—377.

Chaudhry, M. Nawaz, 1971. Feldspars from the Meldon aplite, Devonshire, England. Mineral. Mag. 38, 179—185.

Chorlton, Lesley B. & Martin, Robert F., 1978. The effect of boron on the granite solidus. Can. Mineral. 16, 239—244.

Dana, Edward S. & Wells, Horace L., 1889. Description of the new mineral, Beryllonite. Am. J. Sci. 37, 23—32.

Deer, W. A.; Howie, R. A. & Zussman, J., 1967. Rock-forming minerals. Vol. 3. Sheet silicates. 6th impression. Longmans, Green and Co. Ltd., London. 270 p.

Dubois, J.; Marchand, J. & Bourguignon, P., 1972. Données minéralogiques sur la série amblygonite–montebrasite. Ann. Geol. Soc. Belg. 95, 285—310.

Erämetsä, Olavi & Sihvonen, Marja-Liisa, 1973. Studies on the extraction of cesium from pollucite. Suomen Kemistilehti (Finn. Chem. J.) 46, 13—17.

Fischer, D. Jerome & Volborth, Alex, 1960. Morinite–apatite–whitlockite. Am. Mineral. 45, 645—667.

Fleischer, Michael, 1965. Summary of new data on rock samples G-1 and W-1, 1962—1965. Geochim. Cosmochim. Acta 29, 1263—1283.

— , 1975. 1975 Glossary of mineral species. Mineral. Rec. Bowie, Maryland. 145 p.

Fransolet, André-Mathieu, 1974—1975. Etude minéralogique et pétrologique des phosphates de pegmatites granitiques. Fasc. I: Texte, Fasc. II: Illustrations et annexes. Dissertation, Université de Liège, Faculté des Sciences. 317 p.

Frosterus, Benj., 1900. Vuorilajikartta, [Pre-Quaternary rocks], C2 Mikkeli. Suomen geologinen yleiskartta [General geological map of Finland] 1 : 400 000.

— , 1903. Vuorilajikartan selitys, C2 Mikkeli. Suomen geologinen yleiskartta [Explanation to the General geological map of Finland, sheet C2 Mikkeli], 1 : 400 000. 126 p.

Gaubert, M. Paul, 1907. Sur les indices de réfraction de quelques minéraux. Bull. Soc. Min. Fr. 30, 104—108.

Getmanskaja, T. I., 1966. O psevdomorfozakh bertrandita i fljuorita po piritu. Akad. Nauk SSSR, Tr. Mineral. Muz. 17, 190—195 (in Russian).

Giuseppetti, G. & Tadini, Carla, 1973. Refinement of the crystal structure of beryllonite, $NaBePO_4$. Tschermaks Mineral. Petrol. Mitt. 20, 1—12.

Goldsmith, Julian R. & Laves, Fritz, 1954a. The microcline–sanidine stability relations. Geochim. Cosmochim. Acta 5, 1—19.

Goldsmith, Julian R. & Laves, Fritz, 1954b . Potassium feldspars structurally intermediate between microcline and sanidine. Geochim. Cosmochim. Acta 6, 100—118.

Gossner, B. & Besslein, J., 1934. Über kristallographische Beziehungen zwischen Silikaten und Phosphaten. Ein Beitrag zur Kenntnis des Beryllonites. Centralbl. Mineral. Geol. Paläontol. Jg. 1934, A, Mineral. Petrol. 144—151.

Govorov, I. N. & Stunzhas, A. A., 1963. Mode of transport of beryllium in alkali metasomatism. Geochem. 1963, No. 4, 402—409.

Graham, J. & Thornber, M. R., 1974. The crystal chemistry of complex niobium and tantalum oxides. II. Composition and structure of wodginite. Am. Mineral. 59, 1040—1044.

Grice, J. D.; Černý, P. & Ferguson, R. B., 1972. The Tanco pegmatite at Bernic Lake, Manitoba. II. Wodginite, tantalite, pseudo-ixiolite and related minerals. Can. Mineral. 11, 609—642.

Grigor'ev, Iv. F. & Dolomanova, E. I., 1955. Bertrandit iz olovorudnogo mestorozhdenija v Tsentral'nom Zabajkal'e. Akad. Nauk. SSSR, Tr. Mineral. Muz. 7, 151—154 (in Russian).

Haapala, Ilmari, 1966. On the granitic pegmatites in the Peräseinäjoki—Alavus area, South Pohjanmaa, Finland. Bull. Comm. géol. Finlande 224. 98 p.

— , 1977. Petrography and geochemistry of the Eurajoki stock, a rapakivi-granite complex with greisen-type mineralization in southwestern Finland. Geol. Surv. Finland, Bull. 286, 128 p.

Haapala, Ilmari & Ojanperä, Pentti, 1972. Genthelvite-bearing greisens in southern Finland. Geol. Surv. Finland, Bull. 259, 22 p.

Hanson, A. W., 1960. The crystal structure of eosphorite. Acta Cryst. 13, 384—387.

Heinrich, E. Wm., 1953. Zoning in pegmatite districts. Am. Mineral. 38, 68—87.

— , 1965. Holmquistite and pegmatite lithium exomorphism. Indian Mineral. 6, 1—13.

Heinrich, E. Wm. & Levinson, Alfred A., 1953.

Studies in the mica group; Mineralogy of the rose muscovites. Am. Mineral. 38, 25—49.

Henderson, Wm. A., Jr., 1975. The bertrandites of Connecticut. Min. Rec. 6, 114—123.

Hietanen, Anna, 1975. Generation of potassium-poor magmas in the northern Sierra Nevada and the Svecofennian of Finland. J. Res. U.S. Geol. Surv. 3, 631—645.

Hirsjärvi, V. P., 1933. Eräjärven Uiherlan pegmatiitti. Manuscript, Department of Geology and Mineralogy, University of Helsinki. 16 p.

Hurlbut, Cornelius, S., Jr., 1950. Childrenite–eosphorite series. Am. Mineral. 35, 793—805.

Ibers, James A. & Hamilton, Walter C. (Editors), 1974. International tables for X-ray crystallography. Vol. IV. Revised and supplementary tables to volumes II and III. Kynoch Press, Birmingham. 366 p.

Jahns, Richard, H., 1955. The study of pegmatites. Econ. Geol. Fiftieth Anniv. Vol., 1025—1130.

Jahns, Richard H. & Burnham, C. Wayne, 1969. Experimental studies of pegmatite genesis: I. A model for the derivation and crystallization of granitic pegmatites. Econ. Geol. 64, 843—864.

Jahns, Richard H. & Tuttle, O. Frank, 1963. Layered pegmatite-aplite intrusives. Mineral. Soc. Am., Spec. Pap. 1, 78—92.

JCPDS Powder Data File. Sets 1—29. JCPDS International Centre for Diffraction Data, Pennsylvania, U.S.A.

de Jesus, Amílcar Mário, 1933. Pegmatites mangano-litiníferas da região de Mangualde. Comun. Serv. Geol. Port. XIX, 65—210.

Kallio, Pekka & Alviola, Reijo, 1975. Triphylites and lithiophilites from the granite pegmatites of the communes of Kitee and Tohmajärvi, E. Finland. Fortschr. Mineral. 52, Spec. Issue: IMA-papers 9th meeting, 279—283.

Khvostova, V. A.; Pavlova, V. N.; Aleksandrov, V. B. & Maksimova, N. V., 1966. Pervaja nahodka vodzhinita v SSSR. Dokl. Akad. Nauk. SSSR 167 (5), 1135—1138 (in Russian).

von Knorring, O. 1970a. Mineralogical and geochemical aspects of pegmatites from orogenic belts of equatorial and southern Africa. Pp. 157—184 *in* African magmatism and tectonics, ed. by T. N. Clifford and I. G. Gass. Oliver & Boyd, Edinburgh.

— **, 1970b.** Some mineralogical and geochemical aspects of phosphorus in granite pegmatites. Res. Inst. Afr. Geol., University of Leeds, Department of Earth Sciences. 14th Ann. Rep. 30—32.

von Knorring, Oleg; Sahama, Th. G. & Lehtinen, Martti, 1969a. Ferroan wodginite from Ankole, south-west Uganda. Bull. Geol. Soc. Finland 41, 65—69.

von Knorring, O.; Sahama, Th. G. & Lehtinen, Martti, 1969b. Wodginite from Karibib, South West Africa. Indian Mineral. 10, 105—108.

Korsman, Kalevi, 1977. Progressive metamorphism of the metapelites in the Rantasalmi—Sulkava area, southeastern Finland. Geol. Surv. Finland, Bull. 290. 82 p.

Kouvo Olavi & Tilton, G. R., 1966. Mineral ages from the Finnish Precambrian. J. Geol. 74, 421—442.

Koval', P. V.; Kovalenko, V. I.; Kuz'min, M. I.; Pisarskaja, V. A. & Jurchenko, S. A., 1972. Mineral'nye paragenezisy, sostav i nomenklatura sljud redkometal'nykh al'bitsoderzhashchikh granitoidov. Dokl. Akad. Nauk. SSSR, Geol. 202, 1174—1177 (in Russian).

Kupfer, Donald H., 1960. Pegmatite–granite relationships in the Calamity Peak area Black Hills, South Dakota, U.S.A. Int. Geol. Congr. XXI Sess. Norden 1960, part XVII, Minerals and Genesis of Pegmatites, 77—93.

Lahti, Seppo I., 1974. Eräjärven seudun pegmatiittien mineralogiasta ja rakenteesta. Manuscript, Department of Geology and Mineralogy, University of Helsinki. 116 p.

— **, 1975.** Eräjärven seudun pegmatiittien mineraaleista. Summary: On the minerals of the pegmatites of the Eräjärvi area. Geologi 27, 83—86. English summary also in Zentralbl. Mineral. 1975 I (1), 228.

Laitakari, Ilkka, 1969. On the set of olivine diabase dikes in Häme, Finland. Bull. Comm. géol. Finlande 241. 65 p.

Lappalainen, Rolf & Neuvonen, K. J., 1968. Trace elements in some Finnish pegmatitic potassium feldspars. Bull. Geol. Soc. Finland. 40, 59—64.

Laves, Fritz, 1950. The lattice and twinning of microcline and other potash feldspars. J. Geol. 58, 548—571.

Lawrence, Edmond F., 1963. Antimony deposits of Nevada. Nev. Bur. Mines, Bull. 61, 248 p.

Lima de Faria, J., 1964. Identification of metamict minerals by X-ray powder photographs. Junta de Investigações do Ultramar. Estud. Ens. Doc. 112, 74 p.

Lindberg, M. L.; Pecora, W. T. & Barbosa, A. L. de M., 1953. Moraesite, a new hydrous beryllium phosphate from Minas Gerais, Brazil. Am. Mineral. 38, 1126—1133.

Lonka, Anssi, 1961. Eräjärven pegmatiittitutkimukset talvella 1960—1961. Manuscript, Suomen Malmi Oy. 3 p.

Luna, Jan, 1965. Wodginit a kolumbit–tantalit z pegmatitu od Krasonic. English Abstract. Acta Univ. Carol., Geol. 3, 157—162.

Luoto, Kosti, 1967. Eräjärven maasälpälouhoksesta. PK [Newsletter of Paraisten Kalkki Oy] 81 (2), 7-8.

Luth, William C.; Jahns, Richard H. & Tuttle, O. Frank. 1964. The granite system at pressures of 4 to 10 kilobars. J. Geophys. Res. 69, 759—773.

Luth, W. C. & Tuttle, O. F., 1969. The hydrous vapor phase in equilibrium with granite and granite magmas. Geol. Soc. Am., Mem. 115, 513—548.

Mäkinen, Eero, 1913. Die Granitpegmatite von Tammela in Finland und ihre Minerale. Bull. Comm. géol. Finlande 35, 101 p.

Mårtensson, C., 1960. Euklas und Bertrandit aus dem Feldspat–pegmatit von Kolsva in Schweden. Neues Jahrb. Mineral., Abh. 94, 1248—1252.

Matisto, Arvo, 1964. Kallioperäkartta (Pre-Quaternary rocks), lehti - sheet - 2141 Kangasala. Suomen geologinen kartta (Geological map of Finland) 1 : 100 000.

— , 1976. Kangasalan kartta-alueen kallioperä. Summary: Precambrian rocks of the Kangasala map-sheet area. Suomen geologinen kartta 1 : 100 000. Kallioperäkartan selitykset, 2141 Kangasala. 27 p.

— , 1977. Tampereen kartta-alueen kallioperä. Summary: Precambrian rocks of the Tampere map-sheet area. Suomen geologinen kartta 1 : 100 000. Kallioperäkartan selitykset, 2123 Tampere. 50 p.

Matrosov, I. I., 1978. The banded structures of the rare-metal pegmatites. Int. Geol. Rev. 20, 177—186.

Mikkola, Toini & Wiik, H. B., 1947. Petalite, a mineral new to Finland. Bull. Comm. géol. Finlande 140, 281—286.

Mikkonen, Antti, 1963. Pollusiittia Viitaniemen pegmatiitissa. Geologi 15, 66.

Moore, Paul B., 1971. Crystal chemistry of the alluaudite structure type: Contribution to the paragenesis of pegmatite phosphate giant crystals. Am. Mineral. 56, 1955—1975.

— , 1973. Pegmatite phosphates: Descriptive mineralogy and crystal chemistry. Mineral. Rec. 4, 103—130.

Moore, Paul Brian & Ito, Jun, 1978. I. Whiteite a new species, and a proposed nomenclature for the jahnsite–whiteite complex series. II. New data on xanthoxenite. III. Salmonsite discredited. Mineral. Mag. 42, 309—323.

Moss, A. A.; Fejer, E. E. & Embrey, P. G., 1969. On the X-ray identification of amblygonite and montebrasite. Mineral. Mag. 37, 414—422.

Mrose, Mary E., 1952. Hurlbutite, $CaBe_2(PO_4)_2$, a new mineral. Am. Mineral. 37, 931—940.

— , 1971. Lacroixite: its redefinition and new occurrences. Progr. and Abstr. 20th Clay Min. Conf., Black Hills, South Dakota, p. 10. Abstract also in Am. Mineral. 57, p. 1914, 1972.

Mrose, Mary E. & Appleman, Daniel E., 1962. The crystal structures and crystal chemistry of väyrynenite, $(Mn,Fe)Be(PO_4)(OH)$, and euclase, $AlBe(SiO_4)(OH)$. Z. Kristallogr. 117, 16—36.

Mrose, Mary E. & von Knorring, Oleg, 1959. The mineralogy of väyrynenite, $(Mn,Fe)Be(PO_4)(OH)$. Z. Kristallogr. 112, 275—288.

Munoz, J. L., 1971. Hydrothermal stability relations of synthetic lepidolite. Am. Mineral. 56, 2069—2087.

Neuvonen, K. J., 1960. The pegmatite of Kaatiala. *In:* Nomentallic mineral deposits. Guide to excursions nos. A 37, A 38, C 32, C 33, compiled by Erkki Aurola and K. J. Neuvonen. Int. Geol. Congr., XXI Session, Norden 1960, 50—51.

Nickel, E. H.; Rowland, J. F. & McAdam, R. C. 1963a. Wodginite — a new tin-manganese tantalate from Wodgina, Australia and Bernic Lake, Manitoba. Can. Mineral. 7, 390—402.

Nickel, E. H.; Rowland, J. F. & McAdam, R. C., 1963b. Ixiolite — a columbite substructure. Am. Mineral. 48, 961—979.

Nieminen, Kalervo, 1978. On the geochemistry of Kaatiala pegmatite, western Finland. Bull. Geol. Soc. Finland. 50, 1—14.

Nriagu, J. O. & Dell, C. I., 1974. Diagenetic formation of iron phosphates in recent lake sediments. Am. Mineral. 59, 934—946.

Nykänen, Osmo, 1975. Kerimäen ja Kiteen kartta-alueen kallioperä. Summary: Precambrian rocks of the Kerimäki and Kitee map-sheet areas. Suomen geologinen kartta 1 : 100 000. Kallioperäkartan selitykset, 4213 Kerimäki, 4231 Kitee. 45 p.

Ödman, Olof H., 1941. Minerals of the Varuträsk pegmatite. XXVIII. On »stibiomicrolite» and its decomposition products. Geol. Fören. i Stockholm, Förh. 63, 289—294.

Orville, P. M., 1960. Petrology of several pegmatites in the Keystone district, Black Hills, South Dakota. Bull. Geol. Soc. Am. 71, 1467—1490.

Palache, Charles & Shannon, Earl V., 1928. Beryllonite and other phosphates from Newry, Maine. Am. Mineral. 13, 392—396.

Palache, Charles; Berman, Harry & Frondel, Clifford, 1957. System of Mineralogy of James Dwight Dana and Edward Salisbury Dana. Yale University 1837—1892. Vol. 11, 7th ed. John Wiley and Sons, New York. 1124 p.

Parsons, Ian, 1978. Feldspars and fluids in cooling plutons. Mineral. Mag. 42, 1—17.

Parsons, Ian & Boyd, Rognvald, 1971. Distribution of potassium feldspars polymorphs in intrusive sequences. Mineral. Mag. 38, 295—311.

Pehrman, Gunnar, 1945. Die Granitpegmatite von Kimito (S. W.-Finland) und ihre Minerale. Acta Acad. Abo. Math. Phys. XV, 2, 84 p.

Quensel, Percy, 1955. The paragenesis of the Varuträsk pegmatite including a review of its mineral assemblage. Ark. Mineral. Geol. 2, 9—125. 1957.

Ramdohr, Paul, 1969. The ore minerals and their intergrowths. Pergamon Press, Ltd. 1174 p.

Richardson, Steven M., 1975. A pink muscovite with reverse pleochroism from Archer's Post, Kenya. Am. Mineral. 60, 73—78.

Richardsson, Steven M., 1976. Ion distribution in pink muscovite: a reply. Am Mineral. 61, 1051—1052.

Roberts, Willard L. & Rapp, George, Jr., 1965. Mineralogy of the Black Hills. South Dakota School of Mines and Technol., Bull. 18, 268 p.

Rodrigues da Silva, R., 1975. Phosphate minerals from pegmatites of northeastern Brazil. Fortschr. Mineral. 52, Spec. Issue: IMA -papers 9th meeting, 293—301.

Saltikoff, B.; Honkamo, M. & Kallio, P., 1976. BASIC-kielisiä tietokoneohjelmia petrokemiallisiin laskutehtäviin. Summary: BASIC computer programs for petrochemical calculations. Geologinen tutkimuslaitos, Tutkimusraportti (Geol. Surv. Finland, Report of investigation) 13. 11 p.

San Miguel, Alfredo, 1969. The aplite-pegmatite association and its petrogenetic interpretation. Lithos 2, 25—37.

Schneiderhöhn, Hans, 1961. Die Erzlagerstätten der Erde. Bd. II. Die Pegmatite. Gustav Fischer Verlag, Stuttgart. 720 p.

Seitsaari, Juhani, 1954. The Västilä area in the Tampere schist belt. Bull. Comm. géol. Finlande 166, 95—106.

Simonen, Ahti, 1952. Kallioperäkartan selitys. Lehti - sheet - 2124 Viljakkala-Teisko. Summary: Explanation to the map of rocks. Suomen geologinen kartta 1 : 100 000. 74 p.

— **, 1971.** Das finnische Grundgebirge. Geol. Ruhdsch. 60, 1406—1421.

Skinner, Brian J., 1965. The system arsenic-antimony. Econ. Geol. 60, 228—239.

Slavik, F., 1914. Neue Phosphate vom Greifenstein bei Ehrenfriedersdorf. Bull. Int. Acad. Sci. L'Empereur François Joseph XIX, 108—123.

— **, 1915.** Notiz über den Lacroixit. Bull. Int. L'Acad. Sci. Bohême 20, 1—2.

Solov'eva, L. P. & Belov, N. V., 1961. Kristallicheskaja struktura bertrandita $Be_4Si_2O_7(OH)_2$. Dokl. Akad. Nauk. SSSR 140, 685—688 (in Russian).

Staatz, Mortimer H. & Trites, Albert F., 1955. Geology of the Quartz Creek pegmatite district, Gunnison Country, Colorado, U. S. Geol. Surv., Prof. Pap. 265, 1—111.

Stewart, David B., 1978. Petrogenesis of lithium-rich pegmatites. Am. Mineral. 63, 970—980.

Stewart, David B. & Wright, Thomas L., 1974. Al/Si order and symmetry of natural alkali feldspars, and the relationship of strained cell parameters to bulk composition. Bull. Soc. fr. Minéral. Cristallogr. 97, 356—377.

Strunz, Hugo, 1974. Excursion A1. Granites and pegmatites in Eastern Bavaria. Excursion Guidebook, I.M.A. 9th meeting, ed. by G. Troll. Fortschr. Mineral. 52, Beih. 1, 1—32.

Strunz, H. & Fischer, Margarete, 1957. Childro-Eosphorit, Tavorit, und Fairfieldit von Hagendorf. Neues Jahrb. Mineral., Monatsch. Jg. 1957, 78—88. 1958.

Swanson, Howard E.; Fuyat, Ruth K. & Ugrinic, George M., 1954. Standard X-ray diffraction powder patterns. Antimony, Sb (hexagonal). Nat. Bureau of Standards, Circ. 539, III, 14—15.

Syritso, L. F.; Sokolova, E. P.; Zorina, M. L. & Tomilov, S. B., 1977. Miscibility in the isomorphous series biotite–muscovite. Dokl. Akad. Nauk. USSR, Earth Sci. Sect. 227, 123—127.

Tuttle, O. F. & Bowen, N. L., 1958. Origin of granite in the light of experimental studies in the system $NaAlSi_3O_8-KAlSi_3O_8-SiO_2-H_2O$. Geol. Soc. Am., Mem. 74. 153 p.

Uebel, Peter-Jürg, 1977. Internal Structure of pegmatites, its origin and nomenclature. Neues Jahrb. Mineral., Abh. 131, 83—113.

Vaasjoki, Matti, 1977. Rapakivi granites and other postorogenic rocks in Finland: their age and the lead isotopic composition of certain associated galena mineralizations. Geol. Surv. Finland, Bull. 294, 64 p.

Velde, B., 1965. Phengite micas: Synthesis, stability, and natural occurrence. Am. J. Sci. 263, 886—913.

— , 1967. Si^{+4} content of natural phengites. Contrib. Mineral. Petrol. 14. 250—258.

Vernon, R. H. & Williams, K. L., 1960. Bertrandite from Mica Creek, Queensland. Am. Mineral. 45, 1300—1303.

Vesasalo, Arvo, 1959. On the petalite occurrences of Tammela, SW-Finland. Bull. Comm. géol. Finlande 184, 59—74.

Virkkunen, Marjatta, 1964. Pegmatiittitutkimuksia Eräjärvellä. Manuscript, Geol. Surv. Finland. 65 p.

Vlasov, K. A. (Editor), 1966. Geochemistry and mineralogy of rare elements and genetic types of their deposits. Vol. II. Mineralogy of rare elements. Wiener Bindery Ltd., Jerusalem. 945 p.

— , 1968. Geochemistry and mineralogy of rare elements and genetic types of their deposits. Vol. III. Genetic types of or rare-element deposits. Wiener Bindery Ltd., Jerusalem. 916 p.

Volborth, A., 1951. Montebrasit (Amblygonit) von Eräjärvi. Bull. Comm. géol. Finlande 154, 171—176.

— , 1954a. Phosphatminerale aus dem Lithiumpegmatit von Viitaniemi, Eräjärvi, Zentral-Finnland. Ann. Acad. Sci. Fennicae, Ser. A III, Geol.-Geogr., 39, 90 p.

— , 1954b. Väyrynenit $BeMnPO_4(OH)$, ein neues Mineral. Österr. Akad. Wiss. Anz., Math. Naturwiss. Kl. 91, (2), 21—23.

— , 1954c. The pegmatite quarry of Viitaniemi. *In:* The mines and quarries of Finland, ed. by Erkki Aurola, Geotekn. Julk. 55, 91—92.

— , 1956. Die Mineralparagenese im Lithiumpegmatit von Viitaniemi (Zentralfinnland) vom geochemischen Standpunkt. Tschermaks Mineral. Petrol. Mitt. 5, 273—283.

— , 1960. Gediegen Wismutantimon und andere Erzmineralien im Li-Be-Pegmatit von Viitaniemi, Eräjärvi, Zentralfinnland. Neues. Jahrb. Mineral., Abh. 94, 140—149.

Vorma, Atso, 1971. Alkali feldspars of the Wiborg rapakivi massif in southeastern Finland. Bull. Comm. géol. Finlande 246. 72 p.

— , 1976. On the petrochemistry of rapakivi granites with special reference to the Laitila massif, southwestern Finland. Geol. Surv. Finland, Bull. 285. 98 p.

Vorma, Atso & Siivola, Jaakko, 1967. Sukulaite — $Ta_2Sn_2O_7$ — and wodginite as inclusions in cassiterite in the granite pegmatite in Sukula, Tammela in SW Finland. Bull. Comm. géol. Finlande 229, 173—187.

Warry, N. D. & Kramer, James R., 1976. Some factors affecting the synthesis of cryptocrystalline strengite from an amorphous phosphate complex. Can. Mineral. 14, 40—46.

Wehrenberg, John P., 1954. A corrected unit cell for beryllonite. Am. Mineral. 39, 397.

Whitney, James A. & Stormer, John C., Jr., 1977. The distribution of $NaAlSi_3O_8$ between coexisting microcline and plagioclase and its effect on geothermometric calculations. Am. Mineral. 62, 687—691.

Whitten, E. H. Timothy, 1966. Structural geology of folded rocks. Rand McNally & Co., Chicago. 663 p.

Windley, B. & Bridgwater, D., 1965. The layered aplite–pegmatite sheets of Kînâlik, South Greenland. Medd. Groenland 171, 37 p.

Winkler, Helmut G. F., 1976. Petrogenesis of metamorphic rocks. 4th edition. Springer-Verlag, New York. 334 p.

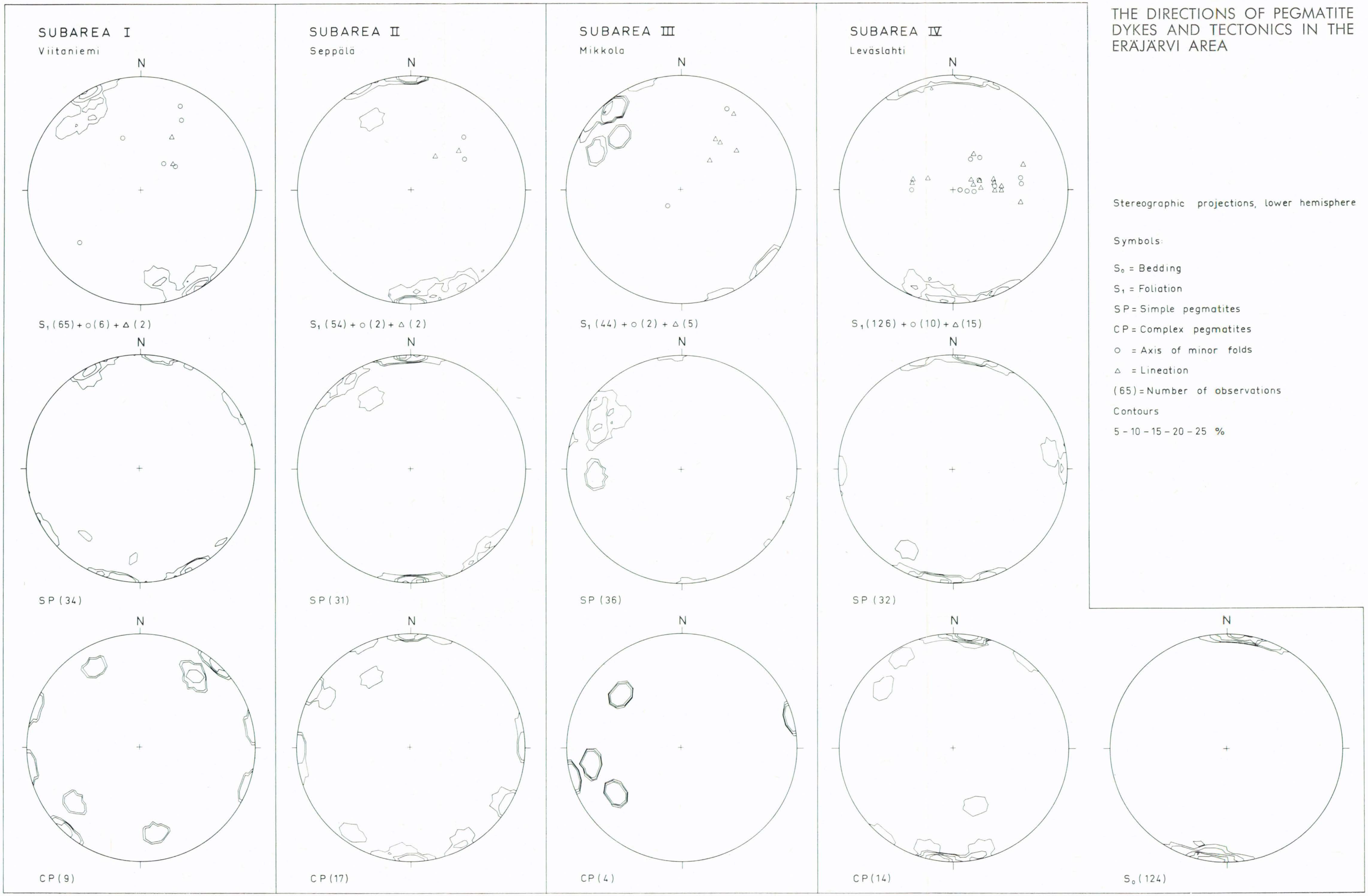
Wretblad, P. E., 1941. Minerals of the Varuträsk pegmatite. XX. Die Allemontite und das System As-Sb. Geol. Fören. i Stockholm Förh. 63, 19—48.

Wyllie, P. J. & Tuttle, O. F., 1961. Experimental investigation of silicate systems containing two volatile components. Part II. The effects of NH_3 and HF, in addition to H_2O on the melting temperatures of albite and granite. Am. J. Sci. 259, 128—143.

Wyllie, P. J. & Tuttle, O. F., 1964. Experimental investigation of silicate systems containing two volatile components. Part III. The effects of SO_3, P_2O_5, HCl, and Li_2O, in addition to H_2O, on the melting temperatures of albite and granite. Am. J. Sci. 262, 930—939.

Zelt, George A. D., 1975. Data for some African columbite–tantalite specimens. Bull. Geol. Soc. Finland 47, 117—125.

Zwart, H. J., 1967. The duality of orogenic belts. Geol. Mijnbouw 46, 283—309.



THE DIRECTIONS OF PEGMATITE DYKES AND TECTONICS IN THE ERÄJÄRVI AREA

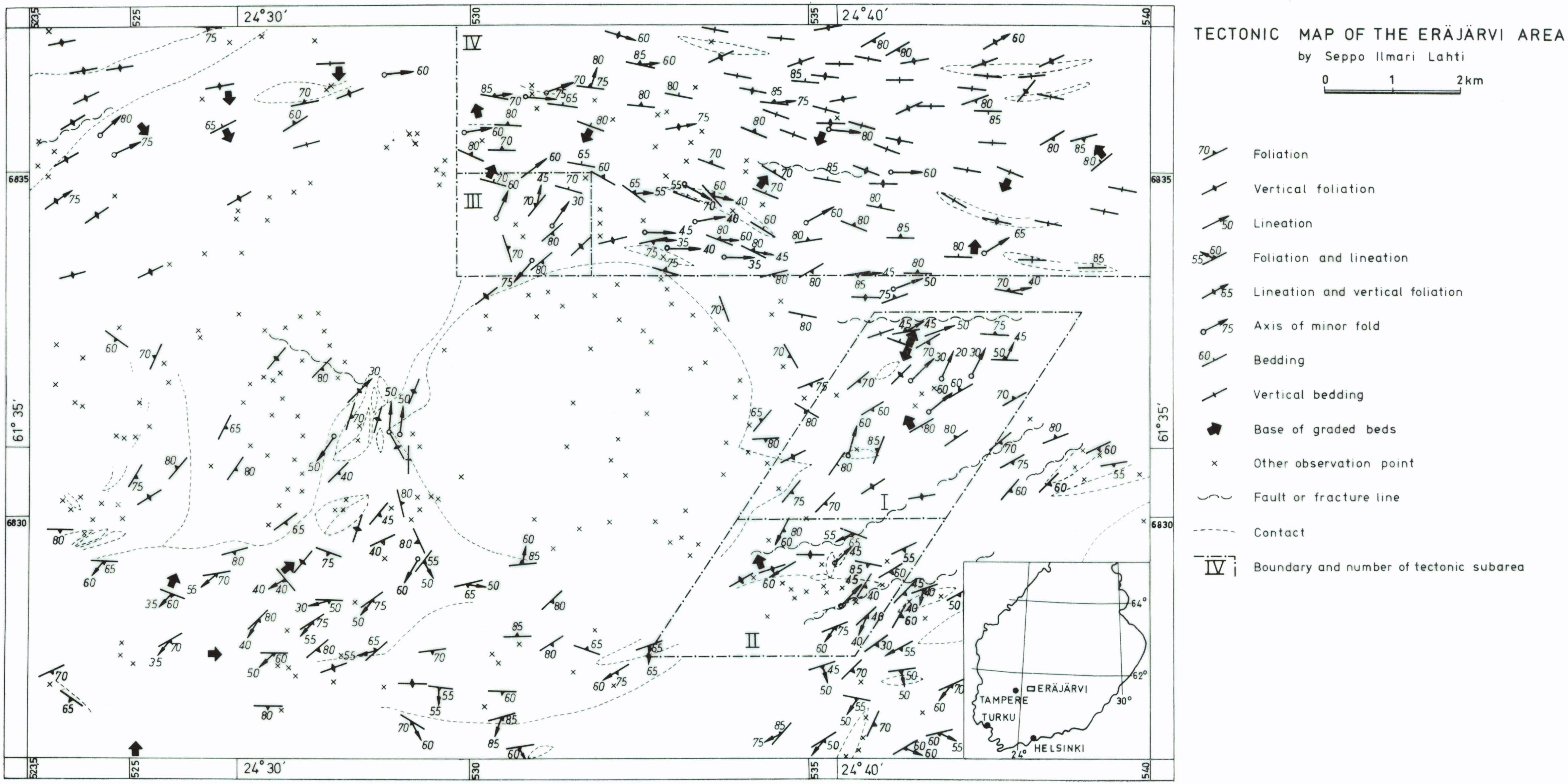
Stereographic projections, lower hemisphere

Symbols:

S_0 = Bedding
S_1 = Foliation
SP = Simple pegmatites
CP = Complex pegmatites
○ = Axis of minor folds
△ = Lineation
(65) = Number of observations
Contours
5 – 10 – 15 – 20 – 25 %

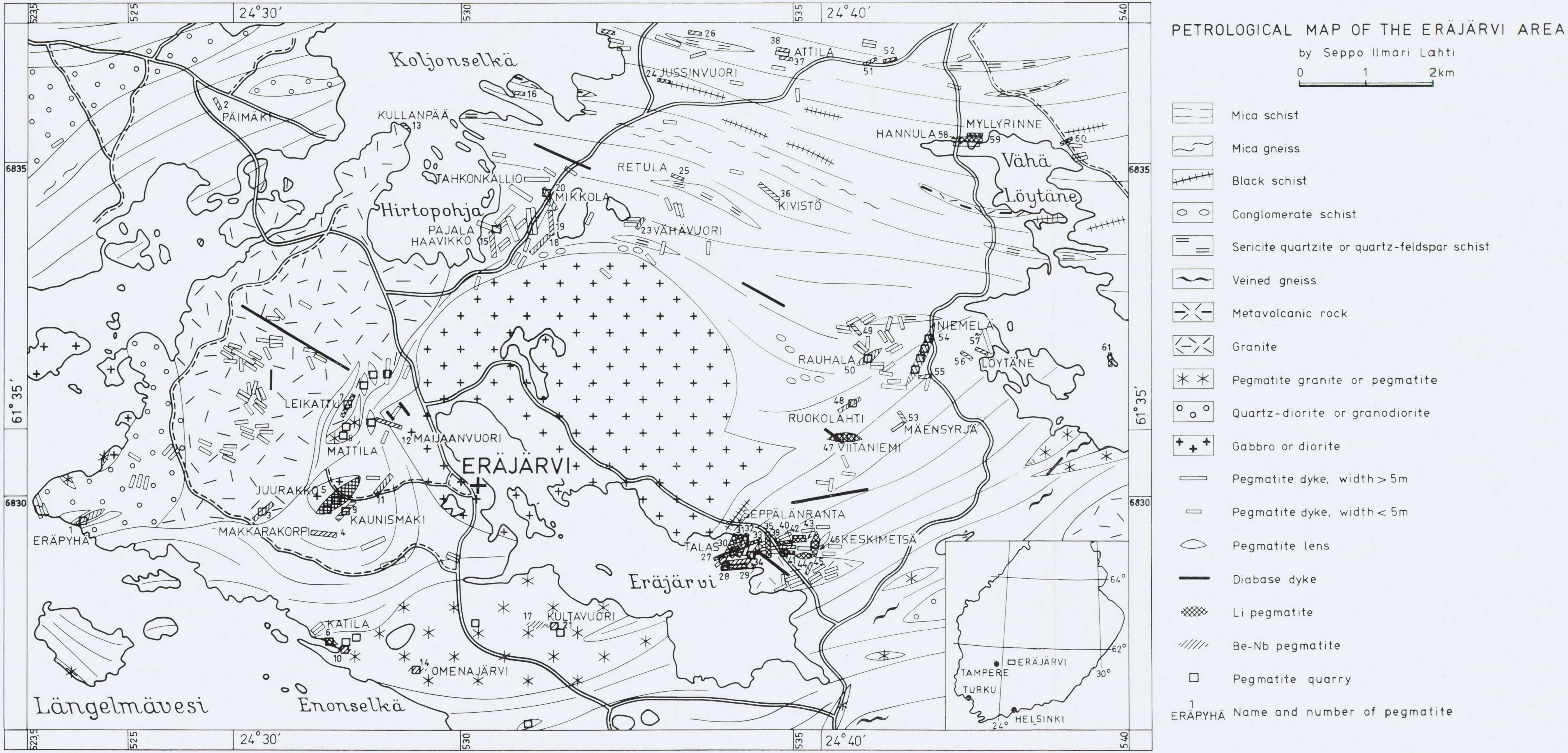
TECTONIC MAP OF THE ERÄJÄRVI AREA

by Seppo Ilmari Lahti



PETROLOGICAL MAP OF THE ERÄJÄRVI AREA

by Seppo Ilmari Lahti



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Postimyynti
PL 516
00101 HELSINKI 10
Puh. 90-539 011

Kirjakauppa
Annankatu 44
00100 HELSINKI 10
Puh. 90-17341

Denna publikation säljes av

STATENS TRYCKERICENTRAL
MARKNADSFÖRENINGSAVDELNINGEN

Postförsäljning
PB 516
00101 HELSINGFORS 10
Tel. 90-539 011

Bokhandel
Annegatan 44
00100 HELSINGFORS 10
Tel. 90-17341

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ISBN 951-690-137-9
ISSN 0367-522X