GEOLOGINEN TUTKIMUSLAITOS

BULLETIN
DE LA
COMMISSION GÉOLOGIQUE
DE FINLANDE

N:o 140

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

PENTTI ESKOLA

65-VUOTISJUHLAJULKAISU
PUBLICATION IN CELEBRATION
OF HIS 65th ANNIVERSARY

HELSINKI
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Helsinki 1947. Valtioneuvoston kirjapaino

Pentti Eskola

PENTTI ESKOLALLE

HÄNEN 65-VUOTISPÄIVÄNÄÄN
8. 1. 1948

OMISTAA TÄMÄN JULKAISUN

SUOMEN GEOLOGINEN SEURA

TO
PENTTI ESKOLA

ON HIS 65th ANNIVERSARY,
JANUARY 8, 1948

THE GEOLOGICAL SOCIETY OF FINLAND

DEDICATES THIS VOLUME

SUOMEN GEOLOGINEN SEURA

kiittää kaikkia niitä yksityisiä henkilöitä, laitoksia, seuroja, yhtiöitä ja muita yhteisöjä, jotka ovat osallistuneet tämän julkaisun avustamiseen.

Erityisesti haluaa Suomen Geologinen Seura lausua kiitollisuutensa Outokumpu Oy:n Säätiölle, tohtori Ernst Cloosille sekä muille amerikkalaisille geologeille, joiden taloudellista tukea vailla tämän julkaisun toimittaminen ei olisi ollut mahdollista.

THE GEOLOGICAL SOCIETY OF FINLAND

begs to thank all those persons, establishments, societies, associations, and other bodies that have contributed to this publication.

Especially the Society wishes to express its gratitude to the Foundation of the Outokumpu Co., to Dr. Ernst Cloos and to other North American geologists whose economic support has made the publication of this volume possible.

To

THE GEOLOGICAL SOCIETY OF FINLAND

It is my very great pleasure to extend to Professor Pentti Eskola the congratulations of the United States Geological Survey in recognition of his notable contributions to the science of geology. It is natural, that his friends all over the world, and those who have come to know him through his work, should join in a sincere tribute to his fruitful career in geological science.

Geologists are wont to meet from time to time, often in field conference, the better to understand and appreciate the work of a colleague. It is even more fitting that they should render in lasting form a token of their appreciation of the work of a noted co-worker.

The United States Geological Survey, like the Geological Society of Finland, is more than an institution; it is made up of men who have in common with each other and with their counterparts in all nations, an intense interest in discovering the secrets of nature and applying them to the betterment of mankind. Co-ordination into such units in no wise lessens the personal interest of scientists in each other's work.

Sincerely yours,

W E Wrather

Director

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SISÄLLYSLUETTELO

CONTENTS

1.

TECTONIC TRANSPORT AND FABRIC IN A MARYLAND GRANITE

BY

ERNST CLOOS

CONTENT

INTRODUCTION

The Ellicott City Granite, Maryland, intrudes an odd-shaped and mostly vertical contact zone between schist and gabbro-amphibolite in the center of a synclinal area. Its structures harmonize with those of the wall rocks and reflect their movements. Contacts are conformable, lit-par-lit injection of schist is the rule, but schist inclusions and occasional granite dikes establish the age of the pluton.

Lineation is shown to indicate a girdle or B-axis and converges toward the deepest portion of the syncline from all sides, posing a difficult problem of interpretation. The question arises in what way are these axes related to the direction of tectonic transport?

TECTONIC ENVIRONMENT

Four anticlinal domes of Baltimore gneiss surround a synclinal depression on all sides (see Broedel, 1937, and Figure 1).

The *Towson dome* to the northeast is almost complete except for its faulted west end. The *Chattolanee dome* to the north is an elongate asymmetrical anticline, steeper on the north than the south side. The *Baltimore dome* to the southeast is largely covered by coastal plain sediments but its western slope dips gently to the northwest under the gabbro complex. The *Woodstock dome* consists of three rather widely differing

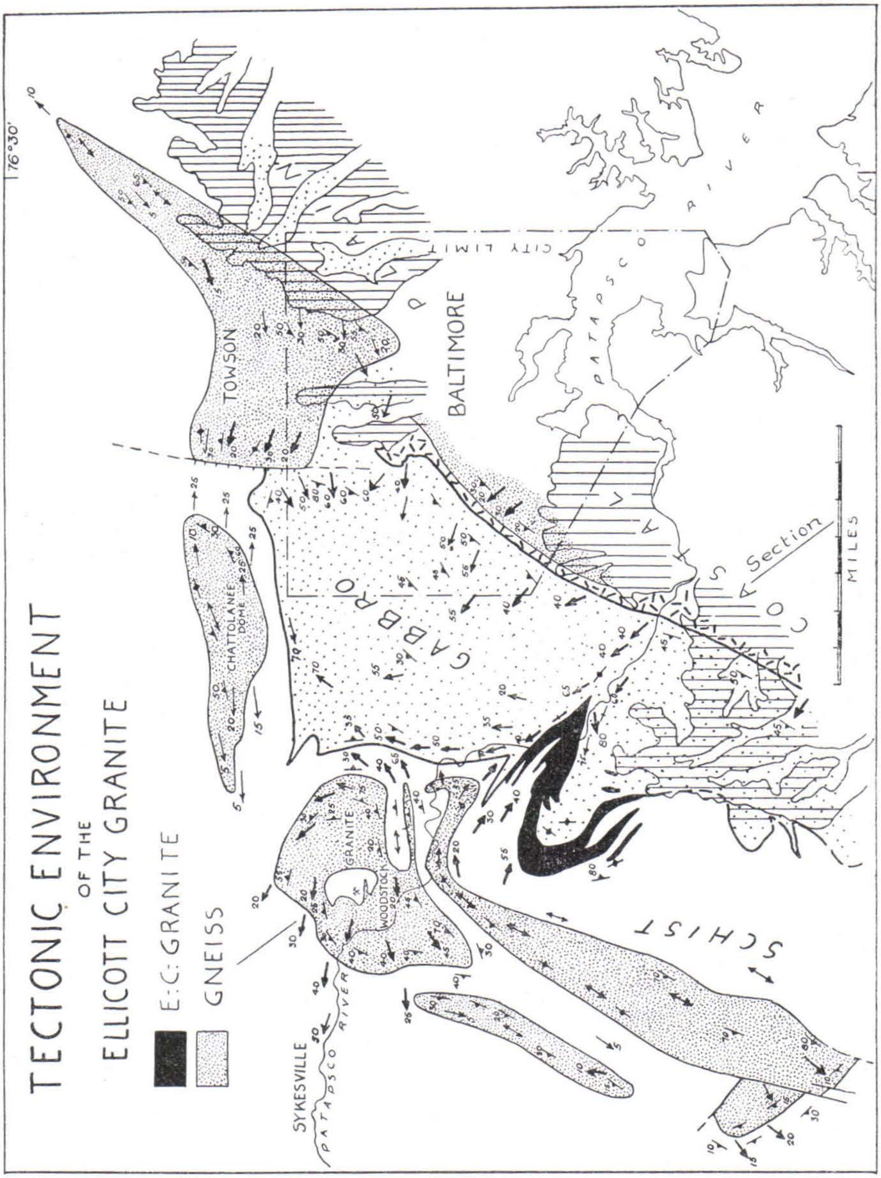

Fig. 1. Tectonic environment of the Ellicott City Granite. Arrows: lineation with dip.

units: the almost circular Woodstock area, a long protrusion to the south and a smaller anticline to the northwest which are overturned toward the syncline.

The western two areas and the northwest side of the Towson dome are overlain by the normal sequence of the Glenarm series (Setters quartzite, Cockeysville marble, and Wissahickon schist). Setters quartzite also rests on the southwest flank of the Towson dome, but the Glenarm sediments are lacking above the Baltimore gneiss in the Baltimore dome and the southeast flank of the Towson dome. A thin sill-like body of quartz diorite is between the gabbro complex and the gneiss along the southeast side of the syncline.

The largest portion of the synclinal area is occupied by the Baltimore gabbro complex which consists largely of amphibolites with some serpentine and ultrabasics. To the northwest the gabbro is in contact with schist (see Cohen, 1937).

A strong foliation dips into the basin from the northeast and southeast roughly following the gneiss dome outlines. In the center and along the western margin the foliation is steep or vertical.

Lineation within the foliation planes points into the syncline and is vertical in the vicinity of the granite.

The granite occupies a conspicuous location in the synclinal area in that the gabbro complex is narrowest here and almost dissected by a wedge of schist which points across the general strike to the southeast; the Baltimore gneiss also curves normal to the general strike; foliation in gabbro, schist, and gneiss trends northwest-southeast and is vertical; and finally, the granite forms the contact between gabbro and schist, following its outline. It is not a ring dike but a boundary pluton (Grenzpluton, Hans Cloos, 1928).

If the Woodstock anticlinorium were moved back to the northwest as far as it has been thrust southeastward, the granite would appear about in the center of a funnel, whose eastern half consists of the gabbro complex and whose western portion is schist.

CONTACTS

The contacts between granite and schist or gabbro are sharp. A zone of lit-par-lit injections, however, occurs between granite and schist, whereas the gabbro begins abruptly and only three granite dikes have been observed in the gabbro area. The granite affected the schist very intensely and it seems as if its intrusion occurred largely at the expense of the schist. Gabbro inclusions in granite are rare, schist inclusions are legion. The contact zone toward the schist is several hundred feet wide.

All observed contacts are vertical or dip steeply.

GRANITE STRUCTURES

Flow structures in the granite have been described and discussed in an earlier paper (Ernst Cloos, 1933) and are briefly summarized here (Figures 2 and 3).

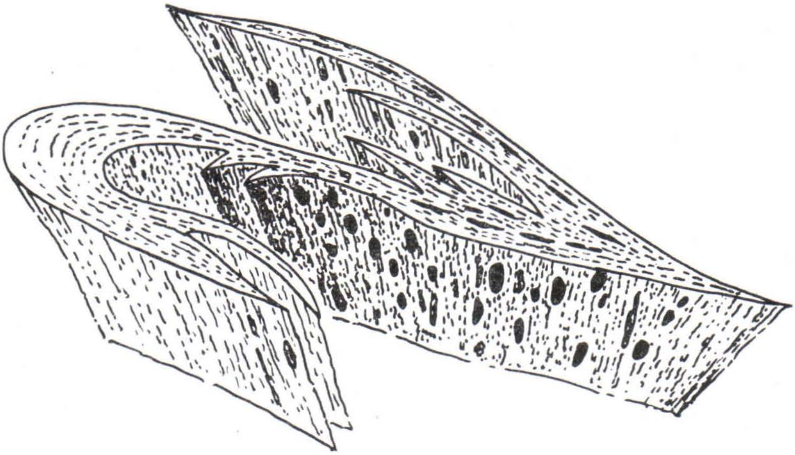

Fig. 2. Diagram of the Granite and its megascopic structures.

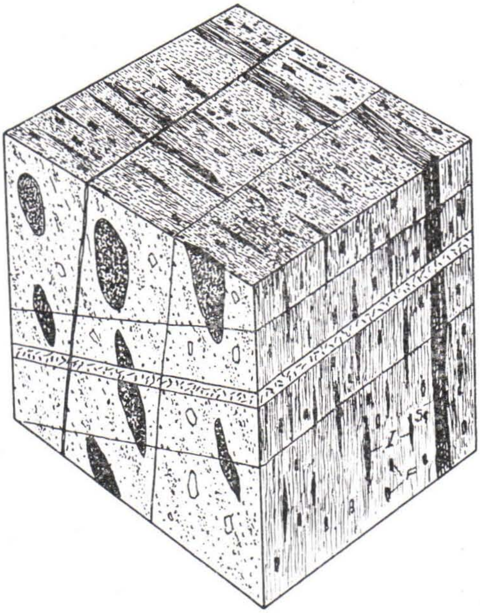

Fig. 3. Flow structures in center of granite showing inclusions, foliation, feldspar plates, schist inclusion, pegmatite dike, and joints.

An intense foliation is due to parallel arrangement of mica flakes, feldspar phenocrysts, pancake-like inclusions, and flat slabs of schist near the contacts.

Lineation is distinct in most of the exposures but not always. It becomes visible due to alignment of the longest axes of inclusions (Figure 3), chains of inclusions or biotite clusters.

Foliation is steep and parallels the contacts, lineation is in the foliation planes and vertical or pointing toward the center of the northwest-trending portion of the pluton. Lineation has not been seen in the rest of the pluton due to lack of good exposures. Figure 2 is an attempt to show these conditions schematically and simplified.

FABRIC

Additional information has been obtained by statistical analysis of mineral orientations of several specimen of which three are here described: Specimen 1 from the center of the granite pluton (Weber quarry at Ellicott City), Specimen 2 just inside the contact at the southeast end of the pluton, and Specimen 3 outside of the contact in a thin layer of quartzite at the same locality.

The granite has been described by Knopf and Jonas (1929, p. 134) in their report on Baltimore County:

> »The rock is a massive biotite quartz monzonite . . . It often shows a porphyritic texture and the phenocrysts of flesh-colored feldspar are sometimes one inch to one and one half inches in length. The microscopic texture is coarsely granitic with slight evidence of strain. The constituents are quartz microcline, microperthite, and plagioclase (a sodic oligoclase) with considerable myrmekite. Biotite is the most abundant dark colored mineral and shows fine pleochroic haloes around inclusions of zircon. Epidote occurs frequently in large crystals with allanite centers. Titanite in large crystals is an accessory constituent. Other accessories are zircon and apatite.»

In *specimen 1* large biotite blades are well aligned in a foliation which is parallel within and outside of the inclusions. Quartz is fractured or undulatory and some post-crystalline deformation with reference to quartz seems evident. There is however no mortar or granulation. Some large quartz grains contain feldspar inclusions. Epidote is common in broken grains.

The biotite diagram, fig. 4, shows 2 areas of maximum concentration of cleavage poles coinciding with the poles of the foliation (S_1). An incomplete girdle shows that biotite orientation fluctuates about an axis which coincides with the lineation (1).

Quartz orientation (Fig. 5) is less regular. One maximum is near the pole of the foliation, a group is in the vicinity of the lineation, and several are near the periphery. The lineation is not as well indicated as an axis in this diagram as in figures 7 and 11.

In *specimen 2* foliation is distinct and due to parallel light and dark bands which also parallel the contact. A strong lineation within the foliation plane is almost vertical and consists in long streaky mica clusters and elongate mica grains. The average biotite is two to three times as long parallel to the lineation than across it within the foliation plane. In cuts normal to the lineation the granite is almost massive but for few widely spaced folia.

Fig. 4. 125 biotite grains, 1—2—3—4—5—(6—7) percent. *l*: lineation. Granite, center of pluton. Lower half of projection net.

Fig. 5. 125 quartz grain 1,5—2—3—4—(5—7) percent. Loc. same as 4.

Microscopically the texture is even grained, rather fine. Biotite and muscovite are fresh, some of the biotite is slightly discolored and bent.

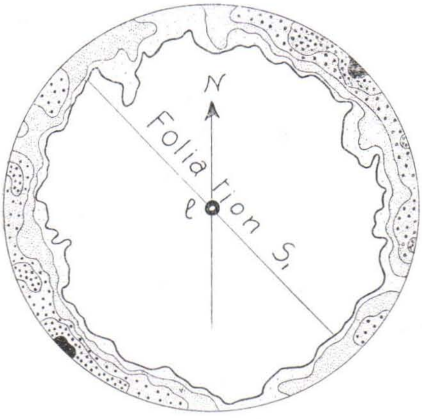

Fig. 6. 200 biotite grains, 1—2—4—6—9 percent. 30 cm inside granite at contact.

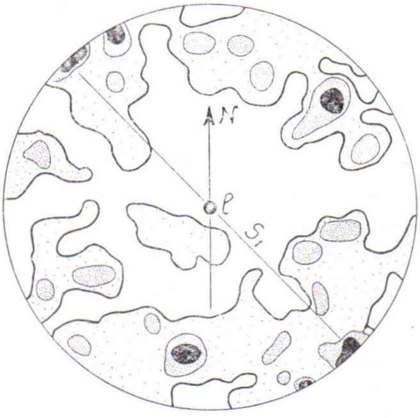

Fig. 7. 300 quartz grains, 1—2—3—3½ percent. Granite same loc. as 6.

Muscovite is entirely undeformed and grains are fewer but larger than biotite. Quartz shows weak undulatory extinction. There are no large quartz grains, granulation is lacking. The foliation is hardly detectable under the microscope.

The biotite diagram (Fig. 6) shows a perfect girdle with *l* as axis. The quartz diagram (Fig. 7) is less regular but also indicates a girdle with a maximum in the foliation plane and one almost normal to it. More important than the concentration of quartz axes in weak maxima is the lack of grains in the center of the diagram, thus indicating an axis. 50 grains of Sphene (Fig. 8) show a concentration of their longest dimension near the center of the diagram at *l*.

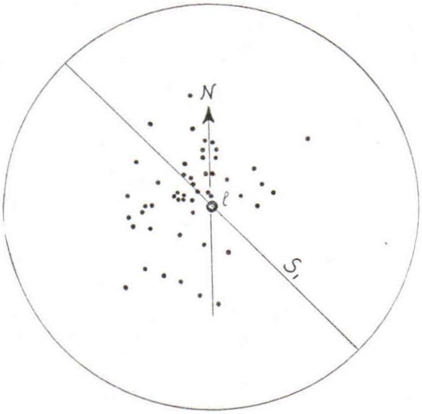

Fig. 8. Apatite, 50 grains, longest dimensions, same loc. as 6.

WALL ROCK FABRIC

Specimen 3 is a few inches outside the granite. Biotite occurs as small blades, brown, strongly pleocroic, grains about 1/10 of that of muscovite. Sections normal to foliation show fluctuation of orientation about *l*. Muscovite is in large fresh blades with many inclusions. Resorption seems considerable and droplet-shaped quartz in muscovite is common. »Worm-eaten» grains are abundant. Quartz is even grained mostly small with some undulatory extinction. Post-crystalline deformation of quartz is mild, all other minerals are undeformed.

The biotite diagram (Fig. 9) shows a maximum in the pole of the foliation and tendency toward an incomplete girdle. Muscovite (Fig. 10) also tends toward a girdle but its maximum is rotated counter clockwise 10—20 degrees in respect to the biotite and foliation maximum. Similar rotation has been described by the author (Cloos and Hietanen, 1941) and may indicate a rotation of fabric with respect to certain minerals representing stages of crystallization. The quartz diagram (Fig. 11) indicates an incomplete girdle normal to lineation *(l)* and a maximum in the position of maximum II (Sahama, 1936, p. 58).

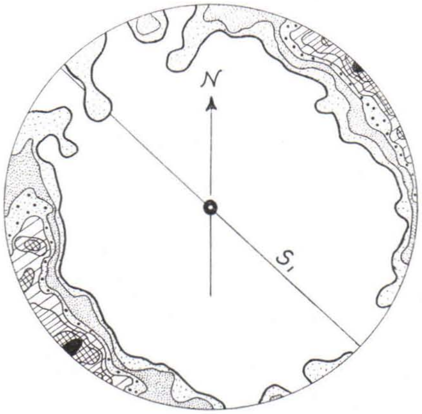

Fig. 9. 100 biotite grains, 1—2—4—6—8—10—12 percent. Quartzite, just outside contact opposite 6.

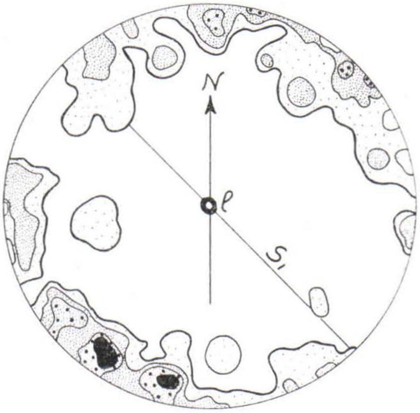

Fig. 10. 100 muscovite grains, 1—2—4—6 percent. Same loc. as 9.

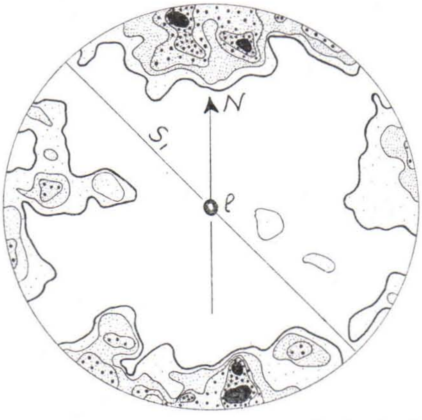

Fig. 11. 200 quartz grains, 1—2—3—4—5 percent. Same loc. as 9. All diagrams are in the horizontal plane with North indicated and projected into the lower hemisphere.

A very intense lineation in the adjacent amphibolites parallels that in the granite. Schistosity and lineation in the schist also parallels the granite structures. With increasing distance the lineation flattens out gradually until it is horizontal in the gneiss domes to the northwest and dips under the amphibolite in the southeast (Fig. 12). Beyond the Woodstock dome lineation dips to the west and is again vertical in the so called Sykesville Granite 3 miles to the west of the western border of the Woodstock dome.

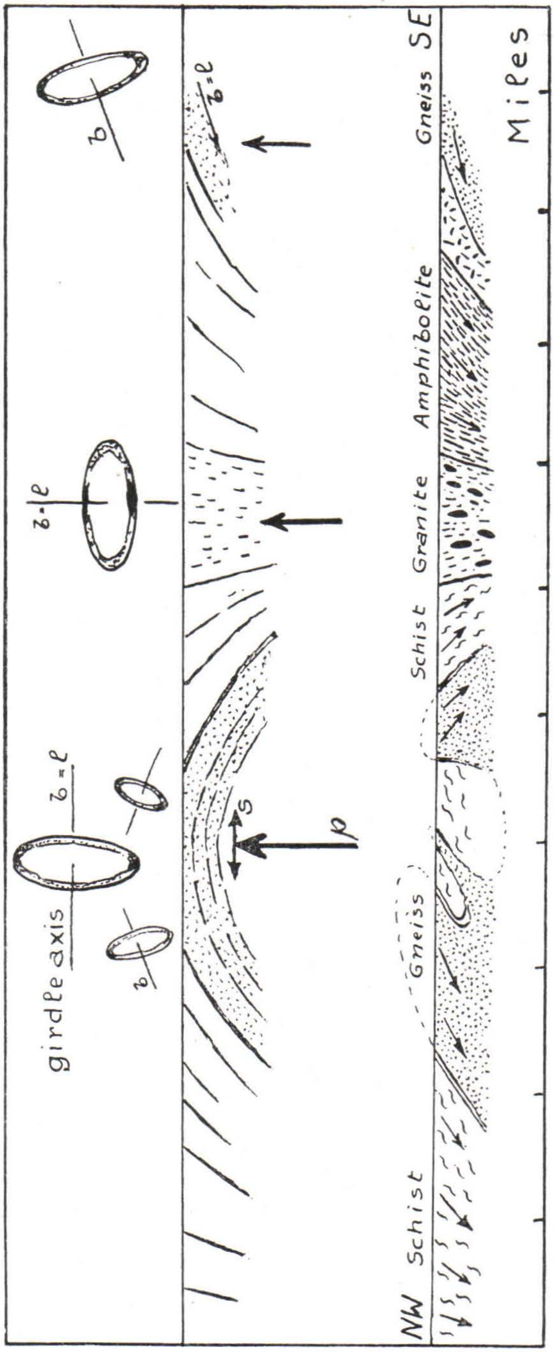

Fig. 12. A. Diagrammatic cross section showing girdles and axes. p: principal direction of movements equal tectonic transport; s: subordinate direction; b = 1 girdle axes.
B. Section through granite area and domes. Arrows indicate lineation. Woodstock granite pluton omitted.

Stefansson made 79 fabric diagrams of quartz, biotite, muscovite, and amphibole orientation in the Baltimore gneiss (1943). These show that the lineation shown as arrows in figure 1 is a girdle axis for mica and quartz. The amphibole maxima are in the center of the girdles.

The harmony of the fabric inside and outside of the granite is here interpreted as an additional indication of its common origin. The granite is syncinematic and registered movement traces common to the entire region.

INTERPRETATION OF DIAGRAMS

The mica diagrams confirm the megascopic foliation as accumulations of mica cleavage poles in the pole of the foliation. The lineation is shown in the center of the diagrams as a girdle axis. Since both planar and linear structures are flowage structures in the granite as indicated by the arrangement of inclusions, the mica orientation is thought to be also an alignment of flaky units in a magma stream (Formregelung). The apatite orientation also suggests orientation according to shape.

The quartz orientation fits into this very simple arrangement except the scattered maxima in figure 5. In both contact specimen, however, the quartz girdle is visible and increasingly prominent as the contact is reached. The girdle axis in the center co-incides with the mica girdle axis and the megascopic lineation.

The quartz maxima in figure 11 can be interpreted as maximum II (Sander, see: Sahama, 1936, Fig. 2).

The diagrams are identical in granite and wall rocks and confirm a simple major plan consisting of a lineation within a foliation and a girdle normal to the lineation. According to general usage (Sander 1930) the lineation is a b (B) axis and the diagram represents the ac plane of Sander's coordinates. The girdles are ac girdles $\perp$ b and $l = b = B$. Orientation in space varies with location and is independent of contacts.

DIRECTION OF TECTONIC TRANSPORT

»Aus einem einwandfreien an allen zugänglichen (nicht etwa grundsätzlich nur an Quarz) Mineralen gemessenem Gürtel $\perp$ B kann man jedenfalls summierbare Teilbewegungen $\perp$ B erschliessen (also einen B-Tektonit). Diese können, müssen aber keineswegs »tektonischem Transport» $\perp$ B entsprechen (Sander, 1936, p. 301)».

Girdles $\perp$ B indicate partial movement $\perp$ B, but this direction is not necessarily the direction of tectonic transport. The problem thus narrows down to the deduction of the direction of tectonic transport from girdle axes and lineations.

The same problem has lately been dealt with by several investigators who found that the tectonic transport as seen in the field coincides with the lineation *(l)* and that statistical investigation of the lineation produced a girdle $\perp$ *l*, thus tectonic transport normal to the girdle plane and parallel with its axis.

Some of the authors identify Sander's »Teilbewegungen» with tectonic transport and conclude that a second phase of movement resulted in the lineation and girdle arrangement. This interpretation seems somewhat forced in view of the obvious and widely recognized field evidence. The present author has lately discussed this problem and collected the literature on this subject (E. Cloos, 1946).

F. C. Phillips concluded that »there is throughout the Moine area evidence that movements along a line approximating to southwest and northeast have been associated with the general metamorphism» (1937, p. 595). As long as this movement is not meant to represent the tectonic transport it may be applicable to the field geology.

E. F. Osborn (1939) describes the Val Verde tonalite fabric and his case resembles the present one in many details: »Lineation in the tonalite parallels the dip of the gneiss planes, and flow structures in the tonalite parallel the schistosity in the wall rock near the contact.» Statistical analysis shows girdles $\perp$ to lineation and the author assumes this orientation to be the result of movement normal to lineation. In order to accommodate the girdles it is even suggested that maybe the flow structures in the tonalite be a metamorphic structure. The author is rather disturbed, however, by the arrangement of inclusions which render it difficult to date the entire fabric as secondary.

The assumption that tectonic transport must be perpendicular to girdle axes leads to the difficulty encountered by Osborn and Phillips. Sander's interpretation is very much less dogmatic and only demands partial movement (Partialbewegung) normal to the girdle axis which may but does not have to correspond with the direction of tectonic transport.

If field evidence points to tectonic transport parallel to a lineation which upon examination turns out to be a girdle axis it would seem that overemphasis were placed on the orientation of quartz grains and »Partialbewegung» versus geologic field evidence if the tectonic transport were interpreted on the basis of the still hypothetical mechanism of quartz orientation. To take refuge in a second act of deformation seems rather unjustified.

Anyone who has observed an oncoming train will have seen that the front of the engine moves in circles normal to the tracks and the beginner who tries to throw a shovel of coal into the fire door will miss by a large margin because of this motion. The train moves sideways and up and down whereas forward motion is barely noticeable unless the velocity changes. Wear and tear due to this corkscrew motion is considerable and, depending on the part of the undercarriage or wheels examined it would seem the only motion. To assume, however, that the train moves only sideways would of course be absurd.

The displacement of rock masses along the Scandinavian overthrust and the observable path of movement have been contrasted with fabric by Kvale (1945) and Strand (1945). These authors did not follow Phillips interpretation in the rather similar Scottish Highlands. Kvale recognizes the girdle as partial movements normal to the major tectonic transport. Strand comes to a similar conclusion and states that (p. 25) »The chief feature of the diagrams here presented is the presence of girdles in a direction normal to the chief direction of movement and, as interpreted by the writer the partial movement producing the girdles and the main transport were integral parts of the same act of deformation.»

CONCLUSION AND DISCUSSION

The direction of tectonic transport in the Ellicott City granite is thought to be in the direction of lineation and largely vertical because 1. the granite intruded the schist lit-par-lit along vertical planes, 2. the shape of the pluton necessitates intrusion in the direction of its vertical axis if discordant or transgressive intrusion of which there is no evidence is excluded, 3. inclusions are aligned in the direction of the axis at a time when the granite permitted such orientation, that is in a mobile state, 4. elongate apatite crystals are aligned in the flowage direction with their longest axes, 5. tabular feldspar phenocrysts are aligned in the foliation planes and 6. mica orientation also parallels foliation and lineation.

The only difficult point is the quartz orientation which cannot be ascribed to form-orientation. The author has attempted an interpretation (1946, p. 38) of quartz growth fabric in an earlier flow structure. If granite flow structures gradually emerge as the result of crystallization of mica and feldspar with inclusions participating, quartz will finally fill the interstices which are, however, not unoriented but represent an oriented mesh due to crystallization of the earlier components. If this mesh is mainly linear it may well be that growth of quartz axes may be influenced by this earlier fabric and a girdle may result. The maxima in figure 7 may thus be interpreted as quartz growing in the foliation direction and normal to it with other grains fluctuating in a girdle. (See also: Sander, 1930, p. 159, »Einflüsse anisotroper Aufwachsungsflächen oder Einflüsse von Gefügeanisotropien eines Starrgefüges in welchem der Kristall kristallisiert»).

The deformation plan is so similar in all rocks examined that a common explanation is called for.

The common factor in all diagram locations is movement in the direction of the girdle axis. Partial movements normal to these axes are probable but cannot be identified as tectonic transport in the granite.

As the domes grew upward during uplift their surfaces suffered stretching and elongation. In the elongated domes (Chattolanee) this distention coincided with the long axis of the dome. In the essentially circular Woodstock dome distention was radially outward, with due allowance for asymmetry to the south. In the Baltimore dome it was down the dip of the dome surface. The syncline between the domes was an area of relative subsidence and became restricted to a reduced area in which tectonic transport was upward (or downward) as the domes rose.

This mechanism is comparable to the rise of the granite dome of the Sierra Nevada in California where distention is evident in the center and movements parallel to the wall result in intense elongation down dip (E. Cloos, 1937). The author has suggested the use of the terms principal and subordinate directions of movement (E. Cloos, 1946). In the domes, girdle axes are normal to the principal direction of movement and parallel to the subordinate direction, on the flanks the girdle axes are parallel to the principal direction of movement.

All orientations are thus believed to have originated from one mechanism due to cylindrical flow (E. Cloos, 1946, p. 34) in the direction indicated by the lineation. Amphiboles in the gneiss and apatite in the granite show form orientation in the center and parallel with *l*. Micas also parallel *l* with emphasis of a foliation plane but forming a girdle. Quartz is the last crystallized mineral and may have been oriented mainly by the influence of an existing fabric (and growth) and some fracturing by partial movement at a late state.

The study reveals a master plan which comprises the entire area rather uniformly and is shown in figure 1 and reaches beyond its borders. The granite plays only a subordinate role conforming with the plan and not transgressive like a discordant youthful intrusion. It seems that the gneiss domes are the dominating units which may have grown like tumors whilst the synclinal area was sucked in by submergence.

THE JOHNS HOPKINS UNIVERSITY, BALTIMORE 18, MD.

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2.

MINERAL ASSEMBLAGES WITH SAPPHIRINE AND KORNERUPINE

BY

THOROLF VOGT

(TRONDHEIM, NORWAY)

The purpose of the present paper is first to point out some features of the mineral facies relations of sapphirine. As indicated by A. Lacroix (1929), this mineral occurs in metamorphic sedimentary rocks and in eruptive rocks rich in magnesia and alumina, and poor in silica; lime may be present or not.

THE SYSTEM SiO_2 — Al_2O_3 — MgO.

In the first instance, only considering the system SiO_2 - Al_2O_3 - MgO (with insignificant quantities of FeO and Na_2O), the mineral assemblages eventuate in the diagram presented as fig. 1. For the sake of comparison, the SiO_2 - Al_2O_3 - MgO diagram of the hornfels or gabbro facies of Pentti Eskola (1920, 1939) is added, according to C. E. Tilley (1925), see fig. 2. Some remarks on the single phases in the diagram fig. 1 will first be made.

The sapphirine is interpreted, in the usual way, as a subsilicate of magnesia and alumina. The low content of boron, 0.16 and 0.75 pct. B_2O_3 in sapphirine respectively from Sakeny, Madagascar (Lacroix 1940) and Itrongay, Madagascar (Lacroix and Gramont 1919) is not considered constitutional. Chemical tests on sapphirine from Fiskerneset, Greenland, performed by H. Bergh at the request of the author, gave nil B_2O_3 (with chinalizarine in strongly acid solution), as also nil F (method of Kühnel Hagen), and BeO (chinalizarine in alkalic solution). X-ray spectrograms taken with rock salt crystal at our institution in Trondheim, displayed only a small amount of calcium.

As emphasized by Gossner and Mussgnug (1928), the old formula for sapphirine of Joh. Lorenzen (1884), $Mg_5\ Al_{12}\ Si_2\ O_{27}$, is not conformable to the X-ray data. These authors propose the formula $Mg_2\ Al_4\ SiO_{10}$, with $Z = 8$, the departure from this composition being accounted for by the common substitution $Mg\ Si = Al_2$. A review of the analyses of sapphirine performed in the eighties and later confirm the latter formula. The six analyses (from Itrongay, Vizagapatam, Sakeny, St. Urbain, two from Fiskerneset, details see below), the points of which appeared close together when plotted in the SiO_2 - Al_2O_3 - MgO diagram, gave the average composition

$$Mg_{14\cdot86}\ Al_{34\cdot79}\ Si_{6\cdot48}\ O_{80}.$$

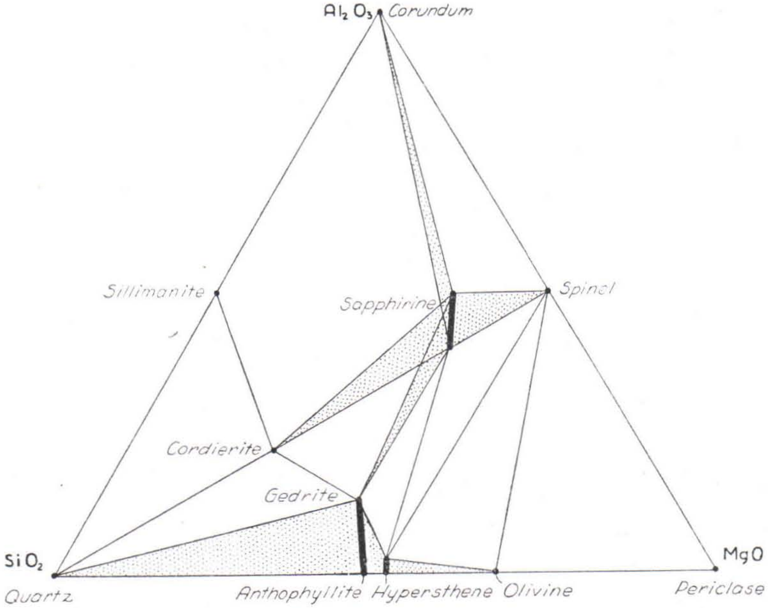

Fig. 1. SiO_2 - Al_2O_3 - MgO diagram of sapphirine bearing mineral facies.

A substitution of 1.35 Al_2 = MgSi gives closely the formula $Mg_{16}\ Al_{32}\ Si_8\ O_{80}$ or $Mg_2\ Al_4\ Si\ O_{10}$. A substitution in the opposite direction of 2.65 MgSi = Al_2, gives closely $Mg_{12}\ Al_{40}\ Si_4\ O_{80}$ or $Mg_3\ Al_{10}\ Si\ O_{20}$. Seven of the analyses dealt with appear more near to the former formula, the eigth is nearer to the latter. The general cell formula for the sapphirine may be expressed:

$$Mg_{16-n}\ Al_{32+2n}\ Si_{8-n}\ O_{80}.$$

The analyses give the following values for n:

Itrongay, Madagascar (Lacroix and Gramont 1919)	0.68[1]
Vizagapatam, Madras, India (Walker and Collins 1907)	0.94
Sakeny, Madagascar (Lacroix 1940)	1.26[1]
Dangin, West Australia (Prider 1945)	1.48
St. Urbain, Quebec, Canada (Warren 1912)	1.61
Fiskerneset, Greenland (Lorenzen 1884)	1.74[1]
» » (Ussing 1889 a, b)	1.88
Blinkwater, Transvaal (Mountain 1939)	2.23[1]

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[1] A small amount of Fe_2O_3 is reckoned as FeO.

It seems most natural to accept the general formula for sapphirine mentioned above, with n varying from nil to about 2.5. These values are used in the diagram fig. 1.

The gedrite from Fiskerneset, Greenland, is extremely rich in alumina, as it appears from the analysis of N. V. Ussing (1889 a, b), corresponding approximately to the composition Na $Mg_{10\cdot5}$ Al_7 $Si_{12\cdot5}$ O_{45} $(OH)_3$, which is used in the diagram. The analysis of Ussing is obviously very good for its time, the rock being also coarse and easy to separate in its pure components. For many years this was the most aluminous gedrite known. The gedrite from Masons Mt., North Carolina, examined by E. P. Henderson (1931), is, however, somewhat more aluminous, corresponding approximately to the same composition.

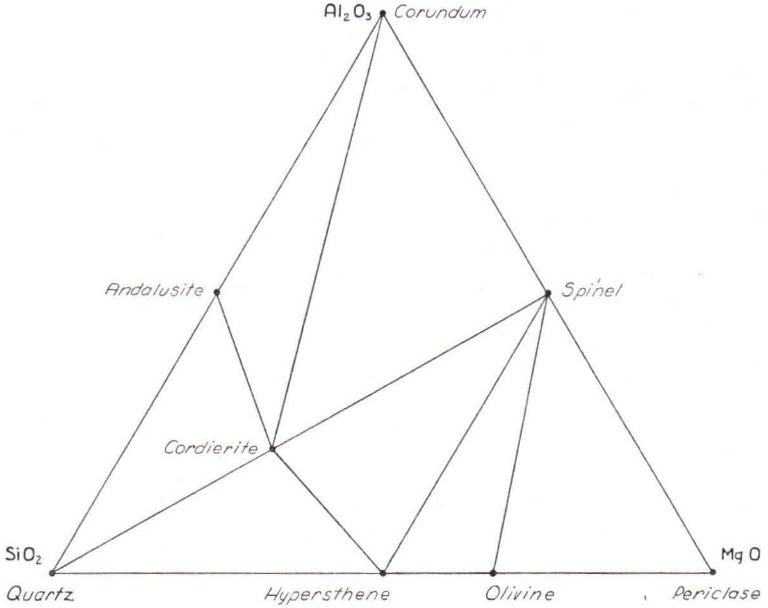

Fig. 2. SiO_2 - Al_2O_3 - MgO diagram of gabbro facies, according to C. E. Tilley (1925).

The analysis of Joh. Lorenzen (1884) of hypersthene from Fiskerneset, Greenland, corresponds approximately to the composition $Mg_{15\cdot5}$ Al $Si_{15\cdot5}$ O_{48}, which is used in the diagram fig. 1.

The following mineral assemblages with sapphirine in the SiO_2 - Al_2O_3 - MgO system have been described. From Fiskerneset, Greenland, N. V. Ussing (1889 a, b) furnished a detailed report of the associations sapphirine - gedrite (- pargasite) and sapphirine - bronzite - pleonast, representing his type I and type II of paragenesis. Further he mentions the association sapphirine - cordierite (- kornerupine), and O. B. Böggild (1905) the association gedrite-cordierite. The association of the two rhombic minerals gedrite and hypersthene surely should be searched for at

Fiskerneset. The rocks from Fiskerneset are rich in magnesia and poor in iron. According to the analyses of Lorenzen and Ussing, the minerals from rocks of type I contain the following percentages of FeO-components: sapphirine 4.0 and 4.5, gedrite 5.8, and pargasite 6.2. Of minerals from rocks of type II, the bronzite contains 8.6 pct. ferrosilite and the pleonast 20.7 pct. hercynite. A considerable enrichment of FeO has taken place in the pleonast.

From the hill tracts of Vizagapatam, Madras, India, C. S. Middlemiss (1904) and T. L. Walker and W. H. Collins (1907) described the association sapphirine-hypersthene- a spinel mineral, corresponding to type II from Fiskerneset. This rock is rather rich in iron, the sapphirine containing 20.0 pct. FeO-component and the spinel mineral being a pleonast - hercynite or a hercynite. At the same locality the khondalite, a quartz-garnet-sillimanite schist, occurs. The sapphirine and the sillimanite do not seem, however, to appear together. The author has only seen abstracts of the papers mentioned.

From St. Urbain, Quebec, Ch. H. Warren (1912) described an ilmenite ore, urbainite, i. a. with sapphirine-spinel. The sapphirine from this locality contains 25.0 pct. FeO-component, being the sapphirine richest in iron known.

From Sakeny, Madagascar, A. Lacroix (1929, 1939, 1940) and H. Besairie (1933) described an interesting rock series, the sakenites, with the associations sapphirine-spinel, sapphirine-corundum and spinel-corundum. Sillimanite gneiss with garnet occurs in the same locality; Besairie also mentions the association sillimanite-cordierite. The sakenites are poor in iron, the sapphirine containing 8.5 pct. FeO-component.

From Blinkwater, Transvaal, E. O. Mountain (1939) described sapphirine with 14.5 pct. FeO-component, and the association sapphirine-corundum.

The sapphirine from Dangin, West Australia, described by R. T. Prider (1945) does not occur with any mineral in the system SiO_2 - Al_2O_3 - MgO; the mineral contains 19.7 pct. FeO-component.

The crystals of sapphirine from Itrongay, Madagascar, described by A. Lacroix (1912, 1929) and Lacroix and Gramont (1919), are found detached from other minerals in soil. The source was first supposed to be a pegmatite, but after the discovery of the sakenites, the origin is considered to perhaps be a paragneiss; the low content of FeO-component, 5.1 pct. in the sapphirine, emphasizes the similarity with sapphirine from the sakenites, which are paragneisses.

The question of the stability in the present mineral facies either of cordierite - corundum or of sapphirine - sillimanite must be left undecided. The sakenites from Madagascar and the rocks from Madras may be favorable for settling this point of uncertainty.

PART OF THE SYSTEM $SiO_2 — Al_2O_3 — MgO — B_2O_3$.

Modern analyses of kornerupine display a rather constant content of boron, first discovered by A. Lacroix and A. de Gramont (1919) in kornerupine from Itrongay, Madagascar, (3.59 pct. B_2O_3). Kornerupine from the new occurrence at Port Shepstone, Natal, contains 3.5 pct. B_2O_3 according to J. E. de Villiers (1940), and kornerupine from the new occurrence in gem gravel in Ceylon contains 2.5, 2.8 and 3.6 pct. B_2O_3, according to M. H. Hey, B. W. Anderson and C. J. Payne (1941). The latter authors also found 3.0 pct. B_2O_3 in kornerupine (»prismatine») from the old occurrence at Waldheim, Saxony. Lacroix and Gramont (1919) further established a boron content in kornerupine from the occurrence first known, at Fiskerneset, Greenland. It seems very probable that boron enters into the mineral in definite proportions as especially emphasized by de Villiers (1940). The kornerupine may, therefore, be regarded as a boron mineral.

The old and simple formula for kornerupine, $Mg\,Al_2\,SiO_6$, was proposed by Joh. Lorenzen (1884) a long time before the boron content in the mineral was known. The question of the chemical composition of kornerupine on the new basis was discussed closely by M. H. Hey, B. W. Anderson and C. J. Payne (1941), who found the oxygen content of the empirical unite cell lying between 82.9 and 87.5. These authors elected the value 86, and proposed the formula

$$[(Al, Fe)_{20+x+y+z}\ (Mg, Fe)_{20-x-2y-z}\ Na_y]\ \ [Si_{18-x-z}\ Al_z\ B_x]\ O_{86}.$$

The author has tried calculations with different possible values for oxygen content in the unite cell, and found the best agreement to the data of the analyses with O = 88, which also gives a simple formula. The four analyses (Ceylon I and III, Waldheim and Itrongay) the points of which appeared close together when plotted in the SiO_2-Al_2O_3-MgO diagram, gave the average composition:

$$Na_{0.86}\ Mg_{17.83}\ Al_{23.32}\ B_{2.60}\ Si_{15.43}\ O_{88}.$$

After the substitution$_{0.86}$ NaSi = MgAl we get

$$Mg_{18.69}\ Al_{24.18}\ B_{2.60}\ Si_{14.57}\ O_{88}.$$

The substitution 1.38 Al_2 = MgSi gives

$$Mg_{20.07}\ Al_{21.42}\ B_{2.60}\ Si_{15.95}\ O_{88}.$$

or very near to $Mg_{20}\ (Al, B)_{24}\ Si_{16}\ O_{88}$.

The boron content, corresponding to this formula, from the analyses of kornerupine from Ceylon I, II and III, Waldheim, Itrongay and Natal are 2.45, 3.11, 2.21, 2.66, 3.09 and 3.18 B. Presuming B = 4 and Al = 20, we get $Mg_5\ Al_5\ B\ Si_4\ O_{22}$. The general cell formula may be written

$$Mg_{20-n}\ Al_{20+2n}\ B_4\ Si_{16-n}\ O_{88}.$$

Different values for *n* are given below, the sum $(Al, B)_{24+2n}$ being used:

Itrongay, Madagascar (Lacroix and Gramont 1919[1])	1.30
Ceylon I (Hey, Anderson and Payne 1941)	1.30
Ceylon III (» » » » »)	1.41
Waldheim (» » » » »)	1.50

The value n = 0—2 is used in the diagram fig. 3. On this diagram are also plotted the composition of dumortierite, according to the formula

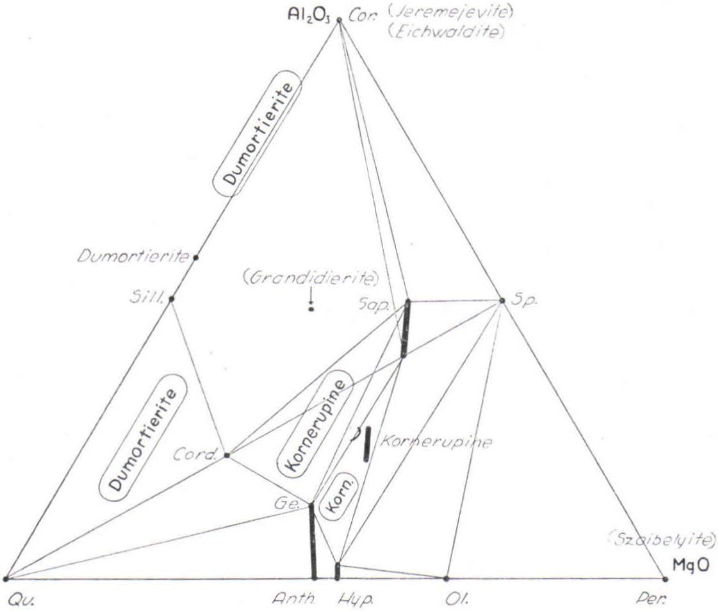

Fig. 3. SiO_2 - Al_2O_3 - MgO diagram of sapphirine bearing mineral facies with boron minerals. Names of boron minerals in their areas of equilibrium are encircled by a closed line.

of Schaller, and the composition of grandidierite, according to the analysis of A. Lacroix (1922). The minerals with the names in brackets are not known to occur in the present mineral facies.

The minerals sapphirine and kornerupine are attached for mineralogists through their mutual occurrence at Fiskerneset, Greenland. According to N. V. Ussing (1889 a, b) the minerals kornerupine-sapphirine-cordierite appear together at this locality, associated with a basic plagioclase and biotite. Whether gedrite or pargasite are found in this association is not quite clear. In a specimen with kornerupine from this locality belonging to the Mineralogical Museum, Oslo, and examined by the author, no amphibole minerals have been found.

The occurrence of kornerupine (»prismatine») from Waldheim, Saxony, is described by A. Sauer (1886), Ernst Kalkowsky (1907) and J. Uhling

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[1] Correction for Fe_2O_3 and FeO by Hey, Anderson and Payne (1941).

(1910). The kornerupine bearing rock appears within an area of sillimanite granulite. A specimen from this locality examined by the author contained albite, garnet, kornerupine, tourmaline, rutile and zircon; formed at a later stage are minerals in reaction rims around the garnet, i. a. quartz, further traces of biotite and muscovite, and a hydrous mineral »kryptotile» with small individuals of dumortierite, derived from the kornerupine. This is largely in accordance with the statements of the authors mentioned, who also report corundum, sillimanite and other minerals associated with the kornerupine. The kornerupine granulite from Waldheim belongs, as is well known, to the granulite facies of Pentti Eskola.

The kornerupine from Itrongay, Madagascar, is, according to A. Lacroix (1912, 1929), found in soil detached from other minerals. At the same locality sapphirine occurs, and the two minerals may presumably have been associated in the rock. According to J. E. de Villiers (1940), the kornerupine at Port Shepstone, Natal, is associated with quartz, tourmaline, biotite, garnet, and grandidierite enclosed in the kornerupine. Finally, the kornerupine latest reported by M. H. Hey, B. W. Anderson and C. J. Payne (1941) from Ceylon, occurs in gravel detached from other minerals except mica and zircon, which appear as inclusions.

The diagram fig. 3 represents the system SiO_2-Al_2O_3-MgO belonging to the sapphirine-gedrite-sillimanite bearing mineral facies, with addition of so small amounts of B_2O_3 that none of the existing phases disappears. This introduced boron will react with the existing minerals, and small amounts of a single boron mineral, in equilibrium with the other minerals, will be produced. The names of the minerals, which may be formed in this way, are encircled by a closed line and placed in their areas of equilibrium.

Applying the formulas $Mg_2\,Al_4\,Si\,O_{10}$ for sapphirine, $Mg_5\,Al_5\,B\,Si_4\,O_{22}$ for kornerupine, and $Na\,Mg_{10}\,Al_8\,Si_{12}\,O_{45}\,(OH)_3$ for gedrite, one may present the following reaction:

$$4 \text{ Sapphirine} + 2 \text{ Gedrite} + 2 \text{ Hypersthene} + 3\, B_2O_3 =$$
$$= 6 \text{ Kornerupine} + 2 \text{ Albite} + 3\, H_2O.$$

The kornerupine, being formed by addition of boron to the association sapphirine-gedrite-hypersthene, is supposed to exist in equilibrium with this assemblage.

The composition of sapphirine as well as of kornerupine being changeable, the latter mineral may be produced merely at the expense of sapphirine, gedrite and boron. As an example may serve the following reaction, where the cell formulas for sapphirine (n = 2) and kornerupine (n = 0.842) are used:

$$6 \text{ Sapphirine} + 28 \text{ Gedrite} + 38\, B_2O_3 =$$
$$= 19 \text{ Kornerupine} + 28 \text{ Albite} + 42\, H_2O.$$

Kornerupine may also be formed in the association sapphirine-gedrite-cordierite, as displayed by for instance the following reaction. Here is used the formula $Mg_{12}\,Al_4\,Si\,O_{10}$ for sapphirine, and the cell formula (n = 2) for kornerupine:

$$211 \text{ Sapphirine} + 48 \text{ Gedrite} + 17 \text{ Cordierite} + 104\ B_2O_3 =$$
$$= 52 \text{ Kornerupine} + 48 \text{ Albite} + 72\ H_2O.$$

The stability of kornerupine in the assemblage sapphirine-gedrite-cordierite is, therefore, probable. This is corroborated through the association kornerupine-sapphirine-cordierite found at Fiskerneset, Greenland. The stability of the kornerupine is surely not restricted to the assemblages mentioned above, as will be seen from the associations observed. It may be mentioned that the mineral evidently may be stable with quartz, provided that a certain chemical composition of the surroundings exists.

The dumortierite is placed in the area quartz-sillimanite-cordierite, and on the line sillimanite-corundum, according to several statements in the literature. Dumortierite is not stable in the presence of both cordierite and plagioclase.

Associations stable in the present mineral facies may then be:

1. kornerupine-sapphirine-gedrite-cordierite
2. kornerupine-sapphirine-gedrite-hypersthene
3. dumortierite-quartz-cordierite-sillimanite
4. dumortierite-sillimanite-corundum.

CONCLUDING REMARKS

When comparing the present facies with the gabbro facies of Eskola, see figs. 1 and 2, the following reaction may be advanced:

$$\text{Cordierite} + 8 \text{ Spinel} = 5 \text{ Sapphirine}.$$

The formation of sapphirine according to this reaction is accompanied by a contraction, viz. from 100 to about 92 pct. in volume. The transformation of the gabbro facies to the present facies is, therefore, favored by pressure.

It may be clear that the present metamorphic facies is closely related to the amphibolite facies of Pentti Eskola (1914, 1915, 1920, 1939). Hornblende is a characteristic mineral both at Fiskerneset, Greenland, and at Sakeny, Madagascar. Further, the associations anorthite-cordierite and hornblende-gedrite occur at Fiskerneset, and anorthite-aluminous monoclinic pyroxene at Sakeny. The difference known between the amfibolite facies and the present facies, in the case of rocks saturated with silica, is the occurrence of gedrite, augite and sillimanite in the latter, instead of anthophyllite, diopside and andalusite in the former. To this may be added that sapphirine occurs in undersaturated rocks in the present facies; this mineral is, at least as yet, not known from the amphibolite facies.

It may be noted that more water is bound in a gedrite rock than in an anthophyllite-cordierite rock of the same chemical composition. The formation of gedrite may, therefore, be favored by pressure. The transformation andalusite to sillimanite is supposed to be favored by increasing temperature and pressure. The formation of aluminous augite from diopside may be favored by rising temperature. It is consequently presumed that the present metamorphic facies belongs to somewhat higher pressure and also higher temperature than the amphibolite facies.

The name sakenite facies would suit well in several respects, and may perhaps be taken into consideration. The name of a metamorphic facies should, however, preferentially be taken from rocks saturated with silica. Here it is difficult to find a short name, sillimanite-gedrite-gneiss facies could, perhaps, be proposed. It may be discussed if this facies should be classed as a subfacies under the amphibolite facies, or as an independent facies.

The sapphirine bearing rocks are accompanied, at Madras and Sakeny, Madagascar, by sillimanite-garnet rocks. The knowledge of the chemical composition of these garnets would be of great importance for the question of this facies. Rocks belonging to the facies discussed above probably occur at Sørlandet and at Modum in Norway.

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3.

A FRONT OF METASOMATIC METAMORPHISM IN THE DALRADIAN OF CO. DONEGAL

BY

ARTHUR HOLMES AND DORIS L. REYNOLDS
GRANT INSTITUTE OF GEOLOGY, UNIVERSITY OF EDINBURGH

CONTENTS

I. INTRODUCTION

The Donegal *Memoirs* of the old Geological Survey of Ireland, published well over half a century ago, provide a wealth of invaluable clues for those in search of field evidence to illustrate critical stages in the operation of petrogenetic processes. In 1935 the *Inishowen Memoir* stimulated us to spend a fortnight in the neighbourhood of Malin Head, the most northerly part of Ireland. A few brief but dramatic sentences had led us to anticipate that we should find there well exposed examples of the effects of metamorphic diffusion, metasomatism, granitization and rheomorphism. The exciting discoveries actually made far exceeded our expectations, in consequence of which we paid another visit to this fascinating area in 1937.

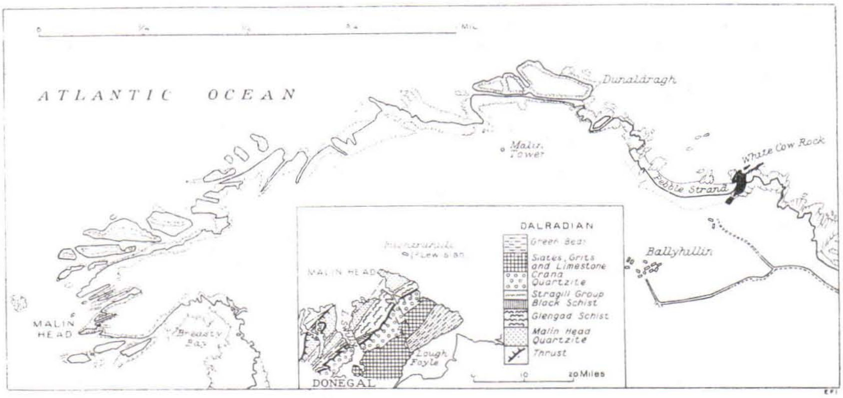

Fig. 1. Map of the Malin Head district, Inishowen, Co. Donegal, Eire, showing the locality of the White Cow Rock. The outcrop of the main epidiorite sill is indicated in black. *Inset:* Index map of the Dalradian rocks of Northern Donegal (after McCallien, 1937).

One sentence that aroused our interest reads as follows: »At Pebble Strand, a large dyke occurs enclosing fragments of gneiss, and so highly altered as to pass in places into hornblende schist». (Nolan, 1890, p. 29). The so-called »dyke» is really a thick sill of epidiorite which makes a conspicuous feature known locally as the White Cow Rock (Fig. 1), because of the occurrence within the sill of a large mass of white vein-quartz (enclosing an »eye» of amphibole-biotite-skarn) with an outline that simulates the form of a cow's head, as seen from certain view-points (Fig. 2). Each of the »fragments of gneiss» was found to be a thin sheet of highly metasomatized quartzite, sheared off by the intrusion from the floor-rock, into the undisturbed part of which it visibly passes without interruption — like a shaving curled through the stock of a smoothing plane, but still attached to the surface of the wood where the blade has come to rest. The progressive transformation of these quartzite sheets into rock-types that converge towards the composition of the enclosing epidiorite appears to have occurred during regional and metasomatic metamorphism that brought about various changes in the original basic rock of the sill. This striking example of metamorphic diffusion will form the subject of a later paper. The present communication, however, is concerned with the effects of geochemical migrations that have affected the epidiorite itself and the country-rock quartzite in its vicinity.

Fig. 2. The White Cow Rock (looking north at low tide), eastern end of Pebble Strand, Malin Head district, Co. Donegal. The White Cow Rock, part of the nearest of the small islets, and the elevated rock in the middle foreground form part of the main epidiorite sill. For the upper and lower contacts compare with Fig. 3.

The sill of the White Cow is accompanied by a narrow underlying sill, from which it is separated on the south-west by a foot and a half of quartzite. Elsewhere, the separation increases owing to transgression by the main sill. Parts of the intervening quartzite and considerable volumes of the underlying quartzite have been transformed into mica-schist, while the margins of the main sill and locally the whole of the thin lower sill have been even more conspicuously enriched in biotite. The evidence of these and related mineralogical and chemical changes — as revealed in the field, under the microscope and by chemical analysis — is so clear and unequivocal that we have selected this study as a particularly appropriate one for the present purpose. It is a pleasure and a privilege to dedicate it with respectful homage to our friend Professor Eskola.

II. GEOLOGICAL SETTING

The peninsula of Inishowen lies between Lough Foyle and Lough Swilly and is almost entirely built of metamorphosed sediments belonging to the Dalradian Series. The strata have been invaded by a swarm of basic sills and dykes which shared in the regional metamorphism and are now in the condition of metabasites, variously described as epidiorites, amphibolites and hornblende-schists. A few irregular masses of a reddish post-tectonic granite also occur, notably in the country immediately to the south-west of the Malin Head area depicted in Fig. 1 (Nolan, 1890).

The oldest sub-division of the Dalradian Series exposed in Inishowen is the Malin Head Quartzite, which forms not only the bold and precipitous cliffs of Malin Head itself, but also the greater part of the Malin promontory at the extreme north of Inishowen. The Dalradian sequence and the structure of the region (and of part of the country to the west of Lough Swilly) have recently been investigated by McCallien (1935 and 1937). The small-scale inset map of Fig. 1 is copied from a map given in the second of these publications by McCallien. It serves to indicate the general sequence and the regional strike of the Dalradian Series.

Formations older than the Malin Head Quartzite outcrop to the west of Lough Swilly and are indicated on the inset map of Fig. 1 by absence of shading. Here the Malin Head Quartzite overlies the Fanad Boulder Bed or Tillite, which in turn overlies the Fanad Limestone. These three formations can be correlated respectively with the Islay Quartzite, the Portaskaig Boulder Bed or Tillite and the Islay Limestone (McCallien, 1935, p. 437). Peach recognised that the upper strata of the Islay Quartzite have certain features *(e. g.* the »Pipe Rock») in common with the Lower Cambrian succession of Durness in the North-West Highlands of Scotland (Peach and Horne, 1930, p. 210) and he therefore regarded the Islay Quartzite as being probably of Lower Cambrian age. This correlation has recently been strongly supported by Pringle (1947) who has drawn attention to the fact that the Portaskaig Tillite »presents the same features that characterise the corresponding deposits in Norway, Finnmark, Spitzbergen and Greenland, namely the occurrence of two distinctive tillites separated by interglacial sediments.» Moreover, the Islay Limestone includes beds of dark oolitic limestone, characterised by an abundance of *Osagia* and *Stromatolites*, algal forms identical with those found in the limestone that underlies the tillites in the northern lands mentioned above. In these regions the glaciation occurred just before the first appearance of typical Cambrian faunas. It is therefore reasonable to infer that the age of the Malin Head Quartzite, like that of its equivalent, the Islay Quartzite, ranges from the extreme end of the Pre-Cambrian into the Lower Cambrian.

The Malin Head Quartzite is largely made up of massive to flaggy quartzites with local partings and occasional thicker intercalations of mica-schist which is commonly garnetiferous. Near Dunaldragh (Fig. 1), where schists are particularly strongly developed, they have been conspicuously enriched in albite. The quartzites are generally white, cream or grey in colour, and of fine grain. Albite is the dominant feldspar, and biotite, muscovite, garnet and calcite are common accessories, together with black ores that are locally concentrated along bedding planes. Mottled pink feldspathic varieties of coarser grain also occur. Some of these have structures suggesting that they were originally feldspathic grits

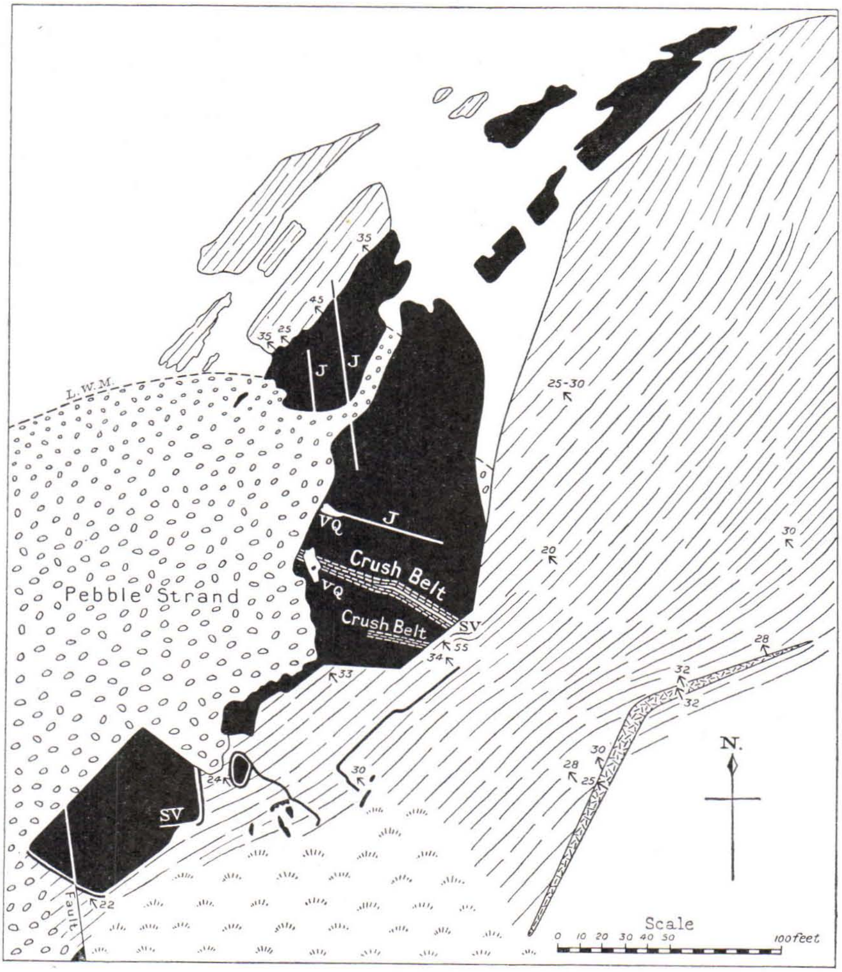

Fig. 3. Geological map of the eastern end of Pebble Strand, Malin Head district, Co. Donegal. Outcrops of the main epidiorite sill and of the altered lower sill shown in black. Strike of the country rocks (quartzite and mica-schist) indicated by lines. The stippled rock to the south-east is a lamprophyre dyke. J = joint; SV = »vein» of altered sedimentary rock caught up by the intrusion; VQ = vein quartz.

approaching arkose. Others, though only mildly feldspathic to begin with, have been strongly feldspathised at the contacts against the local granite (Reynolds, 1946, p. 411).

As may be seen from Figs. 2 and 3, the main sill of the White Cow Rock has been dissected by marine denudation into a number of isolated segments. The upper contact is exposed on one of the small islets situated at about the level of low-water mark, while the lower contact is well exposed along the coast of the mainland, both on the south-eastern side of the White Cow Rock itself, and in a few isolated inland exposures which outcrop to the south-west. Exposures of the lower sill and its relation to parts of the main sill are indicated in the sketch-map, Fig. 4, which also serves as an index to the localities A, B and C of the specimens described. The lower sill at locality C is illustrated by Fig. 10.

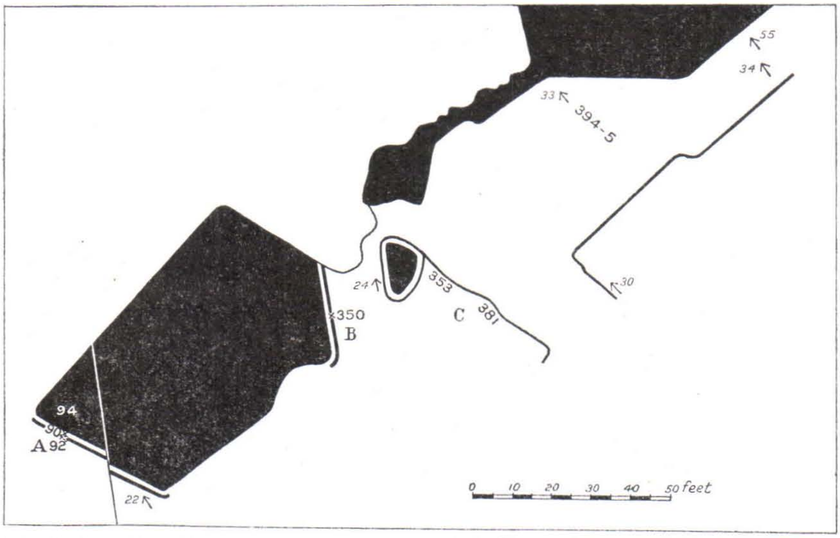

Fig. 4. An enlarged representation of part of Fig. 3 showing the outcrop of the lower sill in relation to the main epidiorite sill, and the positions of localities A, B and C. The numbers indicate the points at which analysed specimens were collected.

III. EPIDIORITE OF THE MAIN SILL AND ITS ALTERATIONS

The main sill has a platy structure and a planar foliation parallel to its upper and lower surfaces, the foliation being due to the presence of fleck-like quartzo-feldspathic lamellae. For a short distance from both its upper and lower contacts the sill becomes noticeably enriched in biotite. The lower zone of enrichment, 4 to 6 inches thick, is not only particularly rich in biotite, but also shows a distinct increase in grain size, as compared with the normal rock of the sill.

(A) EPIDIORITE

In thin section (Fig. 5) the normal rock of the main sill, as exemplified by the analysed specimen (No. 94, near locality A, Fig. 4), exhibits a marked foliation, dependent on the alternation of narrow melanocratic and leucocratic bands. The former are rich in hastingsite, and contain some iron ore, sphene, a little epidote and occasional biotite; while the latter are composed of feldspar, small crystals of epidote and occasional garnet. The amphibole is identified as a member of the hastingsite series from its iron-rich composition (inferred from Table I) combined with its optical properties, which are as follows: Z = bluish green, Y = yellowish green, X = pale straw colour; optic axial plane parallel to (010); $Z \wedge c =$ ca. 18°; $\gamma = 1.678$; 2V large (—ve). The feldspar is very fine grained, the individuals being intimately intergrown, with highly sinuous margins. It is found to be oligoclase, with $\gamma = 1.547$. This result is consistent with the chemical composition of the rock (Table I), for the K_2O content is no more than can be readily accounted for by the amount of biotite present.

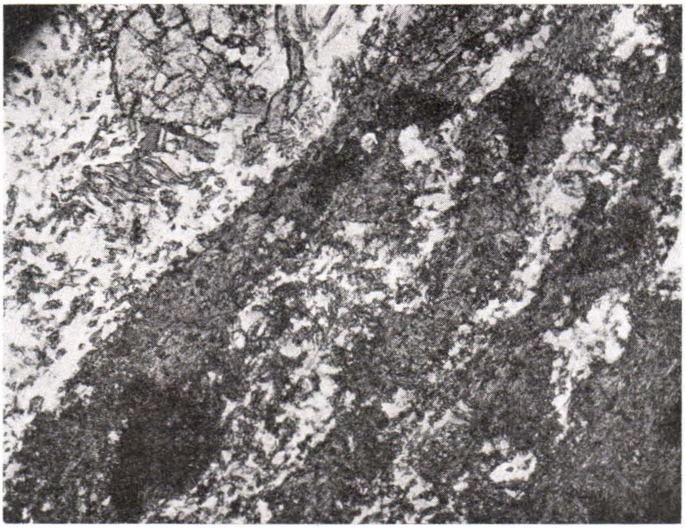

Fig. 5. Epidiorite (No. 94) of the main sill, showing the alternation of narrow melanocratic bands rich in hastingsite, and leucocratic bands rich in oligoclase. A garnet appears in the upper left hand corner of the field. Ordinary light, x 33.

The assemblage of minerals is typical of that of epidiorite in the garnet zone of regional metamorphism. Wiseman (1934, p. 378) has described a similar epidiorite from the garnet zone of South Knapdale in the northern part of the Kintyre peninsula, Argyllshire, an analysis of which is quoted for comparison in Table I (B). In the facies classification of Eskola (1939, p. 334) the mineral assemblage is that of the *epidote-amphibolite* facies, except for the presence of oligoclase instead of albite. However, Ramberg (1945, pp. 53 and 72) has recently proposed to define this facies by the association epidote-oligoclase, the *green-schist* facies being characterised by albite and the *amphibolite* facies by plagioclase more calcic than oligoclase. According to the schemes of Th. Vogt (1927, p. 442) and Kvale (1945, p. 109) the White Cow epidiorite would belong to the *hornblende-almandine-epidote* facies.

The chemical analysis of a representative specimen of the main sill (No 94) is recorded in Table I, together with the normative composition. Although chemically the rock is obviously of olivine-basaltic composition, there appear to be few analysed igneous rocks with which it can be closely matched. This is because of the relatively high Al_2O_3 and low MgO. One notable rock that shares these characters is a glassy variety of olivine-dolerite from the Hällefors Dyke (Krokström, 1936, p. 149) the composition of which is quoted for comparison in Table I (A). It should, of course, be kept in mind that the present chemical composition of the White Cow epidiorite does not necessarily represent that of the original igneous rock of the sill; possible changes brought about during the early stages of the regional metamorphism remain to be investigated. Amongst epidiorites of comparable mineral composition a fairly close chemical analogue (B in Table I) is the example from Knapdale, to which reference has already been made.

Table I

	Percentages			*Mol. Props.*
	No. 94	*A*	*B*	*of No.* 94
SiO_2	47.32	47.69	49.85	*.7879*
Al_2O_3	17.39	17.04	16.90	*.1706*
Fe_2O_3	2.36	3.77	1.74	*.0148*
FeO	10.91	9.39	11.16	*.1519*
MgO	4.78	4.56	3.93	*.1186*
CaO	10.08	9.17	8.72	*.1797*
Na_2O	2.73	2.89	3.45	*.0440*
K_2O	.54	1.50	.31	*.0057*
H_2O+	.96	1.20	1.00	*.0533*
H_2O-	.05	—	—	—
CO_2	tr.	—	—	—
TiO_2	2.68	2.20	2.32	*.0335*
P_2O_5	tr.	.20	.27	. —
MnO	.20	.20	.30	*.0028*
BaO	.01	—	—	*.0001*
	100.01	*99.81*	*99.95*	

Normative Composition of No. 94			
	Orthoclase		3.17
	Albite		23.07
	Anorthite		33.65
Diopside	$CaSiO_3$	6.84	13.63
	$MgSiO_3$	3.12	
	$FeSiO_3$	3.67	
Hypersthene	$MgSiO_3$	4.98	10.85
	$FeSiO_3$	5.87	
Olivine	Mg_2SiO_4	2.68	6.15
	Fe_2SiO_4	3.47	
	Magnetite		3.42
	Ilmenite		5.08
	Water		1.01
			100.03

Epidiorite, No. 94. Main Sill of White Cow Rock, Pebble Strand, Malin Head District, Co. Donegal, Eire.
Analyst: Agnes Gibbs.

A. Glassy olivine-dolerite, Hällefors Dyke, S. E. of Jacobsberg, Södermanland, Sweden (Krokström, 1936, p. 149).
Analyst: Naima Sahlbom.

B. Biotite-epidote-albite-epidiorite (65/4); 0.2 mile N. 51° E. from northern end of Loch-na-Craige, near Achahoish, South Knapdale., Scotland.
Analyst: W. H. Herdsman.

(B) BIOTITE-ENRICHED EPIDIORITE

The biotite-rich marginal portions of the main sill are composed of large biotite flakes and rarer hastingsite, set in a somewhat sparse matrix of relatively fine-grained quartz, epidote and a little feldspar. The matrix minerals also occur as inclusions distributed in sieve-like fashion within the biotite. Throughout the rock calcite is a common and garnet a sparsely distributed constituent.

In places the rock at the base of the sill exhibits an irregular spheroidal weathering. The cores of the spheroids are characterised by a greenish-brown biotite, but towards the margins of the spheroids a golden variety (lepidomelane) takes its place and appears to have developed from it. Similar biotites, having the respective properties stated on pp. 36 and 41 occur in the lower sill. In both occurrences the golden variety of biotite is relatively friable and has been mechanically disintegrated by weathering. It is important to notice, however, that not the slightest sign of any chemical weathering is detectable in thin sections. Both biotites are associated with a little chlorite (ripidolite, see page 44) which seems to have developed from the greenish brown biotite, as in the lower sill, where ripidolite has formed on a considerable scale (page 44).

IV. ALTERATIONS OF EPIDIORITE IN THE LOWER SILL

In the present investigation it is the lower sill that is of special interest, because in it the effects of the biotitisation and associated transformations that have been superimposed on the epidiorite are particularly well displayed.

(A) BIOTITE-EPIDIORITE

At locality B (Figs. 4 and 6) the lower sill is composed of an upper epidiorite zone 7.5 inches thick, and a lower zone of lesser thickness, but having the macroscopic appearance of a biotite-rich schist (No. 348). The boundary between these two divisions is sharp. The epidiorite zone, moreover, has biotite-enriched margins: 2 inches thick (No. 349) against the »biotite-rich schist» and 1.5 inches thick (No. 351) at the upper contact against quartzite. The intervening epidiorite is represented by specimen No. 350, which has been analysed (Table II).

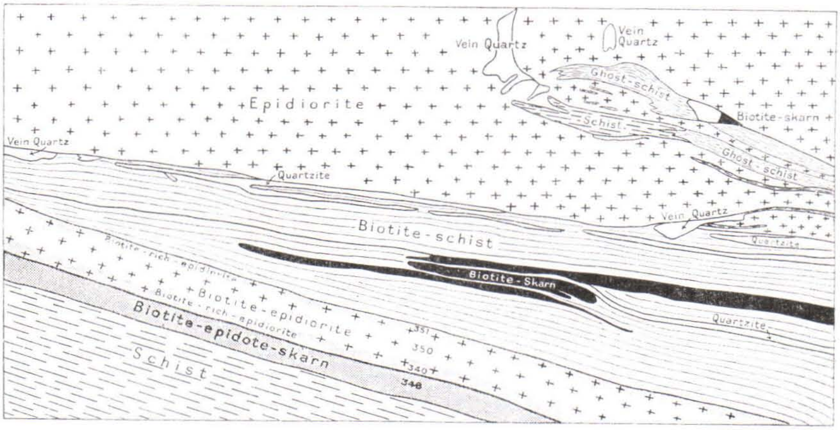

Fig. 6. Elevation of the exposure at locality B (see Fig. 4), showing the lower sill, underlying the main sill, and the disposition of the various rock-types. Numbers refer to described specimens (340 in the above figure should be 349). The section here depicted is 11 feet long.

Fig. 7. Biotite-epidiorite (No. 350) from the lower sill at locality B, showing the development of biotite with cross-cutting relations to the foliation. The foliation is dependent on the parallelism of prisms of hastingsite. Ordinary light, x 33.

This rock, No. 350, is a fine-grained biotite-epidiorite, with a somewhat foliated structure due to the parallelism of abundant elongated prisms of hastingsite (Fig. 7). The other minerals present are quartz, oligoclase ($\gamma = 1.546$) with abundant associated sericite, finely granular epidote, relatively large flakes of biotite, and a little sphene, black iron-ore and pyrite. The biotite, part of which has been altered to chlorite (ripidolite, see page 44) replaces the original constituents of the epidiorite. Its flakes, commonly measuring 1 mm., and rarely 2 mm. across, are developed both parallel and at a high angle to the foliation of the rock. Prisms of hastingsite — large compared with the amphibole forming the main part of the rock — also cut across the foliation and are evidently of late development. The hastingsite in this rock has optical properties identical with those of the hastingsite of the main sill (No. 94).

Table II

BIOTITE-EPIDIORITE FROM THE LOWER SILL LOCALITY B, NO. 350.

	Percentages	*Molecular Proportions*
SiO_2	48·34	*.8049*
Al_2O_3	15·49	*.1520*
Fe_2O_3	2·14	*.0134*
FeO	10.60	*.1476*
MgO	5·56	*.1379*
CaO	9·42	*.1680*
Na_2O	1·98	*.0319*
K_2O	1·36	*.0144*
H_2O+	1·54	*.0855*
H_2O-	.16	
CO_2	none	
TiO_2	2·48	*.0310*
P_2O_5	·32	*.0023*
MnO	·34	*.0048*
	99·73	

Normative Composition			
Quartz			1.02
Orthoclase			8.01
Albite			16·72
Anorthite			29·40
Diopside	$CaSiO_3$	6.12	12·14
	$MgSiO_3$	2·97	
	$FeSiO_3$	3·05	
Hypersthene	$MgSiO_3$	10.87	21·98
	$FeSiO_3$	11.11	
Magnetite			3·10
Ilmenite			4·70
Apatite			1.00
Water			1·70
			99·77

Analyst: W. H. Herdsman

The chemical composition of No. 350 is recorded in Table II. Compared with the composition of No. 94 (Table I) the significant differences are increase of MgO, K_2O and H_2O, and decrease of Al_2O_3 and Na_2O. Silica increases slightly and iron oxides remain about the same. The differences correspond to increase of biotite and quartz at the expense, mainly, of amphibole and albite.

(B) BIOTITE-RICH EPIDIORITE

The rock, No. 351, from the upper contact of the lower sill at locality B, resembles No. 350 just described, but is richer in biotite, which builds flakes up to 4 mm. across, contains more epidote relative to hastingsite, and more quartz relative to albite. It also contains apatite as a noticeable accessory mineral. The biotite-rich epidiorite, No. 349 from the horizon in contact with the lower part of the sill, also resembles No. 350, but again it is richer in biotite, which in this case builds flakes up to 3 mm. across. The biotite crosses the foliation and is obviously a replacement mineral. It is itself cut and replaced by prisms of hastingsite which are large in comparison with those forming the body of the rock. Garnet is a rare constituent of this horizon (No. 349) of the sill, and some of the iron-ore is margined by sphene.

The rock with the appearance of a biotite-schist, No. 348, forming the lower zone of the sill, consists essentially of biotite, epidote and quartz, with accessory garnet and hastingsite. The latter occurs as elongated prisms, 1 mm. or more in length, that cut through all the other constituents, and are evidently of late development. Apart from the presence of accessory hastingsite, this rock is very similar to No. 92 from locality A, an analysis of which is given in Table III.

Throughout the sill the epidote is colourless, has 2V close to 90°, and exhibits marked dispersion $\nu > \varrho$. It shows variation in its optical sign, however, being sometimes + ve, and sometimes — ve. It is more correctly described as clinozoisite in the former case and epidote in the latter, but such a distinction is without chemical significance, since in either case the mineral falls very close to the boundary between clinozoisite and epidote (Winchell, 1933, p. 313). Moreover, zoning with slight increase of birefringence towards the rim is not uncommon. Accordingly, throughout the paper, the term epidote will be used to describe this mineral without further discussion of its optical properties.

(C) BIOTITE-SKARN

At locality A (Fig. 8) the lower sill is rich in biotite throughout. The rock at the upper contact of the sill, No. 95, has a distinctly schistose structure, due to the parallelism of the component minerals. It is composed of biotite, epidote, quartz and hastingsite. The rock No. 96, from the lower contact of the sill, is characterised by the same minerals, but is not schistose.

The rock forming the middle of the sill, No. 92 (Fig. 9) closely resembles the biotite-rich rock, No. 348, from the lower part of the sill at locality B. It is composed of biotite, epidote and quartz, with some calcite, aggregates of sericite, scattered granules of sphene within both the biotite and the quartz, and rare haematite. The biotite has X = straw colour < Y = Z = old gold to greenish brown; 2V small; and γ ranging from 1.643 to 1.645.

The rocks composing the sill at localities A and B are thus not only highly variable, but the distribution of the varieties differs markedly at the two localities. The evidence is clear that the biotite-rich varieties have developed from an amphibole-rich rock, the composition of which is most closely approached by the biotite-epidiorite, No. 350, from the middle of the sill at locality B (Table II). As biotite increases in amount at the expense of amphibole, epidote shows a corresponding increase both in amount and in grain size. The epidote-quartz-biotite rock is, in fact, a skarn replacing epidiorite. At locality A there is every gradation between biotite-rich epidiorite and skarn, whereas at locality B there is a sharp contact between biotite-rich epidiorite and skarn.

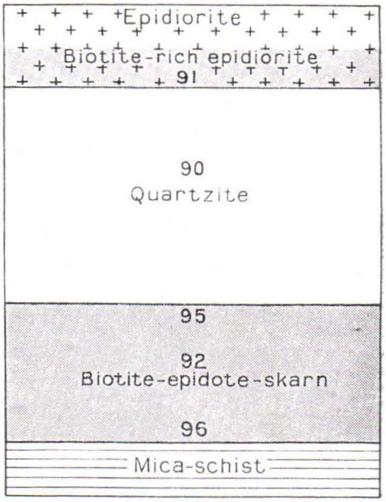

Fig. 8. Elevation of the exposure at locality A (see Fig. 4) showing the disposition of the various rock-types. Numbers refer to described specimens. The height of the section depicted is 3.5 feet.

The chemical composition of No. 92 is recorded in Table III. Disappearance of Na_2O and marked increase of K_2O and H_2O are the most notable changes as compared with the previous analyses, but there is also increase of total iron oxide and decrease of CaO and MgO. An attempt has been made to estimate the composition of the biotite of No. 92 (Table IV). By micrometric measurement the percentage of quartz was found to be about 18, and that of sericite about 1.3. Epidote and biotite could not be satisfactorily measured because of the abundance of minute inclusions of sphene and epidote in the biotite. However, assuming that all the TiO_2 is accounted for by sphene, the amount of epidote can be assessed by taking its composition to be 90 per cent of $HCa_2Al_3Si_3O_{13}$ and 10 per cent of $HCa_2Fe_3Si_3O_{13}$, and allotting to it all the CaO left over after making calcite, sphene and apatite. The final balance, representing the approximate composition of biotite, is given in the last three columns of Table IV, and corresponds to:

SiO_2	R_2O_3	RO	$(K, Na)_2O$	H_2O
600	215	530	89	176
(600)	(176)	(514)	(89)	(150)

This composition falls within the field of basic micas plotted by Hallimond (1926, p. 29). If all the TiO_2 were in the biotite, the composition would be as indicated above in brackets. The colour of the biotite and the profusion of sphene shows, however, that no more than a small proportion of the TiO_2 can actually be present in the biotite (Hall, 1941, p. 32). Allotting some TiO_2 to the biotite would slightly increase the amount of epidote at the expence of the sphene and biotite, but would obviously have little effect on the calculated composition of the biotite as given in Table IV.

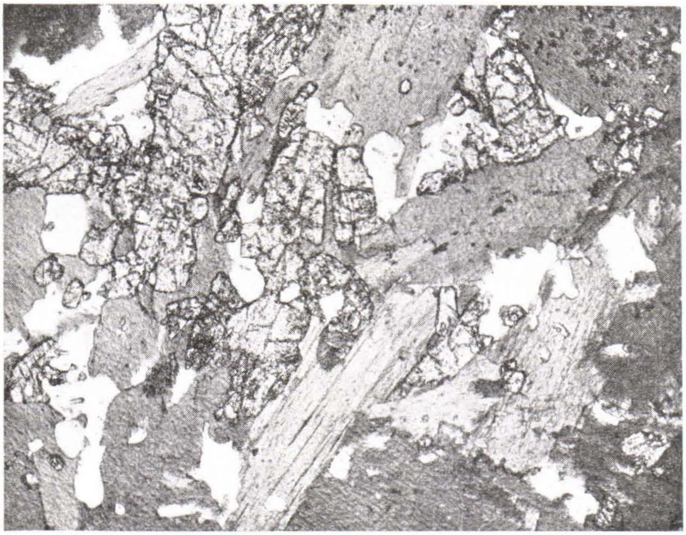

Fig. 9. Biotite-skarn (No. 92) from the lower sill at locality A, showing the association of biotite, epidote and quartz. The colourless areas are quartz. Ordinary light, x 33.

Table III

EPIDOTE-QUARTZ-BIOTITE SKARN, FROM THE LOWER SILL, NO. 92, LOCALITY A.

Percentages		*Mol. Props.*	*Normative Composition*			*Mode from Table IV*	
SiO_2	45·65	*.7601*	Quartz		2·26	Biotite	55·53
Al_2O_3	15·56	*.1526*	Orthoclase		28·16	Epidote	14·16
Fe_2O_3	2·55	*.0160*	Albite		·11	Quartz	18.00
FeO	11·21	*.1560*	Anorthite		24·89	Sphene	5·65
MgO	5·01	*.1243*	Corundum		1·25	Calcite	4·59
CaO	7·94	*.1416*	Hypersthene	$MgSiO_3$ 12·48	27·57	Sericite	1·30
Na_2O	·01	*.0002*		$FeSiO_3$ 15·09		Apatite	·71
K_2O	4·77	*.0506*	Magnetite		3·70		99·94
H_2O+	2·35	*.1304*	Ilmenite		4·37		
$H_2O—$	·10		Apatite		·71		
CO_2	2·02	*.0459*	Calcite		4·59		
TiO_2	2·30	*.0288*	Water		2·45		
P_2O_5	·30	*.0021*			*100.06*		
F	none fd.						
NiO	tr.						
MnO	·23	*.0032*					
BaO	·02	*.0001*					
	100.02						

Analyst: Agnes Gibbs

Table IV

MINERAL COMPOSITION OF EPIDOTE-QUARTZ-BIOTITE-SKARN, FROM LOWER SILL, LOCALITY A, NO. 92.

Molecular Proportions		Quartz (18 %)	Calcite	Apatite	Sphene	Epidote	Sericite (1.3 %)	Biotite		
								M. P.	% in rock	Composition %
SiO_2	*7601*	*2997*	—	—	*288*	*900*	*98*	*3318*	19.93	35.89
Al_2O_3	*1526*	—	—	—	—	*405*	*49*	*1072*	10.93	19.68
Fe_2O_3	*160*	—	—	—	—	*45*	—	*115*	1.84	3.31
FeO	*1560*	—	—	—	—	—	—	*1560*	11.21	20.19
MnO	*32*	—	—	—	—	—	—	*32*	.23	.41
MgO	*1243*	—	—	—	—	—	—	*1243*	5.01	9.02
CaO	*1417*	—	*459*	*70*	*288*	*600*	—	—	—	—
Na_2O	*2*	—	—	—	—	—	—	*2*	.01	.02
K_2O	*506*	—	—	—	—	—	*16*	*490*	4.62	8.32
H_2O+	*1304*	—	—	—	—	*300*	*33*	*971*	1.75	3.15
CO_2	*459*	—	*459*	—	—	—	—	—	—	—
TiO_2	*288*	—	—	—	*288*	—	—	—	—	—
P_2O_5	*21*	—	—	*21*	—	—	—	—	—	—
Total (% of rock)		18.00	4.59	.71	5.65	14.16	1.30	55.53	55.53	99.99

As may be seen from Fig. 10, much of the sill at locality C has undergone a mechanical disintegration, cores of solid rock remaining within the crumbled material. In the following descriptions, it is shown that some of the cores of solid rock are composed of biotite-skarn and others of chlorite-skarn, while the crumbled material is lepidomelane-skarn. The positions of the described specimens are indicated on Fig. 11.

Fig. 10. The lower sill at locality C (see Fig. 4), with quartzite above and below. Cores of »solid» rock (biotite-skarn and chlorite-skarn) can be seen in the »crumbly» rock (lepidomelane-skarn) which here makes up the greater part of the lower sill. The section photographed is 4 feet long.

No. 382 represents the solid rock from the middle of the sill, towards the south-east (on the right-hand side of Fig. 11). It is a calcite-quartz-epidote-biotite-skarn (Fig. 12), very similar to that described from locality A, No. 92. It contains rather less quartz, however, than any of the other sectioned specimens of skarn. The biotite is lace-like in its development,

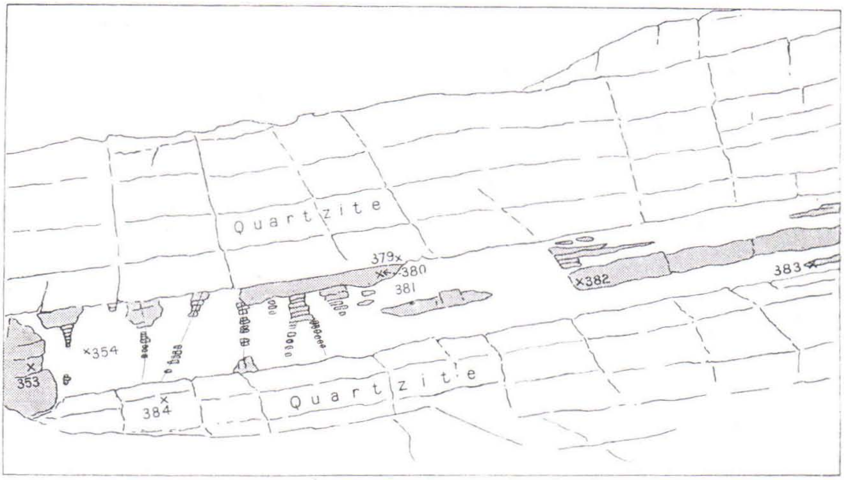

Fig. 11. Elevation of the exposure at locality C (see Fig. 4) showing the distribution of »solid» rock (mainly biotite-skarn and chlorite-skarn) indicated by shading, and of »crumbly» rock (lepidomelane-skarn) left unshaded. Numbers refer to described specimens. The section here depicted is 8 feet long.

Fig. 12. Biotite-skarn (No. 382) from the »solid» part of the lower sill at locality C, showing the association of biotite, epidote and quartz. The dark area towards the centre of the field is a basal section of biotite, and the colourless areas are quartz. Ordinary light, x 33.

owing both to the presence of abundant inclusions of epidote and to the highly irregular margins that have resulted from its growth around individuals of epidote. The biotite evidently developed at the expense of the minerals of the original epidiorite other than epidote. It has a more definitely greenish colour than the biotite characterising the analysed biotite-skarn, No. 92, having X = pale yellow < Y = Z = olive-green or greenish brown; 2V small; and $\gamma = 1.642$. Calcite is of late development and contains relics of the other minerals which it replaces. In addition to the constituents already mentioned, the rock also contains a little amphibole and scattered granules of sphene, the latter being less abundant than in the analysed biotite-skarn, No. 92.

No. 383, collected about 24 inches north-east of No. 382, from the lowest horizon of the sill represented by solid rock, resembles No. 382, but contains in addition fairly abundant potash feldspar, and a very little plagioclase. In this specimen the epidote is in part replaced by a greenish yellow chloritic or serpentinous mineral which forms a meshwork structure similar to that commonly seen in partly serpentinised olivine.

(D) LEPIDOMELANE-SKARN

The crumbled rock differs from the solid varieties described above. The analysed specimen, No. 381, is epidote-quartz-lepidomelane-skarn (Fig. 13), characterised by a different biotite from that occurring in the solid biotite-skarn of localities A, B and C. Macroscopically, the biotite has rather the colour of »old gold». In thin section it has X = light golden yellow < Y = deep gold < Z = reddish gold; 2V *ca.* 20°; $\gamma = 1.681$; and marked dispersion $\nu > \varrho$; it is therefore by no means a normal variety, and is shown below to be lepidomelane. Associated with the golden biotite is a small amount of a greenish brown or distinctly green biotite (with 2V small and $\gamma = 1.646$), which resembles the biotite of the solid biotite-skarn, No. 382, of locality C. The greenish biotite occurs in intimate intergrowth with the golden variety, both as cores within it and as irregular patches or wedge-like lamellae extending parallel to the cleavage. In places it is also clearly cut by the golden variety. From these relationships it is inferred that the golden variety has been formed at the expense of the greenish variety of the solid biotite-skarn. This inference is confirmed by the association of these two varieties of biotite in the main sill. There, as already mentioned (page 33), the cores of the spheroids are characterised by the greenish biotite, while towards the rims of the spheroids this variety is partly replaced by the golden biotite. As in the lower sill, the rock characterised by the golden biotite is friable and has been differentially picked out by mechanical weathering. To a very small extent the golden biotite of No. 381 shows alteration to chlorite (ripidolite).

Another specimen of the crumbly skarn, No. 354, resembles that just described, but contains some hastingsite of the early generation in addition, thereby retaining a trace of its epidiorite ancestry. As in the biotite-skarn, the lepidomelane-skarn is characterised throughout by a profusion of minute granules of sphene occurring within both mica and quartz.

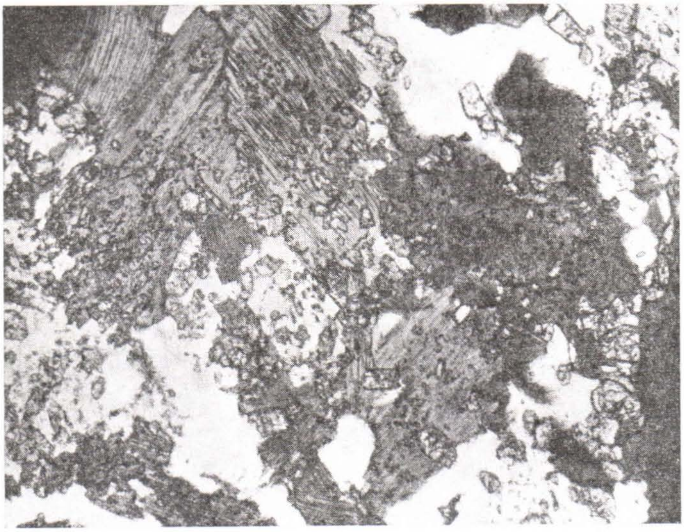

Fig. 13. Lepidomelane-skarn (No. 381) representing the »crumbly» rock of the lower sill at locality C. The field shows the association of the golden coloured lepidomelane with epidote and quartz (colourless). The dark coloured basal section of lepidomelane towards the right hand side of the field has a darker central core of greenish-brown biotite. Ordinary light, x 33.

The chemical composition of No. 381, a typical example of the crumbly rock containing the golden lepidomelane, is given in Table V. The chief differences compared with No. 92 (Table III), apart from silica and calcite, are the marked increase in Fe_2O_3 at the expense of FeO; increase of MgO, H_2O and Na_2O, and decrease of Al_2O_3 and K_2O. These differences are almost wholly accounted for by the change in composition from biotite to lepidomelane. As before (page 37) the composition of the latter can be estimated approximately by assuming it to be free from TiO_2. The colour shows that in any case the proportion of TiO_2 is relatively low (Hall, 1941, p. 32). Epidote (of the same composition as in No. 92) is calculated from the CaO left over after making apatite and sphene. Quartz is estimated at about 20 per cent. The balance, representing the biotite, is listed in the last three columns of Table VI, and corresponds to:

SiO_2	R_2O_3	RO	$(K, Na)_2O$	H_2O
600	250	472	83	222

Table V

EPIDOTE-QUARTZ-LEPIDOMELANE-SKARN, FROM THE LOWER SILL, LOCALITY C, NO. 381.

	Percentages	*Mol. Props.*	*Normative Composition*				*Mode from Table VI*	
SiO_2	48.13	*.8014*	Quartz			8.95		
Al_2O_3	14.76	*.1448*	Orthoclase			21.20	Lepidomelane	55.41
Fe_2O_3	7.34	*.0460*	Albite			3.62	Quartz	20.00
FeO	6.38	*.0888*	Anorthite			27.76	Epidote	17.12
NgO	6.61	*.1639*		$CaSiO_3$	.44		Sphene	6.33
CaO	6.18	*.1102*	Diopside	$MgSiO_3$	.35	.83	Apatite	.68
Ma_2O	.43	*.0069*		$FeSiO_3$	.04			*99.54*
K_2O	3.59	*.0381*	Hypersthene	$MgSiO_3$	16.10	18.04		
H_2O+	2.82	*.1565*		$FeSiO_3$	1.94			
H_2O-	.39	—	Magnetite			10.65		
CO_2	none	—	Ilmenite			4.90		
TiO_2	2.58	*.0323*	Apatite			.68		
P_2O_5	.22	*.0016*	Water			3.21		
MnO	.32	*.0045*				99.84		
	99.75							

Analyst: W. H. Herdsman

Table VI

MINERAL COMPOSITION OF EPIDOTE-QUARTZ-LEPIDOMELANE-SKARN, FROM LOWER SILL, LOCALITY C, NO. 381.

Molecular Proportions		Quartz (20 %)	Apatite	Sphene	Epidote	Lepidomelane		
						M. P.	% in rock	Composition %
SiO_2	*8014*	*3330*	—	*323*	*1089*	*3272*	19.65	35.46
Al_2O_3	*1448*	—	—	—	*490*	*958*	9.77	17.63
Fe_2O_3	*460*	—	—	—	*54*	*406*	6.48	11.69
FeO	*888*	—	—	—	—	*888*	6.38	11.51
MnO	*45*	—	—	—	—	*45*	.32	.58
MgO	*1639*	—	—	—	—	*1639*	6.61	11.93
CaO	*1102*	—	*53*	*323*	*726*	—	—	—
Na_2O	*69*	—	—	—	—	*69*	.43	.78
K_2O	*381*	—	—	—	—	*381*	3.59	6.48
H_2O+	*1565*	—	—	—	*363*	*1202*	2.18	3.93
TiO_2	*323*	—	—	*323*	—	—	—	—
P_2O_5	*16*	—	*16*	—	—	—	—	—
Total (% of rock)		20.0	.68	6.33	17.12	55.41	55.41	99.99

The golden biotite is evidently a variety of lepidomelane with high Al_2O_3 and nearly equal percentages of Fe_2O_3, FeO and MgO. Although no similar lepidomelane appears to have been analysed, the nearest examples having higher Fe_2O_3 and lower Al_2O_3 and MgO, the estimated composition falls well inside Hallimond's (1926, p. 29) field of basic micas. Comparison with the greenish brown biotite (Table IV) shows that MgO and H_2O have increased in the lepidomelane, half of the FeO has been oxidized to Fe_2O_3, and Al_2O_3 and K_2O have decreased.

(E) CHLORITE-SKARN

The biotite-skarn (No. 382) has been transformed not only to the friable lepidomelane-skarn (No. 381), but also, in places at locality C, to a chlorite-skarn. This rock has a grey colour and, like the biotite-skarn from which it is derived, it forms part of the solid rock and cores of the lower sill. No. 380, representing the residual solid rock from the upper contact of the sill, provides a transitional type between biotite-skarn and chlorite-skarn. It is composed of biotite, chlorite, epidote, quartz, calcite, and a little amphibole. The amphibole forms large individuals of late formation, which are largely replaced by calcite. The biotite is the greenish brown variety already described from the biotite-skarn No. 382. All stages of conversion of this greenish-brown biotite to chlorite are represented in No. 380, so that it can be concluded that the chlorite-skarn of the lower sill is a metasomatic derivative from biotite-skarn.

A specimen of chlorite-skarn, No. 353, from the left-hand side of locality C (Figs. 10 and 11), shows the rock to be surprisingly different from the varieties of skarn so far described. The rock is composed of chlorite, epidote, quartz, feldspar, sphene and calcite (Fig. 14). The chlorite forms individuals which commonly measure 2.5 mm. across. It is optically -ve; with $X =$ straw coloured $< Y = Z =$ green; $\gamma = 1.627$; and anomalous ultramarine to purple interference colours. These properties and the estimated chemical composition (Table VIII) show that the chlorite is ripidolite. A colourless mineral is intimately intergrown with the chlorite. The contacts between the two minerals are elongated parallel to the cleavage of the chlorite, while at right angles to the cleavage, they are toothed and dovetailed, wedge-like portions of both minerals extending along the cleavage direction of the chlorite. The intergrowth resembles that so commonly seen between chlorite and biotite. The colourless mineral is optically + ve; with a large optic axial angle; $\gamma = 1.524$; and a low birefringence. These properties suggest that it may be iso-orthoclase, but its intergrowth with chlorite is so anomalous for a feldspar that further investigation is necessary. The colourless mineral never extends beyond the crystal boundaries of the chlorite, and both minerals are speckled with minute inclusions of sphene. These facts suggest that the chlorite and the colourless mineral together take the place of the biotite of the biotite-skarn. Epidote occurs mainly in association with a quartzose matrix resembling that of the biotite-skarn, and to a lesser extent as inclusions within the chlorite. A little albite is associated with the quartz, the latter mineral being speckled with small inclusions of sphene, as in the biotite-skarn. Calcite builds large individuals that show replacement relationships to all the other constituents, relics of which, including abundant small granules of sphene like those found within the other minerals, occur within it.

The chemical composition of the chlorite-skarn No. 353 is recorded in Table VII. Compared with No. 92 (Table III) the most significant change

Fig. 14. Chlorite-skarn (No. 353) from the »solid» part of the lower sill at locality C, showing the intergrowth of ripidolite (grey) with a colourless mineral provisionally identified as iso-orthoclase (crowded with inclusions of granular sphene that appear as short dark lines and patches in the photomicrograph). Ordinary light, x 33.

Table VII

EPIDOTE-QUARTZ-RIPIDOLITE-SKARN, FROM THE LOWER SILL, LOCALITY C, NO. 353

	Percentages	*Molecular Proportions*		*Mineral Composition from Table VIII*
SiO_2	44.98	.7489	Ripidolite	45.07
Al_2O_3	14.44	.1417	Quartz	19.00
Fe_2O_3	3.02	.0189	Epidote	15.09
FeO	10.98	.1528	Alkali Feldspar	8.66
MgO	6.08	.1508	Sphene	7.17
CaO	8.24	.1469	Calcite	3.96
Na_2O	.16	.0026	Apatite	.67
K_2O	1.23	.0131		99.62
H_2O+	5.19	.2881		
H_2O-	.18			
CO_2	1.74	.0396		
TiO_2	2.92	.0366		
P_2O_5	.29	.0020		
MnO	.35	.0049		
	99.80			

Analyst: W. H. Herdsman

Table VIII

MINERAL COMPOSITION OF EPIDOTE-QUARTZ-RIPIDOLITE-SKARN FROM THE LOWER SILL, LOCALITY C, NO. 353

Molecular Proportions		Quartz (19 %)	Calcite	Apatite	Sphene	Epidote	Feld-spar	Ripidolite		
								M. P.	% in rock	Composition %
SiO_2	*7489*	*3164*	—	—	*366*	*960*	*942*	*2057*	12.36	27.42
Al_2O_3	*1417*	—	—	—	—	*432*	*157*	*828*	8.44	18.73
Fe_2O_3	*189*	—	—	—	—	*48*	—	*141*	2.25	4.99
FeO	*1528*	—	—	—	—	—	—	*1528*	10.98	24.36
MnO	*49*	—	—	—	—	—	—	*49*	.35	.78
MgO	*1508*	—	—	—	—	—	—	*1508*	6.08	13.49
CaO	*1469*	—	*396*	*67*	*366*	*640*	—	—	—	—
Na_2O	*26*	—	—	—	—	—	*26*	—	—	—
K_2O	*131*	—	—	—	—	—	*131*	—	—	—
H_2O+	*2881*	—	—	—	—	*320*	—	*2561*	4.61	10.23
CO_2	*396*	—	*396*	—	—	—	—	—	—	—
TiO_2	*366*	—	—	—	*366*	—	—	—	—	—
P_2O_5	*20*	—	—	*20*	—	—	—	—	—	—
Total (% of rock)		19.00	3.96	.67	7.17	15.09	8.66	45.07	45.07	100.00

is a marked decrease in K_2O and a corresponding increase in H_2O. As in the change from biotite-skarn to lepidomelane-skarn there is increase of MgO and decrease of Al_2O_3. Total iron oxides have slightly increased, but only a little oxidation to Fe_2O_3 has taken place. There is an increase of TiO_2 in the chlorite-skarn, correlatable with an increase in the amount of granular sphene as compared with that in the biotite-skarn. The composition of the chlorite is approximately estimated in Table VIII, by allotting all the TiO_2 to sphene and all the alkalis to feldspar, taking the composition of epidote as before, and the percentage of quartz at 19. The chlorite is evidently ripidolite (Winchell, 1933, p. 279) with the constitution:

Antigorite	*Ferro-antigorite*	*Amesite*	*Daphnite*
19	21	29	31.

V. ALTERATIONS OF QUARTZITE

(A) QUARTZITE TO MICA-SCHIST

The country rock within which the epidiorite sills are emplaced is white quartzite with interlaminated layers of silvery grey muscovite-schist. To the south-west of the White Cow Rock, a vertical section of the rocks underlying the sills is exposed in an embayment. Here it can be seen that the quartzite has been metasomatically replaced by biotite-muscovite-schist. The relationships between the two contrasted rock types are clearly illustrated by Fig. 15. Mica-schist is the main rock exposed in the section, but within it irregular relics of quartzite remain in undisturbed position. The following observations have a bearing on the interpretation to be placed on these relationships:

Fig. 15. Part of the country rocks exposed below the sills, southwest of the White Cow Rock, showing undisturbed relics of quartzite (light) in mica-schist (dark). The mica-schist has been formed in situ at the expense of the vanished quartzite. The section photographed is about 5 feet long.

(a) The whole quartzite—mica-schist formation at this locality is foliated parallel to what appears to be the original bedding. The dip is uniform throughout and perfectly concordant with the base of the lower sill, a fact indicating that the bedded structure is a genuine one which has been followed faithfully by the later foliation.

(b) The bedding planes can be followed without interruption or disturbance of any kind from schist to quartzite, through the quartzite, and into the schist beyond.

(c) No sign of folding or fracturing can be detected on dip slopes or in sections at right angles or otherwise inclined to the bedding.

(d) The forms of the quartzite masses are extraordinarily irregular, ranging from small lenticles and long narrow wisps to large masses which may be either jagged and frayed out at their ends, or rounded and discordant with blunt and abrupt terminations against the schist.

(e) The junctions between quartzite and schist are everywhere sharp.

(f) The schist contains accessory iron ores and apatite, whereas the quartzite in direct continuity with it (between the same pair of closely spaced bedding planes) is free from these minerals.

The structural features (a), (b), (c) and (d) show that the relationship cannot be accounted for by either tectonic involutions or brecciation. Similarly (d), (e) and (f) show that the relationships cannot be accounted for as a result of original differences in sedimentation from point to point.

Simultaneous deposition of sand and mud would give gradations and a common suite of heavy minerals. It would be impossible for a hollow of the »cut and fill» type *(cf.* Fig. 15) to be filled on one side with mud and on the other with sand, with a sharp junction between them. Moreover, there is no sign of differential compaction structures such as would subsequently result from deposition of sand and mud side by side. Careful and repeated study of the section in all its aspects has convinced us that the only tenable interpretation is that the greater part of a sheet of bedded quartzite has been metasomatically replaced by mica-schist.

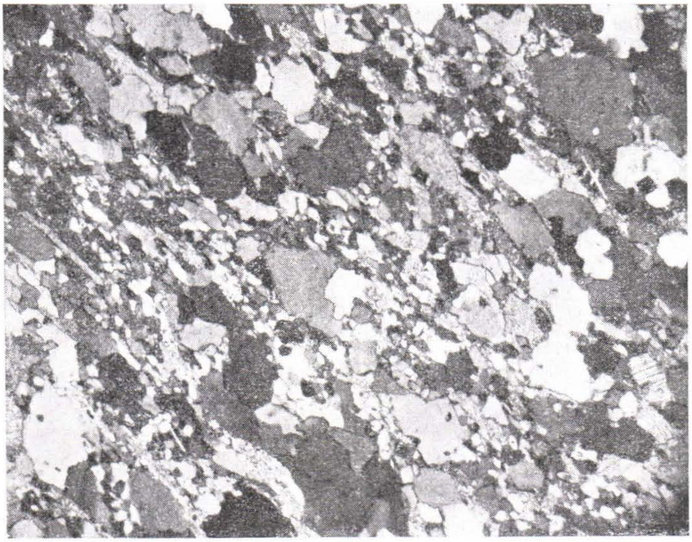

Fig. 16. Quartzite (No. 394) from the locality illustrated in Fig. 15, showing the association of quartz with relatively fine grained albite, occasional flakes of muscovite, and calcite (in the right hand bottom corner). Crossed nicols, x 33.

Microscopically the quartzite, as exemplified by the analysed specimen No. *394*, is seen to be composed of quartz, albite, a little potash feldspar, muscovite, a little biotite, rare garnets and a considerable proportion of calcite (Fig. 16). Between crossed nicols, the quartz, individuals of which measure about *.3* mm. or a little less in diameter, stands in marked contrast to the albite ($\gamma = 1.538$). The latter, in association with a little potash feldspar and quartz, occurs in aggregates within which the individuals measure approximately .07 mm. across — that is, only about a quarter the width of the quartz grains. The rock has a slightly foliated structure dependent on the parallelism of the muscovite flakes and the elongation of the albite-aggregates in the same direction. The rare garnets have a highly irregular form. Biotite is of relatively late development, for it replaces garnet and commonly cuts across the foliation direction of the rock.

The biotite-muscovite-schist, exemplified by the analysed specimen No. 395, differs from the replaced quartzite in being richer in feldspar and biotite and in containing iron ore and apatite as common accessories (Fig. 17). As compared with the quartzite, moreover, there is a great ncrease in the proportion of potash feldspar both relatively to albite and ibsolutely. Like the quartzite, the schist contains rare garnets; but, unlike the quartzite, calcite is absent or very rare. Chlorite is not uncommonly developed at the expense of biotite.

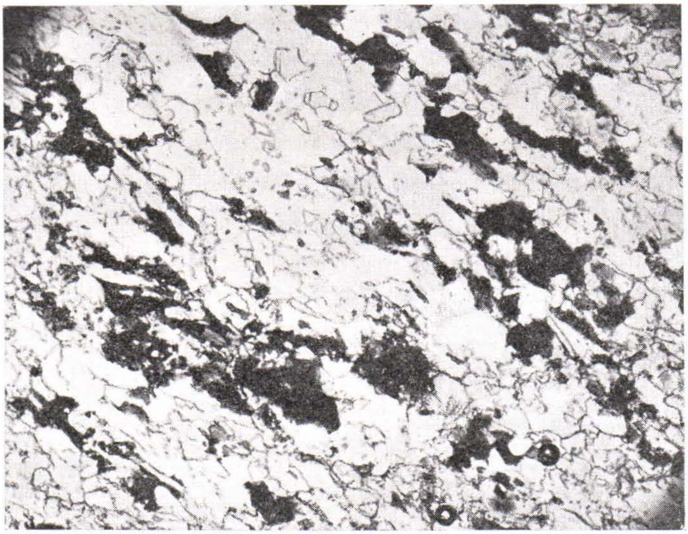

Fig. 17. Mica-schist (No. 395) from the same horizon as, and immediately adjoining, the quartzite of Fig. 16. The biotite appears dark. Ordinary light, x 33.

Table IX

CHEMICAL COMPOSITIONS OF QUARTZITE (NO. 394) AND MICA-SCHIST (NO. 395) FROM THE COUNTRY ROCK UNDERLYING THE WHITE COW SILLS

	Quartzite		*Mica-schist*			*Normative Composition*			
	No. 394		*No.* 395				*No.* 394		*No.* 395
SiO_2	72.06	*1.1998*	65.36	*1.0883*	Quartz		46.94		28.90
Al_2O_3	10.92	*.1071*	15.68	*.1538*	Orthoclase		6.85		26.82
Fe_2O_3	.04	*.0003*	1.49	*.0093*	Albite		24.78		18.25
FeO	1.40	*.0195*	4.32	*.0601*	Anorthite		4.70		3.03
MgO	.31	*.0077*	1.79	*.0444*	Corundum		3.12		6.11
CaO	6.02	*.1074*	.83	*.0148*	Hypersthene	$MgSiO_3$ .77	3.30	4.46	10.13
Na_2O	2.93	*.0473*	2.16	*.0348*		$FeSiO_3$ 2.53		5.67	
K_2O	1.16	*.0123*	4.54	*.0482*	Magnetite		.07		2.15
H_2O+	.84	—	2.32		Ilmenite		.15		1.64
H_2O-	.26	. —	.34		Apatite		none		.55
CO_2	3.98	*.0905*	none		Calcite		9.05		none
TiO_2	.08	*.0010*	.86	*.0108*	Water		1.10		2.66
P_2O_5	none		.18	*.0013*			*100.06*		*100.24*
MnO	.07	*.0010*	.21	*.0030*					
	100.07		*100.08*						

Analyst: W. H. Herdsman

The chemical compositions of Nos. 394 and 395 are recorded in Table IX. Exact correlation between the two specimens is ensured by the fact that both came from a particularly well marked bed about three inches thick, and were originally adjacent on each side of an abrupt junction between quartzite and schist.

It will be seen that the change from quartzite to schist has involved increase of Al_2O_3, iron oxides, MgO, K_2O, H_2O, TiO_2, P_2O_5 and MnO; and loss of SiO_2, CaO, CO_2 and Na_2O.

(B) QUARTZITE TO SKARN-ROCKS

Between the sills, and at and near their outer contacts, the quartzite is usually very much altered. The alterations vary considerably from point to point and include marked enrichments in biotite, garnet or epidote respectively, as well as in various combinations of these minerals. The following examples illustrate the variability and complexity of the mineral changes.

BIOTITE-SKARN

At locality B (Fig. 6), lenticular masses of dark glistening biotite-schist have developed from the quartzite between the two sills, and a similar phenomenon is to be seen above the upper contact of the main sill, well exposed on one of the little islets shown in Figs. 2 and 3. Specimen 313 (Fig. 18) from one such lenticular mass at the upper contact of the main sill is found, in microscope section, to be a calcite-quartz-biotite-skarn with a little fine-grained epidote, some muscovite and granules of sphene. It differs from the analysed mica-schist, No. 395, not only in the much greater abundance of biotite, but also in the relatively large size (up to 5 mm. across) of the biotite flakes, in the presence of epidote and calcite, and in the absence of feldspar. As in the analysed mica-schist, apatite is common as an accessory. This biotite-skarn derived from quartzite actually resembles the biotite-skarn developed from the epidiorite of the lower sill more closely than it resembles the mica-schist. Indeed, it resembles the former not only in its essential minerals, but down to such detail as the presence of small granules of sphene within the biotite. It differs from it, however, in the presence of muscovite, in the relative paucity of epidote, and in its highly schistose structure.

GARNET-SKARN

In microscope section many of the specimens of quartzite collected from the upper and lower contacts of both sills are found to be considerably enriched in garnet, which has in turn been commonly changed in part to biotite. In No. 91 (Fig. 19) from the lower contact of the main sill (locality A), garnets, measuring .5 to 1 mm. across, form about 30 per cent of the rock. No. 318, from one of the small islets immediately above the main sill is also very rich in garnets; whilst No. 342, collected slightly

Fig. 18. Biotite-skarn (No. 313) developed from the quartzite at the upper contact of the main epidiorite sill. The field shows the association of biotite, calcite (left centre of field) and quartz. Ordinary light, x 33.

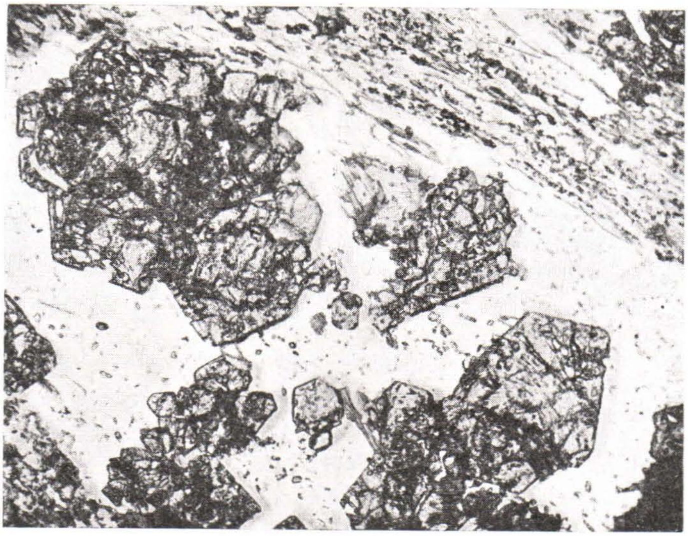

Fig. 19. Garnet-skarn (No. 91) developed from the quartzite at the lower contact of the main sill, locality A. Ordinary light, x 33.

farther away from the same contact, still contains numerous small garnets; and No. 344, collected about 4 or 5 yards further away from the main sill, is distinctly less rich in garnets, being intermediate, in this respect, between the quartzite of the contact zone and the normal quartzite of the area.

EPIDOTE-SKARN

In addition to enrichment in garnet, some parts of the quartzite in contact with the sills contain epidote, a mineral not observed in the quartzite away from the contact. The quartzite, No. 379, immediately above the narrow sill at locality C resembles the normal quartzite of the area, as exemplified by the analysed specimen, No. 394, except that it is free from calcite, contains epidote, is richer in biotite and considerably richer in garnet. The garnets have a highly crystalloblastic habit, and may be quite lace-like in their development on account of their highly irregular boundaries, and the presence of abundant inclusions of quartz and albite. Some of these inclusions can be seen to be optically continuous with grains of the corresponding mineral occurring outside the boundaries of the garnets. From such evidence it is clear that the garnets are of late development. The epidote, occurring as small rounded individuals, is abundant, and is distributed fairly evenly throughout the rock. The quartzite, No. 384, collected directly below the narrow sill at locality C resembles the rock just described, except that the biotite is partially converted to ripidolite. Veinlets of ripidolite, of replacement origin, also cut the schistosity at a high angle. No. 390, from the base of the main sill, is also rich in epidote, in addition to biotite and garnet.

Even more advanced stages of enrichment of the quartzite in garnet and epidote are found in inclusions of quartzite occurring within the sills. Such an inclusion, No. 320 from the main sill, is composed of garnet, quartz, biotite, epidote and alkali feldspar, with iron-ore as a common accessory. Garnet, biotite and epidote jointly form about 50 per cent of this rock, garnet being exceedingly abundant. This rock, except for the presence of epidote, is very like the garnet-skarn described by von Eckermann (1922, p. 306 and Plate XXIV, Fig. 41) as occurring in a zone between amphibolite and crystalline schist. An inclusion of quartzite, No. 385 from the lower sill, is converted to epidote-skarn composed essentially of epidote and quartz with some alkali feldspar, greenish-brown biotite and bluish green amphibole.

Although no chemical analyses have been made of the skarn-rocks developed from quartzite, it is clear that the abundance of garnet and epidote implies local increase of CaO and Al_2O_3. These constituents need not have migrated far, since they could have been supplied partly from CaO in the original and neighbouring quartzite and partly as CaO and

Al_2O_3 liberated during the transformation of epidiorite into skarn. The biotite-skarn which replaces quartzite is essentially an intensification of the quartzite to mica-schist transformation, combined with slight development of epidote.

VI. DISCUSSION OF RESULTS

(A) QUARTZITE TO MICA-SCHIST

As indicated in Table X, the metasomatic metamorphism of quartzite to mica-schist involved increase of

Al_2O_3, total FeO, MgO, K_2O, H_2O, TiO_2, P_2O_5 and MnO

and decrease of

SiO_2, $CaCO_3$, CaO and Na_2O.

A chemically similar metasomatism has been described by Billings (1937) from Loon Mountain, New Hampshire. In that area argillaceous sediments were recrystallised during regional metamorphism into sillimanite-biotite-oligoclase-schist. Subsequently the sillimanite porphyroblasts were propressively replaced by aggregates of muscovite, and at the same time biotite increased in amount while quartz and oligoclase decreased. The compositions of typical examples of these two rocks are given for comparison in Table X. The chemical changes involved increase of

Al_2O_3, total FeO, MgO, K_2O, H_2O, TiO_2 and MnO

and decrease of

SiO_2 (CaO) and Na_2O.

These changes are qualitatively identical with those in the quartzite to mica-schist example, apart from the loss of $CaCO_3$ from quartzite and the gain of P_2O_5 by the mica-schist.

Billings suggests that »moving solutions introduced a great deal of potash into the rock, altering the sillimanite to muscovite». As a working hypothesis he considers certain masses of quartz-monzonite and granite (members of the so-called New Hampshire »Magma» Series exposed several miles away) as the most probable sources for the »potash-rich solutions». These plutonic rocks contain muscovite, and it is thought that juvenile water released during the final stages of crystallisation hydrolysed initial orthoclase to muscovite and so liberated K_2O and six times as much SiO_2, thus providing «potash-rich solutions» that migrated through the surrounding country rocks and locally effected the metasomatism described.

If this were the real explanation, a similar or genetically related metasomatic metamorphism should have transformed the country rocks even more conspicuously in the immediate neighbourhood of the plutonic

Table X

GAINS (+) AND LOSSES (—) IN EXAMPLES OF METASOMATIC METAMORPHISM

	DONEGAL			*NEW HAMPSHIRE*		
	Quartzite No. 394	*Mica-schist No. 395*		*Sillimanite-schist No. 2*	*Muscovitised schist No. 3*	
SiO_2	72.06	65.36	—	66.68	57.43	—
Al_2O_3	10.92	15.68	+	18.17	21.68	+
Fe_2O_3	.04	1.49	+	.90	.08	+
FeO	1.40	4.32		4.98	6.82	
MgO	.31	1.79	+	1.42	2.02	+
CaO	6.02	.83	—	.76	.72	
Na_2O	2.93	2.16	—	.74	.53	—
K_2O	1.16	4.54	+	2.89	6.32	+
H_2O+	.84	2.32	+	1.75	2.80	+
$H_2O—$	.26	.34		.40	.20	
CO_2	3.98	none	—	none	none	
TiO_2	.08	.86	+	.87	.94	+
P_2O_5	none	.18	+	tr.	tr.	
MnO	.07	.21	+	tr.	.28	+
	100.07	100.08		99.90	99.82	

$\left.\begin{array}{l}\text{No. 394}\\ \text{No. 395}\end{array}\right\}$ See Table IX. *Analyst:* W. H. Herdsman.

No. 2. Sillimanite-schist. 0.2 mile N. W. of summit of South Peak of Loon Mountain, New Hampshire, U. S. A. (Billings, 1937, Table I, No. 2, p. 292). Total includes S = .05; C = .29.
Analyst: W. H. Herdsman.

No. 3. Muscovitised schist. Summit of South Peak of Loon Mountain, N. H. (Billings, 1937, Table I, No. 3, p. 292). *Analyst:* W. H. Herdsman.

masses. Moreover, addition of SiO_2 to the country rocks should have been a more noticeable feature than increase of muscovite and biotite. Neither inference is matched by the field evidence described from the Loon Mountain region. Certainly no such explanation is tenable for our Donegal example. The granite exposed at Portronan, south of Malin Head, is a potassic granite; yet at its contact against quartzite the latter has gained Na_2O and lost FeO and MgO (Reynolds, 1946, p. 415): changes which are the reverse of those found in the transformation of quartzite into mica-schist. Moreover, where the same granite is in contact with epidiorite there is no sign of the conspicuous biotitisation that has affected the White Cow epidiorite two miles away. The geochemical migrations under discussion cannot be related to hydrothermal solutions or other »emanations» liberated during the emplacement of the Malin Head granite. Moreover, this granite is post-tectonic and appears to be of later age than the metamorphic changes described in this paper.

The assemblage of introduced constituents is, however, one that can be correlated with part of the stream of geochemical migration responsible for granitisation and its attendant phenomena. Two relevant examples may be given; one syntectonic and regional, the other post-tectonic and local.

The albite-schists of the Dalradian Series in Scotland and Antrim bear witness to streams of migrating materials that were driven ahead from regions characterised by the development of migmatites (Reynolds, 1942). Chemical analyses of correlated specimens show that the introduction of Na and Si into the schists (with formation of albite porphyroblasts) led to the driving out of

Al, Fe, Mg, Ca, K, H [1] and Ti.

These constituents must have migrated beyond the region of albitisation and so have enriched in biotite the rocks in which they became fixed.

The general similarity of this assemblage to that required to account for the formation of mica-schist from quartzite is obvious. As already mentioned (p. 29), albite-schist appears in force near Dunaldragh, north-west of the White Cow Rock, and it is therefore possible that in the latter locality we see the fixation of constituents driven ahead from a region of albitisation further back.

In Goraghwood Quarry (Newry Complex, Co. Armagh) irregular but completely isolated bodies (up to 2 or 3 feet across) of granitic rocks, including trondhjemite, have developed metasomatically within a thick band of hornfelsed Silurian sediments (Reynolds, 1943). In every case these bodies cut abruptly across the bedding of the hornfels, without disturbance of the normal dip and strike, and small residual patches of hornfels, also in undisturbed position, commonly occur within them. Wherever the granitic replacement bodies are in contact with hornfels they are bounded by dark rims up to about half an inch thick, in which the enclosing hornfels (or the included hornfels) is conspicuously enriched in biotite. Each rim is a »basic front», representing the material driven out of the hornfels during its transformation into a granitic rock. Chemical analyses of carefully correlated specimens (Reynolds, 1943) show that the development of trondhjemite from biotite-hornfels involved fixation of incoming

Na, Ca and Si

and the driving out of

Al, Fe, Mg, K, H, Ti, P and Mn.

Similarly, it is found that the development of granodiorite, adamellite and granite pegmatite involved the driving out of

Fe, Mg, Na, H, Ti, P and Mn.

Analysis of one of the dark rims shows that the elements introduced were

Al, Fe, Mg, Na, K, H, Ti, P and Mn.

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[1]) See footnote on page 59.

These constituents (with the exception of some or all of the soda) are precisely those that migrated from hornfels during the granitisation.

Here again, we have the same assemblage of constituents as that required for the quartzite to mica-schist transformation, and, moreover, one that can be correlated with an indubitable example of granitisation in the solid state. In Goraghwood the expelled constituents were fixed immediately around the granitic bodies, from which it can be inferred that the gradients controlling migration were steep, as might be expected on the margin of a post-tectonic complex. The migrations involved in the metasomatic metamorphism of the Malin Head quartzite (and presumably in that of the Loon Mountain schists), though involving the same constituents, were of the far-travelling type characteristic of migmatite fields and the surrounding zones of regional metamorphism.

(B) EPIDIORITE TO SKARN

The epidiorite of the main sill of the White Cow Rock has been enriched in biotite near its contacts with the country rocks. Biotite-epidiorite has been changed more drastically in the lower sill: (1) to epidote-quartz-biotite-skarn, characterised by a greenish brown biotite; and subsequently either to (2 a) epidote-quartz-lepidomelane-skarn, characterised by a golden biotite; or to (2 b) epidote-quartz-ripidolite-skarn, with or without a mineral that is tentatively identified as iso-orthoclase. The successive alterations involved considerable changes in chemical composition, the nature and sequence of which can be readily followed by reference to Table XI. It should be noticed that the complete sequence is as follows:

Epidiorite → Biotite-epidiorite → Biotite-skarn ↗ Lepidomelane-skarn. No. 381
No. 94 No. 350 No. 92 ↘ Chlorite-skarn. No. 353

The last two items may be in part simultaneous, with the change to chlorite-skarn probably continuing a little longer than the change to lepidomelane-skarn. The final sequence of minerals in the closing stages of alteration was (a) chlorite, (b) hastingsite and (c) calcite. Because of these late changes and the resultant fluctuations in the serial variation of some of the constituents, the gains and losses in this series are necessarily more complex than in the relatively simple change from quartzite to mica-schist.

Taking the series as a whole there has been a decrease of

$$Al_2O_3,\ CaO \text{ and } Na_2O.$$

SiO_2 has been gained at some stages and lost at others, the resultant effect being probably insignificant. The decrease of Al_2O_3 is progressive, except for a slight break from No. 350 to No. 92. Na_2O decreases markedly to No. 92, after which it comes in again on a minor scale, probably in accordance with the growth of late hastingsite.

Table XI

	Epidiorite No. 94	*Biotite-Epidiorite No. 350*	*Biotite-skarn No. 92*	*Lepidomelane-skarn No. 381*	*Chlorite-skarn No. 353*
SiO_2	47.32	48.34	45.65	48.13	44.98
Al_2O_3	17.39	15.49	15.56	14.76	14.44
Fe_2O_3	2.36	2.14	2.55	7.34	3.02
FeO	10.91	10.60	11.21	6.38	10.98
(Total FeO)	(13.04)	(12.53)	(13.52)	(12.99)	(13.69)
MgO	4.78	5.56	5.01	6.61	6.08
CaO	10.08	9.42	7.94	6.18	8.24
Na_2O	2.73	1.98	.01	.43	.16
K_2O	.54	1.36	4.77	3.59	1.23
H_2O+	.96	1.54	2.35	2.82	5.19
H_2O-	.05	.16	.10	.39	.18
CO_2	tr.	none	2.02	none	1.74
TiO_2	2.68	2.48	2.30	2.58	2.92
P_2O_5	tr.	.32	.30	.22	.29
MnO	.20	.34	.23	.32	.35
	100.01	99.73	100.02	99.75	99.80

No. 94. Epidiorite, Main Sill (Table I). Total includes BaO = .01.
Analyst: Agnes Gibbs.

No. 350. Biotite-epidiorite, Lower Sill, locality B (Table II).
Analyst: W. H. Herdsman.

No. 92. Epidote-quartz-biotite-skarn, Lower Sill, locality A (Table III).
Total includes BaO = .02; NiO = tr.; F none found.
Analyst: Agnes Gibbs.

No. 381. Epidote-quartz-lepidomelane-skarn, Lower Sill, locality C (Table V).
Analyst: W. H. Herdsman.

No. 353. Epidote-quartz-ripidolite-skarn, Lower Sill, locality C (Table VII).
Analyst: W. H. Herdsman.

The constituents that have increased most conspicuously are K_2O and H_2O. K_2O shows a marked culmination in No. 92, after which it is partly displaced by H_2O, particularly in the change to chlorite-skarn. H_2O rises continuously throughout the series and so does the sum of $H_2O + K_2O$. MgO increases, but not progressively. Total FeO fluctuates, but probably increases slightly on the whole. MgO and total FeO behave antipathetically, one rising as the other falls, except in the change to chlorite-skarn in which they increase together, with MgO rising more than total FeO. The sum of MgO + total Fe increases progressively throughout the series. TiO_2 decreases to No. 92 and then increases again. P_2O_5 and MnO both increase, though not progressively. Taking the series as a whole there has been increase of

MgO, (total FeO + MgO), K_2O, H_2O, P_2O_5 and MnO.

This assemblage has a close family resemblance to that concerned in the quartzite to mica-schist transformation (page 53). The differences suggest that Al and most of the Fe and Ti were used up by the time the sills were reached. In any case, the assemblage is that of a typical basic front. It is evident that the changes in both quartzite and epidiorite are closely related, and that they have resulted from a stream of migrating materials. The constituents of the stream and their relative proportions would necessarily vary from place to place at any given time, and from time to time at any given place, according to the varying nature of the rocks through which they passed and the exchanges of constituents effected by reaction and fixation, as witnessed by the observed evidence of metasomatism.

(c) RELATION TO BASIC FRONTS

It has long been known from field observations that the regional metamorphism connected with major orogenies involved a chemical as well as a mineral and structural transformation of the rocks concerned. Associated with the development of migmatites in depth there is a concomitant basification of the sedimentary rocks, mainly in the peripheral zone. The basaltic rocks (the metabasites), because they are more resistant to granitisation, not only become basified in the peripheral zone, but also remain, as basified relics, within the region of migmatisation itself. The enrichment of the initial rocks in femic or cafemic materials is the expression of a basic front that proceeded in advance of one of migmatisation. It is evidenced by the conversion of limestone to tremolite-skarn, amphibolites to cummingtonite-amphibolites, and leptites to cordierite-anthophyllite rocks (Eskola, 1914); the enrichment of pelitic rocks in cordierite, garnet and biotite (Wegmann, 1935); the conversion of limestones to amphibolite (Backlund, 1936 a); and the progressive conversion of basaltic rocks to epidiorites, amphibolites and eclogites (Backlund, 1936 b), and at any of these stages to skarn (Backlund, 1943).

Recently these field observations, pointing to change of composition with change of metamorphic grade, have received corroboration from a geochemical study by Lapadu-Hargues (1945) of rocks that were originally sedimentary. By statistical comparison of the chemical composition of rocks of different metamorphic grade, irrespective of their age, Lapadu-Hargues has demonstrated that Fe^{++} and Mg attain their maximum concentration in the lowest metamorphic grades, and decrease through the higher grades. Ca increases gradually with increase in metamorphic grade. Total alkalis increase, but in the lower grades of metamorphism Na shows increase and K a slight decrease, whilst in the higher grades (granitisation) Na and K both increase, K attaining its greatest concentration in the final granites. Lapadu-Hargues ascribes the variation in the concentrations of the various elements, in rocks of different regional metamorphic grade, to the difference in mobility of the respective elements under metamorphic conditions. He finds the order of increasing mobility to be K, Ca, Na, Mg, Fe, and he correlates this order with the respective ionic radii of the elements concerned. Fe and Mg, with the smallest ionic radii, are the most mobile ions and, in consequence, they become concentrated in the lowest metamorphic grades, that is in the least altered rocks situated farthest from loci of granitisation.

According to Lapadu-Hargues' statistical study, K, with the largest ionic radius of the elements under discussion, has the lowest mobility, in consequence of which it becomes concentrated in granites representing the highest degree of regional metamorphism. Study of actual examples, however, both in the field and geochemically, shows that this is not the whole story relating to the concentration of K. Not only does K show geochemical culmination in granite, but it also migrates with the vanguard of Fe and Mg, and appears in the biotite of the basic fronts. On a small scale this is exemplified by the biotite-rich rims margining the granitic veins and pod-like bodies of migmatite regions; on a larger scale by the biotite-enriched metamorphic aureoles that commonly surround granitic complexes emplaced within pelitic rocks; and on a still greater scale by the biotite enrichment that appears in advance of regions of albitisation. A telescoped, but particularly instructive example, of the presence of K in a basic front that advanced into limestone is depicted on Plate LXVIII of von Eckermann's (1922) memoir on Mansjö Mountain. The figure illustrates the altered marginal zone of limestone in contact with granite pegmatite. Within this zone of alteration several mineral zones are present. Travelling outwards from the pegmatite to the calcite of the limestone the mineral zones appear in the following sequence: plagioclase, scapolite, diopside, apatite and phlogopite. In company with Si, the various groups of constituents of the basic front have migrated into the limestone with the following increasing order of mobility: Na-Ca; Ca-Mg; Ca-P and Mg-Fe-K-H-Al.

The metasomatic metamorphism of quartzite and epidiorite described in this paper involved introduction of the same group of constituents, including Mg, Fe, K, H [1] and Al, that moved with the vanguard of the basic front in the Mansjö limestone example. It is, moreover, a general phenomenon that K is not only concentrated in granitic rocks, but also migrates with the advancing basic front. A recent study (Reynolds, 1946) of the chemical data relating to the alteration of pelitic and psammitic sediments and basic igneous rocks (occurring in the aureoles surrounding granite masses, and also as inclusions within the granite masses them-

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1) For a discussion of the question whether H^+ or $(OH)^-$ is introduced, see Bugge (1946, pp. 36 and 39). In our example the development of chlorite after biotite suggests introduction of (OH) rather than of H.

selves) has revealed the fact that in every example for which adequate chemical data exist, the emplacement of granite bodies has been accompanied by the introduction of one or more of the cafemic constituents (generally all) and of the alkalis (commonly K) into both aureoles and inclusions. With the further advance of true granitisation these initially basified rocks themselves become granitised. There is a demonstrable sequence both in space and time of (1) basification followed by (2) granitization.

Zones of granite formation and zones of basification are thus complementary phenomena, and the identification of the alterations of quartzite and epidiorite at Malin Head as a basic front implies the coeval development of granite. The basic front of the Malin head area cannot be correlated with the local granite of Portronan, for this post-tectonic granite has caused no such biotitisation and skarn development within the adjoining epidiorites. It is, therefore, more likely that the basic front under discussion will eventually be found to be correlatable with the development of syntectonic migmatites such as occur in the Rosguill peninsula, farther west. The continuation of this migmatite zone into the Malin Head district probably extends underground.

(D) MECHANISM OF MIGRATION

The replacements that have been described in this paper have evidently involved the passage through solid rocks of a stream of interchanging constituents. These materials must either have been transported by and through a gas or liquid phase, or they must have migrated by solid diffusion along crystal boundaries and through crystal lattices. Small-scale solid diffusion is manifestly necessary for the accomplishment of any mineral replacement that requires penetration of crystal lattices by ions. The essential functions of gaseous or liquid solutions at any particular spot are to bring up the incoming ions and maintain their concentration, and to carry away the displaced ions. As between liquid or hydrothermal solutions and gaseous or pneumatolytic solutions, it seems probable that these functions could be carried out more effectively by the former (Graton, 1940, p. 287; see also, Ross, 1935, pp. 46 and 60).

A gas phase would have the advantage of easy penetration along submicroscopic cracks through otherwise impermeable rock, but against this it suffers from several disadvantages. Considerations of relative volatility suggest that a gas phase is most likely to be acid. In the replacement phenomena under discussion, however, the evidence is overwhelmingly suggestive of the dominance of alkaline materials. Moreover, components of widely different volatility would need to be combined in the migrating stream. Graton (1940, pp. 282—4) has reviewed the evidence showing that the vapour pressures of Fe, Mg, Ca, Si and Al are anything from 10^2 to 10^{22} times less than those of K and Na. Against this point, however, Ingerson and Morey (1940, pp. 779—780) cite experimental observations indicating that diffusion through gas is quantitatively a more effective process than transport by gas. However this may be, there is the third difficulty that the degree of ionisation of a concentrated gaseous solution under high pressure and therefore of relatively high density is likely to be extremely low. A gas phase is therefore thought to be a feeble carrier of the ions necessary to initiate solid diffusion through the crystals with which it comes into contact. Crystal linings along the surfaces of cracks and drusy cavities (rather than metasomatic replacements) are the characteristic results to be expected from gaseous transport and deposition.

On the other hand, hydrothermal solutions, if of magmatic ancestry, are likely to be strongly alkaline from the start, as a result of hydrolysis of silicate minerals. They are also likely to have a higher degree of ionisation, though it must not be overlooked that such scanty evidence as is available suggests that ionisation attains its maximum at a little above 200° C and that at higher temperatures it begins to decrease again (Ingerson and Morey, 1940, p. 783). In respect of both composition and degree of ionisation hydrothermal solutions appear to have a definite advantage over gaseous solutions, especially in their capacity for initiating and maintaining metasomatic replacements of appropriate parageneses. But except where adequate passageways are available, e. g. by way of open cracks and fissures, their capacity for penetrating rocks is severely limited. Silica-bearing solutions in particular seem to have a very small power of soaking through rocks, like quartzite, in which the only passageways are the minute interstices between closely fitting mineral grains (Ross, 1933, p. 193).

When the conditions are such that these interstices are isolated and discontinuous, molar transport becomes impossible, and replacement can then be effected only by means of solid diffusion. While tight-packing brings solutions to a standstill and excludes their introduction from outside sources, it favours solid diffusion by ensuring close contact between adjacent crystals. Graton (1940, p. 287) has expressed his surprise »that there has been such widespread neglect of diffusion in geological processes». It is indeed surprising when it is remembered that Greenly (1903) long ago suggested solid diffusion as a process likely to be of importance in the transformation of schists into the rocks we now call migmatites. Only in recent years has there been a systematic exploration of the petrogenetic possibilities of solid diffusion.

Applying the above and other relevant considerations to the example of metasomatic metamorphism under discussion, it can be said that the few criteria available for discrimination suggest that solid diffusion was probably the dominant process concerned.

(a) The intergranular passageways were tight and almost certainly discontinuous, since the rocks had already been recrystallised during an earlier stage of regional metamorphism.

(b) The distribution of the alterations shows that no preferential use was made of more continuous passageways such as may have been afforded by bedding planes and joints. Hydrothermal solutions are demonstrably responsible for potash metasomatism in the rocks underlying the geyser basins of Yellowstone Park. A bore-hole core investigated by Fenner (1936, pp. 262-5) contained an unreplaced »island» of pelitic dacite, and it is clear from adjoining parts of the core that the first alteration proceeded along the cracks, as would naturally be expected. In Donegal, however, the structures betray no sign of any selective attack either along cracks or around crystal boundaries *(cf.* Bugge, 1946, p. 34). Reference to Figs. 10 and 11 shows clearly that cracks were avoided during the alteration from biotite-skarn (»solid» rock) to lepidomelane-skarn (»crumbly» rock). This criterion has also been used by Perrin and Roubault (1941) in connection with a metamorphic front described from the Alps of Savoy. There, a Triassic conglomerate of quartzite pebbles in a sandy matrix appears to rest unconformably on sericite-chlorite-schist. However, they find residual »islands» of the conglomerate and its pebbles remaining within the schist, and »veins» and patches of the schist developed within the conglomerate. Before dying out, the metamorphic front barely penetrated the conglomerate, which evidently constituted a barrier to the further advance of the migrating materials. Perrin and Roubault point out that the propagation of solid diffusion and of the resultant reactions in the solid state (to which they attribute regional metamorphism) is dependent on intimate contact within the affected medium. A conglomerate ,with relatively poor cohesion, would be readily permeable by molar solutions, but would form a barrier against the further advance of solid diffusion.

(c) Returning to Yellowstone Park, which is our standard of reference for hydrothermal metamorphism, it is noteworthy that Fenner (1936, p. 240) observes that »the phenocrysts were little attacked, but the groundmass has been almost wholly metasomatized». In contrast, we find that in our Donegal epidiorites the small matrix crystals, which provide a high surface area, resisted alteration more effectively than the larger crystals, and were the last to be transformed.

(d) In accord with (c) it is found that the contacts (both macroscopic and microscopic) between biotite-skarn and residual biotite-epidiorite are sharp *(e.g.* at locality B, page 33). So are those between biotite-skarn and chlorite-skarn (Fig. 10) and so, again, are those between mica-schist and residual »islands» of quartzite (Fig. 15). In all these examples each rock is definitely of one type or another; there are no gradational stages of the veining or fingering kind that would be expected to result from permeation by liquid or gas.

(e) The demonstrable time-sequence of basification followed by granitization (page 60) has an important bearing on the problem of the mechanism of migration. If an attempt is made to correlate the advance of the basic front with the work of hydrothermal solutions liberated from a consolidating granite magma, the physico-chemical difficulties involved in the alternating displacements and fixations of constituents with highly contrasted properties become fantastically unmanageable. Moreover, the hydrothermal solutions would have to be supplied not only where the supposed granite magma was in contact with the surrounding country rocks, but also around every inclusion within the magma itself. It is impossible in a limited space to discuss the matter systematically, but to us it seems manifest that the whole process is much more elegantly explained by progressive solid diffusion due to migration of ions through the crystal lattices and along the crystal boundaries.

(f) Finally, it may be repeated that the Malin Head region provides no exposures of any rock from which a magmatic source for the migrating materials could be inferred. And, above all, it must be emphasized that, even if there were evidence of co-existing magma, it would still remain possible that the magma was not the source of the metasomatising materials, but was itself a consequence of change of composition and increase of temperature brought about by ionic migrations and exothermic reactions. Recent advances in the study of granitization phenomena and regional metamorphism point to the serious necessity for considering:

(i) whether magma was ever present at all; and, if the evidence is favourable to such an interpretation,
(ii) whether the magma may not have been generated by an intensification of processes akin to those responsible for the changes witnessed by the surrounding rocks.

The minds of most petrologists have been so conditioned by training and practice that »granite» inevitably suggests »granite magma». This psychological tyranny should be consciously realised and deliberately resisted. If in any given occurrence, the magmatic hypothesis is found to be supported by evidence, it should most certainly be adopted. If it is not so supported, then a critical examination of the relevant criteria is likely to lead to further discovery. To avoid any possible misconception as to our own attitude, it should perhaps be added that we are in no way prejudiced against magma in its proper place, but only against facile and unrealised assumptions.

VII. ACKNOWLEDGMENTS

Before her untimely death in 1939 Lady Gibbs collaborated with us in investigating the rocks of the Malin Head area by making a series of chemical analyses, two of which appear in this paper. We wish to place on record our great indebtedness to our late friend for this generous contribution to petrology. The cost of the other new chemical analyses was defrayed by a grant from the Moray Endowment Research Fund of the University of Edinburgh. The award of this grant is gratefully acknowledged.

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4.

THE SYSTEM ANORTHITE—LEUCITE—SILICA

BY

J. F. SCHAIRER and N. L. BOWEN

GEOPHYSICAL LABORATORY, CARNEGIE INSTITUTION OF WASHINGTON,
WASHINGTON, D. C.

INTRODUCTION

The crystalline phases of the system anorthite—leucite—silica are of notable importance as rock-forming minerals, both individually and in the assemblages which they may form at equilibrium. As apart of our general program of equilibrium studies designed to throw light on the genetic relations of minerals, the present study therefore finds a natural place.

The feldspars are undoubtedly the most important rock-forming minerals. Two representatives of the group, lime feldspar and potash feldspar, are among the phases whose relations this investigation seeks to unravel. Potash feldspar is prominent in a great variety of igneous and metamorphic rocks. In Finland such rocks are widespread and Finnish geologists have contributed much to our knowledge of the natural history of potash feldspars. The association of potash feldspar with free silica, one of the phase assemblages of our system, is freely represented in the potash-rich granites and pegmatites of Finland. Among the studies of these rocks those of Pentti Eskola are outstanding, as are also his investigations of the stability relations of potash feldspar in metamorphic rocks.[1]

The association anorthite—silica, another of our phase assemblages and therefore chemically an entirely possible association in rocks, is, nevertheless, not found in igneous types for genetic reasons that have been set down elsewhere.[2] In metamorphic rocks, on the other hand, genetic controls over bulk composition are altogether different, and in these rocks Eskola has studied some interesting examples of the anorthite—quartz association into which he was able to read a broad geological, even a cosmogonic significance.[3]

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[1] P. Eskola, On the Occurrence of Orthoclase and Microcline in the Finnish Granites and Pegmatites. Comptes rendus de la Société géologique de Finlande No. 1, p. 1. 1929.
P. Eskola, The Mineral Facies of Rocks. Norsk Geologisk Tidskrift VI, pp. 143—194. 1920.

[2] N. L. Bowen, The Evolution of the Igneous Rocks. Princeton University Press, 1928, p. 20.

[3] P. Eskola, Conditions during the Earliest Geological Times. Annales Academiae Scientiarum Fennicae, Series A, Vol. 36, No. 4, p. 22. 1932.

The association lime plagioclase—potash feldspar, one of our assemblages, cannot fail to call to mind that remarkable relation between potash feldspar and plagioclase that is displayed in the Rapakivi granites, even although their plagioclase is of very different composition. Brought face to face with the problems of rapakivi in their common homeland, Eskola has made contributions towards the solution of these problems[1].

For these and other reasons it seems not inappropriate that a report on the system embracing these several crystal phases should appear in a volume honoring Pentti Eskola.

Among the ancient rocks of the shield areas of the earth, leucite-bearing types are not represented. In some shield areas recent lavas have been erupted. Central Africa is an example of such an area and some of the lavas there found are leucitic. No such activity has occurred in Finland, but if it had we may rest assured that Eskola would have brought to bear on the problems of leucitic rocks the same powers of painstaking investigation that he has exercised in all his undertakings, and that our understanding of these rocks would also have been greatly increased.

PREPARATION OF COMPOSITIONS

In the preparation of silicate compositions containing K_2O, special care must be exercised on account of the volatilization of K_2O and consequent departure from the desired composition. A series of K_2O—SiO_2 glasses of varying K_2O content was prepared by melting $KHCO_3$ and purified quartz in platinum crucibles. The loss of K_2O was minimized by first sintering at low temperatures (around 700° C.) and then gradually raising the temperature to about 1400° C. The small loss (usually less than 0.0100 g. on a 20.0000 g. sample) was determined by weighing and corrected by adding the appropriate amount of $KHCO_3$ before subsequent fusions. At least three fusions with intermediate crushing of the quenched glasses were necessary to obtain a homogeneous product. A series of glasses between anorthite and silica in composition was prepared by melting together specially purified $CaCO_3$, Al_2O_3 and quartz.

The binary leucite—anorthite compositions and the ternary compositions were prepared by fusing together in platinum crucibles one of the several K_2O—SiO_2 glasses, specially purified Al_2O_3 and anorthite or an anorthite—silica glass. Depending on the composition and consequent viscosity of the glasses, at least three fusions, and sometimes as many as eight, with intermediate crushing of the glasses, were necessary to effect complete solution of the Al_2O_3 and secure a homogeneous glass. Too much

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[1] P. Eskola, On Rapakivi Rocks from the Bottom of the Gulf of Bothnia. Fennia 50, No. 27, p. 1. 1928.

emphasis cannot be placed on the importance of obtaining a homogeneous glass before starting equilibrium studies. The homogeneity of each of the glasses was carefully checked under the petrographic microscope by obtaining a uniform index of refraction of the glass grains. Each of the several compositions studied was crystallized at appropriate temperatures before proceeding with quenching experiments. Only in compositions with less than ten per cent anorthite and with liquidus temperatures below 1200° C. were difficulties encountered in obtaining some crystallization. It was possible to effect complete crystallization only in melts in the binary systems anorthite—leucite and anorthite—silica.

EQUILIBRIUM STUDIES

Investigation of the melting relations was made on twelve binary anorthite—silica compositions, eleven binary anorthite—leucite compositions and on sixty-nine ternary compositions. These studies were made by the method of quenching,[1] which consists in holding a small charge of known composition at a measured temperature for a period of time adequate for attainment of equilibrium between the several phases, and then chilling instantly to room temperature to »freeze» the equilibrium. Quenched charges are examined with a petrographic microscope to determine the nature and number of crystalline and liquid phases. Crystals are identified by their optical properties. Liquids quench to a glass.

In order to hasten the attainment of equilibrium between crystals and liquid, special precautions were taken to have only small crystals present in any mixture before making the quenching experiments. For those compositions near the boundary curves leucite—anorthite and anorthite—tridymite, care was taken to have both kinds of crystals present in the initial charge. It was not possible to obtain crystals of potash feldspar in those viscous compositions in the anorthite field near the boundary curve anorthite—potash feldspar.

Equilibrium between crystals and liquid is reached in a few hours at temperatures near and above 1500° C., and in a few days in all compositions at temperatures between about 1350° and 1500° C. Below 1350° C., equilibrium between leucite crystals and liquid and between anorthite crystals and liquid is attained only after a period of one to two weeks.

In many silicate systems, cristobalite may crystallize and persist metastably for long periods of time at temperatures within the stability range of tridymite. In the binary system anorthite—silica and in the tridymite field in this ternary system, only cristobalite was formed and it persisted for more than a month, except in that portion of the tridymite field

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[1] E. S. Shepherd, G. A. Rankin, and F. E. Wright, Amer. Jour. Sci., Vol. 28, p. 308. 1909.

adjacent to the leucite—silica side line where the liquidus temperature was above 1250° C. Former studies showed that, in the system diopside—leucite—silica,[1] presumably because of the less viscous melts, tridymite always appeared in its range of stability and grew well-developed thin hexagonal plates, flattened parallel to the base (0001).

Quenching experiments were conducted in platinum-wound electric furnaces whose temperature was controlled (± 2° C.) with a temperature regulator—the Geophysical Laboratory furnace thermostat[2] — somewhat modified from the original design but identical in principle.

The temperature was measured with a thermocouple placed nearly in contact with (about 1 mm. distant from) the small charge wrapped in a tiny platinum envelope, which is suspended in that part of the furnace determined as the »hot point.» In a well insulated furnace there is a range of about 5 mm. on either side of the »hot point» where the temperature falls off only about one-half degree. The small platinum envelope containing the charge hangs entirely within this zone.

The Pt—Pt 90 Rh 10 thermoelement was calibrated frequently at several fixed points defined in degrees Centigrade as follows:

Palladium melting point	1549.5°
Diopside melting point	1391.5°
Gold melting point	1062.6°

THE BINARY SYSTEMS

The ternary system anorthite—leucite—silica is bounded by the limiting binary systems anorthite—silica, leucite—anorthite, and leucite—silica There are no binary systems within the ternary system.

The System Anorthite-Silica: This system was studied by Rankin and Wright[3] in their work on the system $CaO—Al_2O_3—SiO_2$. They found a eutectic between anorthite and silica at 1359° C. at the composition anorthite 52.5 % silica 47.5 %. Three additional compositions between anorthite and silica were studied by Andersen,[4] who places the eutectic between anorthite and tridymite at 1353° C. at the composition anorthite 52 % silica 48 %. Andersen called attention to the great viscosity of the melts and difficulties in crystallization. Because our ternary data (given later) did not tie in well with these values for the eutectic temperature and eutectic composition, we prepared twelve compositions between anorthite and silica and made a new investigation of the binary system, allowing much longer periods of time for the attainment of equilibrium between crystals and liquid. The new data[5] for this system are given in Table I and presented graphically as Fig. 1. We obtained the value 1368 ± 2 for

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[1] J. F. Schairer and N. L. Bowen, Amer. Jour. Sci., Vol. 35 A, p. 295. 1938.

[2] H. S. Roberts, Jour. Opt. Soc. Amer., Vol. 11, pp. 171—186. 1925.

[3] G. A. Rankin and F. E. Wright, Amer. Jour. Sci., Vol. 39, pp. 1—79. 1915.

[4] O. Andersen, Amer. Jour. Sci., Vol. 39, pp. 407—454. 1915.

[5] Schairer used these data in Fig. 7, p. 252, of his paper on the system $CaO—FeO—Al_2O_3—SiO_2$ (Jour. Amer. Ceram. Soc., Vol. 25,pp. 241—274. 1942.)

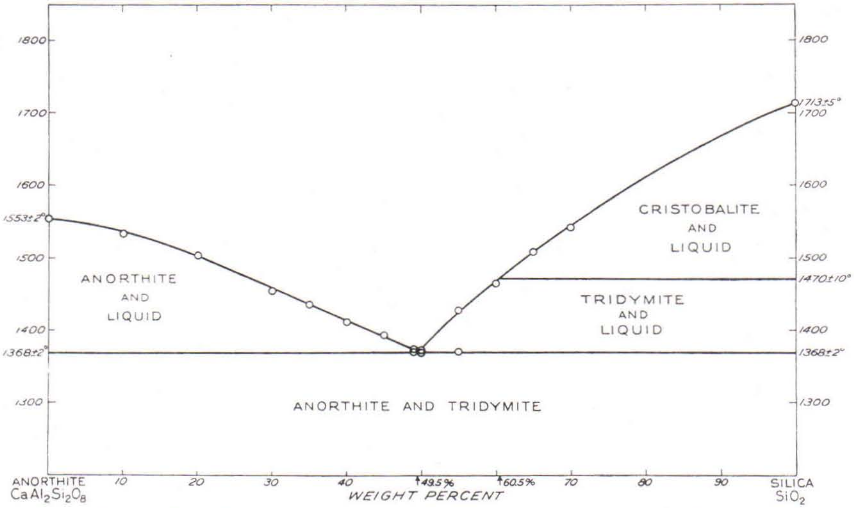

Fig. 1. Equilibrium diagram of the binary system anorthite—silica.

Table I

Quenching Data for the System Anorthite—Silica

Refractive index of glass ±.003	Composition in wt. per cent		Temp. °C.	Time	Phases
	Anorthite	Silica			
1.566	90	10	1530	7 hrs.	Small amount anorthite in glass
			1534	7 hrs.	All glass
1.554	80	20	1500	7 hrs.	Rare anorthite in glass
			1504	7 hrs.	All glass
1.542	70	30	1450	24 hrs.	Small amount anorthite in glass
			1455	24 hrs.	All glass
1.536	65	35	1432	24 hrs.	Rare anorthite in glass
			1435	24 hrs.	All glass
1.529	60	40	1410	2 days	Very rare anorthite in glass
			1415	2 days	All glass
1.524	55	45	1390	2 days	Rare anorthite in glass
			1394	22 hrs.	All glass
1.518	51	49	1363	7 days	Anorthite and cristobalite* (all crystalline)
			1368	7 days	Anorthite and cristobalite* in glass
			1371	14 days	Rare anorthite in glass
			1375	3 days	All glass
1.517	50	50	1363	7 days	Anorthite and cristobalite* (all crystalline)
			1368	7 days	Anorthite and cristobalite* in glass
			1369	7 days	Rare cristobalite* in glass
			1371	14 days	All glass
1.511	45	55	1366	21 days	Anorthite and cristobalite* (all crystalline)
			1370	7 days	Cristobalite* in glass
			1426	24 hrs.	Very rare cristobalite* in glass
			1430	24 hrs.	All glass
1.504	40	60	1460	24 hrs.	Small amount cristobalite* in glass
			1466	24 hrs.	All glass
1.499	35	65	1503	7 hrs.	Small amount cristobalite in glass
			1508	7 hrs.	All glass
1.493	30	70	1540	7 hrs.	Very rare cristobalite in glass
			1545	7 hrs.	All glass

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* In many silicate systems, particularly in viscous melts, cristobalite crystallizes and persists for long periods of time at temperatures within the stability range for tridymite and only the metastable cristobalite equilibrium is obtained.

the eutectic temperature and place the eutectic composition at anorthite 50.5 % silica 49.5 %. Attention is called to the fact that only cristobalite crystals were obtained in the binary system, even with long runs in the temperature stability range of tridymite. Only in less viscous melts in the ternary system (described later) was it possible to realize the stable tridymite liquidus. There can, however, be but little difference in temperature and composition between the stable tridymite—anorthite eutectic and the metastable cristobalite—anorthite eutectic we have measured. The melting point of cristobalite was determined by Greig[1] as 1713 ± 5° and that of anorthite by Osborn[2] as 1553 ± 2° C.

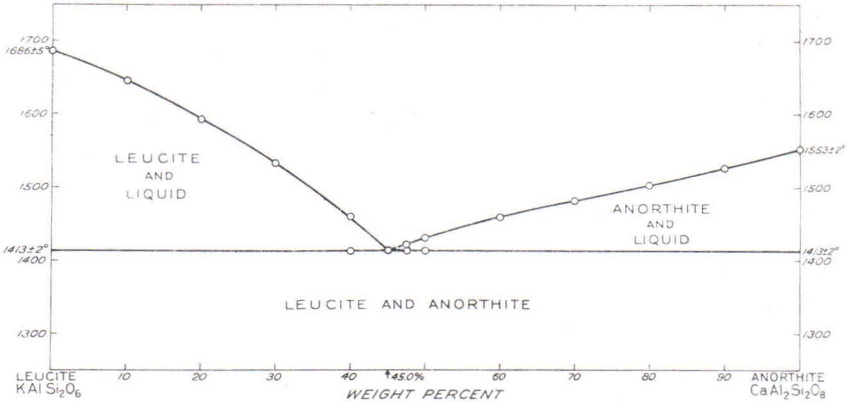

Fig. 2. Equilibrium of the binary system leucite—anorthite.

The System Leucite—Anorthite: Since no data were available on this system, we prepared[3] eleven compositions between leucite and anorthite. The quenching data are given in Table II and presented graphically as Fig. 2. There is no evidence of solid solution, and the eutectic between anorthite and leucite lies at 1413 ± 2° C. and at the composition anorthite 45 % leucite 55 %. Bowen and Schairer[4] determined the congruent melting point of leucite at 1686 ± 5° C.

The System Leucite—Silica: Morey and Bowen[5] showed the incongruent nature of the melting of potash feldspar ($KAlSi_3O_8$) to leucite and liquid and presented a partly hypothetical diagram for the system leucite—silica. As a result of the very great viscosity of compositions between leucite and silica, particularly at temperatures below about 1250° C., direct completion of this system presents almost insurmountable experimental difficulties. The best data for this system have been obtained from quench-

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[1] J. W. Greig, Amer. Jour. Sci., Vol. 13, pp. 7—12. 1927.

[2] E. F. Osborn, Amer. Jour. Sci., Vol. 240, p. 781. 1942.

[3] The authors wish to acknowledge with thanks the assistance of Mr. L. E. J. Brouwer in preparing some of these binary compositions and carrying out some of the quenching experiments.

[4] N. L. Bowen and J. F. Schairer, Amer. Jour. Sci., Vol. 18, pp. 305—306. 1929.

[5] G. W. Morey and N. L. Bowen, Amer. Jour. Sci., Vol. 4, pp. 1—21. 1922.

Table II

Quenching Data for the System Leucite—Anorthite

Refractive index of glass ±.003	Composition in wt. per cent		Temp. °C.	Time	Phases
	Leucite	Anorthite			
1.500	90	10	1642	½ hr.	Small amount leucite in glass
			1647	½ hr.	All glass
1.507	80	20	1590	½ hr.	Rare leucite in glass
			1595	½ hr.	All glass
1.513	70	30	1534	7 hrs.	Very rare leucite in glass
			1540	7 hrs.	All glass
1.521	60	40	1410	6 days	Leucite and anorthite with very small amount glass
			1415	6 days	Leucite in glass
			1460	24 hrs.	Very rare leucite in glass
			1464	24 hrs.	All glass
1.526	55	45	1403	7 days	Leucite and anorthite (all crystalline)
			1410	3 days	Leucite and anorthite with very small amount glass
			1415	3 days	All glass
1.527	52.5	47.5	1410	3 days	Leucite and anorthite with very small (amount) glass
			1415	3 days	Anorthite in glass
			1421	4 days	Very rare anorthite in glass
			1426	7 days	All glass
1.530	50	50	1410	3 days	Leucite and anorthite with small amount glass
			1415	3 days	Anorthite in glass
			1429	3 days	Rare anorthite in glass
			1432	24 hrs.	All glass
1.535	40	60	1455	24 hrs.	Rare anorthite in glass
			1460	24 hrs.	All glass
1.545	30	70	1478	7 hrs.	Moderate amount anorthite in glass
			1485	7 hrs.	All glass
1.556	20	80	1500	7 hrs.	Small amount anorthite in glass
			1504	7 hrs.	All glass
1.565	10	90	1523	7 hrs.	Moderate amount anorthite in glass
			1529	7 hrs.	All glass

ing experiments on those compositions rich in leucite and those rich in silica. Many intermediate compositions were prepared but no satisfactory data have been secured from them. It is nearly if not quite impossible to crystallize them dry. Crystals can easily be obtained by hydrothermal methods, and the product can then be used for the dry quenching experiments. Even then it is impossible to attain or closely approach equilibrium between crystals and liquid in any reasonable length of time. Around their liquidus temperatures, particles of powdered glass of some of these compositions fail to flow together in periods of several weeks duration.

Fortunately, it is possible to obtain information on these highly viscous intermediate compositions of the binary system indirectly by a short extrapolation from data on the very numerous compositions studied by Schairer and Bowen in the ternary system $K_2O—Al_2O_3—SiO_2$. In a previous paper,[1] Schairer and Bowen presented a preliminary diagram for the binary system leucite—silica. On the basis of many new data on the system K_2O—Al_2O_3—SiO_2, particularly on compositions closer to the binary system leucite—silica, a more accurate diagram is now available. This diagram is given here as Fig. 3. Schairer and Bowen [2] have recently published a summary of the data on the ternary system K_2O—Al_2O_3—SiO_2 giving only the final ternary diagrams and not the detailed data on which they are based. A longer manuscript on that system is in preparation which will give diagrams of the binary systems within the ternary system (including the diagram given as Fig. 3 here) and all the data on quenching runs.

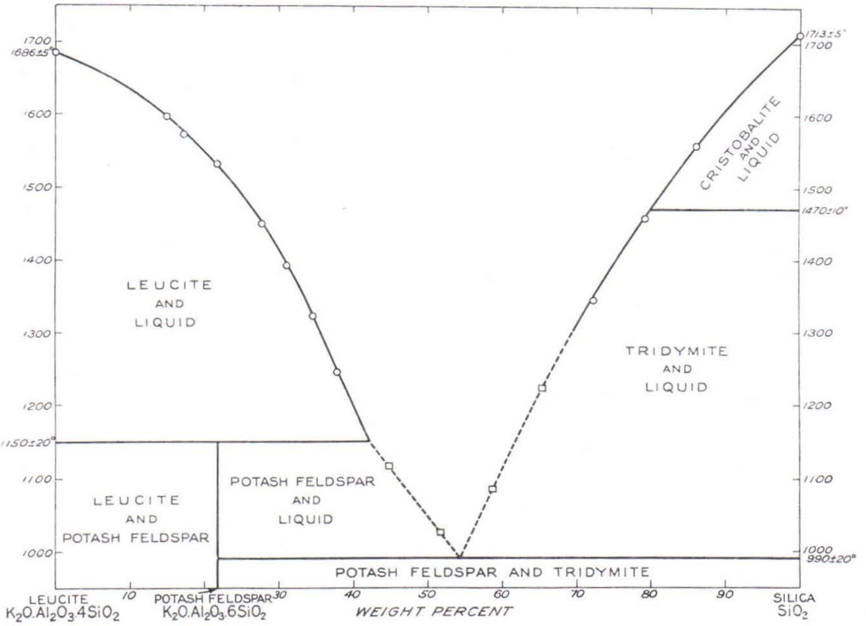

Fig. 3. Equilibrium diagram of the binary system leucite—silica. Broken lines indicate the portion obtained by a short extrapolation from data in the system K_2O—Al_2O_3—SiO_2.

From an inspection of Fig. 3 it may be seen that the incongruent melting temperature of potash feldspar is placed at 1150 ± 20° C. and the composition of the liquid formed is leucite 57.8 % silica 42.2 % (or potash feldspar 73.7 % silica 26.3 %). Morey and Bowen determined the incongruent melting temperature as about 1170° C. but could obtain no accurate composition for the liquid. We have placed the eutectic temperature between potash feldspar and tridymite at 990 ± 20° C. at the composition

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[1] J. F. Schairer and N. L. Bowen, Amer. Jour. Sci., Vol. 35 A, p. 293 (Fig. 3). 1938.

[2] J. F. Schairer and N. L. Bowen, Amer. Jour. Sci., Vol. 245, pp. 193—204. 1947.

leucite 45.6 % silica 54.4 % (or potash feldspar 58.2 % silica 41.8 %). The portions of the liquidus curves of Fig. 3 given as broken lines indicate the part of the diagram based on data from the system $K_2O—Al_2O_3—SiO_2$.

THE TERNARY SYSTEM

In order to determine the liquidus temperatures, the slopes of the liquidus surface and the fields of stability of the primary crystalline phases with their boundary curves, quenching experiments were made on sixty-nine different ternary compositions. These data are assembled in Table III and presented graphically as Fig. 4. All compositions are expressed in weight per cent. Accurately determined boundary curves are indicated by heavy lines. Broken lines indicate some uncertainty as to their precise location. Isotherms are given as lighter lines with the temperature indicated on each. The triangle representing the whole system is divided into two smaller triangles by the tie line joining anorthite with potash feldspar. Dots represent the compositions actually studied, and double open circles

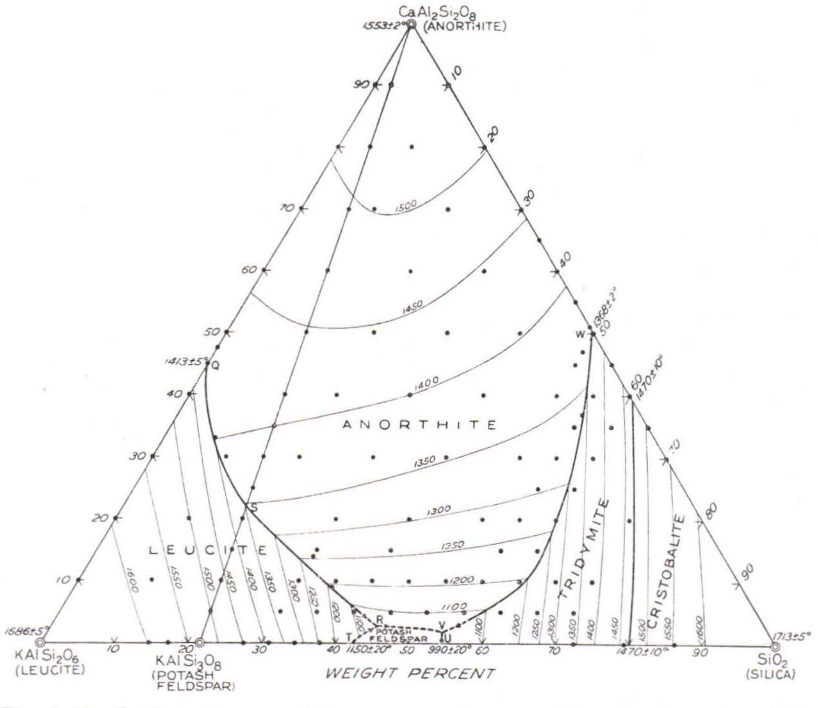

Fig. 4. Equilibrium diagram of the ternary system anorthite—leucite—silica. This supersedes a preliminary diagram given by Bowen: 1937, Amer. Jour. Sci., Vol. 33, Fig. 3, p. 7. 1937.

Table III

Results of Quenching Experiments for the System Anorthite—Leucite—Silica

Refractive index of glass ±.003	Composition in wt. per cent			Temp. °C.	Time	Phases
	Leucite	Anorthite	Silica			
						Points in the Anorthite Field
1.565	7.8	90.0*	2.2	1530	7 hrs.	Moderate amount anorthite in glass
				1534	7 hrs.	All glass
1.554	15.7	80.0*	4.3	1516	7 hrs.	Rare anorthite in glass
				1520	7 hrs.	All glass
1.554	10	80	10	1520	24 hrs.	Small amount anorthite in glass
				1525	24 hrs.	All glass
1.545	23.5	70.0*	6.5	1493	24 hrs.	Small amount anorthite in glass
				1496	7 hrs.	All glass
1.544	20	70	10	1500	4 days	Small amount anorthite in glass
				1504	3 days	All glass
1.543	10	70	20	1490	24 hrs.	Rare anorthite in glass
				1493	3 days	All glass
1.536	31.4	60.0*	8.6	1468	24 hrs.	Small amount anorthite in glass
				1472	24 hrs.	All glass
1.532	10	60	30	1444	3 days	Moderate amount anorthite in glass
				1450	2 days	All glass
1.529	39.2	50.0*	10.8	1445	2 days	Small amount anorthite in glass
				1450	2 days	All glass
1.527	30	50	20	1445	2 days	Small amount anorthite in glass
				1449	2 days	All glass
1.524	20	50	30	1425	24 hrs.	Small amount anorthite in glass
				1430	24 hrs.	All glass
1.520	10	50	40	1400	3 days	Rare anorthite in glass
				1405	24 hrs.	All glass
1.514	3	47	50	1367	4 days	Rare anorthite in glass
				1371	8 hrs.	All glass
1.513	5	45	50	1367	4 days	Rare anorthite in glass
				1371	8 hrs.	All glass
1.521	47	40*	13	1410	3 days	Rare anorthite in glass
				1415	3 days	All glass
1.519	40	40	20	1415	3 days	Small amount anorthite in glass
				1420	24 hrs.	All glass
1.516	30	40	30	1397	2 days	Small amount anorthite in glass
				1401	24 hrs.	All glass
1.513	20	40	40	1378	8 hrs.	Rare anorthite in glass
				1382	7 days	All glass
1.509	10	40	50	1355	8 days	Small amount anorthite in glass
				1360	7 days	All glass
1.505	8	37	55	1335	6 days	Anorthite and tridymite in glass
				1340	7 days	Very rare anorthite in glass
				1345	24 hrs.	All glass
1.516	51	35*	14	1395	8 hrs.	Rare anorthite in glass
				1400	3 hrs.	All glass
1.503	10	35	55	1340	7 days	Rare anorthite in glass
				1345	24 hrs.	All glass
1.515	60	33	7	1395	7 days	Anorthite and rare leucite in glass
				1399	10 days	Rare anorthite in glass
				1403	7 days	All glass
1.513	60	30	10	1385	14 days	Rare anorthite in glass
				1391	14 days	All glass
1.512	54.9	30.0*	15.1	1345	14 days	Anorthite and rare leucite in glass
				1349	14 days	Anorthite in glass

Refractive index of glass ±.003	Composition in wt. per cent			Temp. °C.	Time	Phases
	Leucite	Anorthite	Silica			
				1374	7 days	Small amount anorthite in glass
				1378	8 hrs.	All glass
1.511	50	30	20	1374	7 days	Small amount anorthite in glass
				1378	8 hrs.	All glass
1.508	40	30	30	1367	4 days	Very rare anorthite in glass
				1371	8 hrs.	All glass
1.503	30	30	40	1350	7 days	Rare anorthite in glass
				1355	8 days	All glass
1.500	20	30	50	1330	7 days	Rare anorthite in glass
				1335	7 days	All glass
1.497	15	30	55	1320	7 days	Small amount anorthite in glass
				1325	3 days	All glass
1.507	58.8	25.0*	16.2	1345	14 days	Anorthite and rare leucite in glass
				1349	14 days	Anorthite in glass
				1360	14 days	Rare anorthite in glass
				1362	14 days	All glass
1.494	20	25	55	1303	7 days	Rare anorthite in glass
				1305	24 hrs.	All glass
1.498	50	20	30	1312	6 days	Very rare anorthite in glass
				1315	7 days	All glass
1.494	40	20	40	1300	7 days	Rare anorthite in glass
				1303	7 days	All glass
1.491	30	20	50	1280	4 days	Small amount anorthite in glass
				1285	24 hrs.	All glass
1.489	25	20	55	1280	7 days	Rare anorthite in glass
				1285	6 hrs.	All glass
1.487	20	20	60	1260	14 days	Anorthite and rare cristobalite** in glass
				1265	14 days	Rare anorthite in glass
				1270	10 days	All glass
1.494	55	15	30	1270	10 days	Small amount anorthite in glass
				1275	3 days	All glass
1.489	45	15	40	1265	14 days	Very rare anorthite in glass
				1270	4 days	All glass
1.485	30	15	55	1235	14 days	Rare anorthite in glass
				1240	7 days	All glass
1.484	25	15	60	1240	7 days	Rare anorthite in glass
				1245	7 days	All glass
1.493	56	14	30	1240	7 days	Anorthite and rare leucite in glass
				1245	7 days	Anorthite in glass
				1250	14 days	Small amount anorthite in glass
				1255	14 days	All glass
1.488	55	10	35	1211	14 days	Very rare anorthite in glass
				1215	14 days	All glass
1.486	50	10	40	1205	14 days	Rare anorthite in glass
				1208	14 days	All glass
1.485	45	10	45	1205	14 days	Rare anorthite in glass
				1208	14 days	All glass
1.483	40	10	50	1205	14 days	Rare anorthite in glass
				1208	14 days	All glass
1.481	35	10	55	1195	14 days	Rare anorthite in glass
				1200	14 days	All glass
1.479	30	10	60	1185	14 days	Anorthite and rare tridymite in glass
				1190	14 days	Anorthite in glass
				1195	14 days	Rare anorthite in glass
				1200	14 days	All glass
1.481	50	5	45	1095	28 days	Rare anorthite in glass
				1100	28 days	All glass
1.479	45	5	50	1100	28 days	Rare anorthite in glass
				1105	28 days	All glass
1.478	40	5	55	1095	28 days	Small amount anorthite in glass
				1100	28 days	All glass

Refractive index of glass ±.003	Composition in wt. per cent			Temp. °C.	Time	Phases
	Leucite	Anorthite	Silica			
						Points in the Leucite Field
1.503	70	20	10	1371	14 days	Leucite and rare anorthite in glass
				1375	14 days	Leucite in glass
				1504	7 hrs.	Rare leucite in glass
				1508	7 hrs.	All glass
1.501	62.7	20.0*	17.3	1345	14 days	Leucite and very rare anorthite in glass
				1350	5 days	Leucite in glass
				1366	14 days	Rare leucite in glass
				1371	14 days	All glass
1.497	66.6	15.0*	18.4	1312	7 days	Leucite and very rare anorthite in glass
				1317	7 days	Leucite in glass
				1405	14 days	Very rare leucite in glass
				1410	14 days	All glass
1.496	80.0	10	10	1575	12 hrs.	Very rare leucite in glass
				1580	8 hrs.	All glass
1.494	70.6	10.0*	19.4	1450	2 days	Rare leucite in glass
				1453	2 days	All glass
1.489	60	10	30	1281	14 days	Small amount leucite in glass
				1285	14 days	All glass
1.487	56	9	35	1185	14 days	Leucite and small amount anorthite in glass
				1190	14 days	Rare leucite in glass
				1195	14 days	All glass
1.490	74.5	5.0*	20.5	1492	24 hrs.	Rare leucite in glass
				1500	4 days	All glass
1.486	65	5	30	1335	7 days	Rare leucite in glass
				1340	7 days	All glass
1.485	60	5	35	1240	14 days	Rare leucite in glass
				1245	14 days	All glass
1.483	55	5	40	1075	21 days	Leucite and small amount anorthite in glass
				1080	14 days	Leucite in glass
				1085	20 days	Rare leucite in glass
				1090	14 days	All glass
						Points in the Tridymite Field
1.507	5	40	55	1351	5 days	Cristobalite** and very rare anorthitein glass
				1355	7 days	Cristobalite** in glass
				1367	7 days	Very rare cristobalite** in glass
				1372	7 days	All glass
1.501	5	35	60	1415	3 days	Small amount cristobalite** in glass
				1421	6 days	All glass
1.496	10	30	60	1317	7 days	Cristobalite** and rare anorthite in glas
				1323	10 days	Cristobalite** in glass
				1367	7 days	Very rare cristobalite** in glass
				1372	7 days	All glass
1.492	15	25	60	1295	14 days	Cristobalite** and rare anorthite in glass
				1300	6 days	Cristobalite** in glass
				1320	7 days	Rare cristobalite** in glass
				1325	7 days	All glass
1.485	10	20	70	1455	2 days	Very rare tridymite in glass
				1459	2 days	All glass
1.478	25	10	65	1285	10 days	Very rare tridymite in glass
				1290	2 days	All glass
1.477	20	10	70	1377	7 days	Very rare tridymite in glass
				1382	7 days	All glass
1.476	35	5	60	1150	14 days	Rare cristobalite** in glass
				1158	14 days	All glass
1.474	30	5	65	1235	14 days	Small amount cristobalite** in glass
				1240	14 days	All glass
1.473	25	5	70	1330	7 days	Very rare tridymite in glass
				1335	7 days	All glass

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* Compositions marked with an asterisk in this table lie in the potash feldspar—anorthite join.

** In many silicate systems, particularly in viscous melts, cristobalite crystallizes and persists for long periods at temperatures within the stability range for tridymite and only the metastable cristobalite equilibrium is obtained.

the compositions of compounds. Arrows on the boundary curves indicate the direction of falling temperature.

A marked feature of the ternary system is the extension of the field of anorthite from pure anorthite itself nearly down to the binary bounding system leucite—silica. Even with somewhat less than 5 per cent of anorthite in the total composition of a melt, anorthite may be the first crystalline phase to appear on cooling. For nearly all compositions in this system, anorthite is removed, on cooling, almost quantitatively (with some leucite or some silica also for appropriate compositions), leaving a residual liquid which is largely a mixture of potash feldspar and silica in composition.

The field of potash feldspar is quite small in the ternary system. It has been severely restricted by the crowding down of the anorthite field (to within less than 5 % by weight of that component) towards the leucite—silica side line and the extension of the leucite and tridymite fields towards each other in compositions near the side line leucite—silica. As has been previously noted, compositions in this region of low temperatures are exceedingly viscous,[1] nearly if not quite impossible to crystallize in dry melts, and so slow in reaching equilibrium as to be impractical experimentally. The temperature of the ternary reaction point must be lower than that of the binary reaction point *T* of Fig. 4 (1150 ± 20°) in the system leucite—silica and indeed must be lower than 1078° C., a determined temperature on the boundary curve leucite—anorthite a short distance from the ternary reaction point. It is probably about 1040 ± 20° C. The temperature of the ternary eutectic (*V* of Fig. 4) must be lower than that of the binary eutectic *U* of Fig. 4 (990 ± 20°) in the system leucite—silica, but it cannot be much lower. It is probably about 950 ± 20° C.

The optical properties of anorthite,[2] leucite,[3] tridymite,[4] and cristobalite agreed with those reported in previous papers from this Laboratory. If there were, however, a limited solid solubility of potash feldspar in anorthite, we would have been unable to detect the small change in optical properties on the tiny laths of anorthite obtained in melts in the anorthite field near the side line leucite-silica. No zoning of crystals or other evidence of solid solution was encountered in these studies.

THE JOIN POTASH FELDSPAR—ANORTHITE

This join is a tie line in the ternary system and in it lie all compositions between potash feldspar and anorthite. It is binary only in part because of the appearance of leucite crystals whose composition does not lie in this

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[1] For data on viscosities of silicate glasses, see N. L. BOWEN. Trans. Amer. Geophys. Union, 15 th. Annual Meeting, 249—255. 1934.

[2] H. E. MERWIN, cited by N. L. Bowen. Amer. Jour. Sci., Vol. 33, p. 564. 1912.

[3] N. L. BOWEN and J. F. SCHAIRER, Amer. Jour. Sci., Vol. 18, pp. 308—309. 1929.

[4] C. N. FENNER, Amer. Jour. Sci., Vol. 36, pp. 351—356. 1913.

join. There is a consequent departure of the composition of liquids from this line when leucite is present as a solid phase. The data on compositions between potash feldspar and anorthite (marked with an asterisk in Table III) are plotted in Fig. 5. Heavy curves refer to binary equilibrium and light curves to ternary equilibrium. All compositions between potash feldspar and anorthite become completely crystalline on cooling at the same temperature, 1040 ± 20°, which is the temperature of the ternary reaction point *R* of Fig. 4 or, conversely, on heating begin to melt at this temperature. The crystallization of feldspar mixtures will be described later in more detail, under crystallization.

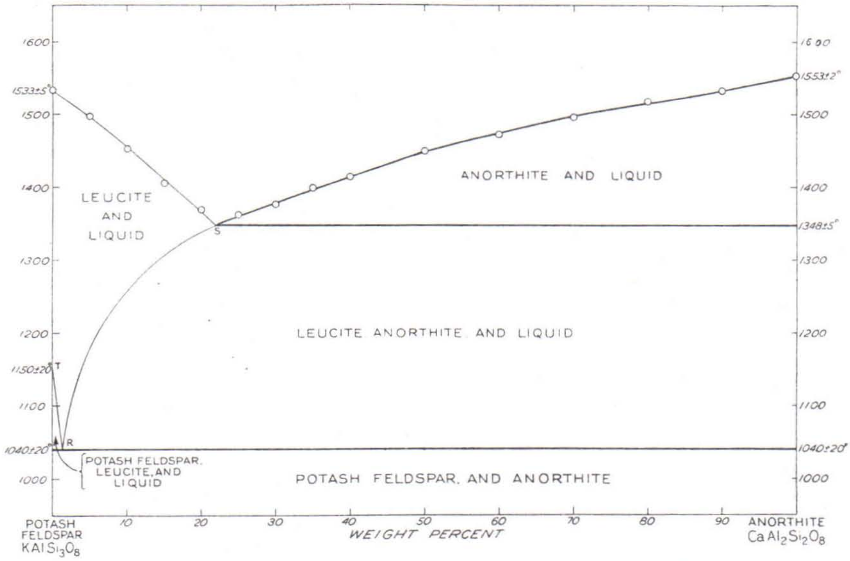

Fig. 5. Equilibrium diagram of the join potash feldspar—anorthite. Heavy curves refer to binary equilibrium and light curves to ternary equilibrium.

INDICES OF REFRACTION OF GLASSES

Homogeneous glasses were prepared for the sixty-nine ternary compositions studied, and the index of refraction of each glass was measured at 25° C. with the petrographic microscope by comparison with standardized immersion liquids. The values obtained are given in the first column of Table III. Similar data were obtained on twelve binary compositions in the system anorthite—silica and eleven binary compositions in the system leucite—anorthite. These data are given in the first column of Table I and Table II. The values are accurate to ± 0.003. No attempts were made to obtain close accuracy, the glasses were not annealed, and the measurements were made in white light. The data are plotted in Fig. 6. The compositions measured are indicated by dots, and isofracts or curves of equal index of refraction are plotted in the composition triangle.

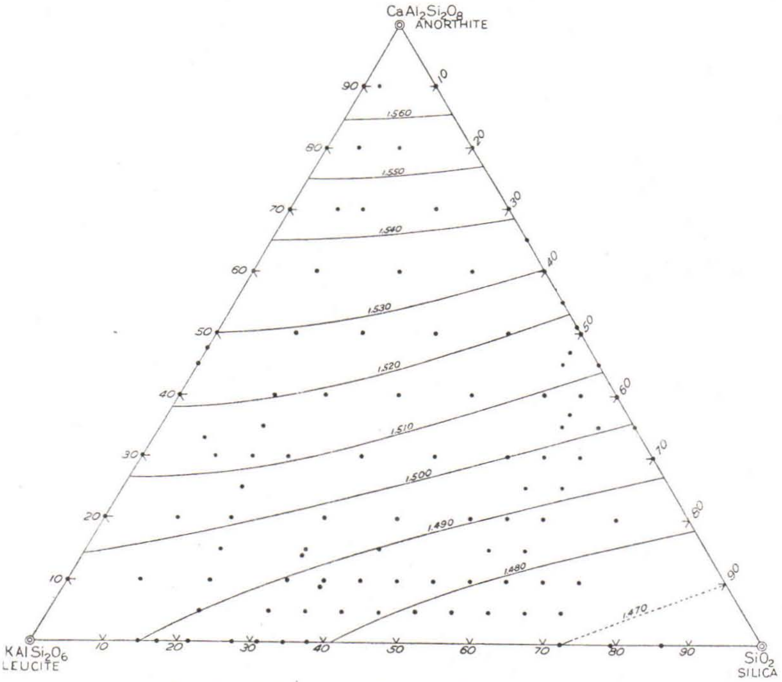

Fig. 6. Refractive indices of glasses at 25° C.

CRYSTALLIZATION IN THE TERNARY SYSTEM

The crystallization of compositions within the ternary system anorthite—leucite—silica (Fig. 4) may be conveniently discussed by considering those compositions in the area leucite—potash feldspar—anorthite, those in the join anorthite—potash feldspar, and those in the area potash feldspar—anorthite—silica.

Compositions within the Area Leucite—Potash Feldspar—Anorthite (Fig. 4). At perfect equilibrium, all compositions within the triangle leucite—potash feldspar—anorthite, when completely crystalline, consist of leucite, potash feldspar, and anorthite. For all these compositions melting begins at the temperature of the ternary reaction point (1040 ± 20°). Some of the compositions in this triangle lie in the anorthite field and the remainder of them in the leucite field: —

(1) Those compositions in the anorthite field, on cooling from a completely molten state, first crystallize anorthite. On further cooling, the composition of the liquid phase follows a path, which is a straight line

joining the total composition and anorthite, until it reaches the boundary curve anorthite—leucite at some point between the binary eutectic between anorthite and leucite *Q* and the point *S* where the join potash feldspar—anorthite cuts this boundary curve. When the composition of the liquid reaches the boundary curve, a second solid phase, leucite, appears and, with falling temperature, crystals of anorthite and leucite continue to separate with the composition of the liquid following the boundary curve to the ternary reaction point *R* where the temperature remains constant until all the liquid has been exhausted by the reaction between leucite crystals and liquid to form potash feldspar. For all mixtures within the triangle leucite—potash feldspar—anorthite, the liquid is exhausted by this reaction while some leucite crystals still remain. Thus, the final completely crystalline product consists of the three solid phases leucite, potash feldspar, and anorthite. The behavior of these compositions on heating follows this process in reverse.

(2) For melts of those compositions within the triangle leucite—potash feldspar—anorthite, which lie in the leucite field, leucite appears, on cooling, as the primary phase and with the separation of this phase crystallization paths are straight lines towards either the boundary curve anorthite—leucite *QR* or leucite—potash feldspar *TR*. Those in the area above a line from leucite to *R* reach the boundary curve *QR* and precipitate anorthite and leucite while the liquid follows this boundary curve to the ternary reaction point *R*, where the temperature remains constant and the liquid is exhausted by the reaction of leucite crystals and liquid to form potash feldspar. Those liquids of composition in the area below a line from leucite to *R* reach the boundary curve leucite—potash feldspar *TR*, and leucite crystals are joined first by potash feldspar and then the liquid follows this reaction curve *TR* to the ternary reaction point *R*, where leucite and potash feldspar are joined by anorthite crystals and the liquid is exhausted by reaction at constant temperature.

Compositions in the Join Anorthite—Potash Feldspar (Figs. 4 and 5). Those compositions which lie in the join anorthite—potash feldspar are a limiting case. With complete equilibrium, all of these, on cooling, become completely solid at the temperature of the ternary reaction point *R* (1040 ± 20°), the last liquid has the composition of this point, and this last liquid and the last leucite crystals are exhausted simultaneously by reaction. With the aid of Figs. 4 and 5 we may examine the crystallization of all mixtures in the join anorthite—potash feldspar: —

(1) From all melts between anorthite and the composition anorthite 22 % potash feldspar 78 % (*S* of Figs. 4 and 5) anorthite appears, on cooling, as the primary phase, the liquid follows this join until it has reached *S* in composition at the temperature 1348 ± 5°, at which point anorthite is joined by leucite crystals. With falling temperature, both anorthite and leucite crystallize and the liquid follows the boundary curve

anorthite—leucite *SR* (Fig. 4) to the ternary reaction point *R*. At the constant temperature of this reaction point (1040 ± 20°), the liquid is exhausted by reaction to convert all the leucite to potash feldspar. Leucite crystals and liquid are exhausted simultaneously and below this temperature the completely solid mass consists of only two phases—anorthite and potash feldspar.

(2) From all melts between the compositions anorthite 22 % potash feldspar 78 % *(S* of Figs. 4 and 5) and anorthite 1.3 % potash feldspar 98.7 % *(R* of Fig. 5), leucite appears, on cooling, as the primary phase and the liquid follows a straight line (drawn through leucite and the total composition) until it reaches the boundary curve anorthite—leucite *SR* (Fig. 4) at some point between *S* and *R*. On reaching the boundary curve, anorthite appears as the second solid phase and the liquid follows the boundary curve *SR* to the ternary reaction point *R*, where, at constant temperature, leucite is converted by reaction between liquid and crystals to potash feldspar. With perfect equilibrium, the leucite crystals and liquid are both exhausted simultaneously and below this temperature the completely solid mass consists of only two phases—anorthite and potash feldspar.

(3) From that small range of melts between the compositions potash feldspar *(T* of Fig. 5) and the composition anorthite 1.3 % potash feldspar 98.7 % *(R* of Fig. 5), leucite appears, on cooling, as the primary phase and the liquid follows a straight line (drawn through leucite and the total composition) until it reaches the boundary curve leucite—potash feldspar *TR* (Figs. 4 and 5), where leucite is joined by crystals of potash feldspar. The liquid follows this boundary curve to the ternary reaction point *R* (Fig. 4), where anorthite appears as a solid phase and leucite is converted by reaction with the liquid at constant temperature (1040 ± 20°) to potash feldspar. With perfect equilibrium, leucite is completely resorbed and the last liquid becomes exhausted simultaneously. Below this temperature, the completely solid mass consists of only two phases—potash feldspar and anorthite.

Compositions in the Area Potash Feldspar—Anorthite—Silica (Fig. 4). With perfect equilibrium, all melts which lie within the triangle potash feldspar—anorthite—silica become completely crystalline, on cooling, only at the temperature of the ternary eutectic (950 ± 30°), where they consist of potash feldspar, anorthite, and tridymite.

(1) If the composition of the melt lies in the leucite field, the course of crystallization is as follows: —

(a) Compositions above a line from leucite to *R* (Fig. 4) first crystallize leucite and reach the boundary curve *SR*, where anorthite appears as the second crystalline phase. The liquid follows the boundary curve *SR* (with leucite and anorthite both separating together) to *R*, where at constant temperature *all* leucite is resorbed by reaction with liquid to form potash feldspar. After this reaction is complete, the liquid follows the boundary curve *RV* to *V*, where potash feldspar and anorthite are joined by tridymite.

(b) Compositions below a line from leucite to *R* first crystallize leucite and go to the boundary curve *TR*. Only those reach the point *R* which lie between this line and a line from potash feldspar to *R*. Those liquids which reach *R* have their leucite completely resorbed at this point and then follow the boundary curve *RV* to *V*. For those compositions below a line drawn from potash feldspar to *R*, the liquid does not reach *R* because leucite resorption along *TR* is complete at some point along this curve and the liquid then leaves the curve *TR* and crosses the potash feldspar field to either *RV* or *UV* (depending on the initial total composition) and thence to *V*.

(2) If their composition lies in one of the fields of silica (cristobalite or tridymite), after an initial separation of cristobalite and/or tridymite crystals, they approach the ternary eutectic *V* along the boundary curve tridymite—anorthite *WV* or tridymite—potash feldspar *UV*.

(3) If their composition lies in the anorthite field, after an initial separation of anorthite crystals, they may approach the ternary eutectic in three ways: —

(a) directly along the boundary curve anorthite—tridymite *WV* to *V*,

(b) directly along the boundary curve anorthite—potash feldspar *UV* to *V*,

(c) along the boundary curve anorthite—leucite between *S* and *R* to reach first the ternary reaction point *R*, where all the leucite is redissolved by reaction with liquid, at constant temperature, to form potash feldspar and then, when this reaction is completed, by following the boundary curve anorthite—potash feldspar *RV* to *V*.

So far, we have discussed the courses of crystallization with perfect equilibrium. Under such conditions, leucite and the forms of silica are incompatible. If, however, leucite crystals fail to react with liquid either partially or completely because of rapid cooling, rising of crystals in melts under control of gravity, inclusion of leucite crystals within other crystals or other similar reasons, they might be found with quartz or tridymite crystals. For a discussion of the potentialities of fractional crystallization with failure of equilibrium, the reader is referred to several previous papers from this Laboratory.[1]

PETROLOGIC CONSIDERATIONS

Of the various assemblages of our system, the anorthite—silica assemblage is found only in metamorphic rocks, as has already been noted in the introduction. The mutual melting relations as given in Fig. 1 are therefore of little significance in petrogenesis.

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[1] N. L. Bowen, J. F. Schairer, and H. W. V. Willems, Amer. Jour. Sci., Vol. 20, pp. 405—455. 1930.
E. F. Osborn and J. F. Schairer, Amer. Jour. Sci., Vol. 239, pp. 738—760. 1941.

Anorthite and leucite, on the other hand, do occur together in some igneous rocks and their melting relations may be regarded as of some importance. The rocks referred to are certain Italian lavas that have been termed leucitites.[1] Other minerals, notably pyroxene and to a lesser extent olivine, are prominent constituents, and for a better approach to the problems of these rocks it is necessary to combine the information furnished by the present investigation with that from the system leucite—diopside—silica.[2] This we expect to do in a subsequent publication.

In view of the abundance of the feldspars in the crust of the earth, the individual and the mutual relations of the three principal members of that group are of much concern to the petrologist. It is only to be expected, therefore, that these questions should have received much attention and that every resource should be brought to bear on their solution. The light shed by the relations of the feldspars displayed in rocks is the first approach. As is so often true, the evidence from rocks is subject to alternative interpretation, and several different diagrams have been drawn up to illustrate the relations between the feldspars.

Laboratory investigation of the effects of the various environmental factors may be a fruitful approach to the solution of this question. The influence of the chemical environment, in other words, of the presence of various other substances, is one aspect that is susceptible of laboratory attack. For example, a cardinal feature of the properties of one of the feldspars, namely, orthoclase, is its incongruent melting point, and the effects on this relation of addition of other substances can be investigated experimentally. It has been found that addition of diopside has substantially no effect.[3] On the other hand, the addition of water has a marked effect. At a pressure of some 2600 atmospheres, water will dissolve in orthoclase liquid to the extent of about 6 per cent, and from this liquid, orthoclase crystallizes directly. In other words, the incongruent relation is destroyed.[4]

The effects of adding other feldspars to orthoclase are naturally of great importance. It has been determined that addition of albite has the same effect as addition of water, though the potency of albite in this respect is not so great as that of water. A little more than 50 per cent albite must be present to insure that potash-rich alkali feldspar shall separate directly from the melt without previous separation of the feldspathoid leucite.[5] Now, as a result of the present investigation, we find that addition of anorthite has no such effect but is analogous to addition of diopside. It

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[1] H. S. Washington, The Roman Comagmatic Region. Carnegie Institution of Washington Publ. No. 57, pp. 128—136. 1906.

[2] J. F. Schairer and N. L. Bowen, Amer. Jour. Sci., Vol. 35—A, pp. 289—309. 1938.

[3] See ref. 24.

[4] R. W. Goranson, Amer. Jour. Sci., Vol. 35—A, p. 88. 1938.

[5] J. F. Schairer and N. L. Bowen, Trans. Amer. Geophys. Union, 16th Annual Meeting, p. 328. 1935.

is of much interest that one end member of the plagioclase series should have an effect opposite to that of the other end member. It raises the question as to where in the plagioclase series the one relation gives place to the other. Since water has such a marked effect, the position of this point may be expected to show notable dependence on the water content of the natural magmas.

We do not find evidence of the early separation of leucite from most orthoclase-rich natural magmas. In most instances the reason for this situation is, no doubt, that albite, water, and other substances, notably free silica, have been present in sufficient quantity to prevent the formation of leucite. In some magmas, however, there is reason to believe that leucite has formed and has later been resorbed. This is the normal fate of leucite that separates in excess of its stoichiometric proportion, as has been shown in the discussion of courses of crystallization on an earlier page. There is also reason to believe that as a result of local accumulation of leucite, a composition may be produced which renders possible the formation of rocks in which the other common feldspathoid, nepheline, is developed. The details of this development and the pseudo-leucite reaction effect whereby it is rendered possible have been discussed elsewhere.[1]

In spite of the complications introduced into the situation by virtue of the incongruent melting of orthoclase, the results of the present investigation render certain facts clear regarding the solid-solution relations between potash feldspar and anorthite. The suggestion had been made that the relation, stripped of its complications, was essentially that of Roozeboom's Type IV double series of solid solutions.[2] If this relation obtained, the point *R* of Fig. 4 would have a higher temperature than the point *T*, and likewise the point Y would be higher than *U*. Although we do not know the exact temperatures of *R* and *V*, it is definitely established that they are lower than *T* and *U* respectively. The suggestion referred to must, therefore, be rejected and other causes must be sought for the development of rock types that were believed to owe their origin to this assumed relation.

The positions of the points *R* and *V* and their relations to the surrounding fields make it clear that the potash feldspar in equilibrium with liquid *R* must have less than 2 per cent anorthite in solid solution and the potash feldspar in equilibrium with *V* must have less than 3 per cent anorthite.[3] These findings are in agreement with the very small anorthite content of natural potash feldspar.[4] Our observations do not enable us to make any

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[1] N. L. Bowen, The Evolution of the Igneous Rocks, pp. 244—257.

[2] N. L. Bowen, The Evolution of the Igneous Rocks, p. 242.

[3] Theoretical considerations show that the point indicating the composition of the potassic feldspar in equilibrium with anorthite, leucite, and liquid *R* must lie below a line joining *R* and leucite, and likewise that the composition of the potassic feldspar in equilibrium with anorthite, tridymite, and liquid *V* must lie below the line from SiO_2 to *V* produced.

[4] E. Mäkinen, Über die Alkalifeldspäte. Geol. För. Förh., Vol. 39, Pt. 2, pp. 164—184. 1917.

statement regarding the amount of potash feldspar taken up in the anorthite crystals. Moreover, our results throw no light on the question of the relation between orthoclase and microcline.

From the manner in which the potash feldspar field is confined to compositions in the immediate vicinity of the leucite—silica side of the diagram, it is clear that all compositions in the whole system will have purged themselves almost completely of anorthite before any crystallization of potash feldspar can occur. If all plagioclases were like anorthite in this respect, such a relation between the feldspars as that shown in rapakivis could presumably never develop. We know, however, that the mutual relations between albite and potash feldspar are altogether different and the association of a soda-rich plagioclase with potash feldspar can therefore exhibit a different paragenesis.

5.

INDICATIONS OF MOVEMENTS OF THE EARTH-CRUST ALONG THE COAST OF NEWFOUNDLAND-LABRADOR

BY

E. H. KRANCK

The boundaries between oceans and continents have been studied mainly from two different points of view. The first starts from the sculpturing of the ocean coasts by the action of waves and by the subaereal erosion in connection with the crustal movements; the second from the geophysical conditions at the edge of the continental blocks, the difference in composition and density between them and the ocean bottom. The problem of the limitation of the continents has for a long time been dominated by the discussions concerning the theories of the continental drift.

The first of these points of view lays stress on the accidental character of the coast-lines; the second on the permanency of the continental blocks.

In the following some observations from the northern edge of the Canadian shield will be presented, which elucidate the connection between the geological structure and the origin and evolution of the continental boarder. With the exception of tectonical investigations carried out mainly along the coasts of Pacific type this side of the problem has been less often studied.

The coast of Labrador has been described both from a geological and from a geographical point of view by several authors. A very complete list of earlier publications concerning this subject is contained in V. Tanner's elaborate book on Newfoundland Labrador (8). The coast-section in question has been studied in the field by Coleman (2), Daly (3), Kindle (6), Tanner (10) and the present author (7). The aereal pictures of Forbes have given important information concerning the physiography of the region (4, 5).

PHYSIOGRAPHY OF THE COAST-SECTION

The middle section of the rocky coast of Newfoundland-Labrador between Domino run and Hopedale can be characterized as a skiergard-coast with a comparatively high relief, reminding of Bohuslän in Western Sweden. The pre-Cambrian surface, sculptured by the glaciers of the continental ice-cap, is generally along the coast very rugged, with numerous rounded hills and small valleys originated from joints and fractures. Farther from the coast, (generally not more than some miles), the coast-mountains attain heights of at an average about 400 feet, exceptionally over 1 000 (Mealy Mountains). A continuous coastal plane or »strandflade» do not occur.

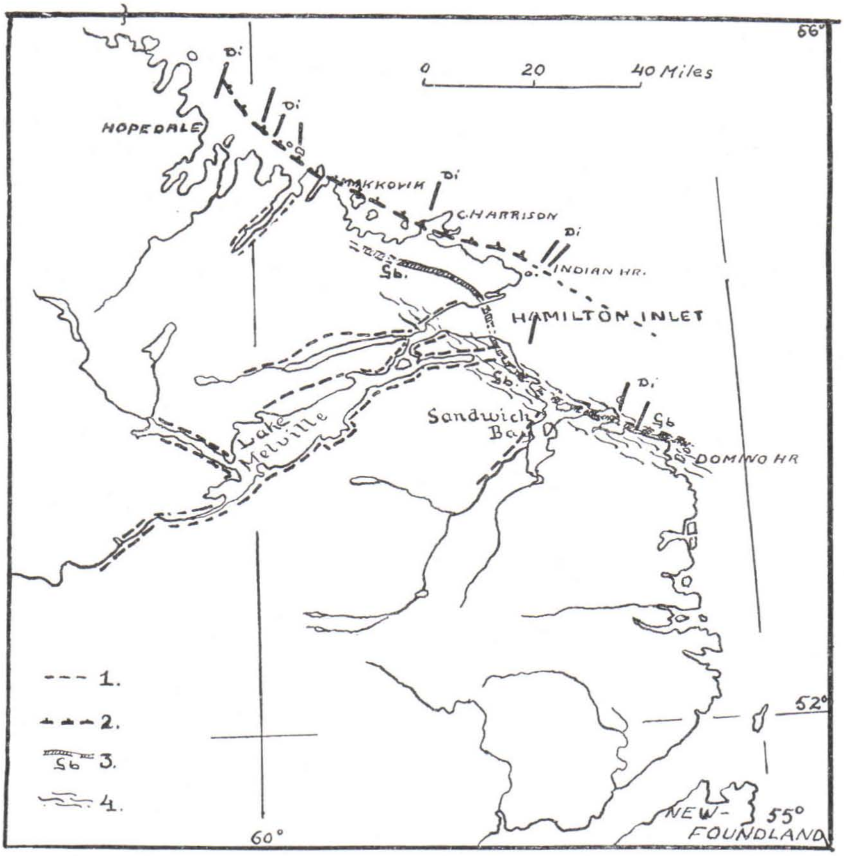

Fig. 1. Cartogram showing zones of dislocation and conspicuous igneous dykes in Eastern Labrador. 1. Fault-line (Tertiary), 2. Southern limit of lamprophyric dykes, 3. Great gabbro dykes, 4. Zone of mylonitization (domino-gneisses). Di = diabasic dykes (eo-Cambrian?).

The topography is generally in the stage of youth and there are many indications of a comparatively recent rejuvenation of the erosion. Only the big rivers have valleys of considerable depth. The inland is frequently plateau-like.

The most conspicuous features of this coast section are the great bay, Hamilton inlet, and its continuation the inland-sea Lake Melville. This great indentation marks a sudden break in the coast-line. North of Hamilton inlet it is displaced towards the north. It appears, however, from the soundings marked on the sea-charts that there is a shelf outside the southern part of the coast-section (Fig. 1), and the northerly coast-line corresponds evidently to the edge of the shelf. The coast is rich in small islands and skerries but a skiergard in the proper sense of the word is met with only in certain places, e. g. outside Sandwhich Bay and between Cape Harrison and Hopedale.

LAKE MELVILLE AND SANDWHICH BAY

Lake Melville has been studied by Kindle (6), who interpreted this remarkable basin as a tectonic depression. This conception has also been adopted by Tanner (10) in his description of the physiography of Newfoundland — Labrador. The steep slopes of the coast-mountains around the basin, and the absence of mature valleys of the stream running down from the highlands, particularly on the south side of the bay, strongly support this conception. The best evidence is, however, delivered by the layers of unaltered sandstone in subhorizontal position (eo-Cambrian) along the northern shore of Lake Melville and its northern arm, Double Mere. These sandstone layers are evidently down-faulted rests of earlier layers of greater extension.

Sandwhich-Bay offers in many respects the same picture as Lake Melville — Hamilton inlet. Particularly the northern shore is rather high and steep and represents probably a fault line. The highland of Mealy mountains, with its comparatively even upland-crest, may therefore represent a »Horst». The old peneplane is here elevated to more than 1000 feet above the sea.

MAGMATIC INTRUSIONS INDICATING CRUSTAL MOVEMENTS

The next feature which interests us in this connection is the abundance of basic dykes, frequently mentioned by visitors of Labrador. The present author has previously pointed out the existence of at least three different types of dykes:

1. Vertical diabase dykes, generally fine grained, sometimes of porphyric texture, running more or less at a right angle to the coast-line.
2. Subhorizontal sills of lamprophyric composition (Odinites, Kersantites etc.).
3. Dike-like instrusions of coarse-grained gabbro-diabase, forming rows of elongated intrusive masses, parallel with the coast. These rocks are often considerably affected by tectonical movements and particularly the marginal parts of the intrusions are highly altered by sheering and recrystallization, due to movements directed N—S. The unaltered rock

seems to be a coarse-grained olivin-bearing or olivin-free gabbro of ophitic texture. It alters into eclogite, garnet-bearing coronitic gabbro, hornblende gabbro or amphibolite.

Furthermore we have to notice the occurrences of completely unaltered young granites and syenites (Makkovik, Strawberry) particularly on the coast-section Cape Harrison — Makkovik. These rocks do not, so far as is known extend farther inland.

TYPES OF CRUSTAL MOVEMENTS ALONG THE LABRADOR COAST

We have mentioned above a number of phenomena connected with movements in the earth-crust within the coastal regions of Labrador. The cartogram fig. 1 shows the general relation between these phenomena.

If we do not take into account the recent post-Glacial elevation, the most recent movements are represented by the faults of the Lake Melville district. Here the traces of displacements can still be seen in the topographic forms.

There have not been found any dike-intrusions corresponding to the faults, connected with this phase of movements, but the remnants of thermal activity at Cape Harrison, found by the author (8) proves that the displacements during late geological times have reached considerable depth.

The next phase of crustal movements which has left visible traces in the geological structures is indicated by the vertical diabas-dykes. So far as the general directions of the dykes are concerned they could be connected with the tertiary movements *e. g.* the dyke-complex in the region of Indian-Harbor N of Hamilton inlet. The fact that there are no traces of superficial volcanic actions corresponding to the dyke-intrusions shows, however, that they have been eroded to considerable depth, and the fissures along which they have intruded must be of much earlier date.

The dykes are younger than any other rock of the region, except possibly the Double Mere sandstone. Ex analogia with the conditions in other pre-Cambrian regions it seems probable that they are of eo-Cambrian age (Jotnian, Keeweanawan), but they may at least partly be younger.

The great abundance of dykes from this period proves that it was a time during which the earth-crust was the subject of strong block-movements, probably connected with tilting and faulting. The trend of the dykes indicates an N-W-erly direction of the main-axis of the deformation, or in other words about the same direction as during most of the orogenic movements since Algonkian time in this part of the Canadian shield.

Cloos (1) has recently pointed out the great similarity of the great rift-valleys of the world, and has in an excellent way been able to demonstrate by means of experiments the formation of tectonic »Graben» of this type in connection with dome-like elevations of the earth crust.

This interpretation seems to be applicable also in our case. Fig. 2 shows the fault-systems of the Lake Melwille depression compared with the Red sea-depression and the Rhin-Graben. The similarity is conspicuous.

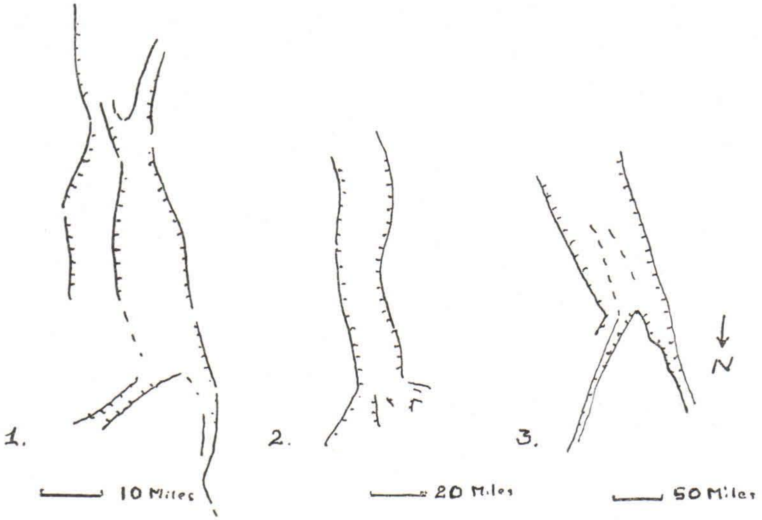

Fig. 2. Lake Melville depression compared with some other great tectonic fault »Graben». 1. Lake Melville, 2. Rhein-»Graben», 3. Red-sea depression. (2 and 3 according to H. Cloos).

The slight progress of the erosion, posterior to the down-faulting, proves, that the latter can hardly be of earlier date than late Tertiary. The movements which have formed the Lake Melville basin are probably of the same age as the elevation of the high mountains in northern Labrador. We evidently have to reckon with a tertiary uplift of the whole Labrador-coast region and the eastern part of Baffin island but this uplift may consist of different local domes giving rise to different joint-systems and Grabens of the same type as Lake Melville. We have in the foregoing already mentioned Sandwhich bay; there may exist still other formations of the same origin (*e. g.* Kipakok Bay, and possibly some of the northern »fjords»).

Dykes belonging to the same phase of movements are found also farther inland as is shown already by the descriptions by Low (9).

The following dyke-system is interesting particularly on account of its subhorizontal position. Dykes of the lamprophyre-group, are cut by the foregoing group and consequently they are older. They occur only along a very limited stripe of the coast-section approximately from Holton harbor to Hopedale, and are mainly restricted to the outermost part of the coast. Farther North, outside Hopedale they are found only on the islands.

The flat position can be tectonically explained either as depending on horizontal beds and joints in the older rock ground, or by the assumption of a slight flexure-like bending of the earth-crust which has given rise to tension-joints (Fig. 3). The first explanation is not applicable on account of the occurrence of lamprophyric dykes in all types of rocks. The later theory seems, on the contrary, to agree well with the mode of occurrence along the edge of a pre-Cambrian block. The slight southerly slope of the majority of the dykes indicates a downwarping of the seaward part of the block. Local faults due to these movements can be studied at Aillik.

The same zone is, as has been pointed out in another connection (8), rich in granitic and syenitic intrusions which are older than the lamprophyres but which evidently belong to the same cycle of crustal movements. In this connection the occurences of similar lamprophyres in the regions characterized by fault tectonics of southern Germany (Schwarzwald etc.) may be pointed out as offering a striking resemblance, so far as the dyke-intrusions are concerned.

The absence of lamprophyres SE of Hamilton inlet may depend on the fact that the flexure-zone continues farther outside the coast.

At last we have the oldest phase of basic intrusions of the coast; the gabbro-diabases, which will bring us in touch with movements characterized by horizontal displacements.

The intrusions of the conspicuous masses of dark gabbroidic rocks all along the coast from Domino run to Hamilton inlet have obviously intruded, a number of great fissures running almost parallel with the coast. They have evidently been formed by tension in connection with the origination of a rather narrow flexure-zone. This zone has, however, later been transformed, by means of shortening of the segment of deformation, into a thrust-zone which has caused the deformation of the gabbros, mentioned above, and also the formation of the mylonitic gneisses called the domino-gneisses. These movements have evidently affected the basic rocks in a higher degree in the eastern section than in the western. The cartogram fig. 1 shows that the dyke at Hamilton inlet is bending northward and lies at Pottles bay outside the strongest zone of deformation. Here the gabbro is therefore almost unaltered. The coincidence between the thrust-movements and the gabbro intrusions is consequently not complete, but until further facts have been presented which should prove the opposite, there seems to be reason to refer both the phenomena to the same orogenetic cycle.

The thrust movements have probably been of a comparatively small extention and cannot be compared with the great thrusts of the Alps. At any rate they represent a very important feature in the morphology of the Labrador coast, as being the earliest structures following the coastline. It is possible that they represent the earliest foundation of the present edge of the continent in this part of North America.

The different stages of the evolution of the coastsection during late-pre-Cambrian is schematically shown by fig. 3.

The absense of dyke-intrusions of other traces of orogenetic and strong epeirogenetic movements between eo-Cambrian time and Tertiary does, of course, not exclude the existence of events of that kind. They have, however, been too gradual and slow to give rise to visible displacements in the earth's crust.

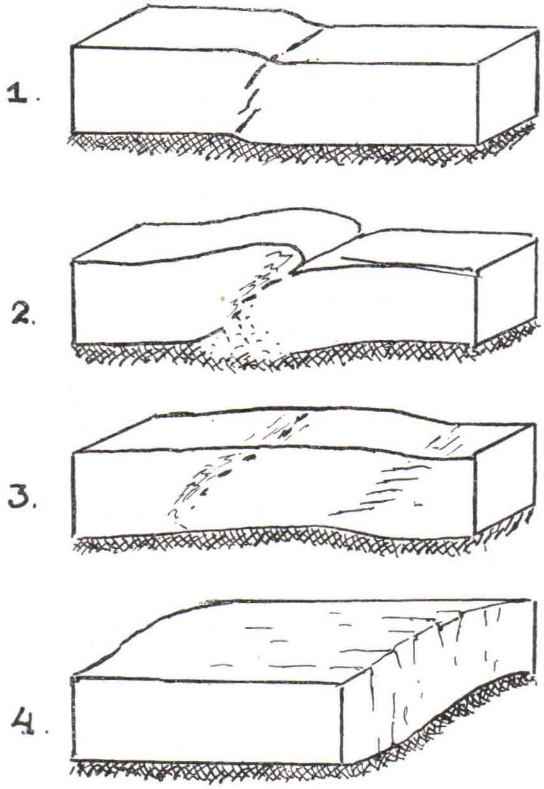

Fig. 3. Types of crustal-movements and magmatic intrusions in Eastern Labrador. 1. Gabbro-intrusions, 2. Thrust and deformation of the gabbros, 3. Origin of Lamprophyric dykes, 4. Origin of vertical diabasic dykes.

CONCLUSIONS

1. The Melville depression is a fault-»Graben» originated in connection with a dome-like uplift of the earth-crust during Tertiary time.

2. The coastal slope has probably existed since late pre-Cambrian time. This does, however, not necessarily mean that Greenland and the Canadian shield represent different blocks altogether. On the contrary, the author believes that they belong to the same resistant block connected by the shelf-areas farther north.

The Labrador sea between S-Greenland and the North American continent has possibly been formed already during latest pre-Cambrian.

3. The coast-zone has repeatedly been affected by movements which have given rise to magmatic intrusions. They will during further research work facilitate the connecting of geological events along the coastal sections of Labrador and Baffin island.

NEUCHÂTEL, SEPTEMBER 1946.

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6.

THE GABBRO-LIMESTONE CONTACT ZONE OF CAMAS MÒR, MUCK, INVERNESS-SHIRE

BY

C. E. TILLEY
(CAMBRIDGE)

ABSTRACT

The great dyke of olivine gabbro intersecting Jurassic limestones at Camas Mòr, Muck, is bordered by narrow zones of pyroxenite with theralite and nepheline gabbro carrying titanaugite, melilite, analcime and zoned iron-wollastonite solid solutions, the last partly in intimate intergrowth with nepheline. Monzonitic and shonkinitic lenses carrying abundant soda-sanidine and a fayalitic olivine, also appear in the border zones. These rocks are associated with clinopyroxene-wollastonite-analcime assemblages which have resulted from the soaking of the limestone by solutions emanating from the modified gabbro. In the broad exogenous contact zone of high temperature minerals the pairs — gehlenite-wollastonite, gehlenite-monticellite, monticellite-periclase, spurrite-gehlenite and monticellite-merwinite are especially characteristic. An assemblage built of gehlenite-larnite-rankinite represents the highest stage of metamorphism attained in the contact zone. The reaction processes in the endogenous zone resemble closely those responsible for the development of the hybrid zones of Scawt Hill, Co. Antrim.

Fifteen years ago a detailed description of the production of basic alkali igneous rocks by the assimilation of limestone by basaltic magma was given by the writer in a paper dealing with the Scawt Hill intrusion of Co. Antrim, Ireland [1]. In that locality the reaction of olivine dolerite and Chalk was shown to lead to the development locally of nepheline bearing rocks of the type of theralite and nepheline dolerite accompanied by the production of wollastonite and melilite. Some account was then given of the probable reaction processes at work responsible for the genesis of these basic alkali rocks, and some discussion devoted to the applicability of these results to the general problem of the origin of alkali rocks.

With the possibility in mind that elsewhere in the British Tertiary province contact zones of basic intrusions and limestone might contribute further data on some aspects of this problem the writer was led to an examination of the borders of the great gabbro dyke intersecting the Jurassic limestones on the south shore of the island of Muck, one of the Small Isles of Inverness-shire. This paper records the first results of such

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[1] Min. Mag., 1931, 22, pp. 439—68.

a study and presents briefly the effects of contact metamorphism and hybridism associated with this intrusion.

The field relations of the Muck gabbro and the associated Oolitic rocks are described by Harker in the Geological Survey Memoir on the Small Isles of Inverness-shire[1]. A broad vertical dyke of olivine bearing gabbro running N. N. W. is mapped as cutting a group of carbonate sediments of the Upper Estuarine Series along the east shore of the great bay of Camas Mòr. In the bay itself appears an undisturbed ascending sequence of Calcareous Sandstone (20 feet), Limestones of the *Ostrea hebridica* group (65 feet) and Cementstones (20 feet) with a gentle dip to the N. W. at 5°—10°. The eastern shore of the bay is formed of gabbro, exposed between tide marks at the northern end and in its southerly continuation forming a bold line of cliffs (fig. 1). In the shore section, the gabbro which is about 30 yards broad is bordered on both sides by the Estuarine Series, here seen dipping in places at high angles to the E. N. E. and in an intensely metamorphosed state. In this broad exogenous contact zone, melilite (gehlenite) is the most important silicate mineral but the occurrence of assemblages rich in monticellite or containing periclase and brucite shows that dolomitic bands were present among the prevailing limestone horizons.

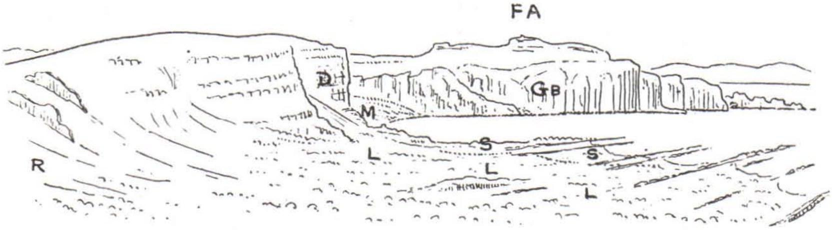

Fig. 1. Camas Mòr looking east, showing the gabbro cliffs (GB) on the east side of the bay and the outcrops of the Estuarine Series. FA = basalt lavas of Fionn-ard, D = basalt making cliff at head of bay, M = mugearite at base. Reefs on shore intersected by numerous dykes are made by the Great Estuarine Sandstone (S) and the *Ostrea hebridica* limestone group (L), R = raised beach.
(Reproduced from a drawing by Dr A. Harker).

The following list presents the chief minerals recorded from the contact zone on both sides of the gabbro dyke on the shore.

Calcite
Grossularite
Wollastonite
Monticellite
Melilite (gehlenite)
Spurrite
Merwinite
Larnite
Rankinite
Cuspidine

Tilleyite
Periclase
Brucite
Spinel
Perofskite
Clinopyroxene
Analcime
Plagioclase
Soda-sanidine
Titanite
} in skarn zone.

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[1] Mem. Geol. Surv. Scotland (Small Isles of Inverness-shire) 1908, pp. 30—33, 96—97.

The more important rock type assemblages built of these minerals are — apart from those of the clinopyroxene skarns — the following

Grossularite-calcite-quartz
Wollastonite-grossularite-quartz
Gehlenite-wollastonite-calcite
Gehlenite-monticellite-calcite-spinel
Gehlenite-spurrite-calcite-spinel
Gehlenite-monticellite-merwinite-calcite
Monticellite-periclase (brucite) -spinel-calcite
Gehlenite-tilleyite-calcite-(spurrite)
Gehlenite-larnite-rankinite.

The earlier stages of metamorphism are seen in the limestones carrying small crystals of grossularite with quartz and calcite still stable together. Some of the *Ostrea hebridica* bands at this stage show the shapes of fossil shell fragments in a calcite base with grains of garnet and chalcedonic quartz. More advanced stages of metamorphism appear in the assemblages built of grossularite and wollastonite but the commonest mineral of the contact zone is a melilite with the optical characters of gehlenite. The white or yellowish hornfels immediately against the gabbro is most frequently a gehlenite-wollastonite assemblage, less frequently one built up of monticellite and gehlenite. Beyond this immediate zone other assemblages appear carrying besides gehlenite, monticellite or wollastonite, the minerals spurrite, tilleyite and less commonly merwinite, larnite and rankinite. Many of these assemblages carry a brown or greenish brown spinel as minute octahedra enclosed in gehlenite, monticellite or calcite.

Gehlenite is developed in tabular crystals of variable dimensions from $^1/_{10}$ millimetre up to $\frac{1}{2}$ millimetre in diameter (Plate II, fig. 4). It is uniformly an optically negative type with $\omega = 1.658 - 1.660$ and low birefringence $= .003$. Microchemical tests show abundant alumina (Cs alum reaction). Apart from the usual fibrous alteration the mineral is frequently replaced by dense aggregates of isotropic grains, slowly soluble in hot hydrochloric acid and of refractive index 1.71. Their properties indicate that these pseudomorphs consist of a garnetoid mineral, a member of the hydro-grossular series.

Monticellite occurs both in the marbles together with periclase or brucite and as a constituent of the carbonate poor assemblages with gehlenite. It is an optically negative type with large optic angle belonging to the magnesian end of the series. In one rock immediately adjacent to the modified gabbro a monticellite of higher iron content (25 wt. % Ca Fe SiO_4) occurs in coarse grains up to 6 millimetres in diameter in a gehlenite-spinel assemblage also carrying perofskite, magnetite and pyrrhotite. This rock passes out into a normal monticellite-bearing marble.

Spurrite occurs in assemblages both rich and poor in calcite, in the former as isolated grains in calcite, and in the latter intimately intergrown with gehlenite, forming irregular spongy crystals enclosing small crystals of this mineral. *Merwinite* occurs in a marble bearing monticellite and gehlenite. It then appears in the characteristic interpenetrant twinned laths 1/4 mm. in length set in calcite and sometimes mantled by a thin rim of spurrite (Plate II, fig. 6).

Rankinite and *Larnite* form important constituents of assemblages found on the west contact of the gabbro- (1) Gehlenite-rankinite-spinel, (2) Gehlenite-rankinite-larnite-spinel. The chief mineral of these assemblages is gehlenite. The associated rankinite has the characters of that previously described at Scawt Hill (Min. Mag., 1942, pp. 190—6). It forms rounded grains devoid of cleavage and may form the host for numerous included laths of gehlenite (Plate II, fig. 5). In the second assemblage, larnite (β Ca_2SiO_4) may be similarly enclosed as drop like grains in larger individuals of rankinite, but it also occurs as independent idioblastic grains. Fine multiple twinning like that developed in the Scawt Hill rocks is characteristic. Struck with a hammer these larnite-rich assemblages dust owing to volume change on inversion of the β phase to γ Ca_2SiO_4. Spinel in these aggregates appears only in the form of minute octahedra, particularly as enclosures in gehlenite.

Cuspidine is a constituent of several assemblages. It occurs as a late mineral investing idioblastic gehlenite and wollastonite in a marble and forming a matrix for a group of such crystals (Plate II, fig. 4). In another marble it is developed in subidioblastic crystals in calcite enclosing rounded grains of calcite and little grains of gehlenite. Twinning is predominantly multiple parallel to the base, Z the acute bisectrix emerging on 001.

Tilleyite forms an important member of a thick band of dark grey somewhat resinous-lustred gehlenite rock exposed on the western contact of the gabbro. It occurs in brightly polarizing aggregates intergrown with fine grained gehlenite so as to form in places at least half of the rock. Typically developed in sponge-like single crystals reaching 1 1/2 mm. in breadth with gehlenite enclosures, it may also be segregated into small pools of inclusion-free crystals with or without calcite. The characters of the tilleyite in these rocks differ somewhat from those given by Larsen and Dunham for the original tilleyite of Crestmore[1].

The mineral behaves as a monoclinic crystal and possesses one good cleavage which is perpendicular to the optic axial plane (010). The refractive indices are $\alpha = 1 \cdot 612$ $\beta = 1 \cdot 633$ $\gamma = 1 \cdot 654$. Optic axial plane sections show an extinction Z $\wedge$ cleavage (taken as (100)) = 13°. Twin structure is an important feature of the Muck tilleyite. It is frequently

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[1] Amer. Min. 1933, 18, pp. 469—473. NOCKOLDS and VINCENT (Min. Soc. meating London, March 27, 1947) have described tilleyite from Carlingford, Eire, and shown by a new analysis that its empirical formula is 3 CaO. 2 SiO_2. 2 $CaCO_3$.

repeated but not with the regularity seen in spurrite. The composition plane (twin plane) is perpendicular to the optic axial plane and makes an angle of 54° from the cleavage, the optical orientation being such that on 010 the characteristic symmetrical extinction in the twinned crystal is X ∧ edge of twin plane = 23° (fig. 2). Like the Crestmore mineral,

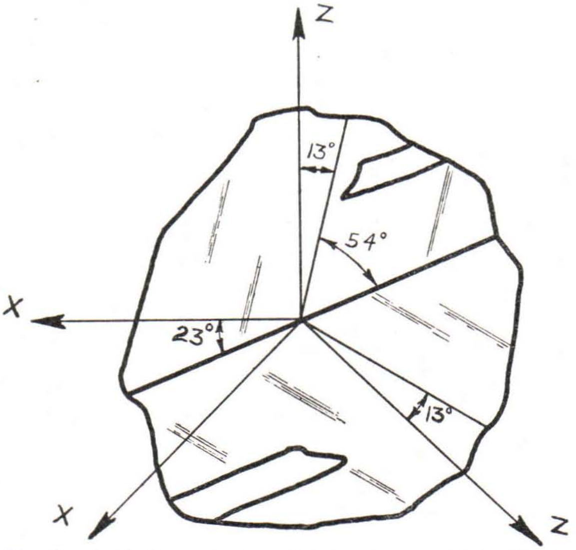


Fig. 2. Optical orientation of twinned crystal of tilleyite section parallel to 010.

the Muck tilleyite has a very high optic axial angle (2 V about 85°) and the sign is positive. I have compared these data with those given by a specimen of tilleyite bearing rock from Crestmore kindly presented by Dr K. C. Dunham. Sections of tilleyite cut perpendicular to the acute bisectrix (Z) show straight extinction against the single cleavage the optic axial plane lying across the cleavage. In sections parallel to the optic plane the extinction is Z ∧ cleavage 13° and against sporadic twin lamellae developed the symmetrical extinction is X ∧ twin edge = 24°. These data are in harmony with those given for the tilleyite of Muck. In these gehlenite-tilleyite rocks of Muck fine twinned crystals of wollastonite and irregular grains of spurrite are found in variable proportion along with the gehlenite and tilleyite.

The mode of development of the tilleyite denotes that it is a late mineral and the typical association of fine irregular granules of spurrite enclosed within it makes it probable that the gehlenite-tilleyite assemblage is ultimately derived from a gehlenite-spurrite assemblage such as has been described above.

Periclase is preserved in a number of the carbonate rocks occurring along with brucite, spinel, monticellite and calcite. It appears in rounded yellow grains showing cubic cleavage and is seen in process of alteration to a brown ferrobrucite eventually being replaced by an opaque iron ore. The development of abundant ferrobrucite and the opaque iron ore gives some of these metamorphosed dolomites a dark colour in hand specimens.

THE SKARN ZONE

CLINOPYROXENE-ANALCIME-WOLLASTONITE SKARNS

The assemblages just described may give place close to the contact with the gabbro to a group of darker coloured rocks which form the skarn zone proper, reaching 1″—3″ in thickness. These skarns are built up essentially of a coloured clinopyroxene (brown, titanaugite type) with wollastonite, but analcime, soda-sanidine and plagioclase appear also as constituents. The lighter coloured members are richer in analcime which forms rounded to trapezohedral areas ¼—1 millimetre in diameter enclosing little granules of clinopyroxene and separated by a dense aggregate of similar clinopyroxene. Wollastonite with normal optical characters is sporadically developed in such skarns. The place of analcime may in part be taken by soda-sanidine (Plate II, fig. 3). Both coarse and fine grained skarns are present, the coarser types being frequently rich in wollastonite, the associated clinopyroxene, a red brown zoned type, showing poeciloblastic texture with enclosed laths or plates of wollastonite.

THE GABBRO

The typical gabbro is a medium (1½ mm.) grained olivine bearing variety rich in labradorite (Ab_{35} An_{65}). This plagioclase shows idiomorphic relations both to the olivine and to the light brown augite which encloses it sub-ophitically (Plate I, fig. 1). A partial analysis showing SiO_2 46.09, Na_2O 3.01, K_2O 0.41, is reported by Harker in the Survey Memoir (p. 100).

THE ENDOGENOUS ZONE

At the contact of the gabbro reaction with the calcareous sediments has produced the following changes which are recognised for a few inches on both sides of the dyke. The rock becomes pyroxene rich and darker in colour. Olivine disappears and plagioclase becomes subordinate. The simplest type of contact relation is that showing a concentration of clinopyroxene without complete disappearance of the olivine. In this stage the pyroxene preserves the colour of the mineral present in the normal gabbro but developing a greener zoned edge to the crystals. Locally the richness in pyroxene may be such as to warrant the term pyroxenite for the assemblage (Plate I, fig. 2). In other examples a pyroxenite zone (½″ or more in width) of strongly coloured titanaugite with minor interstitial plagioclase forms the contact. In this wollastonite may appear as an accessory constituent and olivine has disappeared (Plate I, fig. 3). The grain size of these pyroxenite zones is variable: those of fine grain may reach a thickness of 2—3 inches. These contain subordinate plagioclase as insets crowded with fine pyroxene except for a thin outer shell free from inclusions (Plate II, fig. 2). They form heavy dark rocks readily distinguished in the field from the normal gabbro. More highly modified assemblages are associated with these pyroxenites or pyroxene rich gabbros. These occur as streaks, veins and lenses in the endogenous zone and are found to carry abundant wollastonite, plagioclase, melilite, nepheline and analcime besides subordinate soda-sanidine. Such assemblages correspond to *wollastonite-theralites* or when plagioclase is subordinate *wollastonite nepheline gabbros.* Their grain size is typically coarse; strongly coloured titanaugites reach 3 mm. while individual wollastonites may attain even larger dimensions. Nepheline too appears in large grains 3 mms. or more in diameter.

The titanaugite is strongly zoned passing from a lighter coloured core through to zones of dark red brown tint and then to a shell of a dull green colour much less strongly pleochroic. The pleochroic tints of the darker zones are α and β reddish brown γ yellow and dispersion both of bisectrixes and optic axis is strong, particularly optic axis B. $\varrho > v$. Occasional fringes of aegirine, or acmite occur and are again found in the groundmass with analcime.

The wollastonite — an iron wollastonite solid solution — shows a patchy zoning the refractive index γ rising to 1·660. Extinction in sections perpendicular to the zone of cleavages rises to 45° $(X^1 \wedge 100)$. It is more frequent to find that the exterior zones of the wollastonite show higher refractive index as well as greater extinction angle as seen in the cross sections. Very characteristic is the association of wollastonite in the form of intergrowths with nepheline, the latter acting as host (Plate I, fig. 4). The same type of intergrowth occurs with analcime but there is reason to believe that the analcime has frequently developed from original nepheline. Nepheline in fact is frequently altered being represented in the various assemblages by hydronephelite, analcime and in some cases replaced by sericitic mica. Pectolite too occurs in association with analcime forming radiating spears within this mineral.

Melilite has probably been much more common: now only cores of the mineral are preserved in a fibrous alteration product fringed at the periphery of the original grain by a secondary growth of fibres and granules of green pyroxene in much the same way as recorded in the hybrid assemblages of Scawt Hill (Plate II, fig. 1). Perofskite has not thus far been observed in these assemblages.

The incoming of a potash soda felspar in the alkali gabbro lenses has already been referred to. There remains to be described other assemblages in which this mineral as soda-sanidine forms the preponderant felspar. The fine pyroxenite zone is cut by lenses and streaks up to one inch in width of a rock type of monzonitic facies in which the constituents are augite, large plates of soda-sanidine containing cores of andesine, a fayalitic olivine, apatite and magnetite (Plate I, fig. 5). In some of the smaller veins the felspar may be almost wholly soda-sanidine giving a shonkinitic facies (Plate I, fig. 6). It is noteworthy that the augite is not the strongly coloured titanaugite of the nepheline gabbros but a greenish variety comparable with that of the outer zones of the titanaugites and of the alkali mesostasis of the theralites. Wollastonite has not been found in these assemblages. The soda-sanidine is optically negative with a variable optic angle (moderate to small 2 V).

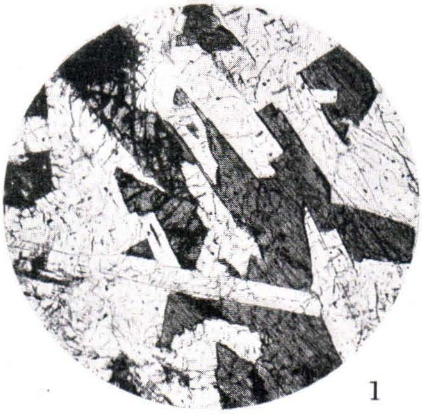
These alkali-felspar assemblages call to mind the segregation vein of anorthoclase-fayalite-dolerite met with in the hybrid zone of the Scawt Hill intrusion (*op. cit.* p. 453). Like that occurrence these monzonitic veins represent a siliceous and iron rich representative of the nepheline-bearing hybrids and there can be little doubt that these monzonitic facies derive through the chain of reactions consequent on assimilation.

CONCLUSION

Three groups of assemblages build up the complex at the borders of the gabbro dyke of Muck: (1) the carbonate-bearing contact zone, (2) the clinopyroxene skarn zone, (3) the hybrid endogenous zone. The nature of the mineral associations in the carbonate contact zone shows that these are typical high temperature facies reaching a high water mark of metamorphism in the larnite and rankinite bearing members then devoid of calcite. The bands of gehlenite and monticellite-rich, carbonate poor type immediately contiguous to the gabbro give clear indication of the introduction of silica-bearing solutions from the gabbro.

In the clinopyroxene-analcime zone there is unmistakable evidence that in addition to silica, iron, titania, alumina and soda were introduced from the gabbro and it is in every way probable that these analcime bearing assemblages derive from a liquid emanating from a gabbro already modified by incorporation of limestone.

The endogenous contact zone, like that of Scawt Hill has a very restricted development and the products — pyroxenites, theralites and nepheline-gabbros present so striking a similarity to the types met with at that locality that we can be confident that they are to be ascribed to a similar origin and to a similar set of reactions. A reaction mechanism has been set forth in detail in the paper devoted to the Scawt Hill rocks.



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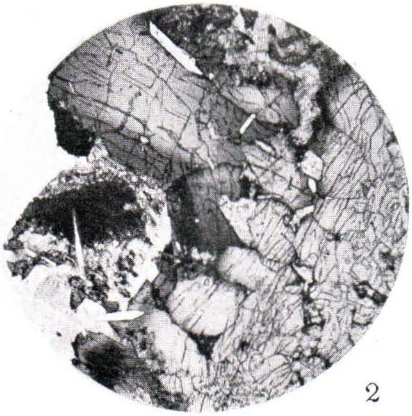

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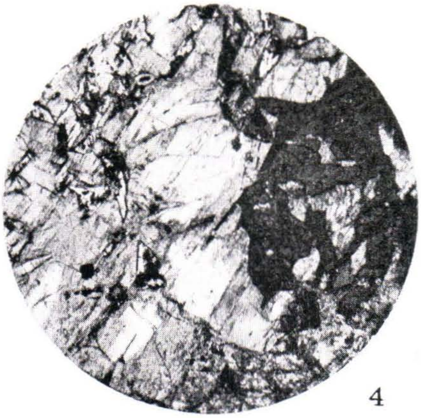

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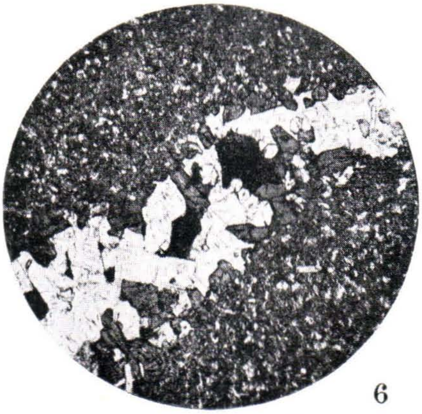

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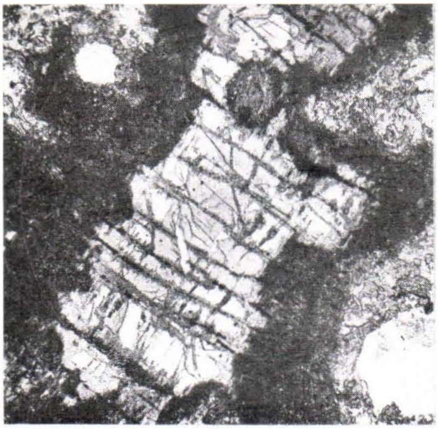
C E. Tilley: The Gabbro-Limestone Contact Zone of Camas Mòr, Muck, Inverness- Shire.



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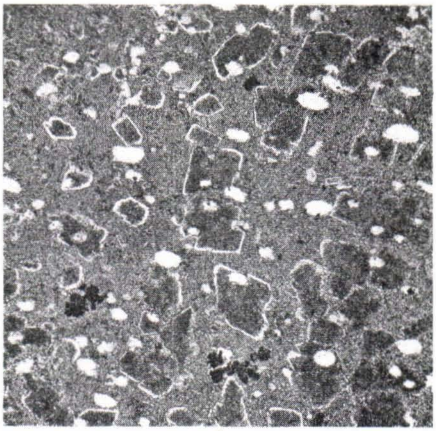

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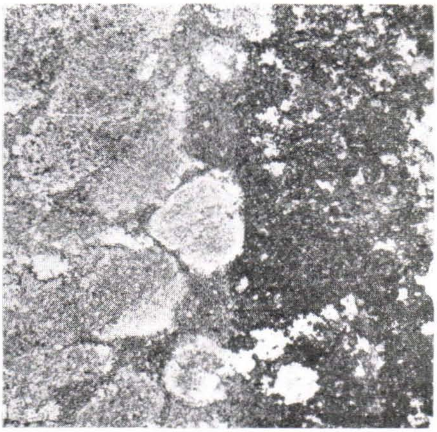

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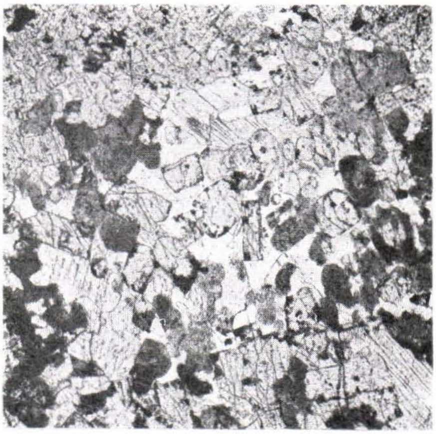

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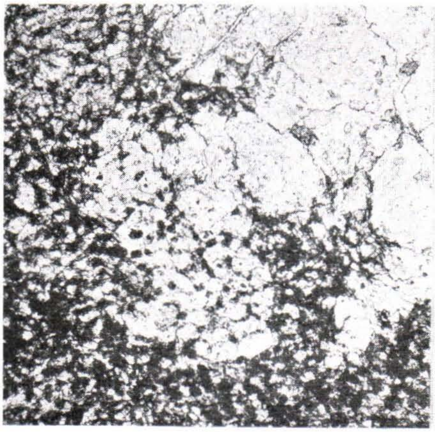

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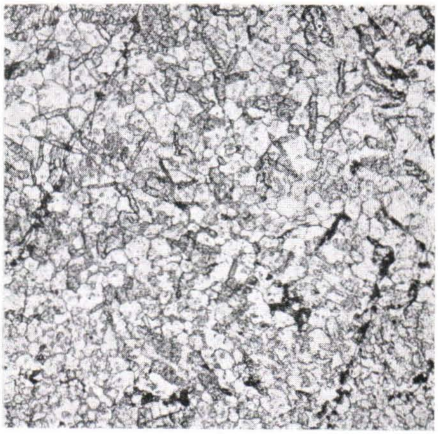

6

C. E. Tilley: The Gabbro-Limestone Contact Zone of Camas Mòr, Muck, Inverness-Shire.

PLATE I

GABBRO AND CONTACT MODIFICATIONS OF CAMAS MÒR, MUCK

Ordinary light × 16 diameters.

Fig. 1. Olivine Gabbro: normal type showing ophitic texture.

Fig. 2. Pyroxenite: at contact with metamorphosed Jurassic limestone — passing (left) into theralitic type with nepheline replaced by analcime.

Fig. 3. Pyroxene rich area in »nepheline gabbro»: wollastonite solid solution (below) and nepheline above.

Fig. 4. »Nepheline Gabbro» with titanaugite, nepheline (clear areas unaltered) intergrown with wollastonite solid solution (left).

Fig. 5. Monzonite lens in the contact zone showing augite, preponderant soda-sanidine with cores of andesine.

Fig. 6. Lens of soda-sanidine-augite rock in fine grained dark pyroxenite (endogenous).

PLATE II

ENDOGENOUS AND EXOGENOUS CONTACT ROCKS OF THE OLIVINE GABBRO OF CAMAS MÒR, MUCK

Ordinary light × 18 diameters.

Fig. 1. Melilite in pyroxenite zone: melilite in process of replacement (late stage) by a dense aggregate of granules and rods of green augite; clear areas, vugs of analcime.

Fig. 2. Dark pyroxenite with insets of plagioclase (clear rims) filled with granules of titanaugite: little vugs of calcite.

Fig. 3. Clinopyroxene-analcime skarn (some soda-sanidine) against fine pyroxenite. Ovoid areas of analcime enclose granules of clinopyroxene.

Fig. 4. Gehlenite-wollastonite rock with calcite and cuspidine. In the centre the idioblastic gehlenite crystals are set in a base of cuspidine: right centre and left lower corner, cleaved wollastonite; above gehlenite in calcite.

Fig. 5. Gehlenite-rankinite-larnite rock. Clear areas, rankinite and gehlenite; dark, larnite. In the centre two larger crystals of rankinite as hosts to numerous grains of larnite and gehlenite, the latter less clearly distinguished owing to similar relief to rankinite.

Fig. 6. Merwinite-gehlenite-monticellite-marble. Elongate sections of merwinite in a base of calcite, bearing gehlenite and monticellite.

7.

ON THE MICROSTRUCTURE OF TITANIFEROUS IRON ORE AT OTANMÄKI

BY

O. VAASJOKI

CONTENTS

INTRODUCTION

The titaniferous iron ore situated south from Oulujärvi at Otanmäki in the parish of Vuolijoki has in the last ten years aroused the interest of Finnish mining people not only as a promising iron ore, but also for its vanadium content. Discussions regarding above all its economic future were carried on in various directions until the Finnish Government finally set up a special committee to investigate the practical possibilities of the occurrence. In the first place the recent war years have been a cause why up to the present, nothing has been published which could clear up this significant occurrence, belonging as a discovery to the youngest ones in our country. The author will in the following present some of its essential features, the recognizing of which also can be a help when planning its economic use. This treatise is grounded on some analytical and especially ore microscopic determinations of the ore in question, which the author carried out at Helsinki University during the Spring of 1945 under the guidance of Prof. P. Eskola. The aim was thus to get a view of the microscopic structure of the ore and of the relations between its ore oxides according to the principles given in the first place by P. Ramdohr (1) and St. Foslie (3) in their remarkably investigations. Owing to the mainly orientative character of this work, all points have not been treated with the thoroughness which often would have been needful. In reference to this, some remarks are to be made particularly regarding the accuracy of the analyses. The author hopes, however, that, in spite of these imperfections, the following study may give a logical preliminary view of the Otanmäki ore, which in a noticeworthy degree, especially in the light of microscopic observations, resembles the Routivare titaniferous iron ore in northern Sweden, most recently described by P. Ramdohr (2).

ANALYSES

The ore at Otanmäki is associated with a deformed amphibolite zone, where the ore occurs as elongated, lenticular and veinous formations. Its quality is very varied and it is technically divided into several different classes on the ground of its silicate content. The best ore is a compact mass of magnetite and ilmenite, containing possibly a little pyrite and in a very few cases also chalcopyrite. In addition to these högbomite is found in minor amounts and, when using higher magnifications, there are brightly reflecting idiomorphic grains very rarely observed, the quality of which has so far not been possible to determine because of the small grain size. Chiefly clinochlorite is present as silicate material in the best ore, furthermore biotite and a little spinel. In the same degree as the quality of the ore becomes poorer, the amount of pyrite grows, and naturally the amount of silicates increases too, these then representing the magmatic amphibolite facies. In this connection must also be stated a fact verified by the determinations on I-table, viz. when the type of the ore declines, the ilmenite becomes more predominant, until the most impure ore is almost pure ilmenite. This change is noteworthy when considering that the ratio between magnetite and ilmenite in the best ore is about 2 : 1. The following analysis, performed of the best ore, gives a view of its composition and of the relations between its elements:

Table I.

	%	1000 × mol. prop.
SiO_2	1.27	21.1
Fe_2O_3	41.87	261.7
FeO	30.94	430.1
$T O_2$	18.86	235.6
MnO	0.24	3.4
MgO	2.92	72.4
CaO	0.0	—
Al_2O_3	1.56	15.3
Cr_2O_3	tr.	—
V_2O_3	0.42	2.8
P_2O_5	0.0	—
S	0.68	21.2
H_2O+	0 74	41.1
	99.61	

Sp. g. 4.55

From the same ore type, of which the total analyses have been made, material was separated also for determinations of magnetite and ilmenite. The results, after which the corresponding mineral components at the same time have been formed, are:

Table II.

The magnetite analysis

	%	1000 × mol. prop.		mol. %
Fe_2O_3	62.09	389.1	Fe_3O_4	90.10
FeO	33.14	460.7	$MgTiO_3$	3.23
TiO_2	2.50	31.2	$FeTiO_3$	0.61
MgO	1.08	26.8	$MnTiO_3$	0.06
V_2O_3	0.68	4.5	FeV_2O_4	1.00
Cr_2O_3	0.20	1.3	$FeCr_2O_4$	0.29
MnO	0.03	0.4	FeO	4.44
	99.72			99.73

Sp. g. 5.05

Table III.

The ilmenite analysis

	%	1000 × mol. prop.		mol. %
Fe_2O_3	60.85	67.8	$FeTiO_3$	77.79
FeO	36.92	513.3	$MgTiO_3$	10.23
TiO_2	48.10	600.0	$MnTiO_3$	0.51
MgO	3.42	84.8	Fe_2O_4	0.35
V_2O_3	0.18	1.2	V_2O_3	0.18
Cr_2O_3	tr.	—	Fe_2O_3	10.61
MnO	0.24	3.4		99.70
	99.71			

Sp. g. 4.67

As to the distribution of elements, it will be seen that the vanadium has concentrated in magnetite, as in general has also been observed in earlier investigated titaniferous iron ores. The obvious cause of this is the analytical relationship and the similarity of lattice constants between Fe’’’ and V’’’, as reported in literature. In this respect attention must be drawn to the paper of C. W. Carstens (4), in which these matters are discussed and in which he also has defined the conception of the vanadium module, which is

$$\frac{100 \times V}{\text{Fe in magnetite.}}$$

The value of the module in question is in Otanmäki ore about 0.67.

From the analyses has furthermore been observed that the chromium follows the magnetite, while both the magnesium and the manganese follow the ilmenite. The magnesium content of ilmenite seems to be comparatively high, which in its turn gives support to the suggestion, set forth later in this paper, that the ilmenite of the ore may have been crystallized in higher temperature conditions than its magnetite.

The excess of FeO, which is characteristic of the magnetite of the titaniferous iron ores, is also in this case considerable in the composition of the magnetite. The ilmenite analysis again, on the other hand, shows an excess of Fe_2O_3, which can be observed also as occurring as own peculiar formations, according to the microscopical description.

ORE MICROSCOPIC OBSERVATIONS

The percentage of ore in polished sections used has varied between 75—95. These specimens show that the magnetite and ilmenite practically have crystallized as separate grains, building among themselves a structure of granoblastic appearance. The magnetite is in its forms essentially very irregular, masslike, while the grains of ilmenite are in general more sharpedged. Sometimes a weak orientation can be observed in regard to ilmenite, although the influence of deformation appears most clearly as brokenness of ilmenite structures. From the occurrence of silicates as inclusions and their corrodedness in ore material can be concluded, that the crystallizing of the ore material is of younger origin. The reciprocal relations between magnetite and ilmenite are discussed later on in this paper.

Magnetite. — In the Otanmäki ore the magnetite and ilmenite do not in general form lamellaric intergrowth structures, though in some few cases, a kind of traces of such can be observed. There is instead the typical phenomenon that a small amount of ilmenite occurs in magnetite as irregular spots, the average size of which is between 0.002—0.05 mm (Fig. 1). On separating the material for the magnetite analysis the grain size applied was between 0.2—0.15 mm and therefore the ilmenite, appearing as flecks, was no longer separable. Thus the relatively high TiO_2-content of the analysis in question could be entirely due to these flecks. An approximative microscopic calculation confirms this suggestion, too.

Here and there in magnetite can be observed a weak inclination to change into martite, which in most cases begins to develop at the borders or cracks of grains, progressing skeletally along the octahedral cleavage towards the inner parts of magnetite grains (Fig. 2). The martite has probably caused errors in the mineral analyses, having regard on the one side to the excess of FeO in magnetite, and on the other to the amount of Fe_2O_3 in ilmenite. However, the possibility in question, which in any other connection could be a noticeable factor, effects in this case, no changes in the treatment of phenomena, which the author has followed in this presentation.

The spinel in Otanmäki magnetite is feebly developed and is not met with in all grains. When occurring, it follows, in typical lath-like structures, the regular direction of the magnetite (Fig. 3).

Ilmenite. — The grain size of the ilmenite is in general smaller than that of the magnetite. Ilmenite grains have angular forms, and such grains can often be met, which are surrounded by small crushed fragments. As a typical microstructure the hematite can be seen, which, due to immiscibility, appears in two different generations (Fig. 4). The first one is seen already with weaker magnification in systematic rows like fragmentary lines or is shaped as irregular larger flecks having a homogeneous orientation. Between these all the buildings of the other generation exist as very small oval flecks, hardly noticeable without recourse to oil immersion. The hematite buildings of both kinds are decreasing towards the grain border and are lacking entirely in the border zone. The flecks of ilmenite included in magnetite, as mentioned above, may sometimes contain a little hematite as product of immiscibility. In most cases, however, the hematite streaks cannot be found.

In the ilmenite component of Otanmäki ore twinning is a common phenomenon, and it appears in places as a strongly developed lamellarity (Figs. 4 and 5). In such a structure the influence of deformations can be verified in an excellent way as flexures and cataclastic brokennes of lamellae. Furthermore, it can be observed that above all in the compact and the best type of the ore the borders of twinning lamellae in respect of most grains are practically evaporated. Thus the inaccuracy of the lamellae borders resembles a new melting. Based on the inverted position of the hematite streaks, belonging to the first generation, the existence of twinning lamellae may best be concluded.

CONCLUSIONS

When considering the Otanmäki ore from a purely analytical point of view, a distinct early magmatic tendency in its composition can be noted. When examining the occurrence as a whole in the light of field observations, it is quite obvious that a considerably metamorphosed formation is in question. Under the ore microscope both features mentioned above can be established, although the immediate consequences caused by movements are often strongly developed and easily identified. The indirect consequences of movements, on the contrary, are more difficult to see at once, and therefore the author attempts in the following to form a logical idea for the interpretation of phenomena appearing in the foregoing studies. In this sense the author assumes that in the Otanmäki area there have been two effective geological periods, both of which have been noteworthy and of significance to the structures of the present ore. They are:

— The genesis of primary titaniferous iron ore and consolidation in connection with the development of the older rock ground.
— The mobilization of ore material during the movements and the recrystallization chiefly in association with same. Due to these processes the ore has got the structural features, which at present can be recognized, preserving, however, the analytical properties of the first phase. The mineral composition of the wall rock has changed at the same time to correspond to the facies of lower temperature.

The events of the first geological period do not require any closer explanation. The ore thus formed has gained some of the typical structures, which are so characteristic of the titaniferous iron ore widely described in literature. Associated to eruptives they have crystallized peacefully without any greater disturbances caused by movements or by metamorphism. Instead, the events of the second geological period require a closer scrutiny. This has chiefly been based upon microscopic determinations.

The masslike ore, associated with primarily basic rocks possibly of hyperitic composition, has owing to tectonical causes entered upon new conditions of temperature and pressure and become mobile. The mobilization has, however, not been so thorough, that it could be considered as a new melting, but the ore material has at least in part been somewhat plastic. Had the melting been complete it would not be possible to see any relic features in the present structures of the ore. The lamellaric ilmenite in magnetite, however, sometimes shows this phenomenon. On the other hand the temperature has been so high that the lamellae of ilmenite in question have largely been transformed into flecks, due to the continuous movement of the mobilized mass. These buildings now represent one part of the flecks which are met in the microstructure of magnetite, as mentioned above. No homogenization has taken place, although this could be possible between 500°—800°, according to the experiments carried out by J. Murdoch (5).

Probably the magnetite of primary ore has also been relatively poor in ilmenite. Perhaps the magnetite and ilmenite also have occurred as separate grains. This opinion can be based upon the immiscibility of hematite in ilmenite, which shows, that Fe_2O_3 has appeared in excess compared with TiO_2 and FeO. Thus the magnetite is poor in titanium, according to the excellent statement of St. Foslie. He supposes namely that the titanium occurs in high temperatures as Fe_2TiO_4, and thus, if there is excessive Fe_2O_3, there occurs the reaction:

$$Fe_2TiO_4 + Fe_2O_3 \rightarrow FeTiO_3 + Fe_3O_4$$

Fe_2TiO_4, very widely isomorphicaly miscible with magnetite, is in the lower temperatures not stable and decomposes:

$$Fe_2TiO_4 \rightarrow FeTiO_3 + FeO$$

Thus the excessive FeO, typical of the magnetites of titaniferous iron ores, can be explained, and thus, probably in a corresponding way, the excessive FeO has also been produced in the Otanmäki magnetites. In spite of the mobile phase the excess of ferrous oxide has remained in the lattice of magnetite and in the same way the ferric oxide has followed the ilmenite phase.

When the ore material during the mobile period had reached the maximum of temperature and the consolidation of material was beginning, the miscibility between magnetite and ilmenite must have been near zero. Diagrammatically this means that there has been a large gap in the range of miscibility, and the minerals practically have been crystallized separately. Besides, as mentioned already in connection with analyses, the ilmenite, owing to its great magnesium content, had a higher melting point and therefore could begin its crystallization earlier. Along with the changes in temperature, the tectonical forces have imparted to the ore mass a continuous movement, chiefly in one direction. Thus it has been possible, for both the ilmenite and the magnetite to have aggregated to their respective phases. The separation of mineral phases has not been thorough, only partial, in the first place because the tectonical movement had weakened and the temperature fallen. Towards the border parts the ilmenite has gradualy become predominant in the ore because of the differences in specific gravities and of the obviously greater mobility of ilmenite material. These border parts have also entered earlier into the conditions of crystallization, in connection with which characteristic features of granoblastic structure have developed, such as breaks, twinning, zonarity and recrystallization (Fig. 5). This manner of crystallization affects the whole mass of ore, however, the chief part of the ore, in which the magnetite becomes predominant, has finally been crystallized later and has also been mobile later on. In the light of the processes described above the superiority of ilmenite is understood, when the ore becomes poorer and at same time also the amount of sulphides increases. Likewise it is comprehensible, that the very smallest individuals crushed from the borders of early crystallized ilmenite grains, having been included in magnetites, resemble the ilmenite flecks of primary origin. Probably such flecks, their nature being that of inclusions, can be distinguished from the primary ones, due to buildings caused by immiscibility of hematite. According to the theory of St. Foslie, mentioned above, such a phenomenon cannot be possible in primary flecks.

In connection with microscopic studies, martite associated with magnetite was found. In many cases it follows the cracks of magnetite, but is, however, most strongly developed around many ilmenite grains. Starting from the border between the magnetite and ilmenite, the martite grows towards the inner parts of magnetite. If the martite is supposed to be secondary, its occurrence around the ilmenite grains establishes in an excellent way the inclusive nature of ilmenite grains (Fig. 2). It is quite clear, that the joins with the host mineral had been loose and thus also were most disposed to secondary changes. The martite may also be a product of the mobile period. This means, that the excess of Fe_2O_3 originally contained in the ilmenite of primary ore, had no longer been preserved in the ilmenite phase during the mobilization, but that immiscibility had in part taken place. The building of martite as an event after the mobile state seems more probable.

In choosing the material attention was mainly paid to the different types of ore, and not to the mode of appearance from the field geological point of view. Therefore it is purposeless to draw more extensive conclusions in this connection. As mentioned in the beginning of this paper, the study performed has been mainly orientative. The author has attempted to give a general view only so far as required by the explanation of the features observed by means of the ore microscope. In the light of this the iron ore of Otanmäki, though it in many details resembles the titaniferous iron ore of early magmatic origin, has gained the prevailing features of its structure as a mobilization product, caused by disturbances of a later tectonical period and thus in its whole, the ore resembles also the Routivare occurrence, explained by P. Ramdohr.

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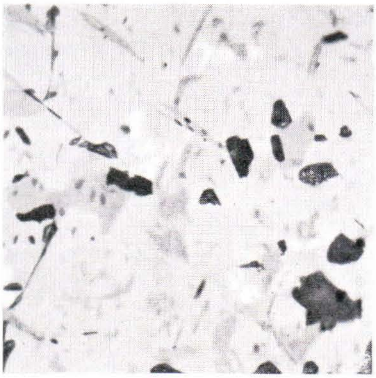
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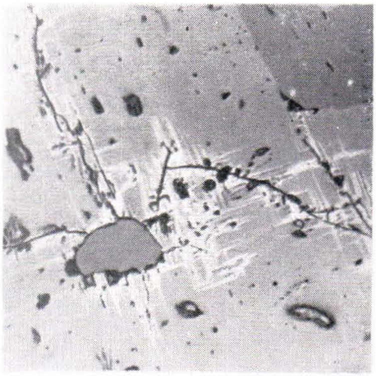
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O. *Vaasjoki:* On the microstructure of titaniferous iron ore at Otanmäki.

EXPLANATION OF PLATE I.

Fig. 1. Ilmenite in magnetite. Magnetite = light gray; ilmenite = slightly darker gray; silicate = black. Oil immersion, magn. 580 x.

Fig. 2. An ilmenite inclusion and martite in magnetite. The lighter sceletal structure is martite. Oil immersion, magn. 580 x.

Fig. 3. Laths of spinel in magnetite. Oil immersion, magn. 580 x.

Fig. 4. Twin lamellae in ilmenite. In the picture there can be noticed the streaks of the first and the flecks of the second generations of hematite in ilmenite. Oil immersion, magn. 580 x.

Fig. 5. Deformed ore. Twin lamellae of ilmenite bent and broken. Undulatory extinction in the ilmenite also noticeable. + Nic., magn. 100 x.

8.

THE SCAPOLITE OCCURRENCE OF PUSUNSAARI

BY

AARNE LAITAKARI

Soboleffski, in 1839, was the first to mention the scapolite occurrence of Pusunsaari in literature. Having commented briefly on the discovery of graphite deposits at Pusunsaari, and on its dolomite, he proceeds in the following manner: (1) »Enfin on trouve encore a Pouzo une roche particuliere, qui parait subordonnée au gneis. Par son extérieur elle se rapproche du feldspath. À des joints naturels tres-marqués, elle unit une couleur violette et un éclat un peu graisseux. Cette roche dans les mains d'un ouvrier habile doit devenir remarquablement belle.» The foregoing is clearly a reference to the scapolite occurrences along the eastern shore of Pusunsaari.

H. J. Holmberg (2), referring to Soboleffski, mentions the scapolite of Pusunsaari in his mineral catalogue of the years 1857—1858, and continues by saying that the scapolite there is regarded as forming a certain material of the Pusunsaari granites.

In 1858 there was published by Axel Gadolin (3) a detailed description of the geology of Pusunsaari, together with an accurate account of the relation of the scapolite fels to the other rocks. This description by Gadolin has been preserved in its main points unchanged in the work by O. Trüstedt on the Pitkäranta area (4). Across the northern part of Pusunsaari there runs in an E-W direction a dolomite limestone layer, in connection with which there appear in two or three places the scapolite fels in question. Along with the main bulk the rock contains in small amounts the following minerals: hornblende, titanite, pyrite, quartz, apatite and molybdenite. Scapolite is coarse-grained and light in color, more generally reddish or violet. Along the northern edge of the limestone layer there is in one place a very impure deposit of graphite in the form of a thin layer in a mica-rich hornblendeschist. Graphite was quarried here already before the year 1838, as according to Soboleffski, the quarry was already then old and abandoned.

Gadolin considers the scapolite as belonging to the under part of the limestone layer but refrains from setting forth any views as to its origin

This deposit of scapolite was trial-quarried in 1936, with the result that new light was thrown on the question. The sketch drawn by me (Fig. 1) and the photographs (Figs. 2 and 3) give a good idea of the scapolite and its coarse-grainedness. Separate scapolite individuals can be up to 50 cm long and 20—30 cm wide. The scapolite rock proper is to a large extent free from foreign inclusions, among which mention can be made of diopside, light oligoclase and quartz, as well as titanite. In the border parts there is also some pyrite and a few flakes of molybdenite. (Close to this scapolite occurrence in the central part of the island, a considerable amount of chalcopyrite has been found.)

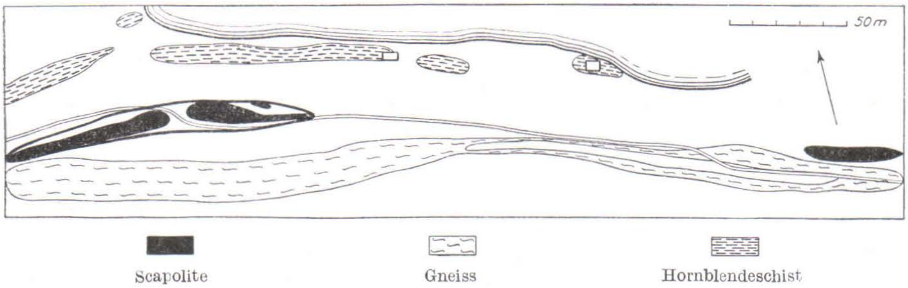

Fig. 1. The scapolite occurrence of Pusunsaari.

The following is an analysis of the Pusunsaari scapolite, performed by Dr. L. Lokka in 1936:

	%	Mol. prop.		Atomic prop.
SiO_2	47,85	0,7935	Si	7935
TiO_2	—	—		
Al_2O_3	27,02	2644	Al	5288
Fe_2O_3	0,27	17		13223 = 12 × 1102
FeO	—	—		
MnO	0,01	1	Mn	1
MgO	0,16	40		
CaO	15,77	2811	Ca	2811
Na_2O	3,88	626	Na	1252
K_2O	0,93	99	K	198
CO_2	2,40	545		4262 = 4 × 1063
SO_3	0,23	29		
Cl	1,55	437	CO_3	545
H_2O+	0.44		SO_4	29
$H_2O—$	0,07		Cl	437
				1011

$(Ca, Na, K)_4 (Si, Al)_{12} O_{24} (CO_3, SO_4, Cl)$

Scapolite thus contains about 25 % marialite and 75 % meionite.

Its density is 2,66, the indexes of refraction being $\varepsilon = 1,560$, $\omega = 1,5960$, $\varepsilon — \omega = 0,0305$.

There are no crystal forms present in this scapolite, but in some places, on cleavage surfaces, there are seen lines parallel to pyramidal faces, which shows that in certain phases of the process of crystallization there have been crystal forms also.

Fig. 2. The largest scapolite outcrop of Pusunsaari.

The scapolite fels presented in the sketch measure: the first 150× 15 and the second 50 × 10 meters. The larger one has formed small knolls some few meters in height (Fig. 2) while the smaller one has formed correspondingly lower ones.

Fig. 3. The scapolite of Pusunsaari. The quarried part, in which can be seen several scapolite crystal individuals.

On the southern side the occurrences are bounded by granite-gneisses, the direction of which coincides with the direction of elongation of the scapolite fels. In the region of contact the granite-gneiss is rich in quartz and strongly deformed. Besides quartz with an undulatory extinction, it contains microcline, oligoclase, hornblende, diopside and calcite with grains of ore, apatite, titanite, chlorite and epidote as accessory materials. The large content of diopside, together with the hornblende which has originated from it, is worthy of notice.

On the northern side of the scapolite deposit the rock is a strongly schistose hornblende-schist of medium grain size. The hornblende of the rock is typical of the amphibolite facies and brownish green in color (γ bluish green $>$ β brownish green $>$ α yellowish brown, $c \wedge \gamma$ 15—22°, $2V_{\alpha} = 60°$, $\gamma - \alpha$ 0,025). As other minerals there are present andesine or labradorite — plagioclase, titanite, quartz, calcite, apatite, chlorite, ore and zircon. Here and there microcline appears in place of plagioclase.

Besides the two scapolite deposits presented in the sketch there are at least a couple of similar formations in the same limestone layer. They are all closely related to the limestone layer, none having been found elsewhere on the island. In this connection there is reason to note that nowhere in the whole Pitkäranta area, in which there are numerous similar limestone deposits, can be found such occurrences of scapolite. Trüstedt in the above-mentioned paper says in regard to the appearance of scapolite in the Pitkäranta area, »Im eigentlichen Erzgebiete ist Skapolith ziemlich selten und nur in geringen Mengen nachgewiesen. Gewöhnlich ist er von Plagioklas, Diopsid, Quarz, Titanit, Apatit, Molybdänglanz und Pyrit begleitet.»

As in the Pitkäranta area in general, pegmatite formations are scattered here and there at Pusunsaari. In certain of them microcline and quartz are present in such a coarse form that for these they have been quarried for technical purposes.

In 1931 I presented my views (5) to the effect that the scapolite occurrence in question had originated from pegmatite having penetrated into the limestone in such a way that the felspar and quartz had been replaced, just as is the case in the limestone areas of Korpo (6) and Parainen (7), in which the pegmatite penetrating into the limestone had become scapolitized.

In the scapolite of Pusunsaari there can be found in places, especially in the vicinity of contact, numerous inclusions of light diopside, so that the rock can properly be called a scapolitic-diopside rock. Besides these there are also present, in places even in considerable amounts, the following minerals: titanite, plagioclase, calcite, quartz, apatite and a little microcline. On the basis of the study of its optical properties on the U-stage, the diopside is practically iron-free ($2\,V\ \gamma = 58^\circ$, $c \wedge \gamma = 38^\circ$, $\gamma - \alpha = 0{,}028$) which fits in well with the hypothesis of the mode of origin set forth above.

Pegmatites are present at Pusunsaari in numerous places, and there has been a weak zone in the wall rock, namely the limestone layer, into which the pegmatite material has penetrated. The pegmatite material, along with the liquids and gases accompanying it, has contained sufficient amounts of SiO_2, Al_2O_3 and alkalies, together with the necessary amounts of sulphur and chlorine. The dolomite-limestone has contained enough CaO, MgO and CO_2. Under such conditions there has formed at Pusunsaari, in place of the ordinary pegmatite, the scapolite fels there present, appearing as coarse-grained masses of the same order and grain size as the closely related pegmatites. The surplus of alkali and SiO_2 has wandered further or changed the adjacent rock, and a part of the Mg of the dolomite, together with the SiO_2 present, has formed diopside, which in these conditions has remained poor in iron content. At the same time there has formed in the same layer some pyrite and molybdenite in a scattered form, together with the slightly larger amount of chalcopyrite in one place.

The scapolite of Pusunsaari thus seems to have formed in the same way as the scapolite-pegmatites of Korpo and Parainen, by liquid and gas-rich pegmatite magmas having penetrated into limestone. The regional-metamorphic mode of genesis as applied to the scapolitic rocks of Northern Sweden (8,9) and Northern Finland (10) cannot be applied in this case.

Why scapolite fels have not formed elsewhere in the Pitkäranta area, where geological conditions are similar, remains unexplained. Conditions at Pusunsaari in particular have been especially favorable for the formation of scapolite, because, as far as I know, such large occurrences of scapolite are indeed rare.

In comparison with other scapolite occurrences those of Pusunsaari are peculiar in that they are so large in size and also because they consist almost entirely of pure scapolite as large crystal individuals.

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9.

GRANITE-PORPHYRY FROM ÖSTERSUNDOM II

BY

L. H. BORGSTRÖM

THE OUTCROPS OF GRANITE-PORPHYRY

The author described already in 1907 an occurrence of granite-porphyry at Östersundom, 20 km east of Helsinki—Helsingfors. In later years, however, several new outcrops of this rock, which is quite rare in this part of the country, have been located. Because of the nearness to Helsinki it is probable that the granite-porphyry vein will often be visited by students interested in geology. A guide to the outcrops may therefore be of use.

The granite-porphyry emerges at several places along a line, which from the little lake Heltingträsk runs in a SSE:ly direction to the island Fårholmen in the neighbourhood of Skyttenskär. The distance is 12 km. It is evident that the fissure into which granite-porphyry intruded was not a continuous open crack of uniform breadth. The width of the dike changes in some places gradually, in others rather abruptly. In a few places the vein splits up into several parallel dikes, as in the vicinity of lake Heltingträsk. It seems that the crack has remained closed along part of its length. The strike as seen in the different outcrops deviates more or less from the general direction of the vein. In Gumbostrand, at Fransman's and at Hartwall's, the strike of the dike is SE. Also on the island Takvedaholmen the pieces of the veins seen in the surface of the rocks run SE or SSE, although the outcrop of Hartwall's lies in a SSE:ly direction from Fransman's and the outcrops of Takvedaholmen nearly S from the vein at Hartwall's. A study of the map (Fig. 2) will make this clear.

The schematic map (Fig. 3) and the insert show a way in which the string of outcrops possibly was made. According to this, the original crack was sinous. At the time of the intrusion, or shortly before it, the walls of the crack were displaced in relation to each other, as shown by the arrows in the insert. The older rocks in which the granite-porphyry intruded are described by Kranck and Wegmann (Bull. de la Comm. géol. de Finlande N:o 89, 1931). The older rocks have a general strike in N60°E, which is proved by the arrangement of the enclosed pieces and of the stripes in the migmatitic granite. These older features are crossed by a few young diabase-veins, which usually are nearly vertical and mostly strike in about N 60° E, or at about right angles to the direction of the granite-porphyry.

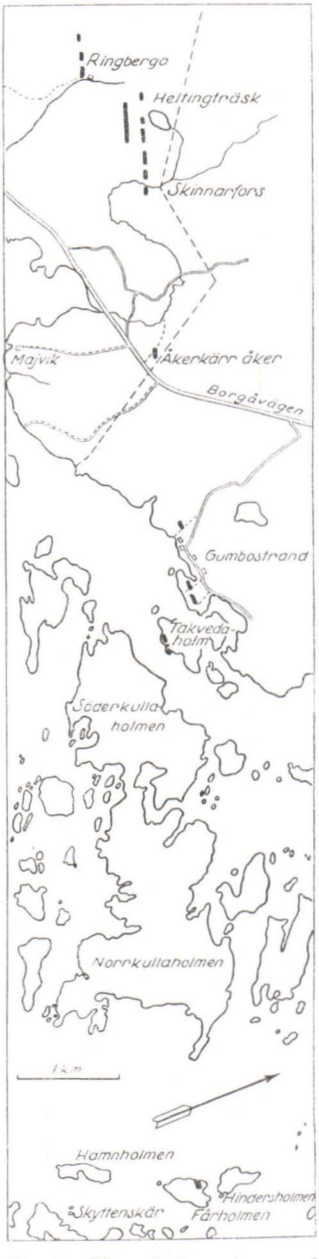

Fig. 1. Map of the outcrops of granite-porphyry at Östersundom.

It is very convenient to visit the outcrop of granite-porphyry which is only 150 m north of the new road between Helsinki and Borgå, 26 km from Helsinki (on the map at Åkerkärr. This place has the number 12 on the map from 1907). — If the visitor proceeds along the highway towards Borgå and takes the road to Gumbostrand 2 km farther on, following it until the sea becomes visible, and then takes a byway to the right along the border of a field, he will pass across the granite-porphyry dike at Fransman's. This is N:o 13 on the map Fig. 2. The vein, which is 7—8 m broad, is here uncovered for a length of about 30 meters. The porphyry has in the middle of the vein a greyish, fine-grained ground-mass. This ground-mass darkens towards the borders, and close to the contact it is vitreous and pitch black. 20—40 cm from the contact, where the ground-mass appears dense, the rock surface cracks and disintegrates under atmospheric influences, forming small lenticular pieces. Both ends of the vein disappear under the overburden of till. The NW end of it points towards the outcrop N:o 12, which is 1.5 km farther away. Although the rock surface is bare on the greater part of the area between N:o 12 and N:o 13, there are no signs of the vein in the interval between these two outcrops. Towards SE the vein probably crosses a diabase vein, which can be seen in front of the main entrance to the building of Fransman's. It is deplorable that the exact crossing of these veins is covered.

The granite-porphyry reappears at the rock surface on the peninsula at Gumbostrand (outcrop N:o 14), 0.5 km from N:o 13. Here the dike attains its greatest width, 10 m. A visitor who walks round the innermost part of the little bay of Gumbostrand, and then ascends the ridge on the peninsula, crosses the dike on the summit of this ridge, and may be able to follow it for about 50 m to the west. The ground-mass of the granite-porphyry is a little coarser than that of the other outcrops, and the different grains of it can be distinguished by the naked eye. (An account of the microscopic investigations is given later on.) Towards SE the vein is soon covered by the moraine, but it reappears 50 m to the east on an area of a few m² as a peculiar, black, lustrous, coal-like coating on the vertical rock which forms the north boundary of the little garden of Hartwall's. This coal-like substance is the contact between the granite-porphyry and the older rock. It stuck to the latter as the clefts to which the little valley owes its origin split up. These clefts show that the crack followed by the porphyry at its intrusion very much earlier still exist.

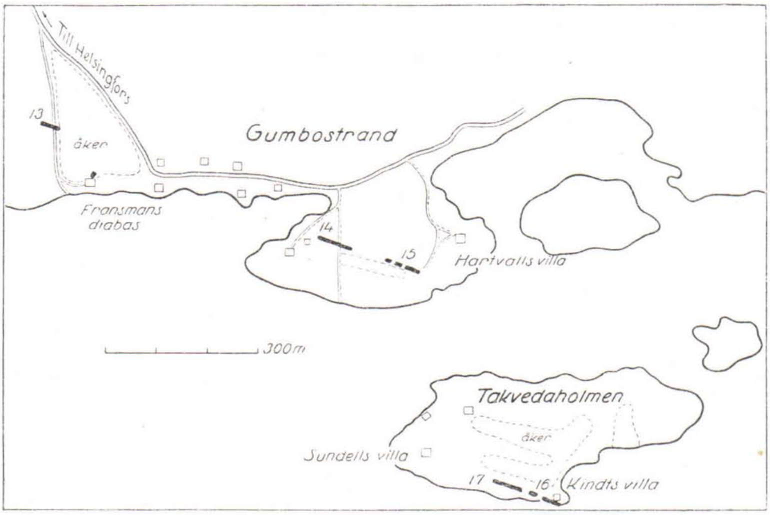

Fig. 2. Map of the granite-porphyry veins at Gumbostrand.

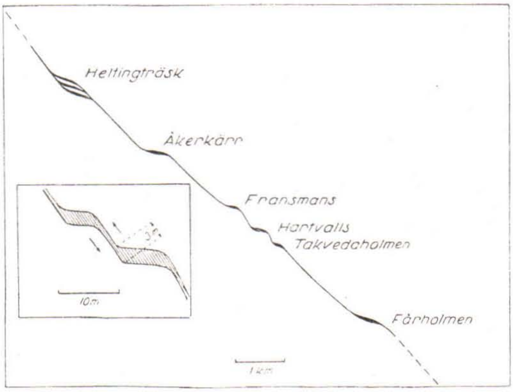

Fig. 3. Schematic drawings of the granite-porphyry veins.

At a distance of a few paces farther to SE the dike continues as a 20 m long and about one meter broad seam, on which runs a path along the foot of the continuation of the above mentioned steep rock-wall.

The granite-porphyry dike appears again at Takvedaholmen, 0.5 km south of the outcrop at Hartwall's garden. As the dikes at Takvedaholmen and at Hartwall's strike in SSE they probably do not form a direct continuation of each other. It may be that the vein has been faulted or that it has been split up into two parallel veins. The outcrops on Takvedaholmen are easy to locate at the cliffs on the southern shore of the island, a little W of the house of Mr. Kindt. A rock surface of migmatitic red granite is here crossed by 1.2 m thick dike of granite-porphyry which can be followed for about 15 m. Near the shore, where the surface of the porphyry is washed clean from lichen, it is possible to discern different systems of cracks in the dike. There are three systems which are nearly perpendicular to the vein and form angles of about 120° with one another. It is possible to find some pieces forming six-sided prisms, which recall the pillars in the famous basalts, Fingal's Cave or Giant's Causeway, and have been formed by the cooling and shrinking of the vein-eruptive. These cracks never cross the boundary line of the dike. Other cracks continue from the dike to the country rock, and bear witness of the stress in the local rock-ground. The outcrop at Takvedaholmen dives towards W under the water and the swampy vegetation of the shore, but the dike appears again 10 m to the west as a small and irregular patch of porphyry. Another 10 m still farther westwards the dike is only half a meter broad, but it can be followed for about 25 m before it disappears under the moraine.

From Takvedaholmen it is 5 km to the next outcrop, on Fårholmen. The rock surface between these two islands, Takvedaholmen and Fårholmen, is for the greater part under the sea level but, anyhow, there are long stretches of rock shores and naked rock surfaces, where a continuation of the porphyry dike would be detected, but so far the search has been in vain. It is probable that the crack has never opened between these last two known outcrops. The dike at Fårholmen was discovered by Wegmann and Kranck, who, however, did not associate it with the granite-porphyry of Östersundom.

The vein appears on Fårholmen as a dark seam 4.5 m in thickness, which crosses the belt of reddish, granitic rocks at the shore on the NW side of the island. On closer inspection of the rock-surface one perceives small porphyric crystals of felspar and quartz in a dark ground-mass, and between them rounded or oblong patches, one to two mm in diameter, the central parts of which look like small cavities in the weathered surface of the rock. Under the microscope these turn out to be miarolitic druses which in the outer part of the vein are rounded and related to the amygdaloid structure. The minerals in the miarolitic druses are quartz, mica, and felspar, with calcite filling the central parts.

Where the surface of the granite-porphyry is fresh, or where it has been cleaned by ice and waves, it is dark grey in colour. A surface which has been longer weathered is lighter grey. In the vicinity of the contact, where the ground-mass is glassy, the rock is split up by the weathering. Small concave depressions, only a few cm across, lie beside one another, and are covered by sharp-edged splinters. This formation, which is less than half a meter broad, is a distinctive feature which may be used when searching for dikes where the migmatite of the country rock is grey or dark. If the migmatite is light-coloured or reddish, the granite- porphyry is easy to distinguish from it on the rock surface.

MICROSCOPIC FEATURES

Where the dike attains a thickness of 10 m the groundmass of the central part of the dike shows a microgranitic structure in the microscopical slides. The porphyric grains are of about the same size and occur in the same number as in the northern occurrences of the dike, which were described in 1907. The porphyry grains vary greatly in number. One sample, 10 by 15 cm, may show a dozen porphyric grains, another only a few. In a usual thin-section 2—5 porphyric crystals are often found, and usually there are a greater number of the miarolitic druses, which, because of the coarse grains of their quartz, appear as light spots like the phenocrysts. Towards the boundaries the microscopic characteristics of the rock correspond to the description of 1907. Nearer the centre the structure varies, the structure of the ground-mass sometimes becoming aplitic. In some places, as for instance in the neighbourhood of the miarolitic druses, highly developed granophyric and micropegmatitic structures appear. A careful search using high magnification reveals a great variety of granophyric structures distributed in all parts of the ground-mass. Sometimes they resemble common micro-pegmatite (Fig. 4d), sometimes they constitute structures in which the felspar forms a system of lamellae between which the quartz is interbedded (Fig. 4c). In some cases the felspar forms small idiomorphic crystals surrounded by an aureole of allotriomorphic quartz (Fig. 4b). Or again an idiomorphic crystal of felspar lies swimming in clear quartz into which a seam of felspar needles grows out from the crystal at right angles to its faces (Fig. 4a). The felspar needles are not

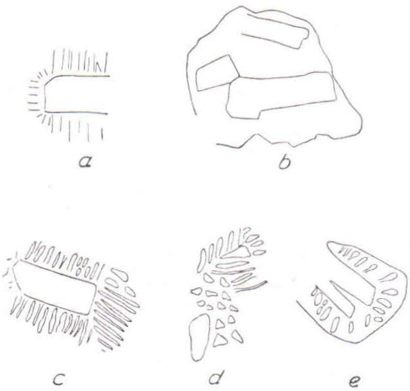

Fig. 4. Granophyric structures × 200.

always straight and may vary in shape (Fig. 4d). The quartz which encloses felspar lamellae or felspar-crystallizations of the above kind is usually quite allotriomorph against the felspar individuals which surround it. In those parts of the dike in which the felspar forms greater and better crystallized grains, also the mica and the hornblende show less jagged shapes, and their tendency to form poikilitic structures and long-drawn crystal skeletons is less apparent than in the outer zones of the dike which have been more quickly chilled. — The grains of the ground-mass in the middle of the thicker dikes are 0.1 to 0.15 mm. The phenocrysts here measure 0.5 to 1.0 mm, and in general they do not differ in size from the phenocrysts in the part with subcrystalline or glassy ground-mass. The quartz phenocrysts are smaller and less numerous than the felspar crysts.

THE AGE OF THE GRANITE-PORPHYRY

The granite-porphyry intruded into and was solidified in comparatively cold surroundings, as shown by its vitreous borders. Dikes of equal thickness do not show chilled borders when traversing rocks which were still warm at the time of the intrusion. Such dikes, *e. g.* the granite dikes in the Onas granite, show smaller grains than the Onas granite proper, but are still fine-grained granites and do not, like the granite-porphyry of Östersundom, show structures which bring to mind those of hypabyssal rocks. It seems that even in comparatively »cool» surroundings only narrow dikes have vitreous borders. Eruptive dikes of 30 meters and more in thickness seem to be able to heat the country rock to such an extent that the solidification of the dike becomes slower and permits of the formation of structures similar to those in abyssal rocks.

The country rock of the dike, the migmatite, is with good reason believed to be generated by the heating and partial re-melting of strata which were downfolded to a great depth. Because of the isostatic balance these strata were lifted back almost to the surface of the earth when the slow process of denudation had removed the overburden. In this way they were brought into »cooler» zones of the earth's crust, where also the static pressure was low enough to permit the rocks to be brittle. That the country rock of the porphyry was brittle is proved, not only by the nature of the fissure into which the rock intruded, but also by the study of thin-sections from the contact between the glassy border of the dike and the country rock, which show that the rock was crushed and to some extent ground to a fine powder by the movements of the walls of the crack against each other before the intrusion happened. Stresses in the brittle part of the earth's crust had produced the cracks into which the porphyry intruded. From these circumstances we may infer that the granite-porphyry is very much younger than the migmatite.

As already mentioned in 1907, the chemical composition of the granite-porphyry does not differ very much from the composition of the Onas granite, which is at a distance of about 6 km from Fårholmen. A continuation of the dike would run into the Onas granite at its southernmost occurrence on the island of Rågskär. This seems to point to the possibility that the granite-porphyry may be an offshoot of the Onas granite.

The Onas granite intruded into brittle and not into plastic strata, as the beautiful breccias along its southwesterly border prove. The country rock was not cool enough to modify the structure of the Onas granite to fine-grained or to quartz-porphyric modifications at it contacts. In the breccias and for a distance of a few hundred meters from the contact the grains of the Onas granite are half their usual size. Now this has been thought to signify that the Onas granite intruded into strata which, because of greater distance from the earth's surface, were warmer than for instance the country rock when the granite-porphyry intruded. If one considers the great bulk of the Onas granite, about 20 km^2, one may draw the conclusion that the differences in structure of the main mass and of a stringer like the granite-porphyry can be ascribed to the differences in bulk of the eruptives. It does not necessarily mean that the porphyry was intruded into decidedly cooler strata than the granite stock. The granite-porphyry may thus belong to the Onas granite. The Onas granite, as also the rapakivi granites and the granite-porphyry, has been spared from folding and differs in this respect from the older rocks in the neighbourhood. The hypothesis that the contacts of the Onas granite would signify that this rock was solidified at a considerable depth and because of that could not be connected with the granite-porphyry is contrary to the conception that the volcanic rocks of Hogland are offshoots of the great Wiburger rapakivi granite of Viborg, which conception has been verified by observations on the island Someri[1], where rapakivi granite with chilled, quartzporphyric borders against the labradorporphyrite occur. In other places the rapakivi contacts are quite similar to the contacts of the Onas granite.

Regarding the relations between the granite-porphyry and the numerous diabases, which occur along the north shore of the Gulf of Finland, it may be stated that they, like the granite-porphyry, form narrow dikes with vitreous borders, and do not show any signs of regional metamorphism. Because of this they are considered to belong to the youngest Precambrian rocks. The author believes them to be older than the Onas granite, because he has in vain looked for diabase veins in the Onas granite, although such veins are common in the surrounding territories. Because of the same reasons the rapakivi of Viborg must be younger than the diabases. Diabase dikes are not found in the rapakivi district but occur in the older rocks to the west and to the east of it. If these diabases are older than the Onas granite, they are also older than the granite-porphyry of Östersundom.

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[1]) W. Wahl: Geol. Fören. Förhandlingar. 1938. p. 88.

PLATE I

Fig. 1. The granite-porphyry dike on the south east of the island Takvedaholm.

Figs. 2 and 3. Microphotographs with crossed nicols × 30. Micropegmatitic structures.

Fig. 4. Microphoto × 200, showing granophyric structures.

Fig. 5. The granite porphyry dike on the north shore of the island Fårholmen.

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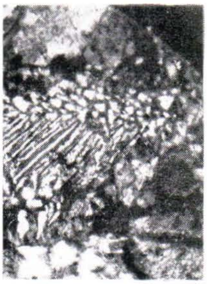

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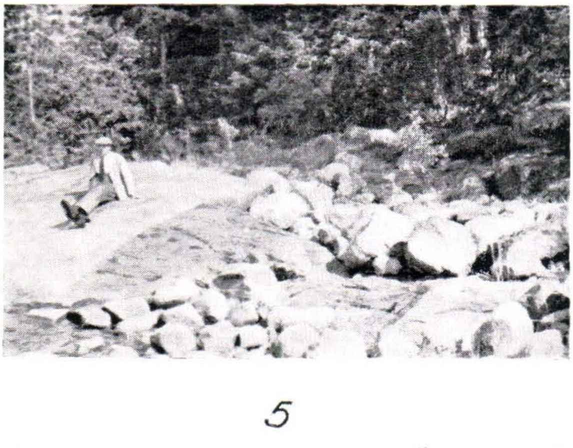

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L. H. Borgström: Granite-porphyry from Östersundom II.

10.

SOME RECENT TRENDS IN THE GEOCHEMICAL INVESTIGATION OF THE LITHOSPHERE

BY

KALERVO RANKAMA

Geochemistry is defined as the doctrine of the chemical composition and chemical evolution of the Earth. The field of geochemistry, compared with its parent sciences, geology and chemistry, is not very old. True enough, the name »geochemistry» was used already a little more than a hundred years ago, or in 1838, by the Swiss chemist Chr. F. Schönbein. However, the classical book »The Data of Geochemistry» written by the American scientist F. W. Clarke in 1908 is to be considered as beginning the evolution of geochemistry in the modern sense. Since Clarke's time, this branch of science has been in a state of constant progress leading, until approx. 1920, to the discovery of much fundamental knowledge on the geochemistry of the different spheres of the Earth. Among the more important contributions to the geochemistry of the lithosphere the work of the Norwegian scientists W. C. Brøgger, J. H. L. Vogt, and V. M. Goldschmidt must be especially mentioned.

Beginning with the third decade of the Twentieth Century a new era started in geochemistry, leading to the discovery of the fundamental laws governing the distribution of the elements, and mainly due to the persistent work of V. M. Goldschmidt. The new progress is found to be partly incorporated with the evolution of modern physics, more especially of atomic and nuclear physics, and, partly, with the development of new chemical and physical methods of determination of the elements, particularly optical and X-ray spectrochemical analysis. Accordingly, the geochemical distribution of most of the elements is known, mainly by the work of V. M. Goldschmidt and George Hevesy and their co-workers, and of Ida and Walter Noddack.

It seems to the author that a new phase of evolution is beginning, once again, in geochemistry, more especially in the study of the lithosphere. As to the remaining branches of geochemistry, there is ample proof of a lively activity. To mention some authors of most recent contributions, there are E. J. Conway and H. Pettersson working on the Oceanic geochemistry, G. E. Hutchinson on biogeochemistry, and R. Wildt on chemistry of the stellar atmospheres.

With especial reference to the investigations dealing with the geochemistry of the different elements in the lithosphere, most of them leave something out of consideration, *viz.* the question of the regional distribution of the elements, as has been pointed out by the present author on a previous occasion (Rankama, 1944, p. 63). The potential value of regional geochemical investigations was emphasized previously by Sahama and Rankama (1938). This branch of geochemistry has, since then, received considerable attention in the working program of the Geochemical Laboratory at the Institute of Geology, University of Helsinki, a laboratory built under the auspices of Professor Pentti Eskola. The work carried out at Helsinki has, thus far, dealt with the distribution of the rare earths (Sahama and Rankama, 1938; Sahama and Vähätalo, 1939), rubidium and caesium (Erämetsä, Sahama, and Kanula, 1941), and tantalum and niobium (Rankama, 1944, 1947), more especially in the Finnish Archean formations. By these investigations it has been found possible to divide the rocks into certain provinces and formations showing marked similarities in their content of certain of the rarer elements, often termed Index or Pilot Elements (*cf.* Goldschmidt, 1932). It further seems to be possible to extend the results of regional geochemistry to establish the order of the Pre-Cambrian granites of Finland according to their ages, with due respect to their connections with the orogenic cycles.

Important recent contributions to regional geochemistry are two papers by Sahama (1945 a and b) dealing with the trace elements in the rocks of South Lapland and in East Fennoscandian rapakivi granites. In Sweden Lundegårdh (1946) has published a rather copious volume on the rock composition and development in Central Roslagen, largely based on geochemical considerations. As a result of the work carried out nearly exclusively in Finland and in Sweden the Pre-Cambrian granites are now decidedly the best-known rocks as to their geochemical features.

The importance of regional investigations in geochemistry is well understood also in many countries outside Fennoscandia. In this connection the large work on the composition of the Earth's crust in the East Indian Archipelago by van Tongeren (1938) must be mentioned.

An attempt at the geochemical characterization of the Pre-Cambrian of Finland on the basis of the granites has been made by the author (Rankama, 1946) together with a discussion on the geochemical migration of the elements on global scale, as illustrated by some geochemical features of granitization.

The application of geochemistry to problems connected with the formation of ore deposits has been studied by Landergren in Sweden on several occasions. Thus far, the most recent of his papers (1943) deals with the geochemistry of iron ores of the Grängesberg Mining Field. This paper contains a series of important results in the interpretation of the origin of these ores. By the study of the distribution of certain elements in the ores and in their country rocks Landergren decides that the ores are of secondary magmatic or palingenic origin. The important role of geochemistry in the study of ore deposits is easily detected by the examination of Landergren's paper cited above, and still more interesting results are to be expected from his work on the geochemistry of all Swedish iron ores which is, at present, close to its completion. It is not possible, in this place, to refer to the large number of papers published in different countries, dealing with the geochemistry of ores and ore-bearing formations. The important recent papers of Oftedal (1940) in Norway and of Gavelin (1943) and Gabrielson (1945) in Sweden can be mentioned only in passing; the many American contributions to these subjects seem, with few exceptions, to be based on information of more or less qualitative character.

By the survey presented above the fact will become evident that there has occurred, during the past ten years or so, a definite trend in geochemistry towards its application to problems of general and regional geology and petrology. Compared with the above contributions, certain other recent work, pertaining to the gathering of more basic knowledge of the laws ruling over geochemistry, thus far, seems to fall behind in number of papers published and in their dimensions. To this very important class belong several notable recent papers, published in Sweden by Wickman (1941, 1943, 1944), the first of which deals with the calculation of the total amount of coal and bitumen with the use of the coal isotopes. In the second paper in the series the geochemistry of igneous rocks is discussed with the use of kinetic reasoning with temperature as an essential feature instead of the static way of treatment previously applied by Fersman and Goldschmidt. The third paper is a contribution to the geochemistry of the elements in sedimentary rocks where the division of the elements is based on the rules governing the hydrogen and hydroxyl bonds instead of the concept of ionic potential previously used by Goldschmidt.

Other theoretical contributions to the geochemistry of the lithosphere will, most certainly, become available by the application of thermodynamic and thermochemical principles to the domain of petrology. Thermodynamic considerations have been offered, in numerous recent papers, by several young Norwegian geologists (J. Bugge, H. Ramberg, I. Th. Rosenqvist). At the present, however, no uniformity in their opinions seems to have been reached. The application of thermochemistry to petrology, on the other hand, is relatively new, mainly due to the fact that suitable data are, thus far, not available. Perhaps the most modern thermochemical contribution is one offered by Sahama (1945 c) and dealing with the abundance relation of fluorite and sellaite in rocks. It is evident that the proceeding exactization of petrology will bring forward contributions important also to the general geochemistry of the lithosphere.

The above examples of some recent work carried out on the geochemistry of the lithosphere might well suffice as evidence of active evolution now prevailing. The classical geochemistry of the lithosphere, best termed, perhaps, as mineralogical geochemistry, is giving ground to modern trends of geological geochemistry on regional, possibly even on a global scale. Along with these trends, more or less descriptive in their nature, theoretical geochemistry is absorbed in detecting the primary causes of geochemical phenomena and the reason of geochemical laws. As pointed out by Wickman (1946, p. 150) the experimental geochemistry is evidently apt to make important new progress with the aid of modern physical tools: separated isotopes and artificially radioactive elements. One of the *lacunae* in our present-day geochemical knowledge is the geochemistry of sediments, as is also stated by Wickman *(loc. cit.)*. It seems to the author, however, that much light will be shed on this problem by the study of the geochemistry of the Ocean bottom, now more easily accessible to science than ever before, due to the new sampling equipment. Other problems, evidently much promising, include the investigation of the geochemistry of granitization and of metamorphism. As to the last-mentioned problem, only very little information is, as yet, available of this vast field of phenomena.

The present author has been privileged to meet recently many of the geochemists from different countries and from different continents, active with problems concerning the geochemistry of the lithosphere. This contact has made evident that many of the new ideas and problems are apt to appear contemporaneously, paying no heed to distances and geographic boundaries. There also seems to be definite proof of future international co-operation in the domain of the geochemical research of the lithosphere.

SARATOGA SPRINGS, NEW YORK,
FEBRUARY, 1947.

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11.

ON THE REACTION QUARTZ GLASS-HIGHCRISTOBALITE FROM THE THERMODYNAMICAL POINT OF VIEW

BY

NILS EDELMAN

The conditions of equilibrium of the minerals are in many ways practically uninvestigated. The doctrine of mineral facies, which is based on field and microscope studies of the mineral paragenesis, does not explain the cause relations, the thermodynamical basis of the variations in the mineral compositions of the rocks. Therefore the temperature and pressure scale is also relative. But if we could calculate thermodynamically the stability fields of the minerals, the limits of the facies fields would be fixed absolutely in degrees and atmospheres. The thermodynamical constants of the silicates are unfortunately very defectively known. An exception is the silicondioxide. The specific heat c_p of its different modifications is determined for all temperatures up to the melting point. Rudolf Wietzel has earlier thermodynamically calculated the conditions of stability of the different SiO_2-minerals, but for the pair quartz glass — cristobalite he gets a result, which does not correspond with the experience as the melting point is 2300—4000 °C instead of 1710° C. In his calculations he started from the absolute zero and the faults of c_p adding up obviously caused, the difference to become so great at high temperatures. I have now calculated the same reaction, basing on newer values of c_p. In order to reduce the faults I have moreover set out from the melting point, because the interesting field is closer to this point than to the absolute zero. In this way I could not calculate the absolute value, but merely the differences of the energy content and the entropy between quartz glass and cristobalite. These are enough for a thermodynamical calculation of the reaction

$$SiO_2\text{-glass} \rightarrow SiO_2\text{-cristobalite}$$

The specific heat c_p of a glass is generally higher than that of the same substance in the crystalline state. Table 1 and fig. 1. show, how c_p of quartz glass and cristobalite varies with the temperature according to different tables. The Roman numerals in the tables and figures refer to the following works: I International Critical Tables (4 p. 105), II Roy W. Goranson (2 p. 231) and III Herman Ulich (5 p. 319). The quite considerable differences between the various values can depend on different methods of determination, but can also partly be connected with the previous history of the substance. Thus for instance the transformation point and the heat of transformation of cristobalite vary according to the temperature of formation (4 p. 106)

Temperature of formation	1100° C	1600° C
Transformation point	230° C	270° C
Heat of transformation	4.2 joule/g	10.5 joule/g

As Table I and fig. 1 show, the c_p of the quartz glass sinks more rapidly with sinking temperature than the c_p of the cristobalite. Below the point where the curves cut each other they are almost parallel. This fact appears better in fig. 2, which shows the difference c_p-glass — c_p-cristobalite. Below the just mentioned point the specific heats of quartz glass and cristobalite are almost equal, according to the experimental values even smaller. Though the curves taken absolutely are quite different, it is, however, remarkable, that they all have such a point.

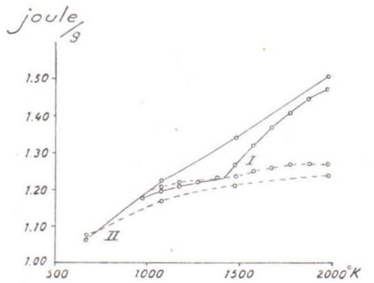

Fig. 1. c_p of quartz glass and cristobalite. Quartz glass — whole line, cristobalite — broken line.

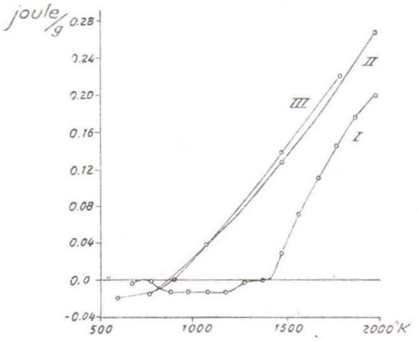

Fig. 2. The difference between c_p of quartz glass and cristobalite.

The cooling of molten SiO_2 can take place in two ways. Either the cristobalite crystallized at 1710°C, where the melting heat or heat of crystallization is emitted, or the melt supercools into quartz glass. The difference between the energy content of the glass and the cristobalite is at the melting point quite equal to the melting heat. As the temperature sinks, the difference diminishes, because the glass emits more heat owing to its greater c_p. The difference between the quantities of heat, which the glass and the cristobalite emit at cooling, is calculated by multiplying the sinking of temperature by the means of the c_p differences, commonly for intervals of 100 degrees, as Table II shows. When we subtract these values from the heat of crystallization, we get the difference between the energy-contents of the glass and the cristobalite, or in other words the heat of crystallization at the temperatures in question. Fig. 3 shows the results. It is notable, that this difference is about zero at the temperature where the c_p-curves cut each other. The energy-content of the crystalline modification is there much the same as that of the amorphous one. Below this point the heat of crystallization of the cristobalite or the heat emitted by the reaction quartz glass-cristobalite is almost zero in a temperature field of many hundred degrees, because the specific heats of both phases there are practically the same.

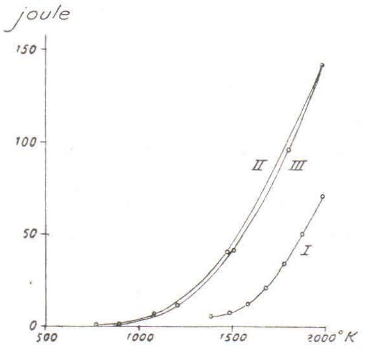

Fig. 3. The difference between the heat content of quartz glass and cristobalite or in other words the reaction heat of the reaction quartz glass → cristobalite.

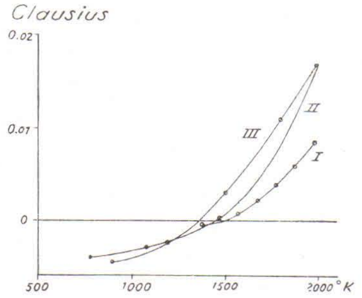

Fig. 4. The difference between the entropy of quartz glass and cristobalite.

To explain the conditions of stability we must besides the heat of reaction know the change of entropy by the reaction. This is calculated by multiplying the sinking of temperature by the means of c_p/T. The results are shown in Table III. Fig. 4, again, shows the difference between the entropies of the quartz glass and the cristobalite or the change of entropy by crystallization of the cristobalite. This is negative at higher temperatures, positive at lower. The temperature, where the change of entropy is zero, is higher than that, where the heat of reaction is zero, which is very natural, because the cristobalite has emitted more heat (the heat of crystallization) at higher temperature than the glass.

The direction of a reaction is determined by the change of the free energy according to the following formula

$$\Delta\ Up = \Delta\ Fp + T\ \Delta\ Sp \quad (5\ p.\ 89)\ \text{or}$$
$$\Delta\ Fp = \Delta\ Up - T\ \Delta\ Sp,$$

where Δ Fp is the change of the free energy, Δ Up the heat of reaction, Δ Sp the change of the entropy, all these at constant pressure, and T the absolute temperature (°K). If Δ Fp is negative, the reaction will take place spontaneously. In the present instance Δ Up is negative in the point, where Δ Sp is zero, and Δ Sp positive i. e. — T Δ Sp negative in the point, where Δ Up is almost zero. Therefore Δ Fp is negative on the whole field, wherefore the cristobalite is more stable than the glass. To the extent that the thermodynamical constants of SiO_2 in the future will be corrected, the result of this calculation can change for instance so that the quartz glass proves to be more stable than the cristobalite in some tempereture interval below the melting point, or in other words that the cristobalite has a second melting point, where the melting takes place by sinking temperature, as suggested by Wilhelm Eitel (1 p. 260). Even though this might not be the case, it is very interesting, that the heat of reaction in a great temperature interval is almost zero. Here it must be pointed out, that high quartz or tridymite are the stable modifications of silicondioxyde at the temperatures in question. That the entropy of the cristobalite is greater than of the quartz glass means, that the energy is more ordered in the »amorphous glass» than in the crystalline cristobalite. The glass seems to lie structurally between the stable highquartz or tridymite and the cristobalite, or in other words the glass is obviously cryptocrystalline. As comparison we can refer to the c_p-curves of glassy and crystalline glycerine, which curves have a similar course (1 p. 280.) The specific heat and the entropy are, however, at all temperatures greater in the amorphous glycerine than in the crystalline.

The results which we have attained touch in the first place the crystal chemistry and the thermodynamic, but if other silicate glasses too behave in the same way as the quartz glass, this fact can illustrate geological questions from a thermodynamical point of view as for instance the granitization, where the pressure will play an important role. It is moreover conceivable, that the silicate chemical industry can make use of the results of this investigation, e. g. by examination of crystallization phenomena in glass pulps.

		Quartz glass		Cristobalite		Difference
I	1973°K	1.472 joule/g		1.27 joule/g		0.202 joule/g
	1873	1.447		1.27		0.177
	1773	1.417		1.27		0.147
	1673	1.371		1.26		0.111
	1573	1.321		1.25		0.071
	1473	1.267		1.237		0.030
	1373	1.229		1.230		— 0.001
	1273	1.221		1.225		— 0.004
	1173	1.208		1.221		— 0.013
	1073	1.196		1.208		— 0.012
	973	1.175		1.187		— 0.012
II	1983	1.508		1.238		0.27[1]
	1473	1.34		1.21		0.12
	1073	1.21		1.171		0.039
	673	1.06		1.074		— 0.014
III		joule/g	cal/mol	joule/g	cal/mol	joule/g
	1800	1.442	20.8	1.220	17.6	0.222
	1500	1.346	19.4	1.207	17.4	0.139
	1200	1.250	18.0	1.180	17.0	0.070
	900	1.138	16.4	1.138	16.4	0.000
	600	1.013	14.6	1.033	14.9	— 0.020

Table I. The specific heat c_p of quartz glass and cristobalite and the difference c_p glass — c_p cristobalite.

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[1] Calculated according to the formula $cp = a + bT - c/T^2$, quartz glass : $a = 0.892$, $b = 0.311 \cdot 10^{-3}$, $c = 0.021 \cdot 10^5$, cristobalite: $a = 1.191$, $b = 0.032 \cdot 10^{-3}$, $c = 0.625 \cdot 10^5$ (erratum in the table (2 p. 231), where the value is $0.0625 \cdot 10^5$).

	I		II		III
	joule/g		joule/g		cal/mol
1983°K		1983°K		1983°K	
1973	2.02	1473	102.0	1800	659
1873	18.95	1073	33.8	1500	780
1773	16.20	773	5.9	1200	450
1673	12.90		141.7	900	150
1573	9.10				2039 = 141.5 joule/g
1473	5.05				
1373	1.45				
	65.67				

Melting heat I 71 joule/g (4 p. 106)
II 142 » (2 p. 238)

Table II. The difference between the heat content, which the quartz glass and the cristobalite emit at cooling in mentioned temperature intervals. As comparison the melting heat of the cristobalite.

		Quartz glass		Cristobalite	
		c_p/T.	Clausius/g	c_p/T	Clausius/g
I	Melting entropy				0.0086
	1983°K	0.000177		0.000153	
	1873	185	0.0199	162	0.0173
	1773	191	188	171	167
	1673	196	194	180	176
	1573	201	199	190	185
	1473	206	204	200	195
	1373	214	210	214	207
			0.1194		0.1189
II	Melting entropy				0.0171
	1983°K	0.000182		0.000.149	
	1473	217	0.1017	196	0.0879
	1073	270	0.0973	261	0.0914
	773	340	0.0915	340	0.0902
			0.2905		0.2866
III	Melting entropy				0.0171
	1983°K	0.000184		0.000148	
	1800	192	0.0344	162	0.0284
	1500	215	611	192	531
	1200	249	696	235	640
	900	302	826	302	806
			0.2477		0.2432

Table III. c_p/T calculated from Table 1 and the difference of entropy in Clausius/g in the mentioned temperature intervals calculated from the means of c_p/T (c_p in cal/g)

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12.

DIE SYSTEMATIK DER GESTEINE UND MINERALLAGERSTÄTTEN

VON

PAUL NIGGLI
(ZÜRICH)

1. EINLEITUNG

Jede natürliche Klassifikation der Gesteine und Minerallagerstätten muss verschiedenen Gesichtspunkten Rechnung tragen:

1. Sie muss übersichtlich sein und so angelegt werden, dass die in ihr enthaltene Grossgliederung, ohne grundlegende Aenderung des allgemeinen Klassifikationsprinzipes, verfeinert werden kann.

2. Sie muss die Ergebnisse der Forschung widerspiegeln, ohne durch allzu starre Platzanweisung problematischer Bildungen in den Fehler zu verfallen, an sich zum Streitobjekt zu werden. Es wird jedoch immer so sein, dass im Einzelfall hinsichtlich der Zuordnung eines Gesteins die Meinungen auseinanderfallen können. Umfasst die Systematik alle wissenschaftlich wirklich begründeten Bildungsmöglichkeiten, so bedeutet dies nur Platzwechsel im gleichbleibenden Rahmenwerk. Nimmt ausserdem die Klassifikation Rücksicht auf die Schwierigkeiten der Abtrennungen, indem sie das verschiedenartig Deutbare nicht allzu sehr auseinanderreisst, so bleibt sie neuen Erkenntnissen am besten anpassbar. Jede gute Klassifikation, die grosse Erfahrungs- und Wissenskomplexe zusammenfasst, ist ja für den Fortschritt einer Wissenschaft sowohl förderlich wie hinderlich. Günstig wirkt sie, indem sie das Augenmerk auf oft beobachtbare Zusammenhänge lenkt; ungünstig kann ihr Einfluss sein, weil die notwendige Schematisierung Beziehungen anderer Art zurücktreten lässt, deren Bedeutung vielleicht lediglich zu einer gegebenen Zeit unterschätzt wird. Trotzdem darf daraus nicht der Schluss gezogen werden, die pädagogisch massgebende Klassifikation habe sich überhaupt aller Deutungsversuche zu enthalten, d. h. sie dürfe nur äussere Kennzeichen verwerten. Eine solche Klassifikation, sofern sie nicht für ganz bestimmte engere Zwecke aufgestellt wird, könnte das lebendige Interesse am Gegenstand abtöten und daher gefährlicher wirken als ein Einteilungsversuch, der gewisse genetische Möglichkeiten anderen gegenüber zu stark betont. Das rein wissen-

schaftliche Interesse der Gesteinskunde und Lagerstättenlehre wird immer den Bildungsbedingungen dieser Grosseinheiten der Lithosphäre gelten, so dass jede der Forschung dienende Systematik in erster Linie das genetische Moment zu berücksichtigen hat.

3. So sehr es den an Physik und Chemie geschulten Mineralogen und Petrographen lockt, die Gliederung von Anfang an nach rein physikalisch-chemischen Prinzipien vorzunehmen, so sehr muss er dieser Versuchung Widerstand leisten. Die massgebenden Faktoren treten in einer Kombination auf, die für die äusseren Teile der Erde charakteristisch ist. Die Geologie sucht ihrerseits diese Vorgänge zu klassifizieren und an dieses Vorgehen muss angeknüpft werden. Dann und nur dann kann es sich um etwas handeln, was den Namen »natürliche» Klassifikation verdient.

4. Hat einerseits die Systematik so zu erfolgen, dass in ihr die Erkenntnisse der Forschung verarbeitet sind und dass durch sie neue Untersuchungen angeregt werden, so muss sie anderseits gestatten, den Aufbau eines Erdrindenstückes in zweckmässiger Weise darzustellen und abzubilden. Gewiss stellt die geologische oder petrographisch-lagerstättenkundliche Kartierung ihre besonderen Anforderungen, aber es darf nicht so sein, dass eine wissenschaftliche Systematik der Objekte für eine Gliederung und Zusammenfassung derselben im Kartenbild völlig unbrauchbar wird. Anderseits ist zu bedenken, dass gerade die in Kartenwerken benutzten Einteilungen und Namengebungen besonders hartnäckig beibehalten werden, während jede natürliche Systematik entwicklungsfähig sein muss. Das führt wieder dazu, mit grosser Sorgfalt abzuwägen, wie weit man in der genetischen Interpretation gehen darf.

Nun wissen wir, dass in einer Einheit, wie dem Erdmantel, durch das Zusammenspiel der Kräfte Uebergänge zwischen verschiedenen Bildungsarten auftreten werden, somit jede auf genetischen Prinzipien beruhende Einteilung irgendwo künstliche Grenzlinien zu ziehen hat, d. h. mit Definitionen arbeitet, die konventionellen Charakter tragen.

Bei dieser Sachlage ist es selbstverständlich, dass jeder Klassifikationsversuch angreifbar wird und bis zu einem gewissen Grad eine persönliche Färbung besitzt. Doch entbindet dieser Umstand nicht, sich zu jeder Zeit die Frage zu stellen, ob Modifikationen gebräuchlicher Versuche sowohl als Einführungen in die Lehre von den Gesteinen und Minerallagerstätten wie auch als Rahmenwerk für die wissenschaftliche Weiterforschung geeigneter sind als das Überlieferte, weil sie Missverständnisse, die im Laufe der Zeiten entstanden sind, beseitigen helfen. Es ist nämlich merkwürdig, wie oft in der Rückschau durch jüngere Forscher ältere Darstellungen simplifiziert werden und von gegensätzlichen Schul- und Lehrmeinungen gesprochen wird, die als solche nie existiert haben. Es mag dann sein, dass eine Umgruppierung, die an sich wenig ändert, Streitigkeiten ohne reale Grundlagen zum Verschwinden bringt.

2. AUSGANGSPUNKT

Seit Ende des letzten Jahrhunderts hat sich eine an sich logische Gliederung in Bildungen sedimentären, magmatischen und metamorphen Ursprungs durchgesetzt und erhalten. Alle drei Bildungsweisen stützen sich auf direkte Beobachtungen in der Natur und auf Experimentaluntersuchungen (Beobachtung und Nachahmung von Verwitterungsvorgängen, Untersuchungen an rezenten Sedimentbildungen, Beobachtungen an Vulkanen und Experimentaluntersuchungen an Schmelzlösungen und wässerigen Lösungen hoher Temperatur, direkte Beobachtungen der Metamorphose von Schnee, Eis und organogenen Ablagerungen und mannigfache Tastversuche in verschiedenen Systemen.) Es *gibt* somit in der Natur Mineralvergesellschaftungen, die ihren jetzt beobachtbaren Charakter durch Sedimentation, magmatische Erstarrung oder Metamorphose erhalten haben, und man kann höchstens die Frage stellen, ob andere prinzipiell verschiedene und gleichberechtigte Grossvorgänge übersehen wurden. Auch bleibt fraglich, wie die Abgrenzungen vorgenommen werden müssen.

Dabei ist selbstverständlich, dass der Komplex der Metamorphose innerhalb der Lithosphäre die geringsten Möglichkeiten direkter Beobachtungen darbot und daher nur auf spekulativem Wege erschliessbar war. Von fundamentaler Bedeutung bleibt, dass von 1870 an in den ältesten Gesteinen archäischer Komplexe Relikte gefunden wurden, die zeigten, dass manche Gesteine früher Sedimente waren. Die skandinavischen und russischen Forscher wie Törnebohm, Reusch, Kolderup, Inostranzeff, De Geer, Sederholm, später Goldschmidt, Eskola, Mäkinen u. A. konnten ebenso wie Van Hise, Leith und viele Forscher in Amerika nachweisen, dass Konglomerate, Fossilien, Deltabildungen, Ripplemarks oft so gut erhalten sind, dass Zweifel an einer ursprünglich sedimentären Entstehungsweise unmöglich sind. Zugleich ergab die chemische Untersuchung, dass in manchen Fällen Stoffverschiebungen wesentlichen Masses fehlen, also eine Umbildung *in sich* erfolgte. Diese Entdeckungen bewiesen:

1. dass im tiefsten Grundgebirge Gesteine aufgeschlossen sind, die primär erst entstehen konnten, als Lithosphäre, Hydro- und Atmosphäre bereits existierten. Es lassen sich daher bis zu einem gewissen Grad die heutigen Zustände auf diese Gesteinsbildungsprozesse übertragen und es musste die Vorstellung, man könnte in Urgebirgen eine primäre Erstarrungskruste der Erde studieren, aufgegeben werden;

2. dass die Aufschlüsse nicht in wesentlichem Masse in Tiefen reichen, die nicht zu gewissen Zeiten oberflächennah waren. Dann hat aber der Geologe und Petrograph bei der Erklärung späterer Prozesse der Metamorphose vorerst nachzuweisen, wie gross eine allfällige Ueberdeckung in einem darauffolgenden Stadium sein konnte; es ist ihm nicht gestattet, ohne weiteres anzunehmen, es hätte sich um sehr grosse Erdtiefen gehandelt.

Es ist vom geologischen Standpunkte aus unwahrscheinlich geworden, dass wir irgendwo einen direkten Einblick in Erdtiefen erhalten, die zur Zeit der Gesteinsum- oder neubildung an Ort und Stelle auf mehr als maximal 50 km zu schätzen sind. Durch geophysikalische Untersuchungen sind wir über die in diesen Tiefen vorhandenen Druckverhältnisse und Zustandsformen im grossen, sowie über die heute herrschenden Normaltemperaturen noch recht gut orientiert. Neuerdings ist es gelungen, bis zu derartigen Drucken und Temperaturen Experimentaluntersuchungen durchzuführen, deren Ergebnisse nicht mehr gestatten anzunehmen, es würden sich hiebei neuartige, unbekannte Phänomene einstellen. Dadurch haben die Schlussfolgerungen wesentlich an Sicherheit gewonnen und es ist nur noch die Frage zu stellen, ob das Temperaturgesetz in archäischen Zeiten wesentlich von demjenigen der Gegenwart abwich;

3. dass in den ältesten Erdrindenteilen neben Mineralaggregaten, die Stoffzu- oder wegfuhr erlitten haben, Gesteine vorkommen, die trotz mehrfacher Metamorphose praktisch chemisch unverändert blieben. Damit steht, wie auch die Gesamtgliederung zeigt, fest, dass im Laufe der Zeiten in diesen Teilen der Erdkruste kein allgemeiner Stoffaustausch, keine vollständige Metasomatose stattgefunden hat.

Dadurch wird die Petrographie und Lagerstättenkunde weitgehend zu einer Wissenschaft, die mit experimentellen und theoretisch bekannten Wirkungsweisen bei der Bildung und Umbildung von Mineralvergesellschaftungen rechnen kann. Wohl weist vieles darauf hin, dass Material aus noch grösserer Erdtiefe in diese Kruste eingedrungen ist, der Schauplatz der nachträglich in ihren Produkten zu Tage tretenden Umformungsprozesse jedoch reicht nirgends in Tiefen, die der Phantasie freien Spielraum geben.

Durch die Untersuchungen der alpinen Petrographen, z. B. von Becke und Grubenmann, konnten die von skandinavischen Forschern gefundenen Ergebnisse weitgehend bestätigt werden, indem es gelang, in einem jungen Faltengebirge alle Uebergänge von wenig veränderten magmatischen und sedimentären Gesteinen zu »kristallinen Schiefern» zu finden, die sich oft in keiner Weise von analogen Gesteinen des skandinavischen Grundgebirges unterscheiden. Im Gegensatz zu allerdings nur sehr vereinzelten Anschauungen waren sowohl Van Hise wie Becke und Grubenmann und ihre Schüler überzeugt, dass die natürliche Gesteinsmetamorphose nicht nur über die schon lange bekannten Störungszonen an den Berührungsflächen von Kristallkörnern erfolgt, sondern sehr häufig auf dem Wege über Lösungsumsatz. Mit Recht hat daher Eskola (Kristalle und Gesteine, 1946) Becke, Van Hise und Grubenmann als Begründer und Verfechter der Umkristallisation durch »Porenlösung» bezeichnet, wobei darunter sowohl flüssige, wie gasförmige Phasen zu verstehen sind, und es ist unrichtig, wenn Glangeaud (Les états de la matière dans la pétrogenèse profonde) annimmt, die Einführung dieser Begriffe sei erst durch die Sederholmsche Schule erfolgt. Die Beweise für die Mitwirkung mobiler Phasen, selbst bei

Metamorphosen in Epi- und Mesozone, waren leicht zu erbringen und wurden z. B. von Niggli 1915 in einer Diskussion mit Tammann (Gleichförmige Pressung, Stresspressung und Gesteinsmetamorphose, Z. f. anorganische Chemie) scharf herausgearbeitet. Reine »Solidisten» gab es unter den Petrographen damals so wenig wie heute. In dem bereits 1923 erschienenen Buch »Die Gesteinsmetamorphose I» von Grubenmann und Niggli wurden im Kapitel über die Kinetik der Umwandlungsvorgänge sowohl der strukturellen Betrachtung eines Kristallaggregates, seiner Grenzflächenbeschaffenheit und der Diffusionsmöglichkeit im festen Zustand wie der Imbibition durch eine mobile Phase gebührende Beachtung geschenkt[1]. Ein heute konstruierter Gegensatz zwischen diesen Vorstellungen und späteren programmatischen Darstellungen bestand ebenso wenig wie er zwischen Umwandlungen im festen Zustand und über molekulardisperse Phase empfunden wurde, da feststund, dass am Gesamtbild *beide* Phänomene beteiligt sind. Damit war die Metamorphose Vorgängen im magmatischen System näher gerückt.

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[1] z. B. Seite 159: »Die Oberfläche eines jeden Kristalles stellt einen Zustand dar, der nichts in sich Abgeschlossenes bietet, denn jeder Kristall hat das Bestreben, weiterzuwachsen, sofern ihm nur genügend Stoff zur Verfügung steht. Es kommen ihm, um die Sachlage bildlich chemisch auszudrücken, freie, nach aussen strahlende Valenzen zu, die sich abzusättigen suchen. Der wohldefinierte raumgitterartige Aufbau ist nach aussen gestört, noch nicht vollendet; die äusseren Schichten befinden sich in einem besonderen Zustand. Grenzen zwei Kristalle der gleichen Art aneinander, so werden bei nicht vollständiger Parallelorientierung die Kraftfelder an der Grenze andere sein als im Kristallinnern. Ein absolutes Gleichgewicht ist nicht vorhanden und so wird sich ein von der gegenseitigen Kraftwirkung und Orientierung abhängiger Zwischenzustand ausbilden ...... Ob sich an einer Berührungsfläche zweier beliebig orientierter Kristalle durch Umgruppierung der Atome, bzw. Moleküle, ein neues Kristallgitter bildet, oder ob der eine Kristall auf Kosten des andern seine Grenze nach aussen verschiebt, oder ob schliesslich der Zustand, wie er gerade herrscht, erhalten bleibt, wird von der Grösse und Natur der Kristalle, von der Gestalt der Berührungsfläche und der Beweglichkeit der Massenteilchen abhängig sein. Die letztere steht im direkten Zusammenhang mit der Temperatur, so dass der Einfluss der Temperaturerhöhung auf den Ablauf dieser Prozesse ohne weiteres verständlich wird» ...... und weiter Seite 162: »Einig scheinen sich die Forscher darin zu sein, dass aus den früher genannten Gründen die Kristalloberflächen beliebig aneinander grenzender Kristalle prädisponierte Stellen der Reaktion und Neuorientierung sind. Man führt dies auch etwa auf Oberflächenspannungen zurück und bedient sich dabei gewissermassen nur eines anderen Ausdruckes, die Oberflächenspannungen lassen sich jedoch zwischen kristallinen Phasen nicht lediglich auf Krümmungsverhältnisse zurückführen, der wesentliche Faktor liegt in der Strukturorientierung, so dass die atomistische Betrachtungsweise einzig die variabeln Verhältnisse überblicken lässt.

Früher hat man es als notwendig erachtet, eine intermediäre Dampfphase anzunehmen, neuerdings ist eine wohldurchdachte und gut ausgearbeitete Theorie der Rekristallisation von W. Rosenhain entwickelt worden, nach welcher amorphe Häute um die Kristalle und zwischen den Kristallen die Träger der Ummineralisation seien. So sehr sich viele Metallographen energisch gegen diese Vorstellungen wenden, so bedeutsam ist doch der innewohnende Wahrheitskern. Die Oberflächenschichten der Kristalle sind notwendigerweise von anderer Beschaffenheit als das Kristallinnere, wie man ihren Zustand bezeichnet, ist eine rein terminologische Frage. Er entspricht weder dem gewöhnlichen flüssigen noch dem gasförmigen Aggregatzustand, hat aber mit diesen Zuständen gemein, dass an der Grenze zweier verschiedener Kristallarten eine zu einer gewissen Durchmischung führende gegenseitige Beeinflussung vorhanden ist.»

So wurde 1923 die Lehre vom Intergranularfilm auf die Gesteinsbildung übertragen, die »interaction des particules à l'échelle atomique» und daran anschliessend (S. 163 ff) die »action de la pression à l'échelle atomique» besprochen, die neuerdings Glangeaud unter Berücksichtigung der Arbeiten von Mérigoux und Wegmann jüngeren Zeiten zuschreiben möchte. Im gleichen Abschnitt der »Gesteinsmetamorphose» wurde die Diffusionsmöglichkeit im festen Zustand, die Plastizität und die Erhöhung der Reaktionsfähigkeit durch Porenlösung und Imbibition besprochen.

Das soeben erwähnte Buch über die Gesteinsmetamorphose, die dritte Auflage des Werkes von Grubenmann über »Kristalline Schiefer», hat auch, wie die Namensänderung zeigt, als eines der ersten Bücher die Gesamtheit der metamorphen Prozesse betrachtet und die Vorstellungen der französischen Schule (Fournet, Lacroix, Michel-Lévy) mit denjenigen von Duparc, Goldschmidt, Gutzwiller, Lehmann, Mrazec, Niggli, Schwenkel über Injektionen und denen der skandinavischen Forscher (Goldschmidt, Sederholm u. A.) über Einschmelzungen, Palingenese und pneumatolytische Kontaktmetamorphose in die Lehre eingebaut. Bereits waren grundlegende Arbeiten von Backlund, Eckermann, Eskola und Quensel erschienen. 1920 wurden dann in dem Werk »Die leichtflüchtigen Bestandteile im Magma» eingehend und unter voller Berücksichtigung der skandinavischen, französischen und nordamerikanischen Beobachtungen und Begriffsbildungen die Begleiterscheinungen der plutonischen Erstarrung, die Metasomatose, und die tiefstmagmatische Wirkung der leichtflüchtigen Bestandteile (Injektionsgesteine und Migmatite) physikalisch-chemisch zu deuten versucht. Der Austausch der Erfahrungen zwischen den bahnbrechenden Forschern Eskola, Laitakari und Sederholm einerseits und der »Zürcherschule» anderseits, sowie gemeinsame Exkursionen ergaben Uebereinstimmung in den Grundanschauungen und vermittelten wertvolle Ergänzungen, da einige Phänomene besser im Alpengebirge, andere im Grundgebirge Skandinaviens studiert werden konnten.

In der »Gesteinsmetamorphose» von 1923 wurden auf über 100 Seiten die gewöhnliche, pneumatolytische bis hydrothermale Kontaktmetamorphose, die Injektion bis Einschmelzungsmetamorphose in Deckgebirge und Grundgebirge behandelt und begrifflich in einer Art und Weise gegliedert, die auch den neuesten Anschauungen genügt.[1] In beiden Werken findet man wenigstens versuchsweise eine Behandlung, die durch die Weiterentwicklung an Bedeutung gewonnen hat, so dass sich deren Ausbau aufdrängt. Aus physikalisch-chemischen Erwägungen und Beobachtungsinterpretationen wurde zu ergründen gesucht, welches in tieferen Regionen die Aequivalente von Vorgängen sind, die im äussersten Teil der Erdrinde leichter deutbar sind. So wurde zunächst die Umbildung des Stoffes durch Metamorphose im Deckgebirge behandelt und erst daran anschliessend diejenige älterer Gesteinsserien des Grundgebirges. Seite 347/48 der »Gesteinsmetamorphose» wurde dieses Vorgehen damit begründet, weil die schicksalsreichen Erdrindenteile oft Polymetamorphose zeigen, die Intensität und Extensität der Erscheinungen in älteren Zeiten von denen der jüngsten Vergangenheit etwas verschieden sein können und sehr oft im

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[1] Das erste zusammenfassende Studium des Verhaltens leichtflüchtiger und leicht beweglicher Phasen in der Lithosphäre hatte indessen eine Entwicklung über die Rolle solcher Emanationen und der durch sie verursachten Metasomatosen zur Folge, die dem ersten Bearbeiter der Theorien zu weit zu gehen scheint. Doch ist das ja eine der Fragen, die Klassifikationsversuche nicht zu ändern brauchen, sondern lediglich die Deutung im Einzelfall betrifft.

Grundgebirge tiefere (nicht epidermale) Krustenteile sichtbar werden (Seite 353: Das Grundgebirge gestattet uns eben viel tiefere Einblicke in die Erdrinde, als bei der Untersuchung der Deckgebirge jemals zu erreichen möglich ist). Es wurde versucht, das Aequivalent der tertiär alpinen Metamorphose als Tiefenfacies zu rekonstruieren, umso u. a. zwangsläufig die Erscheinungen der Anatexis, Palingenese, der Ausschwitzungen und Migmatisierungen zu erhalten (siehe auch das gleiche Werk Seite 334/335).

Backlund, Barth, Broegger, Eskola, Goldschmidt, Harker, Holmquist, Hommel, Sederholm, Wegmann, haben diese Tiefengliederung vorbereitet und z. Teil weiter ausgebaut, auch traten später hiefür besonders gebräuchlich gewordene Namen auf, wie Metamorphose in der Ober- und Unterkruste (Wegmann), Tiefentektonik und Oberflächentektonik (Argand, Demay). Die Lehre von den magmatischen Gesteinen hatte diesem Gegensatz (plutonisch, vulkanisch) stets Beachtung geschenkt und durch das Studium des Verhaltens der leichtflüchtigen Bestandteile im Magma wurde offensichtlich, dass sich in der »Tiefenfacies» die Abtrennung magmatischer von »ultrametamorphen» Prozessen oft sehr schwierig gestalten wird. Nun ist selbstverständlich, dass petrogenetisch die übrigens durch alle Uebergänge miteinander verbundenen Unterschiede weder durch eine Gegenüberstellung von Deck- und Grundgebirge, noch durch eine solche der Falten- und Bruchtektonik erfassbar sind. Im Grundgebirge erfolgten manche Intrusionen sehr oberflächennah; so ist z. B. die oberkarbonische Granitintrusion im Gotthardmassiv auf Grund sorgfältiger Beobachtungen unter einer eher geringeren Bedeckung erfolgt als die miocänen Intrusionen in das algerische Deckgebirge oder als die Intrusion in die Karooformation, und die Metamorphose von Grundgebirgskomplexen erfolgte oft in einem Niveau, das unzweifelhaft auch in jungen Faltengebirgen infolge nachträglicher Erosion aufgeschlossen ist. Bestehen bleibt jedoch, dass Mineralbildungsprozesse innerhalb der Lithosphäre untereinander grosse Verwandtschaft aufweisen können, wenn sie unter vergleichbaren ähnlichen Temperaturen erfolgten, unabhängig davon, ob der Prozess als Kristallisation aus magmatischen Lösungen oder als Metamorphose zu bezeichnen ist. Physikalisch-chemisch ist dies leicht verständlich, hatte doch bereits Grubenmann die Temperatur als wichtigen Faktor bei der Klassifikation der metamorphen Gesteine in Rechnung gestellt, wobei jedoch sein »Zonenbegriff» oft missverstanden wurde (z. B. neuerdings noch von Read).

Es erscheint daher heute zweckmässiger zu sein (selbst wenn dadurch andere wichtige Zusammenhänge undeutlicher in Erscheinung treten), die sich in der Lithosphäre und in den von dort stammenden Lösungen abspielenden minero- und petrogenetischen Prozesse etwas anders zusammenzufassen und die Hauptgliederung umzugestalten. Diese Anpassung drängt sich besonders auch demjenigen auf, der gesteinsmässige und akzessorische

Minerallagerstätten nach gleichen Gesichtspunkten gliedern möchte, denn manche bis jetzt grundlegende Bezeichnungen haben auf die Bildung akzessorischer Minerallagerstätten gar keine Rücksicht genommen.

3. DER NEUE KLASSIFIKATIONSVERSUCH

In Uebereinstimmung mit dem Vorgehen in der Geologie lassen sich die Mineralbildungsprozesse sinngemäss gliedern in:

1. Prozesse, die sich in der Grenzregion Lithosphäre, Atmosphäre und Hydrosphäre abspielen, wobei deren Verlauf durch die Wechselwirkung dieser drei Bestandsmassen bestimmt wird. Von exogenen Vorgängen und Bildungen spricht man indessen nur, wenn während der Gesteinsbildung das reaktionsfähige Material der Grenzregion selbst entstammt. Vulkanismus ist eine Aeusserung endogener Kräfte, es bilden sich Minerallagerstätten endogener Abstammung.

2. Prozesse, die sich in der Lithosphäre selbst vollziehen oder durch Faktoren bestimmt werden, die auf Innenverhältnisse der Lithosphäre rückgeführt werden müssen und keine spezielle Folge der exogenen Einwirkungen von Atmosphäre oder Hydrosphäre auf die Gesteinsschale darstellen. Diese Prozesse erzeugen Gesteine und Minerallagerstätten von endogenem Charakter.

Die Grossgliederung soll somit lauten:

I. Exogene Gesteine und Minerallagerstätten
II. Endogene Gesteine und Minerallagerstätten

Exogen ist weit besser als sedimentär oder neptunistisch (wie Read vorgeschlagen hat), da manche Mineralbildungen in der äusseren Reaktionszone der Lithosphäre durch den Begriff »Sedimentation» nicht erfasst werden. Auch ist heute bekannt, dass magmatische Bildungen sowohl durch Sedimentation in Magmen und als auch durch Kristallisation aus wässeriger Lösung zustande kommen können.

Endogen als Hauptbegriff nimmt darauf Rücksicht, dass man in manchen Fällen mit Sicherheit feststellen kann, dass Vorgänge innerhalb der Lithosphäre und ohne wesentliche Beeinflussung durch Hydro- und Atmosphäre zum beobachtbaren Strukturbild geführt haben, ohne zunächst entscheiden zu können, ob magmatisch oder metamorph die richtige Bezeichnung sei. Dementsprechend hat auch Eskola seinen Faciesbegriff endogener Gesteine allgemein gefasst. Selbstverständlich muss man, als Folge der stets vorhandenen Wechselwirkungen und Uebergänge, konventionelle Abgrenzungen bzw. Zuordnungen vornehmen.

Schon erwähnt wurde, dass (dem bisherigen Sprachgebrauch gemäss) Lavengesteine endogener Natur sind, ebenso alle Kristallisationsprodukte der unmittelbar aus dem Erdinnern stammenden ascendierenden molekulardispersen Phasen, selbst wenn der Bildungsort der Grenzzone der Lithosphäre nach aussen hin angehört. Werden jedoch die Einflüsse der Atmosphäre oder Hydrosphäre auf die Neuprodukte überwiegend, wie bei manchen vulkanischen Tuffen, subaquatischen exhalativen und zugleich sedimentären Lagerstätten usw., so wird es zweckmässig sein, die Erscheinungen in beiden Grundkapiteln zu erwähnen und in dem einen ausführlicher zu behandeln. Es kann sich z. B. um Sonderfälle exogener Lagerstättenbildung mit endogener Stoffzufuhr usw. handeln. Diagenetisch oder nur sehr schwach (nach Ueberlagerung) endogen veränderte exogene Bildungen werden, wie bisher üblich, immer noch als exogene bezeichnet, wenn die erlittenen Veränderungen nicht im Stande waren, Mineralbestand, Struktur und Textur grundlegend zu verändern. Ob man die organogenen Kohlen- und Kohlenwasserstoffgesteine, die durch Umwandlung von Material, das der Biosphäre entstammt, ihren Sondercharakter erhalten haben, als exogen oder als endogen metamorph bezeichnen will, ist Konvention. Da die Umgestaltungsprozesse meist schon mit der exogenen Ablagerung einsetzen, hält es oft schwer, später von statten gehende Prozesse von den Anfangsstadien abzutrennen, so dass es manchmal zweckmässiger ist, die Umwandlungsprodukte, so lange sie sich noch innerhalb typisch exogener Gesteinsvorkommen befinden, bei dieser Gruppe zu belassen. In der endogenen Gruppe befinden sich zudem naturgemäss viele ursprünglich exogene Bildungen, die jedoch endogen so verändert wurden, dass die neue Zuordnung gegeben ist. Umgekehrt gelangen endogene Bildungen durch Erosion in die Reaktionszone der Verwitterung. Ob sie noch nach ihrer ursprünglichen Bildungsweise klassifiziert werden müssen oder als Böden nach ihrer späteren Umformung, hängt von der Stärke dieser Beeinflussung und dem besonderen Standpunkt (Bodenkarte, Untergrundskarte) ab.

Die Einteilung der exogenen Gesteine und Minerallagerstätten kann, sobald man sich von der Sedimentation als Grundbegriff freigemacht hat, viel zweckmässiger gestaltet werden. Es werden die autochthonen Verwitterungslagerstätten und Gesteine den eigentlichen Sedimenten gegenübergestellt. Im ersteren Falle ist die laterale Stoffwanderung gering, während die vertikale erheblich sein kann, im zweiten Falle wirkten Transporte, Umlagerungen und Neuabsätze in tangentialer Richtung zur Erdoberfläche mit. Innerhalb der Gruppe der Sedimente muss den neuen Erkenntnissen Rechnung getragen werden. Darnach ist ein Grossteil des Materiales der Pelite nicht kurzweg Trümmermaterial der Gesteinsverwitterung, sondern verfrachtetes Material von Neubildungen in der Verwitterungszone bis hinunter zu kolloidaler Grösse. Die Absatzbedingungen hiefür stehen denen kolloidaler, ja chemischer Ausfällungen viel näher als denen der eigentlichen klastischen Materialien. Die Bezeichnung Schlamm für rezente Bildungen, unabhängig vom speziellen Stoff, ist hiefür kennzeichnend (toniger Schlamm, Kalkschlamm, Kieselschlamm usw.)

In einem lehrbuchartigen Versuch hat sich folgende Einteilung bewährt:

I. Exogene Gesteine und Minerallagerstätten

A. *Autochthone Gesteine und Minerallagerstätten*

- a. Die Gesteinsböden mit ihren Ortssteinen oder Ausblühungen und den Verkieselungen usw.
- b. Sonderfälle von Verwitterungsrückständen und Neubildungen in der Oxydations- und Zementationszone; Erzlagerstättenböden, exogen umgelagerte Erzvorkommnisse, Gangbildungen aus descendierenden Lösungen, usw.

B. *Sedimentäre Gesteine und Minerallagerstätten*

- a. Trümmersedimente, klastische Gesteine
  - α. Psephite } mit ihren Erz- und Edelsteinseifen
  - β. Psammite } mit ihren Erz- und Edelsteinseifen
- b. Pelite, Gelite, Humite und verwandte Gesteine u. Lagerstätten
  - α. Tonige und tonig-karbonatige Pelite (inklusive Mergel)
  - β. Sapropelite, Humite und zugehörige Kaustobiolithe
  - γ. Sulfo- bis Sulfosapropelite und zugehörige Erzlager
  - δ. Eisen- und Manganhydrogelite, Fe- und Mn-hydroxydische und oxydische Sedimente und Erze
  - ε. Siliziumhydrogelite und Kieselgesteine im allgemeinen
- c. Karbonatgesteine
  - α. Kalkgesteine
  - β. Dolomite und spezielle exogene Karbonatgesteine
- d. Sulfat- und Chloridgesteine, sowie verwandte Salzgesteine und Salzlager
- e. Schnee und Eis

Die Trennung in chemische und biochemische Sedimente ist in b und c praktisch nicht immer durchführbar, doch schaltet sich in diese zwei Gruppen am häufigsten die Biosphäre ein.

Die endogenen Bildungen umfassen der bisherigen Systematik nach die magmatischen und metamorphen Mineralbildungen. Ein Vorschlag von Read (Meditations on granite. Part two: Proceedings of the Geologist's Association vol. LV, 1944), diese Gruppe in vulkanische und plutonische Gesteine zu gliedern, kann meiner Meinung nach aus zwei Gründen nicht in Frage kommen. Erstens müssen wie bisher auf Grund von Beobachtungen und Erwägungen physikalisch-chemischer Art die magmatischen Gesteine selbst in vulkanische und plutonische gegliedert werden. Reads Einteilung ist ja nur verständlich, weil dieser Forscher auf dem Standpunkt steht, dass neben »orthodoxen Vulkangesteinen» lediglich »minor intrusions» magmatischen Ursprungs sind. Zweitens müssten, da zu den plutonischen Gesteinen alle metamorphen Gesteine gehören sollen, die öfters unter sicherlich geringer Bedeckung metamorphosierten Gesteine wie Sericitphyllite, Epimarmore, Epigneise, Grünschiefer, Glimmerschiefer usw., deren Uebergänge in normale Sedimente oder Eruptivgesteine im Gebirge studiert werden können, Plutonite genannt werden, während kleinere Intrusionen in grösserer Tiefe Vulkanite wären.

Der Begriff der iso- oder homöophysikalischen und iso- oder homöochemischen Gesteinserien, der im Anschluss an Arbeiten von Broegger, Eskola, Goldschmidt, Harker, und Washington von Niggli 1920 in einem Vortrag an der Jahresversammlung der Schweiz. Naturf. Gesellschaft in Neuchâtel aufgestellt wurde und im Faciesbegriff von Eskola (1921) sowie in Untersuchungen von Bowen, Buddington und Tilley eine Ausarbeitung erfuhr, scheint geeignet zu sein, eine systematische Grundlage zu liefern. Das geht schon daraus hervor, dass er sich prinzipiell bereits in der Klassifikation der metamorphen Gesteine durch Grubenmann (Katagesteine, Mesogesteine, Epigesteine) vorfindet und als geeignet befunden wurde, auch die magmatischen Bildungen (katathermale, mesothermale, epithermale) zu gliedern. Die allgemeinste Aussage ist die, dass es möglich erscheint, verschiedene Minerallaggregate mit analogem Bauschalchemismus in Reihen zu ordnen mit dem physikalischen Bedingungkomplex als Variabler, und dass es umgekehrt bei verschiedenem Chemismus angebbar ist, welche Facies dieser Reihen unter einigermassen ähnlichen physikalischen Verhältnissen entstehen konnten. Es ist jedoch selbstverständlich, dass verschiedene Bauschalzusammensetzungen auf Temperatur- und Druckänderung ungleich empfindlich sind, so dass mehr nur die Aufeinanderfolgen und nicht die Einzelfacies streng parallelisierbar werden. Der Faciesbegriff muss dem Stoff gemäss auf die endogenen Gesteine übertragen werden, d. h. es können mehrere oder nur wenige verschiedene Facies abtrennbar sein. Unabhängig davon stellt sich die Frage, ob zu Uebersichtszwecken verschiedene Facies zu einem gleichen Oberbegriff zusammengefasst werden können, was ja Grubenmann durch seine Kata-, Meso-, Epigliederung glaubte bejahen zu dürfen. Eine nähere Untersuchung zeigt, dass fast durchgehend Faciesentwicklungen relativ hoher Bildungstemperaturen gegenüber solchen relativ niedriger Bildungstemperaturen unterscheidbar sind, und dass der Zwischenbereich bereits sehr starken Uebergangscharakter aufweist, daher bald mehr der hochthermalen, bald mehr der niederthermalen Serie zugeordnet werden kann. Zu den hochthermalen Mineralvergesellschaftungen gehören assoziationsgemäss auch die normalen Eruptivgesteine. So dürfte man z. B. wohl Eskolas Sanidinit-Diabas-Pyroxenhornfels-Gabbro-Granulit-Eklogit-Hornblendegabbro-Amphibolit-Facies relativ hochthermal, die Epidotamphibolit-Glaukophanschiefer- und Grünschieferfacies relativ niederthermal nennen, wobei manchmal (jedoch nicht durchgehend) noch nach Druckbedingungen und Stressverhältnissen unterscheidbar ist.

Wir nennen daher, im vollen Bewusstsein, dass dadurch der untern und obern Temperaturgrenze nach sehr Verschiedenartiges zusammengefasst wird, Paragenesen relativ hoher Temperatur von der Schmelzerstarrung an abwärts hochthermale oder katathermale Paragenese und solche relativ niedriger Temperatur von der Temperatur der Erdoberfläche an aufwärts niederthermale oder epithermale Paragenese. Zwischen beiden Gebieten schaltet sich das mittel- oder mesothermale Gebiet ein, das teils besondere Mineralkombinationen liefern kann, teils nur Uebergangscharakter besitzt, und teilen nun zunächst die Gesamtheit der endogenen Gesteine und Minerallagerstätten ein in:

II a Kata- (bis meso-)thermale Gesteine und Minerallagerstätten und in

II b Epi- (bis meso-)thermale Gesteine und Minerallagerstätten.

Dadurch wird die einheitliche physikalisch-chemische Betrachtungsweise gewährleistet und auf die in beiden Gruppen vorhandenen Schwierigkeiten der Zuordnung zu magmatisch oder metamorph Rücksicht genommen. Aber die grundsätzliche und auch für den Geologen wichtige Diskussion darf damit nicht umgangen werden, sie rückt gewissermassen nur an zweite Stelle, was, wie später erwähnt wird, in der Namengebung leichter berücksichtigt werden kann. Denn innerhalb dieser zwei Gruppen II a und II b wird es nun zur Notwendigkeit, zu fragen, ob es sich um magmatische Bildungen handelt, d. h. um Bildungen, die zur Hauptsache als Kristallisationsprodukte aus molekularen Lösungen anzusprechen sind, oder um metamorphe Umbildungen eines Festbestandes, der während der letzten massgebenden Vorgänge nie als Ganzes flüssig war. In verschiedenen neueren Arbeiten[1] hat der Verfasser sowohl für akzessorische Minerallagerstätten der magmatischen Abfolge als auch für metamorphe und ultrametamorphe Gesteine versucht, heute verwendete Begriffe genauer zu definieren. Es sei darauf verwiesen, und lediglich nochmals betont, dass ein durch Wiederaufschmelzung gebildetes Magma naturgemäss für die späteren Prozesse kurzweg als Magma in Rechnung zu stellen ist.

So ergibt sich auf Grund der oben genannten Erörterungen für II a, die kata- (bis meso-)thermalen Gesteine und Minerallagerstätten, im grossen folgende Einteilung:

A. *Katamagmatische Bildungen*, umfassend die normalen, in Plutonite und Vulkanite einteilbaren Eruptivgesteine und die pneumatolytisch-pegmatitischen bis katahydrothermalen magmatischen Erzlagerstätten. Die Gesteine werden Katamagmatite oder, da dies die Hauptform der Eruptivgesteine ist, kurzweg Magmatite genannt und in Familien untergeteilt.

B. *Katametamorphe Bildungen*, umfassend Produkte der Metamorphose eines Festbestandes oft unter teilweisem Umsatz über atom-, ionen-

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[1] Siehe Literaturangaben. Auch Scheumann hat sich darum bemüht und in bezug auf Erzlagerstätten Schneiderhöhn.

oder molekulardisperse Phasen, jedoch ohne starke stoffliche Aenderung und ohne starke Metasomatose durch äussere Stoffzu- und Wegfuhr. Die Produkte sind Katametamorphite.

C. *Ultrakatametamorphe Bildungen.* Hier handelt es sich um Produkte von Vorgängen, die sich unter sehr starker Veränderung des ursprünglichen Festbestandes in der Lithosphäre abspielen, sei es, dass sich gleichzeitig grössere Mengen im flüssigen Zustand befanden oder Stoffzu- und -abwanderungen (und damit Metasomatosen mit Aenderung des Bauschalchemismus) im grossen Masstabe zur Auswirkung gelangten. Es entstund jedoch kein einheitliches Magma, das seinerseits zu neuen Magmatiten Veranlassung gegeben hätte. Die Katamigmatite und Katadiabrochite gehören hieher und leiten ihrem Aussehen nach oft zur Gesteinsgruppe A über.

Ueberall werden mesothermale Bildungen, die unmittelbar in katathermale übergehen, angeschlossen. Paragenesen vom Typus der normalen Pegmatite gehören noch hieher.

Erhält ein Gestein nur nach Mineralbestand und allgemeinen Gefügeeigenschaften seinen Namen, so kann unter Umständen der gleiche Hauptbegriff bei verschiedener Entstehungsweise benutzt werden. Ein relativ helles, holokristallines körniges Gestein beispielsweise, das bei massiger Textur vorwiegend aus Alkalifeldspat, Quarz, Plagioklas und Biotit besteht, ist in sehr vielen Fällen ein Magmatit und führt als solcher den Namen Granit. Sehr granitähnliche Gesteine können jedoch auch durch Ultrametamorphose oder unter Umständen durch gewöhnliche Katametamorphose (z. B. einer Arkose) entstehen. Sind die Unterschiede (z. B. in der Textur, wie das bei den Gneisen der Fall wäre) kaum bemerkbar oder rechtfertigen sie eine Namensänderung nicht, so kann von Metagraniten (durch gewöhnliche Metamorphose entstandene granitähnliche Gesteine) oder von Ultrametagraniten (im speziellen Migmagraniten oder Diabrochograniten) gesprochen werden. Zur näheren Kennzeichnung mag es bei Vorhandensein verschiedener Bildungen in der gleichen Region sogar erwünscht sein, das magmatische Gestein nicht kurzweg als Granit, sondern genauer als Magmagranit zu benennen.

Gleiche Schwierigkeiten treten u. a. bei Eklogiten (Eskola) und Amphiboliten auf und können bei der Namengebung berücksichtigt werden. Es stellt sich zudem oft die Frage, ob eine bestimmte Struktur und Textur nur durch ein besonderes Stress- oder Druckfeld bei der Magmenerstarrung zustande kam (Piezokristallisation nach Weinschenk) oder durch eine der Erstarrung nachfolgende bzw. diese überdauernde Metamorphose.

Für die epi- bis mesothermalen Gesteine und Minerallagerstätten endogenen Ursprungs werden im grossen ähnliche klassifikatorische Prinzipien gelten wie für die katathermalen. Allein es macht sich eine wesentliche Verschiebung bemerkbar. Als Lösungen, aus denen Epimineralbestände auskristallisieren, kommen zur Hauptsache nur sehr wasserreiche Lösungen in Frage, die im Festbestand der Erdrinde wegen ihrer Fluidität und ihrem geringen spezifischen Gewicht kaum dauernd (oder auf längere Zeit hin) grosse Räume erfüllen. Ein normales Gesteinsmagma ist unter diesen Temperaturbedingungen in der Lithosphäre wenig bestandfähig. Eine Grosszahl der epi- bis mesothermalen Gesteine ist daher durch Metamorphose entstanden, ja selbst die Ultrametamorphose tritt gegenüber der gewöhnlichen und schwach metasomatischen zurück. Indessen zeigt ein näheres Studium der meso- bis epithermalen, endogenen Bildungen recht deutlich, wie wertvoll die hier angewendete Gliederung der katathermalen Gesteine und Minerallagerstätten ist, indem die Uebertragung auf Formationen niedriger Temperatur manche Erscheinungen verständlich macht, die bis jetzt oft als fremdartig empfunden wurden. So gibt es gewisse Eruptivgesteine mit ausgesprochenem Epi- bis Mesomineralbestand. Sie sind, wie z. B. die Spilite, aber vielleicht auch manche Grünschiefer, in dieser Form zum Teil aus sehr wasserreichen Magmen entstanden (hydromagmatische Gesteine) oder während der Bildung unter dem Einfluss wässeriger Restlösungen des Magmas umgewandelt worden (Autohydrometamorphose). Andere Gesteine ähnlichen Mineralbestandes haben spätere epithermale Umwandlungen erlitten, der Mineralbestand ist dann die Folge einer Metamorphose.

Neben den relativ seltenen primär-hydromagmatischen Gesteinen treten ferner viele hydrothermale Minerallagerstätten auf, die sich in Klüften und Spalten aus emporsteigenden (ascendierenden) hydrothermalen Lösungen ausgeschieden haben, sowie Erzlager aus wasserreichen Sulfid- bis Oxydschmelzen mit Erstarrungsbeginn bei tiefer Temperatur. Das ergibt eine erste Gruppe hydromagmatischer bis hydrothermal-magmatischer Gesteine und Lagerstätten. Epi- bis Mesometamorphite sind in äusseren Kontakthöfen, vor allem aber in Faltengebirgen, als sogenannte dislokations- bis tektometamorphe Gesteine häufig anzutreffen. Sie zeigen die Uebergänge zu den noch wenig veränderten Ursprungsgesteinen und lassen die Reaktionsabläufe oft gut studieren. Ihrerseits können sie begleitet sein von Ader- und Kluftbildungen, die dartun, dass sich sekretionär, durch »Ausschwitzung», Lösungen gebildet haben, die in offene Räume abwanderten und dort bei Bedingungsänderungen kristallisierten. Gerade dadurch wird offensichtlich, wie gross bereits bei der gewöhnlichen Metamorphose der Lösungsumsatz sein kann. Die entstehenden chorismatischen Produkte erinnern, wenn wir vom veränderten Mineralbestand absehen, stark an gewisse katamigmatitische Bildungen und an Arterite und Venite der Katazone und lassen deren

Bildung leichter verstehen. Im Bereich aufsteigender Hydromagmen oder hydrothermaler Lösungen kann der Stoffaustausch, der teils mehr den Charakter der Auslaugung, teils der Imprägnation und der Metasomatose zeigt, recht grosse Beträge erreichen, so dass es durchaus am Platze ist, von (meist hydrothermal veränderten) Meso- bis Epidiabrochiten zu sprechen, d. h. von meso- bis epiultrametamorphen Gesteinen, die oft den Charakter des Ursprungsgesteins völlig verloren haben. So werden wir auch bei den epi- bis mesothermalen Gesteinen und Minerallagerstätten auseinanderhalten:

A. *Epi- bis mesomagmatische Bildungen*, umfassend die hydromagmatischen Gesteine und meso- bis epithermalen magmatischen Erzlagerstätten.

B. *Epi- bis mesometamorphe Bildungen*, umfassend die Meso- und Epimetamorphite und die ihnen zugeordneten lokalen Ader- und Kluftbildungen.

C. *Ultraepi- bis mesometamorphe Bildungen* mit starkem Stoffaustausch, zur Hauptsache als Diabrochite entwickelt, jedoch bei starker Ader- und Kluftbildung auch direkt aus B hervorgehend als Epi- bis Mesomigmatite. Ferner können, wie bei einzelnen Schalsteinbildungen, starke Vermischungen von hydromagmatischen Phasen mit sedimentärem Material erfolgen und so gleichfalls zu einer Art Migmatitbildung führen.

Geht man bei Betrachtung der Katagesteine zweckmässig von der Bildung der Eruptivgesteine aus, so wird man bei dem Studium der Epigesteine von der sogenannten säkulären (d. h. nicht unmittelbar an die äussere Lithosphärengrenze gebundenen) Verwitterung und Gesteinsumwandlung ausgehen. Im ersteren Falle kommt man über die Kontaktmetamorphose im Bereich der Magmaherde und über die Anatexis und Assimilation zur gewöhnlichen Katametamorphose, im zweiten Falle über die Metamorphose in nur dislokations- und tektometamorphen Gebieten (ohne Magmenintrusion) zur Epi- und Mesometamorphose im ganzen. Anderseits zeigt das Studium der Magmenentwicklung bis zu den Thermalbildungen die Zusammenhänge zwischen liquidmagmatischen und hydrothermalen Phasen, zwischen kata- und epithermalen Prozessen.

Es ergibt sich ja von selbst, dass bei geänderter Stoffgruppierung neue Querverbindungen geschaffen werden müssen, damit die nicht benutzten wertvollen Zusammenhänge auch weiterhin beachtet werden. Als Ganzes aber möchte der neue Versuch (der in gewissem Sinne an Eskolas Faciesbegriff anschliesst) die Konsequenzen aus der Entwicklung der Lehre von der Wirkung leichtflüchtiger Substanzen in der Lithosphäre und aus den Parallelisationsversuchen hoch- und niederthermaler Prozesse ziehen. In diesem Sinne wird er zur Diskussion gestellt, wobei allerdings erst die Ausarbeitung in einem im Druck befindlichen Werke »Gesteine und Minerallagerstätten» (Bd. I, 1947) Bewährungsprobe sein kann.

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13.

RAPAKIVI AMPHIBOLE FROM UUKSUNJOKI, SALMI AREA

BY

TH. G. SAHAMA

The most coherent igneous rock group of any greater areal distribution in the Finnish Pre-Cambrian, the rapakivi granites, are texturally very widely varying. True, some extremely conspicuous features, like the potassium feldspar ovoids, often surrounded by oligoclase mantles, are completely unknown in other granitic rocks. However, due to the fact that these ovoids are entirely lacking in some other types of granites of the rapakivi areas, they cannot be considered as essential when defining the whole rapakivi conception. According to the current terminology now used in Finland, the classic country of these rocks, the term »rapakivi granite» is to be extended to all granite types occurring in the well-known »rapakivi areas» of this country.

In spite of the considerable textural variations, the rapakivi granites show some peculiarities in their chemical composition, justifying their inclusion in one single petrographical province. A general view of the chemistry of the Finnish rapakivi granites has been recently given by Sahama [1].

Depending upon the chemical behavior of the rapakivi granites, the mineral composition is also characteristic of these rocks. Leaving the light constituents aside in this connection and taking into consideration only the principal dark ones, biotite and hornblende are left, having a very great distribution in almost all types of rapakivi granites. They occur either together in the rock or alone; some rapakivi granites are only biotite-bearing types, others carry hornblende without biotite, and in many varieties these two minerals can be detected.

Now, of biotite as well as of hornblende occurring in rapakivi granites some data are presented in literature pointing to the suggestion that these two minerals differ in their chemical composition from those »biotites» and »hornblendes» which usually are found in rocks of the other granite groups of the Finnish Pre-Cambrian. As especially emphasized by Popoff[2], the optical properties of hornblende in rapakivi granites from the Wiipuri area are surprisingly uniform in some respects. The color of the mineral in thin

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[1] TH. G. SAHAMA: On the Chemistry of the East Fennoscandian Rapakivi Granites. Bull. Comm. géol. Finlande. N:o 136. C. R. Soc. géol. Finlande. N:o XVIII. p. 15, 1945.

[2] BORIS POPOFF: Mikroskopische Studien am Rapakivi des Wiborger Verbreitungsgebietes. Fennia, Vol. 50, No. 34, 1928.

section is very dark and the refringence extremely high. On the other hand, considerable variations in the optic axial angle were found. Popoff did not make any chemical analysis of the hornblende type in question; he only suggested, on the basis of the optical properties found, that the mineral represents an extremely iron-rich variety of the type called hastingsite.

Because of the fact that the type of hornblende which was optically investigated by Popoff in his paper cited above really seems to be very characteristic of rapakivi granites in Finland in general, an exact knowledge of its chemical composition is highly desirable. So far as I know, no analysis of the mineral has, however, ever been made. From the collections of the Institute of Geology, University of Helsinki, I chose a suitable specimen, which happened to be derived from the Uuksunjoki River, Salmi area, where this river crosses the main road from Koirinoja to Uomaa. I myself collected the specimen, which contained well-developed potassium feldspar ovoids without oligoclase mantles. The amphibole was separated out of the rock in a fraction of some 15 g. and analysed chemically by Mr. Oleg v. Knorring, M. A.

The optical determinations were made by Mrs. Toini Mikkola, M. A. and by me. The results are given in Table I.

Table I.

	%	Mol. prop.
SiO_2	38.24	634
Al_2O_3	10.17	99.5
TiO_2	2.00	25
Fe_2O_3	5.00	31
FeO	26.64	371
MgO	1.07	27
MnO	0.28	4
CaO	10.64	190
Na_2O	1.50	24
K_2O	1.57	16.7
H_2O+	1.88	104
F	1.06	56
Cl	0.51	14
H_2O-	0.08	—
	100.64	
—O	0.57	
Sum	100.07	

Specific gravity 3.447 (Pycnometer by Mr. Oiva Joensuu, M. A.)
$b \parallel \beta$; $c \wedge \gamma = 12°$.
$\gamma = 1.730$.
$\alpha = 1.702$.
$2V$ very small; looks almost uniaxial.
Absorption: $\gamma > \beta > \alpha$.
γ bluish green.
β dark greenish brown.
α pale greenish brown.

The analysed fraction of the amphibole showed no other impurities than minute amounts of inclusions of a brownish matter. The identification of these inclusions is very difficult. During the separation the observation was made that those grains which contained more of the inclusions in question were enriched in fractions of higher specific gravities than the analysed purest amphibole fraction. Thus, the specific gravity of the inclusions seems to be higher than that of the amphibole. For a possible identification with an X-ray powder diagram no apparatus was available. — It is especially to be noted that no fluorite or apatite could be detected in the analysed fraction with a binocular microscope.

The analysis given in Table I was calculated according to the principles given by Foslie[1].

The results of the calculation, are presented in Table II.

Table II.

Si	6.16		Z = 8.00	Si 6.16 Al 1.84
Al	1.93	1.84 0.09		
Ti	0.24		Y = 0.93	XY = 4.83
Fe^{III}	0.60			
Fe^{II}	3.60		X = 3.90	
Mg	0.26			
Mn	0.04			
Ca	1.84		W = 2.63	
Na	0.47			
K	0.32			
OH	2.04		(OH, F, Cl) = 2.72	
F	0.54			
Cl	0.14			

As is seen from Table II, the analysis of the mineral agrees very well with the demands regarding the general amphibole composition put forward by Foslie *(op. cit.)*. The water content and especially the content of fluorine as well as of chlorine are remarkably high. This is in complete harmony with the prediction expressed already by Popoff *(op. cit.)* who found the rapakivi amphibole always in a close association with fluorite in thin section and thus concluded a high fluorine content in the hornblende. In this place it is again worth noting that no fluorite could be detected in the analysed amphibole fraction from Uuksunjoki.

When comparing the composition of the amphibole mineral from Funta, Tysfjord, given by Foslie and by him called ferrohastingsite, with the present mineral, a close parallelism will be found. However, the ratio between ferrous iron and magnesium is greater in the Uuksunjoki variety.

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[1] STEINAR FOSLIE: Hastingsites, and Amphiboles from the Epidote-Amphibolite Facies. Norsk Geol. Tidskr., Vol. 25, p. 74, 1945.

Due to the general tendency of magnesia to remain extremely low in rapakivi granites, the analysed rapakivi amphibole is extremely iron-rich. This fact fits very well with the dark colors of the rapakivi amphiboles in general and, in this special case, where exact determinations of two refractive indexes are available, also with the unusually high refringence.

The data given above seem to justify the designation ferro-hastingsite for the amphibole variety from Uuksunjoki. The conseption of hastingsite still being far from uniform in literature, the meaning of this term is here used in the sense defined by Foslie. A very characteristic feature in the optical properties is the small axial angle.

As to the other of the principal dark constituents of the rapakivi granites, the biotite, no modern chemical analyses are present of it. However, Wahl [1] has presented two old analyses. They surely are no longer to be regarded as reliable for modern purposes. In any case they show, together with the extremely high refractive indexes of rapakivi biotites, that this mineral also is very rich in iron. Of the biotite occurring in the same rock as the amphibole reported above the following properties were measured:

$$\gamma \sim \beta = 1.730.$$
$$2V_\alpha = 22^\circ.$$

The mineral was not obtained in a completely pure fraction. An analysis, made of an impure fraction, showed almost the same ratio between ferrous iron and magnesium as found in the amphibole. Because of the impurity of the fraction in question, the analysis will not be given.

The results reported above of the amphibole from Uuksunjoki give an idea of the mineral in question. Although it seems highly probable that the amphiboles, occurring in various types of the rapakivi granites, in their chemistry and optical properties approach the type just described, further work is needed until a general characterization of the rapakivi hornblende can be presented, based upon sufficient observations and measurements. Because of the fact that exact data in the matter at present are almost entirely absent, the publication of this contribution seems to be justified. I am happy to report these results of my co-workers and myself in the Jubilee Volume dedicated to Professor Pentti Eskola, a famous expert on the Finnish rapakivi granites.

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[1] WALTER WAHL: Die Gesteine des Wiborger Rapakiwigebietes, Fennia Vol. 45, No 20. 1925.

14.

ON THE DIDACTIC FORM OF THE SYSTEM OF CRYSTAL CLASSES

BY

HEIKKI VÄYRYNEN

The division of crystal forms into 32 classes has been treated originally by Hessel (1830), Bravais (1848), and more thoroughly by A. Gadolin (1867). At that time crystal classes were presented in the order beginning with the classes of highest symmetry and progressing to the classes of lowest symmetry. The classes belonging to the same system were derived from the one of the highest symmetry or one from another as hemihedric, hemimorphic, and tetartohedric forms (e. g. Groth 1876 and Liebisch 1896). Already in 1883, however, Tschermack writes in the preface of the first edition of his »Lehrbuch der Mineralogie»: »Das Fortschreiten vom Einfachen zum Zusammengesetzten erfordert in der Krystallographie eine Anordnung, deren ich mich seit Jahren bediene, die aber manchem auffallen wird, weil die Reihe der Krystallsysteme mit den triklinen Formen anhebt.» In the year 1895 P. Groth writes in the third edition of his »Physikalische Krystallographie»: »Die bisher übliche, besonders von Naumann begründete, Systematik der Krystallformen, d. h. deren Eintheilung in holoëdrische, hemiëdrische, hemimorphe u. s. w. Gruppen, leidet (wie namentlich Gadolin zeigte) an einer Reihe von Unklarheiten und Inconsequenzen, welche dadurch entstanden sind, dass man sich für die Beurtheilung der möglichen Gruppen an gewisse, nur partiell gültige Analogien der nach und nach an den krystallisierten Körpern, besonders den Mineralien, beobachteten Formen hielt, ohne hierbei ein allgemeines Princip zu verfolgen. Nachdem nun aus dem empirischen Grundgesetze der geometrischen Krystallographie, dessen Notwendigkeit sich auch aus dem physikalischen Verhalten der Krystalle ergiebt, die Gesammtheit aller möglicher Krystallformen streng mathematisch abgeleitet und nachgewiesen worden ist, dass es nur 32 Klassen von Formen, welche sich durch ihre Symmetrie von einander unterscheiden, geben könne, erweisen sich die bisherigen Krystallsysteme als mehr oder weniger willkürliche Zusammenfassungen je einer Anzahl jener Klassen, deren Beibehaltung lediglich einen praktischen Werth, behufs leichterer Uebersicht, besitzt. Der logisch richtige Weg der systematischen Behandlung der 32 möglichen Klassen von Krystallen ist offenbar derjenige, bei welchem man von den einfachsten Formen, d. h. diejenigen mit dem niedrigsten Grade der Symmetrie, ausgeht und durch Einführung der möglichen Symmetrieelemente nach und nach zu immer complicirteren Formen gelangt, statt, wir bisher, mit den letzteren zu beginnen und die einfacheren aus ihnen abzuleiten. Durch die Einführung dieser Methode fallen nun die Begriffe der Hemiëdrie, Tetartoëdrie, Hemimorphie etc. und mit ihnen alle Unklarheiten und die Schwierigkeiten, welche diese bisher dem Anfänger, wie dem Vorgeschritteneren bereitet haben, vollständig fort.»

The principle mentioned above, which starts from simple forms and progresses towards complicated forms, has been consistently used by Niggli in his »Lehrbuch der Mineralogie», published in 1920. With the aid of the planes and axes of symmetry the main types of crystal forms, i. e. pedion (Einflächner), pinakoid, dome, and sphenoid (Zweiflächner) a well as prism and pyramid have been defined without reference to th crystallographic axes. The main features of the lattice structure of crystals, known already in that time, were also taken into account in the exellent survey by Niggli.

But while presenting the crystal forms along the inductive principle mentioned above, the physical properties of crystals have not been adequately observed. These, especially the optical properties, obey the simplest laws in crystals with the highest symmetry, while the laws become more complicated with the reduction of symmetry. Groth has already in 1875 drawn attention on this point of view. The crystals of the highest possible symmetry, i. e. of the cubic system, are isotropic. The crystals with principal axis and principal plane of symmetry are anisotropic, but have an axis of isotropy. The crystals with three unequal crystal axes are fully anisotropic. Therefore the requirement can be introduced that the system of crystal classes should offer a basis for the presentation of the physical properties of the crystals. Besides, the physical properties in general and the optical properties especially are most directly dependent upon the lattice structure. Thus a system dealing with the physical properties must at the same time be in the best accordance with the crystal structure. Indeed, the simplest structures exist in the cubic and hexagonal systems representing the highest symmetry, and we are right in asserting that crystals of the lowest symmetry comprise usually more complicated structures.

A special attention should be paid to the possibility that the idea of simplicity or complexity of crystal structures can be taken as a starting point in the deduction of the system of crystal classes. As a matter of fact, it is the lattice structure that is the primary feature, to which all properties of crystals can be referred, including the formation of crystal faces as well as the laws of the constancy of angle and the rational intercepts. On the other hand there is no question of what the simplest structures are, as the primitive lattices presented by Bravais in 1848 always have been regarded as such. It is therefore of interest to attempt a deduction of crystal classes from this point of view. The following treatment is intended to present a preliminary investigation of this theme.

It has not been possible to observe the theoretically simplest, so called »single primitive» lattices in crystals, but rather simple »double primitive» structures are known to exist in cubic and hexagonal crystals. Such are the cubic body-centred and face-centred lattices, the former of which occurs in alkali metals, in α-iron and many other metallic and non-metallic elements, the latter in copper, silver, gold, lead, platinum etc. Very simple is also the hexagonal body-centred structure of zinc and several other elements. More complicated structures can be deduced from these by inserting new points in these lattices, which are all holosymmetric and which in the following text are called lattices of reference.

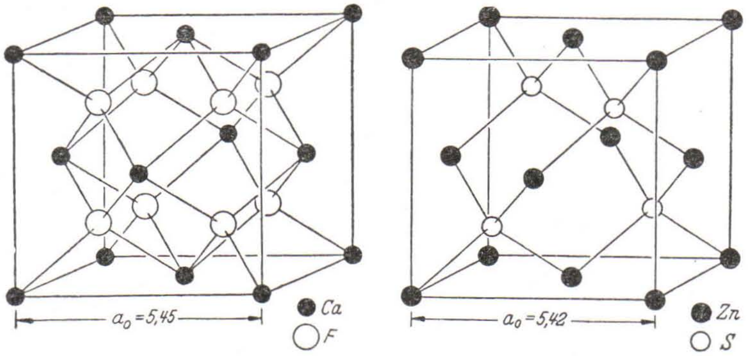

Fig. 1. a. Fluorite and b. Sphalerite.

The structure of rock salt is deduced from the face-centred cubic unit cell by placing Cl-ions between the Na-ions. The symmetry of the lattice has not been reduced thereby, and the crystal form is also holosymmetric. Similarly a F-ion can be placed in every octant of the cubic unit cell in the lattice of metallic calcium, which has the same structure as Na, and the holosymmetric crystal form is preserved. Consequently the insertation of new points in a lattice is without influence on the crystal symmetry, if the symmetry of the lattice has not been reduced.

In the structure of sphalerite, ZnS, the anions have the same places as F-ions in the fluorite, though these sites are only alternately occupied forming a tetrahedron. Thus the symmetry of the new distribution of points does not correspond to the symmetry of the lattice of reference; consequently the symmetry of the crystal form is reduced to that of a tetrahedron.

It is, however, to be noted, that if in fluorite the structure formed by the F-ions is taken as the lattice of reference, the Ca-ions form a tetrahedric system of points in it. The crystal symmetry is apparently determined by F-ions, which are twice as numerous as Ca-ions. In cuprite the Cu-ions form tetrahedra in a body-centred lattice of reference formed by O-ions. According to the former case the crystal form should be tetrahedric, but in the crystals from Cornwall enantiomorphous hemiedry has been noted. On the other hand Strunz considers cuprite belonging to the holosymmetric class.

In quartz trigonal enantiomorphous hemiedry or trapezohedron results from a screw structure formed by SiO_4-tetrahedra linked together.

Complicated structures also arise by replacing single ions with complex groups of atoms. Such groups consist of two, three, or more atoms, which are bound together by forces very much stronger than those binding the group to the rest of the structure. These complex ions usually do not have the spherical symmetry of a single atom but are arranged in a line (diatomic and triatomic groups), in a plane (BX_3- and BX_4-ions), or in a space group (BX_4- and BX_6-ions). If the complex ion is bound loosely enough to be in a free rotation, as is the case with CN-ion in KCN and frequently also with NO_3-ion, the holohedric crystal symmetry is not reduced. Thus KCN has the same symmetry as sodium chloride. Is the rotation prevented, the crystal symmetry is dependent on the space orientation of the complex ion.

In the structure of pyrite the group S_2 occupies in the face-centred lattice of reference (γ-Fe) the same position as Cl in sodium chloride, but is in some way so oriented that only those planes of symmetry coinciding with the axial planes remain. In carbides again two different types of structure appear depending on the direction of the linear group C_2. Here too the complex ion has the same arrangement as S_2 in pyrite, but in one type (CaC_2, LaC_2, UC_2 etc.) the complex ions are directed parallel to one edge of the cube, which is thereby extended so that the lattice becomes in fact tetragonal with the axial ratio $a:c < 1$. In the other type again (ThC_2, ZrC_2) the complex ions occupy the same interstitial sites, but lie parallel to a face of the cube with their axes in two mutually perpendicular directions. Here too a tetragonal lattice results with an axial ratio > 1. Consequently the reduction of symmetry by replacing a single ion with a complex group is not restricted to the hemiedric classes of the same crystal system but in certain cases we get forms belonging to another crystal system with lower symmetry.

The same thing appears from the following well known example. The structure of $NaNO_3$ is similar to that of NaCl with the Cl-ions replaced by the complex ion NO_3, which is planar, and its plane is directed normal to a triad axis. Consequently the cubic structure of the lattice of reference is distorted by compression along the triad axis and a rhombohedric structure results. A similar structure occurs in calcite and in many other carbonates. Frequently, however, the NO_3-group is found in free rotation behaving as a spherical ion; in this case the cubic structure is not distorted.

The structure of potassium chlorate, $KClO_3$, is also closely related to this type. The pyramidal arrangement of ClO_3-ion, however, as contrasted with the plane form of NO_3-group gives rise to a lattice of monoclinic symmetry only.

The ion BX_4 may also sometimes be planar as in K_2PtCl_4, but usually this group has a space co-ordination of either regular tetrahedral or distorted tetrahedral form. It occurs in sulphates, phosphates, molybdates, tungstates and in many other compounds. This ion has also been found in rotation, and many times the regular tetrahedra have a symmetric position and do not have an influence on the symmetry of the lattice of reference. Remarkable is also the role of the SiO_4-tetrahedra in silicates. This group is not a complex ion in the same sense as those mentioned above, the bindings having no stronger values than the usual ionic bonds, but it occurs also as an anion e. g. in olivine, garnets, zircon and in many others. Garnet deserves special attention because it shows how a very complicated structure may have a high degree of symmetry. Especially interesting is the comparison of the structure of zircon and scheelite. These are qualitatively similar, but the SiO_4-tetrahedra in zircon are regular and have positions corresponding to the tetragonal symmetry, while the complex ions WO_4 in scheelite are distorted tetrahedra and have oblique positions causing a lowering of symmetry from holosymmetric form of zircon to the paramorphic hemiredric class in scheelite.

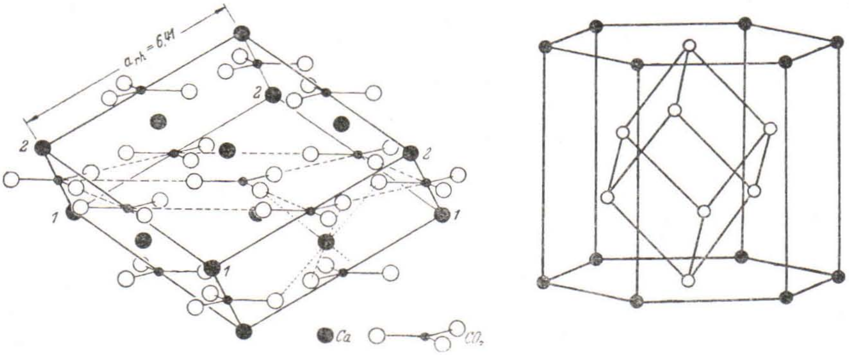

Fig. 2. a. Calcite. b. Rhombohedric position of additional points in the hexagonal unit cell.

Further the lowering of symmetry is caused by replacement of two similar ions or atoms with different ones either in a complex ion or at two independent points. The relation of cobaltite and ullmannite to pyrite is a good example in this respect. In cobaltite and ullmannite the structure has been found to be similar to that of pyrite, one of each pair of sulphur atoms being replaced by an atom of arsenic in cobaltite and by an atom of antimony in ullmannite. In this structure there is no centre of symmetry as in pyrite and a polar tetartohedric crystal form results instead of the paramorphic hemihedric form of the pyrite.

The structure of dolomite is similar to that of calcite except that alternate calcium ions lying in a plane normal to the vertical axis are replaced by magnesium ions. Therefore no plane of symmetry is left, and the hemihedric symmetry of calcite is reduced in dolomite to that of a tetartohedric rhombohedron. The same is the relationship between hematite and ilmenite. It is interesting also to point out the similarity of the structures of sphalerite and chalcopyrite. These are closely alike with the exception that the Zn-ions in sphalerite are alternately replaced by Cu-ions and Fe-ions. Thereby the interval of identity becomes twofold in one direction and remains unchanged in the others: in sphalerite $a_0 = 5{,}42$; in chalcopyrite $a_0 = 5{,}24$; $c_0 = 10{,}30$. The crystal form of sphalerite is tetrahedric, that of chalcopyrite is tetragonal bisphenoidal.

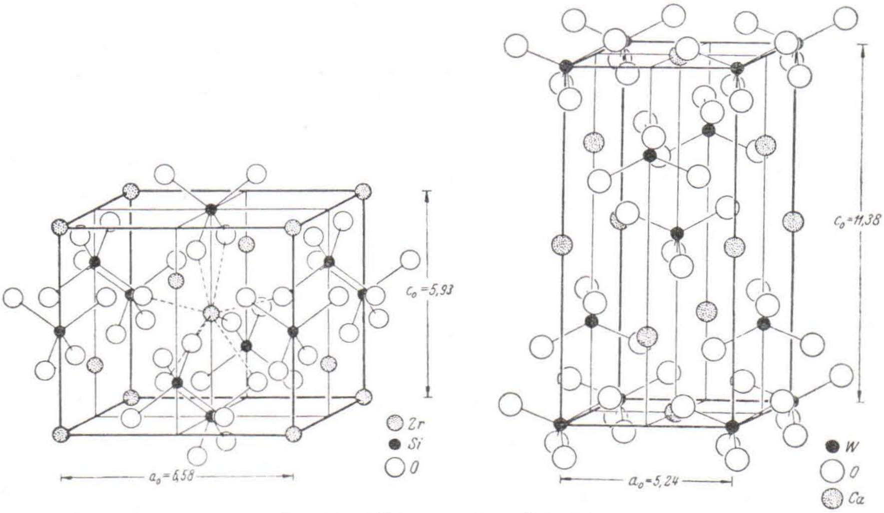

Fig. 3. a. Zircon and b. Scheelite.

The ability of SiO_4-tetrahedra to be linked together in different modes is a farther way to complex structures. The tetrahedra are aggregated to isolated groups, rings, chains, sheets, and frameworks. The structures composed of rings of 3, 4 or 6 tetrahedra possess the highest degree of symmetry, respectively revealing itself in trigonal, tetragonal, or hexagonal forms. On the other hand the frameworks usually have the lowest symmetry but exceptional cases exist.

These selected examples present some of the simplest cases concerning the relationship between lattice structure and crystal form. The material is much more extensive, but these may be sufficient to give an idea of this relationship. It is seen that the simple elements have very simple crystal structures: cubic face-centred or body-centred or hexagonal body-centred lattices, the close-packed cubic and hexagonal structures. The structure of diatomic compounds is defined by the ratio of atomic radii. Therefore their structure is often very simple, accordingly their crystal form often has a very high degree of symmetry such as is shown by the holosymmetric class of the cubic system. With growing complexity of the compound the structure too becomes more complex, because it is dependent upon several rations of ionic radii. Accordingly the symmetry becomes more restricted.

The cubic face-centred lattice occurs in many metals and crystallizes in holosymmetric form. This same structure as a lattice of reference with inserted points gives a pentagonal form in pyrite, a tetrahedral hemihedric form in sphalerite, and a tetartohedric form in cobaltite. In other crystal systems similar instances are met with. In general the complicated structures seem to prevail in the triclinic system.

Thus by starting from the crystal structure and not from the crystal form and by following the inductive principle it is more logical to begin the order of crystal classes from those of highest symmetry, i. e. of the holosymmetric classes and deduce from these the different hemihedric, tetartohedric and hemimorphic forms.

In this deduction a feature deserves special notion. Viz., in certain changes of the structure one set of planes of symmetry disappears, while the other remains, and in some other changes the sets of planes of symmetry behave in the opposite way. We must thus distinguish between two sets of planes of symmetry, those normal to a trigonal, tetragonal, or hexagonal axis of symmetry, and those normal to a digonal axis. We may call these the first and second order planes of symmetry. In this respect the scalenohedron (the rhombohedron) has the same relationship to the holosymmetric dihexagonal (hexagonal) bipyramid as the tetrahedron to the octahedron and the tetragonal scalenohedron (the sphenoid) to the holosymmetric ditetragonal (tetragonal) bipyramid. In all these hemihedric forms the first order planes of symmetry have disappeared, while the second order planes of symmetry or at least half of them remain. As Niggli has pointed out in his textbook mentioned above, the rhombohedric lattice can be deduced from the hexagonal structure by distributing internal points into the hexagonal lattice analogically with that causing tetrahedric hemihedry in the cubic lattice (fig. 1—2). By distributing the points in a different way screwlike chains may be formed, whereby both sets of planes of symmetry disappear and the enantiomorphous class results. If again the changes, causing the two hemihedries named above, appear at the same time, the result is the tetartohedric class (dolomite, ilmenite). The hemimorphic forms originate from the holosymmetric forms and from the first hemihedry through appropriate changes.

The system of crystal classes constructed on the basis of these facts is represented in the accompanying table. In this table three different groups are to be distinguished.

The first group involves the five classes belonging to the cubic system

Degree of symmetry	Cubic system	Tetragonal system	Hexagonal system	Trigonal system	Rhombic system	Monoclinic system	Triclinic system
Holosymmetric class	Six-faced Octahedron Oh	Ditetragonal Bipyramid D_4h	Dihexagonal Bipyramid ← D_6h	Ditrigonal Bipyramid D_3h	Rhombic Bipyramid D_2h	Monoclinic Prism C_2h	Triclinic Pinacoid Ci
The first or Paramorphic hemihedry	Diploid Th	Tetragonal Bipyramid C_4h	Hexagonal Bipyramid C_6h	Trigonal Bipyramid C_3h	—	—	—
The second or Sphenomorphic hemihedry	Six-faced Tetrahedron Td	Tetragonal Scalenohedron D_2d	Hexagonal Scalenohedron D_3d ↖	—	Rhombic Bisphenoid D_2	Monoclinic Sphenoid C_2	Pedion C_1
The third or Enantiomorphic hemihedry	Plagihedron O	Tetragonal Trapezohedron D_4	Hexagonal Trapezohedron D_6	Trigonal Trapezohedron D_3	—	—	—
Tetartohedry	Tetrahedral-pentagonal Dodecahedron T	Ditetragonal tetartohedric Bisphenoid S_4	Hexagonal tetartohedric Rhombohedron S_6 (C_3i)	—	—	—	—
The first or Holosymmetric hemimorphy	—	Ditetragonal Pyramid C_4v	Dihexagonal Pyramid ← C_6v	Ditrigonal Pyramid C_3v	Rhombic Doma C_2v	Monoclinic Doma Cs	—
The second or Hemihedric hemimorphy	—	Tetragonal Pyramid C_4	Hexagonal Pyramid C_6	Trigona Pyramid C_3	—	—	—

and comprising crystals which as to the most of the physical properties are isotropic. Besides the holosymmetric class three hemihedric classes and one tetartohedric class exist. In the first or paramorphic hemihedry, there occur the first order planes of symmetry of the holohedric class with the axes degraded to digonal ones, while the second order planes have disappeared (pyrite). In the second (or sphenomorphic) hemihedry, the second order planes of symmetry are left and the first order planes have disappeared (sphalerite). In the third hemihedry no planes of symmetry exist. In the tetartohedry the planes of symmetry are also nonexisting.

The second group is just as homogeneous as the first concerning the physical properties of the crystals. Crystallographically this group is characterized by a principal axis of symmetry, which is trigonal, tetragonal, or hexagonal. Accordingly, the group is divided into three systems. On the ground of physical (e. g. optical) properties it is not possible to distinguish these from each other. Thus if due justice will be given to the physical properties of crystals and consequently to the structure, full analogy must exist as far as possible between all the systems belonging to this group in their division in crystal classes. Complete analogy between the tetragonal and the hexagonal system is really arrived at, if both of them comprise seven classes including one holosymmetric, three hemihedric, one tetartohedric and two hemimorphic classes. In this case the trigonal system will include only five classes, viz. a holosymmetric, two hemihedric, and two hemimorphic ones. One hemihedric class and the tetartohedric class of the trigonal system are not possible. The hemihedric and the tetartohedric classes are analogous also with those in the cubic system. In the first or paramorphic hemihedry, the principal plane of symmetry corresponds to the first order planes of symmetry, but the other planes of symmetry (the second order planes) have disappeared analogously with pyrite as a result of unsymmetric behaviour of a complex ion (scheelite). This hemihedry occurs in all three systems. On the second or sphenomorphic hemihedry, the second order planes of symmetry exist, but not the principal plane of symmetry. Such a class exists only in the tetragonal and the hexagonal system but not in the trigonal, because 3 is not divisible by 2. The lattice structure contains a tetrahedric or rhombohedric distribution of points (chalcopyrite, calcite). In crystals belonging to the classes of the third or enantiomophous hemihedry, all planes of symmetry are lacking. The screwlike distribution of points occuring in the crystal structure within these classes is represented by the trigonal quartz. The rhombohedron occuring in the crystal forms of quartz can be considered to be formed as an ultimate form of ditrigonal trapezohedron, in such a case that the ditrigonal bipyramid has the same form as the hexagonal bipyramid. The ditrigonal prism can also have the form of a hexagonal prism, well known to occur in quartz. This also explains the fact that forms seemingly belonging to two different classes or systems appear in the same crystal. By the tetartohedric classes, too, no planes of symmetry appear, and such a class is nonexisting in the trigonal system, because 6 is not divisible by 4. In the hexagonal system the crystal form of this class has the form of a rhombohedron and is in this respect analogous with the tetartohedric form in the cubic system where it has the tetrahedric shape. The structure of the tetartohedric class is the result of unification of two different motives, those of first and second hemihedry, as indicated by the structure of dolomite and ilmenite. Of the two degrees of hemimorphy one is deduced from holosymmetric forms and the other from paramorphic hemihedry, because these only have the principal plane of symmetry, the disappearance of which results in hemimorphy.

The third group includes the rhombic, monoclinic, and triclinic crystal systems. As none of them has the first order plane of symmetry, the paramorphic hemihedry as well as the hemimorphy deduced from this are lacking. All these systems can be distinguished from each other also by their physical (optical) properties. The enantiomorphic hemihedry with the screw-like structure and the tetartohedry are also impossible, because all directions are dissimilar. Consequently the rhombic and monoclinic system contain only three classes corresponding to those of the former group. These are the holosymmetric, the second hemihedric and the first hemimorphic classes. In the triclinic system only two classes are possible, because the only element of symmetry possible here is the centre of symmetry. The class without this may be taken either as a hemiedry or a hemimorphy.

The method of treatment presented here presupposes the main features of the crystal structure to be known. Should this method be used in instruction, teaching of crystallography must start with leading lines of the crystal structure. This has, however, the advantage, that the treatment becomes thereby more interesting than the lectures on the crystal forms alone. The system of crystal classes presented above is also more perspicuous and therefore easier to keep in mind than the customary form of the system. For an instructed mineralogist this is perhaps a minor advantage, but for students the easiest system is the best one.

THE MINERALOGICAL INSTITUTE
OF THE TECHNICAL UNIVERSITY IN HELSINKI.
IN APRIL 1947.

15.

THE BIRKELAND GRANITE, A CASE OF PETROBLASTESIS

BY

TOM. F. W. BARTH

CONTENTS

I. DESCRIPTION AND COMPOSITION

Figure 1 represents a map of the pre-Cambrian rock types of the Skagerrak costal region of Southern Norway.

At two places there appear ostensibly younger bodies of granite which, when typically developed, is conspicuously different from the surrounding gneiss. On the shore, just east of the town Grimstad is the Fevik or Grimstad granite, and about 10 km N. W. is the larger area of the Birkeland granite. The two granites are rather similar petrographically and in their contact relations.

A prominent friction breccia, which extends northeastward beyond the area of the map covering a total distance of about 350 kilometers (2), following a deep gorge, cuts through the Birkeland granite, without noticeably displacing one part of it in respect to the other. This shows that there has been no great motion of faulting accompanying the development of the breccia since the time of the formation of the Birkeland granite which—as will be shown presently—took place during the waning stages of the pre-Cambrian orogenesis, the cause of all rock changes encountered in this area.

A detailed map of the Fevik granite has been published by Ivar Oftedal (6), the Birkeland granite has never been mapped before; it is here shown in Fig. 2.

In the southern and western parts of the area the granite is massive, homogeneous, coarse-grained (feldspar crystals of ca. 3 cm) with a typically red color, very similar to the chief type of the Fevik granite.

A chemical analysis of it is listed in Table I. For comparison are entered two other South Norwegian pre-Cambrian granites entirely surrounded by gneisses, and presumably of similar mode of origin. They both are perfectly massive, but occupy much smaller areas than does the Birkeland granite. (2) The Oddersjå granite is located approximately 30 km S. W. of Birkeland, just north of the town Kristiansand. A detailed map and a discussion of its genesis have been published by me (1), but no analyses of it was available at that time. (3) The Grönsfjord granite occupies a small area on the Lindesnes Peninsula just beyond the limits

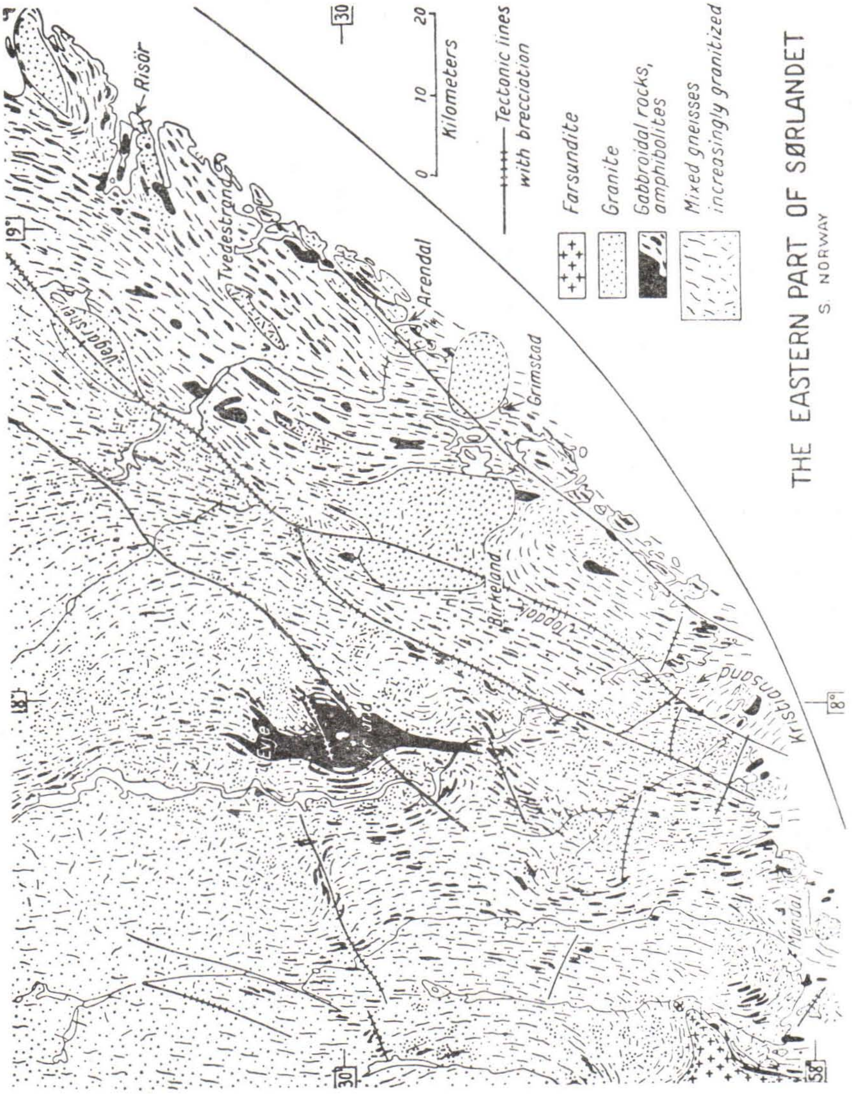

Fig 1.

Table I

CHEMICAL ANALYSES

	1	2	3	4	5
SiO_2	69.27	71.10	72.17	70.63	63.68
TiO_2	0.50	0.81	0.48	0.27	0.36
Al_2O_3	15.53	13.24	13.74	14.25	19.89
Fe_2O_3	0.96	1.21	2.76	0.23	0.70
FeO	1.15	3.00	0.38	1.95	0.60
MnO	—	—	tr	0.04	—
MgO	0.55	0.58	1.04	1.41	0.13
CaO	1.56	1.70	0.31	1.47	1.38
BaO	n. d.	0.11	n. d	0.13	n. d.
Na_2O	4.28	2.27	1.33	1.58	2.35
K_2O	5.50	5.35	5.98	7.40	11.60
P_2O_5	0.03	0.03	0.05	0.18	—
CO_2	0.28	—	—	none	—
H_2O+	0.24	0.36	1.30	0.69	—
H_2O-	0.12	0.10	0.16	0.11	—
S	—	0.05	—	tr	—
	99.97	99.91	99.70	100.35	100.15

1. Birkeland granite, Reddal. Astri Thorkildsen analyst.
2. Granite, Oddersjå, Barth analyst.
3. Granite Grönsfjord, Bruun analyst.
4. Composite granite-pegmatite, Newry. Quoted from Reynolds (8).
5. Microcline from Birkeland-granite, Thorkildsen analyst.

of the map in Fig. 1. It has never been described . (4) In this column is entered also for comparison, a composite granite-pegmatite described by Reynolds (8) as a replacement body of granitic composition consisting of porphyrablastic crystals and nodes of perthitic microcline developed within plagioclase-biotite-hornfels.

The Birkeland granite has such a simple mineral composition that it is not necessary to go into details as to the recalculation of the chemical analysis into mineral constituents. It is enough to mention that the perthitic microcline, which makes up a little less than one half of the rock, has been analyzed, and its analysis is given under (5) in Table I. The plagio-

Table II

MINERAL CONTENTS OF THE CHIEF TYPE OF BIRKELAND GRANITE AND THE CHIEF TYPE OF FEVIK GRANITE.

	Birkeland Equ. Mol %	Fevik Vol. %
Quartz	18	20
Microcline perthite	44	44
Plagioclase (An_{11})	33	32
Biotite	4	3
Ore	1[1])	1[2])

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[1] Contains traces of calcite.

[2] Contains a fraction of titanite and apatite.

clase has been determined optically as albite-oligoclase with 11 An. In addition to quartz a small amount of biotite and ore is present in the analyzed sample. Titanite is seen in the Birkeland granite, but was not encountered in this sample.

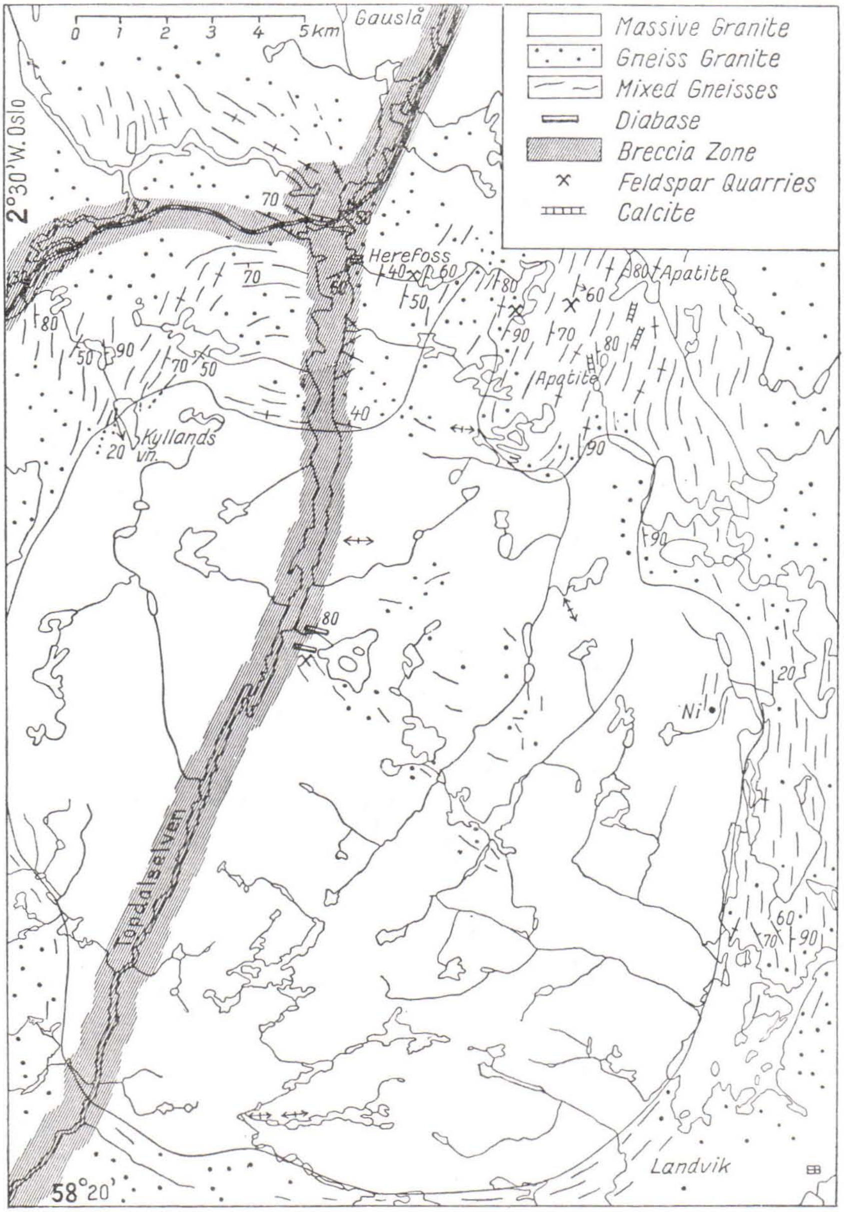

Fig. 2. The Birkeland Granite Area.

By using the Niggli equivalent molecular percentages the minerals can be easily calculated, and come out as listed in Table II. With this is compared Oftedal's estimate of the mineral composition of the Fevik granite. The table demonstrates that the mineral composition of the chief type of the Birkeland granite is identical to that of the chief type of the Fevik granite.

II. INCLUSIONS AND CONTACT RELATIONS

The Birkeland granite is far from uniformly developed within the area. The chief type, which has been analysed and described on the preceding pages dominates in the Southern and Western parts.

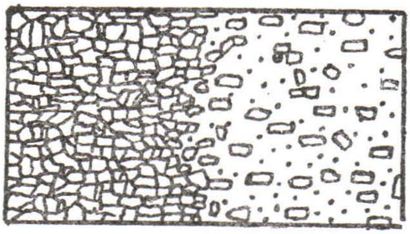

Fig. 3. Coarse-grained and fine-grained types sharply separated. E. of Topsli. Length of picture about 1 m

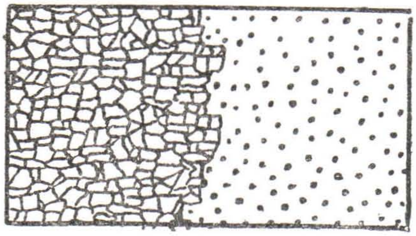

Fig. 4. Relation between a coarse-grained type and a porphyritic type. E. of Topsli. Length of picture about 1 m.

Along the breccia the granite is partly mylonitized and again compressed to a dense, red felsite, it is intensely shattered and full of veins and stringers of quartz or, not unfrequently, fluorite.

In the northeastern parts various granite types are encountered: coarse-grained types alternate with fine-grained types grading into true aplites; nests and lenses of pegmatite are also present. All types are massive but are frequently intersected by dikes of pegmatite or aplite, in most places one can see no indication of either flow or foliation. Two types may grade into each other, or they may show sharp boundaries. See Figs. 3 & 4. Inclusions are abundant. They consist of gneisses in a high state of granitization. Usually the schistosity conforms to the boundaries of the inclusions, but frequently they exhibit no sharp outlines; by transitions granite grades into gneiss; black amphibolite becomes grey and appears as ghost-like relics in the granite. Often diffuse granitic areas have developed within the inclusions (see Fig. 5); pegmatites too may form here. A feldspar quarry has been opened on a pegmatite lens in a gneiss inclusion (marked on the map close to two small diabase dikes cutting the granite).

Mapping of the inclusions has not been attempted; some of them are small, but it seems as others may be one or two kilometers long.

The contact between the granite and the surrounding gneisses is usually sharp. This is always true in the southern part of the area where the typical coarse, homogeneous granite prevails. The strike of the gneisses is mostly north and south, the dip is steep. But it is worthy of notice that the schistosity of the gneisses swings around the granite body and orients itself parallel to the walls of the granite; therefore no typical intrusive contact can be seen. See Fig. 2.

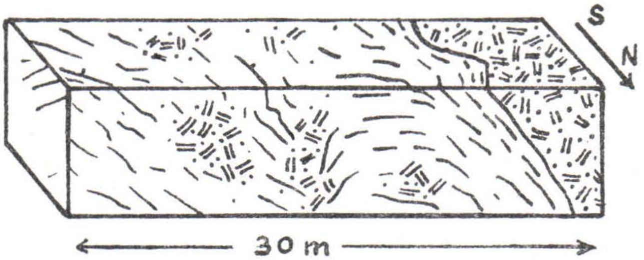

Fig. 5. The western contact of a gneiss inclusion in Birkeland granite at Begervann (east of the feldspar quarry). Strike and dip in the gneiss are shown, the contact towards the granite (right end of section) is concordant, but sharp. Within the inclusion several diffuse patches of petroblastic granite have developed.

The contact in northwest is interesting. North of Kyllandsvann the gneiss complex consists of large bodies and lenses of amphibolite which, as one approaches the granite, become more and more massive; and finally, close to the border, look like massive gabbroic rocks. Now, these massive gabbroic lenses can be traced into the adjacent granite as dark »shadows» which finally become completely assimilated and disappear as ghost-like relics in the ordinary granite.

The most interesting contact phenomena are displayed in the extreme north of the granite area: East of Herefoss it is impossible to draw any boundary line between granite and gneiss. In the homogeneous granite, pegmatites and aplites will begin to develop: Small nests and pockets of pegmatite spread out over large areas, or consolidate into ridges and dikes of a general north-south trend; fine-grained granite develops into aplite which in turn becomes a light massive gneiss, coarse-grained granite becomes augengneis. Thus an area of fairly homogeneous granite by gradual transitions is transferred into a complex of augengneisses and fine-grained gneises interwoven with ridges of pegmatite and lenses of aplite.

It is interesting that similar phenomena have been observed in the Fevik granite. Jens Bugge (3) writes: »In addition to the main rock type we find smaller areas of more basic rocks (which) are situated in the strike directions of the great amphibolite dikes (of the surrounding gneiss) and probably are formed through assimilation of material from these rocks».

Although Ivar Oftedal *(loc. cit)* does not quite agree to this, he, too, considers the Fevik granite as of palingenic origin and describes inclusions and contact rocks very analogous to those observed by me in the Birkeland granite. There is little doubt that the two granite areas are genetically related.

Finally attention should be drawn to the fact that rocks similar to the Birkeland granite occur at various places outside the granite area, obviously representing constituent parts of the gneiss complex. Lenses and layers of massive granite, virtually identical to the typical coarse Birkeland granite, may be seen at innumerable places in the gneiss. Likewise a large granite area with rather sharp boundaries towards the gneiss exists northeast of the apatite mine (northeast corner of the map). And massive granite also extends along the railroad from Herefoss and westward.

The previously mentioned granites at Oddersjå and Grönsfjord (See Table I) belong to the same category of rocks.

III. THE ROCK SERIES: GNEISS-AUGENGNEISS-GRANITE

In order to appreciate better the interrelation of gneiss and granite, it is necessary to describe some of the rock types in the gneiss area.

(1) Augengneisses: There are various types, all transitional into one another and into pegmatite. One of the types is coarse-grained, massive and identical to the Birkeland granite. It often contains gneiss inclusions.

(2) Fine-grained gneisses: Various types, aplitic, reddish, but most commonly grey. The grey gneiss frequently contains granite inclusions and is cut by veins and dikes of augengneiss transitional into pegmatite.

But at other localities the relation is completely reversed: Dikes of grey gneiss cut the augengneiss. A closer inspection reveals the fact that these dikes entered the augengneiss before it had become an augengneiss; for the grey gneiss dikes themselves are in places »augenized».

It can be demonstrated in the field that the formation of the augengneisses corresponds to a regional »augenization» of the area. Some rock types are more susceptible to it than others. The grey gneiss seems to be rather resistant, and has often set a barrier to the penetration of the augens; in other places, it, too, has had to yield; augens begin to grow within it, first small and scattered, later more numerous and bigger.

Consequently the formation of the augengneisses is the youngest petrogenetical event in this region.

Obviously it was connected with a relaxation of the pressure. Although I think it is wrong to call all augengneisses post-tectonic, I agree with Rosenqvist (12) that they are causally related to decreasing pressure and steep gradients. The formation of the breccia and the formation of augengneisses are thus cause and effect. Likewise the Birkeland granite is a product of the same cause. By gradual transitions it shades into augengneisses, and the field evidences indicate that the same agents that made the feldspar augens grow in the gneisses, also made all feldspar crystals grow in the Birkeland granite. Thus the formation of the typical coarse Birkeland granite within the granite body is correlative to the formation of the augen in the gneiss bodies. (See Figs. 3 & 4).

Conclusion: Augengneisses were formed in a broad reaction zone north of the Birkeland granite. Feldspar porphyroblasts grew in the granite and in the surrounding gneisses thereby producing the transition Birkeland granite→gneiss.

IV. DISCUSSION OF ORIGIN

The preceding description suffices to show that the Birkeland granite did not take birth simply by congealing of »younger» granitic magma. The peculiar contact relations and the gradual transition granite-gneiss call for another explanation.

In my discussion of the origin of the Oddersjå granite 15 years ago, I came to the conclusion that it generated through a very slow process of metasomatic replacement. Following the ideas of Sederholm which had been further developed by Eskola in his general work on pore liquids and metamorphic differentiation it was assumed that an ichor of granitic composition was formed at great depths in the interstices of the mineral grains; driven by its own buoyancy the ichor slowly rose to higher levels, dissolving, assimilating, or altering metasomatically every molecule of the superjacent rock masses. Thus were formed augengneisses and granites.

Wegmann (13) repeatedly emphasized the difficulties in finding the channels through which the ichor had streamed, the feldspar porphyroblasts are ductless. Often they penetrate the rocks like oil stains . . . .» Oft ist die Struktur des früheren Gesteins durch die Feldspatbestäubung hindurch sichtbar» (p. 321). To explain this Wegmann made the physical nature of the ichor more indefinite by introducing the conseption of an intergranular film . . . .» Im Intergranularfilm können zugeführte Stoffe wandern» (p. 325).

In many other localities it has been shown that granite and allied rocks have developed as replacements of pre-existing rocks.[1] But the temperature and the physico-chemical nature of the »ichor» which soaks through the rocks are still much debated.

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[1] For example: ANDERSON, A. L.: Contact Phenomena Associated with the Cassia Batholith, Idaho. J. Geol. 42, 1934. ANDERSON, G. H.: Granitization, Albitization and Related Phenomena in the Northern Inyo Range of California-Nevada. Bull. Geol. Soc. 48, 1937. BACKLUND, H. G.: The Problems of the Rapakivi Granites, J. Geol. 46, 1938. GOODSPEED, G. E.: Small Granodiorite Blocks Formed by Addition Metasomatism, J. Geol. 45, 1937. Pre-Tertiary Metasomatic Processes in the South Eastern Portion of the Wallova Mountains, Oregon. Proc. Sixth Pacific Sci. Congress, 1939. MACGREGOR, M.: The Western Part of the Critfel-Dalbeattie Igneous Complex. Quart. Jour. Geol. Soc., London, 93, 1937., The Evolution of the Critfel-Dalbeattie Quartz Diorite: A Study of Granitization. Geol. Mag. 75, 1938. QUIRKE, T. T. and COLLIN , W. H.: The Disappearance of the Huronian. Mem. Geol. Surv. Canada No. 160, 1930. REYNOLD , D. L.: The Eastern End of the Newry Igneous Complex. Quart. Jour: Geol. Soc. 90, 1934., The Two Monzonite Series of the Newry Complex. Geol. Mag. 73 1936. The Albite Schist of Antrim and theirPetrogenetic Relationship to Caledonian Orogenesis. Proc. R. Irish Acad. 48 B, No 3, 1942., Granitization of Hornfelsed Sediments in the Newry Granodiorite of Coraghwood Quarry, Co. Armagh. Proc. R. Irish Acad 48 B. No 3, 1943.

Recently evidence has been presented that the ichor is not necessarily a liquid. Ions or atoms may migrate through rocks without being carried in solution. It has been recognized that diffusion is of great importance in petrogenetic processes.

Following the best traditions in the French-school of Geology Perrin and Roubault (7) have developed the idea of large scale reactions in the solid state. Their conclusion is not widely known, so it is given here in their own words: (p. 147) »En examinant les preuves envoquées en faveur du granite liquide, nous avons constaté que les seule qui paraissent vraiment très fortes en premier examen, reposent avant toutes choses sur des impression subjectives, sur des apparances évoquant l'état fondu, par assimilation avec les phénomènes connus de tous; mais le détail des observation prouve que les réactions dans le solide peuvent donner naissance et ont effectivement donné naissance dans certains cas à des apparences toutes semblable, même plus étranges encore.»

In England Reed (10, 11) has developed similar ideas.

Likewise in the Oslo Mineralogisk Institutt the importance of large scale diffusion and migration of atoms in solid rocks has been recognized. According to Jens Bugge (4) large areas of (semi)- massive rocks some tens of kilometers N. E. of the Birkeland granite have originated in this way. According to Ramberg (9) rocks may originate and grow by slow diffusion of rock-forming ions of high energy towards a place where they lose the energy and consolidate in crystals; for such rocks was introduced the name petroblastic rocks. Such rocks are neither magmatic, nor metasomatic, nor metamorphic.

According to the preceding description of the Birkeland granite it is probably a petroblastic rock. It is definitely not magmatic, nor metamorphic. Also it is difficult to call it metasomatic, although it is of course true that it owes its existence to metasomatic processes. There is not much difference between petroblastic and metasomatic: If the diffusible, migrating ions directly crystallize to a rock body, it is petroblastesis, if they react with a pre-existing rock, and then crystallize, it is metasomatism.

In the present area the cloud of migrating ions had a composition corresponding to that of the granite (See Table I, and observe the uniformity of composition of analyses 1, 2, 3, and 4). Whenever a granite is formed it corresponds more or less closely to the consolidated »cloud of ichor» itself.

As a petroblast the Birkeland granite body very slowly grew by contributions from the cloud of ichor. The heat of consolidation might conceivably be large enough to develop »igneous» contacts, locally. None have been observed, however. - - Its force of crystallization was strong enough to push aside the adjacent gneisses and make them concordantly fold around and follow its own body. At the same time the ichor cloud penetrated all the adjacent rocks in the whole area, metasomatically trans-forming them to gneisses; and whenever conditions were right the uncontaminated ichor itself would crystallize, forming in the gneiss terrain countless patches of massive petroblastic rocks of granitic or pegmatite-granitic structure and composition. The largest of these patches is the Birkeland granite.

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16.

ZUM TEKTONISCHEN STIL VON PALINGENGRANIT UND MARMOR IN DEN SVEKOFENNIDEN IN FINNLAND

VON

ADOLF A. T. METZGER

(Mit 4 Abbildungen im Text)

Es kann auf den ersten Anblick eigentümlich erscheinen, wenn zwei so verschiedene Gesteine, wie Palingengranit und Marmor, vom gleichen Standpunkt aus behandelt werden sollen. Tatsächlich aber zeichnen sich beide Gesteine gemeinsam durch hohe Mobilität aus. Da sie ausserdem ihre heutige räumliche Gestalt gleichzeitig erhalten haben, können beide als recht bezeichnende Indikatoren des kinetischen Ablaufs der Tektogenese angesehen werden.

Die heute vorliegenden tektonischen Verhältnisse in den Svekofenniden Finnlands sind, was die Faltentektonik betrifft, in erster Linie das Ergebnis der serorogenen Phase in der Gebirgsbildung. Bezeichnend für den Faltenbau ist das Vorhandensein die verhältnismässig flache, d. h. nur wenige Kilometer tiefgehende Lage der Hauptelemente, verbunden mit oft eng verfältelten und steil stehenden Einzelheiten. Noch vor kaum mehr als einem Vierteljahrhundert hatte man die Auffassung, dass das Grundgebirge innerhalb der Svekofenniden allgemein steile, oft gegen die Teufe praktisch genommen ins Unendliche fortsetzende Elemente aufzuweisen habe. Die seit 1922 vom Verf. betriebenen tektonischen Studien in den Svekofenniden, in erster Linie in den Gebieten von Pargas, Lohja und Sibbo, haben uns jedoch bald zu der Überzeugung gebracht, dass es sich doch um recht flachliegende Grosselemente handelt, eine Meinung die wir erstmals im Zusammenhang mit der Beschreibung des Gebietes von Svartå-Mustio (1928) vertraten. Später haben die schönen Untersuchungen von Wegmann (1931) und Kranck (1931—1937) im Gebiet um Helsingfors und an der Südküste, von Mikkola im Gebiet von Lohja (vergl. Eskola 1941), von Parras im westl. Uusimaa (1941), von Hietanen (1943) im Gebiet von Kalanti und von Hausen (1944) im åboländischen Schärenhof deutlich gezeigt, dass unsere Erfahrungen auch auf andere Teile der Svekofenniden übertragen werden können. An allen Stellen zeigte es sich, dass die Steilstellung der Kulissen zwar wirklich bestand, dass aber die Elemente im grösseren Verbande oft nur geringen Tiefgang zeigen, also gegen die Teufe ausflachen. Weiter zeigte es sich, dass die Spuren der einzelnen Elemente sich gern zu Bögen und Girlanden aneinander reihen, Axialkulminationen oder Depressionen umschliessend. So erscheint es uns heute als ob die Svekofenniden in unserm Lande aus sich stetig ablösende Kulminationen und Depressionen der Hauptaxe aufbauen, wobei das Ganze recht flach liegt.

Mit Wegmann stehen wir auf dem Standpunkt, dass für die erste Deformation der Svekofenniden ein alpider Stil angenommen werden kann. Aus dieser Periode, der primorogenen Phase, haben wir im Allgemeinen wohl nicht viel deutliche Bauformen erhalten. Wir wissen jedenfalls heute mehr vom stofflichen Inhalt als von der Form dieser Phase. Wir wissen, dass das Gebirge aus gewaltigen Sedimentmassen tonigen und tonig-sandigem bis sandigem Material aufgebaut war, wofür uns die tonerdereichen Gneise (Kinzigitgneise), die Parras neuerdings (1946) treffend Lutogenite benannt hat, mit ihrer weiten Verbreitung ein Zeugnis ist. Mit ihnen vergesellschaftet finden sich als sedimentogene Bildungen Diopsidamphibolite (Mergelgesteine) und kalzitische bis dolomitische Marmorlager. Das gesammte Sedimentmaterial wurde jedoch in die Tiefe gefördert und unterlag der regionalen Umschmelzung (Anatexis), deren Hauptergebnis der serorogene Palingengranit wurde. Mit seinem Auftreten ist eine zweite (vielleicht sogar die dritte?) Faltungsphase verbunden. Die während dieser entstandenen Faltungsformen, sind es, die uns heute erhalten sind.

In unserer Beschreibung des Gebietes von Pargas (1945) haben wir versucht, den Ablauf der Palingenese näher zu schildern. Es soll hier nicht näher auf die petrologische Seite des Problems eingegangen werden Was uns hier interessiert, ist die tektogenetische Bedeutung der Bildung des Palingengranits Wie wir bereits betont haben, ist es anzunehmen, dass die reinen petrochemischen Reaktionen der Schmelzung, die ja nach Eskola (1932) als metamorphe Differentiation bezeichnet werden kann, auch von mechanischer Trennung in geschmolzene und ungeschmolzene Massen begleitet worden ist Diese Erscheinung nannten wir metamorphe Segregation. Das ungeschmolzene Material, das heute als Reste im palingenen Migma vorliegt, war entweder rein bauschalchemisch und mineralparagenisch nicht im Stande an der Anatexis teilzunehmen, oder es wurden ihm die schmelzbaren Bestandteile bereits entzogen. Was dann nachblieb, spielte die Rolle der Schlacke.

Diese Trennung in Schmelze und Ungeschmolzenes musste natürlich weitgehende Bedeutung für die Tektogenese bekommen. Hochmobiles palingenes Material durchströmte bereits durch Faltung etwas versteiftes Gebirge. Es liegt in der Natur der Sache, anzunehmen, dass es zuerst die tieferen Teile des Gebirges waren, die der Palingenese anheimfielen. Es wird der Prozess an ganz bestimmten günstigen Stellen begonnen haben, Stellen, die dann als Infektionsherde auf die Umgebung gewirkt haben müssen. Zuerst wird es sich um recht bescheidene Schmelzausscheidungen gehandelt haben, die aber allmählich immer weiter ausgreifend sich in

Schmelzflüssen vereinigt haben werden und so in die Höhe strebend allmählich mit ungeheuerer Kraft in den Bau des Gebirges eingegriffen haben werden. Zweierlei Bewegungen werden hauptsächlich stattgefunden haben, nähmlich einmal ein stetiges Ansteigen des palingenen Migmas verbunden mit Raumschaffung durch Aufschmelzung neuer Massen und seitlichem Zusammenschub nichtschmelzender Massen, und zweitens ein Absinken überliegender unschmelzbarer Gesteinskörper in das in der Tiefe sich entwickelnde und stark erweichte palingene Migma. Den ersteren Bewegungsstil hat Wegmann bezeichned das Aufsteigen der Migmatitfront genannt (1936). Auf den letzteteren Prozess haben wir in der Pargas-Arbeit aufmerksamgemacht.

Die soeben vorgebrachte Ansicht setzt also umfassende Materialtransporte und Bewegungen voraus. Diese betreffen in erster Linie den Palingengranit selbst. Dies steht nun scheinbar im Widerspruch zu der bekannten Tatsache, dass man auch in stark granitisierten Teilen des Gebirges den tektonischen Bau der prägranitischen Elemente wiederfindet. Das Eindringen des Palingenmagmas ist also scheinbar auf vorgezeichneten Bahnen vorsichgegangen und der Erfolg wäre eine Abbildung der älteren Tektonik. Hierzu sei nun zunächst grundsätzlich bemerkt, dass die Annahme, es handele sich um die Abbildung einer älteren Tektonik, durchaus nicht als bewiesen angesehen werden kann. Vielmehr erscheint es uns berechtigt gerade das Gegenteil anzunehmen. Gewiss sind die palingenen Migmamassen in der Form des Ichors (Sederholm) auf durch ältere Tektonik vorgezeichneten Bahnen, vornehmlich älterer Verschieferung, aufwärts gewandert, aber es steht dem nichts im Wege, wenn wir annehmen, dass auch während des Aufdringen des hochmobilen Palingenmaterials Neubewegungen auf diesen Flächen stattgefunden haben. Hierbei wurden die alten s-Flächen neubelebt und konnten sich als Flächen zwischen mobilem und unmobilem Material erhalten, ihre Lage aber konnte sich bedeutend verändern. Als guter Indikator für die während des Aufdringens des Palingengranits stattgefundenen Bewegungen finden wir nun den Marmor, welcher dank seiner ebenfalls hohen Mobilität alle Faltungsbewegungen lebhaft registrierte. Wir gehen nun zu einigen Beispielen über, die uns auch verschiedene Einzelheiten im tektogenetische Geschehen beleuchten sollen.

Zunächst wenden wir uns nochmals, nach unserer früheren Beschreibung (1945), dem Gebiet von Pargas (SW- Finnland) zu. Die Abb. 1 stellt einen Schnitt durch den östlichen Teil des Gebiets dar. Die Breite des in Zentralperspektive gezeichneten Blocks ist 3 km, die Höhe etwa 2,5 km. Das Charackteristikum des Baustils ist die Synklinale des unteren Hauptamphibolits (2). Die Unterlage derselben ist aus einem stark migmatitisierten Lutogenit aufgebaut, in welchem der Palingengranit stark vorherrscht, namentlich unter dem nördlichen Schenkel. Über dem Hauptamphibolit folgt zunächst ein Marmorhorizont und darüber wiederum Lutogenit, diesmal mit nur untergeordnetem palingenen Anteil. In diesen Lutogenithorizont tauchen von Oben her weitere Amphibolitfalten (3), welche wiederum von Lutogenit überlagert werden. Die Form der Synklinale ist typisch für derartige Bildungen in den finnischen Svekofenniden. Die Schenkel stehen ziemlich steil, aber schon in ca 2 km Tiefe biegt der Amphibolit sich zur Stirn der Falte um. Wir haben anderen Ortes (1945) bereits hervorgehoben, dass die jetzige Form der Synklinale wesentlich das Ergebnis des Aufsteigens der palingenen Massen ist. Ursprünglich lag der Amphibolit, ebenso wie die Lutogenite flacher, aller Wahrscheinlichkeit nach in Deckenfalten alpiden Stils. Durch die Palingenese in der tieferen Unterlage kam es zu Massenverschiebungen in der Teufe. Die palingenen Massen trafen nun bei ihrem Aufstieg auf den Amphibolithorizont. Dieser setzte der Palingenese entschiedenen Widerstand entgegen. Die Synklinale von Pargas, die bereits vorher in flacherer Form vorgelegen haben wird, wurde nun nicht einfach aufgeschmolzen, sondern wurde rein mechanisch durch den aufdringenden Palingengranit verschoben. Die palingenen Massen oder das Ichor stieg den Schenkeln entlang gegen Oben, während die Stirn der Synklinale in der erweichten, mobilisierten Unterlage nach Unten absinken konnte. Dieses bedingte einen seitlichen Zusammenschub der Synkline. Die Schenkel stellten sich steiler, das Ganze nahm den Character einer Tauchfalte an. Unter den Schenkeln bewegte sich der Palingengranit auf durch die ältere Verschieferung vorgezeichneten Bahnen. Hierdurch wurden diese s-Flächen übernommen, aber auch sie veränderten ihre

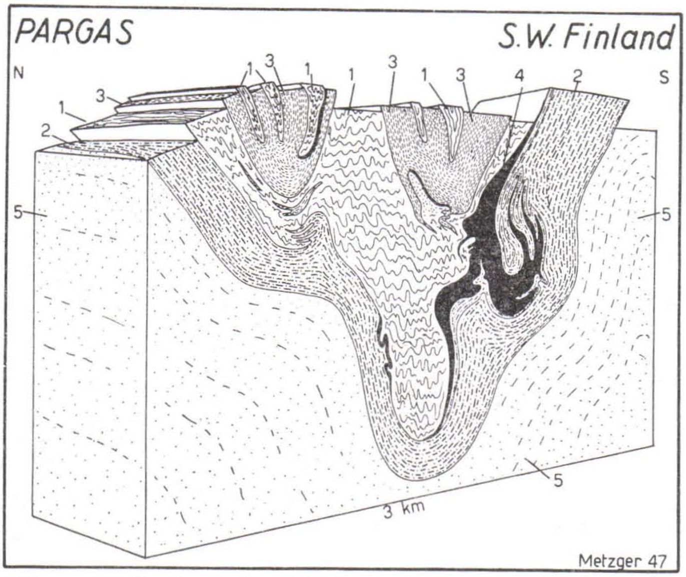

Abb. 1. Schnitt durch das Gebiet von Pargas. 1 = Lutogenit, 2 = Hauptamphibolit, 3 = Oberer Amphibolit, 4 = Kalzitmarmor, 5 = Palingengranit mit migmatitischen Resten von Lutogenit. Orig. Zeichng.

Lage im Raum durch Steilstellung, wobei wohl auch Verschiebungen auf den Flächen selbst stattgefunden haben werden. Deutlich kommen nun diese Verhältnisse im Innern der Synklinale zum Vorschein. Der Lutogenit ist in enge Kleinfalten gelegt, die in ihrer Amplitude oft nur wenige Zentimeter aufweisen. Sie stehen mit ihren Axen der Hauptfaltung parallel. Sie sind aber scheinbar diskordant gegen die Amphibolite, die sich durch starke Verschieferung und teilweise Streckung im Sinne des Haupteinfallens auszeichnen. Nur stellenweise finden wir hier transversale Kleinfaltung. Auch die von oben in das Innere der Synklinale eindringenden Amphibolit- und Lutogenit-Falten sind seitlich zusammengedrückt und zeigen ähnliche Kleinfältelung. Die Amplitude ist recht verschieden in den einzelnen Falten.

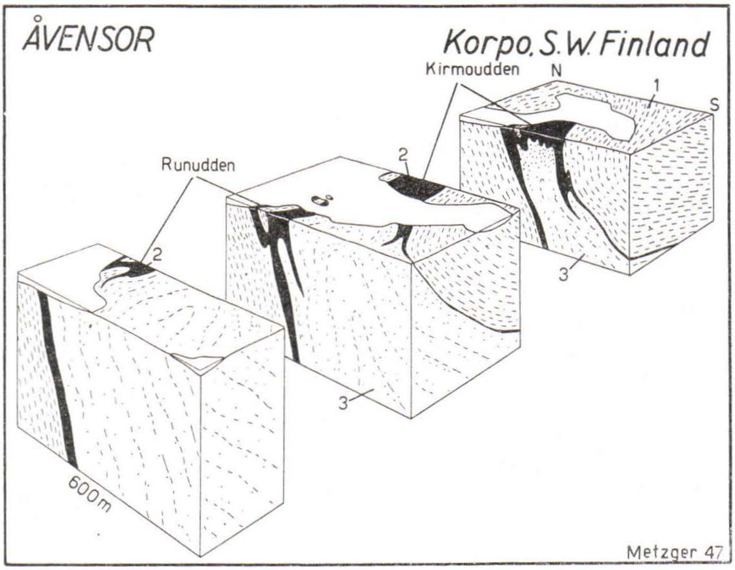

Abb. 2. Stereogramm über das Marmorvorkommen von Åvensor in Korpo, SW-Finnland. 1 = Superkrustalgneis, 2 = Marmor, 3 = Palingengranit und Migmatit. Orig. Zeichng.

Die durch das Empordringen des Palingengranits bedingte Faltung setzte auch den Kalzitmarmor in Bewegung. Er wurde aus dem Nordschenkel (links) fast völlig herausgepresst und sammelte sich unter Schutze der Spezialtauchfalten des Hauptamphibolits im Südschenkel (rechts) zu gewaltigen Massen an, teilweis sich sowohl mit Amphibolit als auch Lutogenit einrollend. Die Bewegung des Marmors überlebte recht lange diejenige des Palingengranits.

Als nächstes Beispiel sei hier die Kalksteinlagerstätte von Åvensor im Kirchspiel Korpo, SW-Finnland vorgebracht. Hier kommt der Marmor (Abb. 2) auf zwei Halbinseln vor, der östlichen Halbinsel von Kirmoudden und der westlichen von Runudden. Die erstere Fundstelle ist von Laitakari (1916) beschrieben worden. Wir hatten Gelegenheit beide Fundpunkte 1938—39 genauer zu untersuchen. Wie uns das hier wiedergegebene Stereogramm, dessen Blockbreite etwa 600 m ist, zeigt, besteht der Gesteinsgrund im wesentlichen aus mehr oder minder migmatitisierten Superkrustalgneisen mit eingelagertem Kalzitmarmor. Der Marmor tritt in der Stirn einer Antiklinalfalte auf, deren Axe gegen E hin (gegen rechts hinten) absinkt. Diese Falte zeigt mehrere Digitationen. Das Vorkommen ist zuerst auf Kirmoudden abgebaut worden. Dies geschah durch Tagebau, der jedoch bald zum Erliegen kam, weil man, infolge völliger Verkennung der Tektonik, gezwungen war in der flachliegenden Faltenstirn mehrere unreine Horizonte mitabzubauen. Später ist man zu einem flachem Pfeilerbau übergegangen. Auf Runudden hat man es nur mit dem einen Schenkel der Falte zu tun. Es ist nun von Interesse, dass in der Stirn der Falte unter dem Marmor eine starke Anreicherung des Palingengranites stattgefunden hat. Der Marmor hat hier offenbar, ähnlich wie der Amphibolit in Pargas, dahin gewirkt, dass der Palingengranit in seiner Aufwärtsbewegung gehindert worden ist. Es ist hier nur umgekehrt wie in Pargas, der Palingengranit hat nicht seitlich ausweichen können, sondern hat sich lediglich als homogene Granitmasse unter dem Kalksteinhorizont ansammeln können. Dass er hierbei aktiv an einer intensiven Kleinfältelung der Faltenstirn innerhalb des Marmors hat mitwirken müssen, erscheint uns als sehr wahrscheinlich. Ob der Palingengranit auch an der Ausbildung der Falte durch Aufpressen aktiv teilgenommen hat, können wir heute nicht entscheiden. Sicher aber ist dies der Fall in anderen Gebieten gewesen. Wir bringen hierzu das Beispiel von kuppelförmigen Axialkulmination von Svartå-Mustio (Abb. 3), die wir hier nach unserer Beschreibung von 1928 mit wesentlichen Änderungen in der Gesteinsbezeichnung nach der Neuuntersuchung von Mikkola (vergl. Eskola 1941) nochmals abbilden. Neubegehungen in dem Gebiet haben uns zu der Überzeugung gebracht, dass der im Zentrum der Kalotte erscheinende Marmor um Bruksträsket in Svartå tatsächlich auf einem Palingengranitstock ruht. Wir kommen daher zu der Auffassung, dass es sich hier um eine diapiritische Ansammlung von Palingengranit in Stirn einer Antiklinale handelt. Dass der Palingengranit hier aktiv an der Ausgestaltung der Axialkulmination teilgenommen hat, ist mehr als wahrscheinlich. Er wird an der Steilstellung der Flanken der Kalotte teilgenommen haben und durch die Diapirbildung auch die räumliche Begrenzung derselben verursacht haben. Ähnliche Bildungen finden sich auch in anderen Teilen der Svekofenniden. So liegen östlich und westlich von Nummi auf der schönen Karte von Parras (1941) zwei derartige Kalotten, von denen die westliche einen schon viel tieferen Schnitt zeigt, der direkt das palingene Diapir angeschürft hat. Im Falle von Svartå ist es wieder von Interesse festzustellen, dass es hier wieder ein Marmorhorizont ist, an dem die Granitisierung haltgemacht hat, wobei die weitere Einwirkung des Palingengranits auf das rein mechanische Empordringen übergegangen ist. Hierbei ist der Marmor ebenfalls in die Höhe getrieben worden und hat sich in der Spitze des Diapirs mächtig angesammelt.

Der Palingengranit kann sich also an hierfür günstigen Stellen ansammeln. Hierbei werden die ungeschmolzenen Reste des E-dukts, welches in erster Linie wohl sandige und sandig-tonige Gesteine bildeten, zurückbleiben, wodurch eine Homogenisierung des palingenen Migmas eintritt. So sind die Gebiete von reinem Palingengranit zu verstehen. Es kommt nun wohl darauf an, wie die Hülle dieser Ansammlungen beschaffen war. Hielt sie dem Druck und der chemischen Einwirkung des palingenen Magmas stand, so verblieben die Granitmassen im gegebenen Horizont, unter Umständen Diapire bildend. Andererseits werden solche Ansammlungen von Palingengranit auch ihre Decken durchbrochen haben und müssen dann als regelrechte Magmagesteine in höhreren Stockwerken erstarren. Hausen (1944) hat neuerdings auf die Möglichkeit hingewiesen, dass die Grobgranite von Nagu derartige aus der Tiefe durchgeschlagene Palingengranite sind. Auch der grosse Körper des Perniögranits, den Eskola (1914) beschrieben hat, kann ähnlich entstanden sein. Uns erscheint es aber auch durchaus möglich, dass die verschiedenen Granitgebiete der III Gruppe

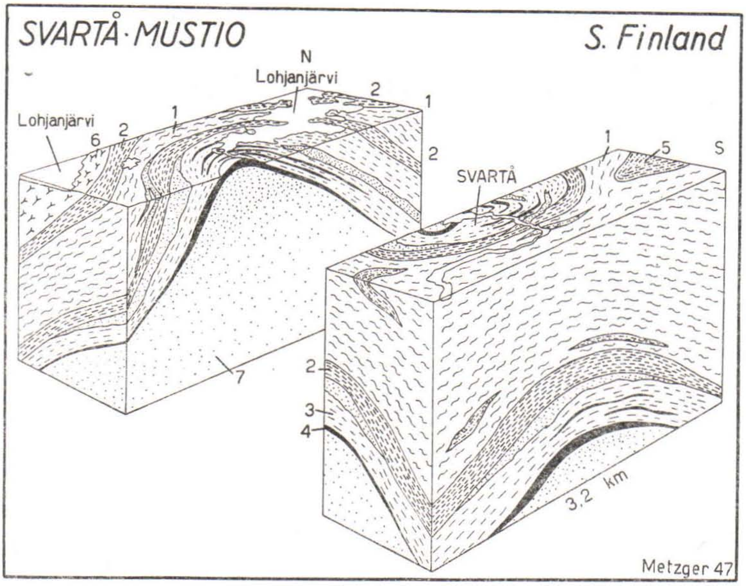

Abb. 3. Stereogramm durch das Gebiet von Svartå-Mustio, nach Metzger 1928 mit wesentlichen Ergänzungen von Mikkola 1941. 1 = Plagioklas-Biotitgneis, 2 = gebänderter Diorit, 3 = Leptitgneis, 4 = Marmor, 5 = Amphibolit, 6 = Granitgneis (primorogen), 7 = Palingengranit mit Migmatit. Orig. Zeichng.

nach Sederholm, die zweifellos postkinematisch sind, als palingene Massen aus tieferen Stockwerken durchgebrochen sind.

Wir haben bisher nur Fälle behandelt, bei denen die palingenen Massen sich nur aufwärts bewegt haben. Es sei hier nun der Fall vorgeführt, dass der Palingengranit auch wesentlich auf die tektonische Gestaltung einwirken kann durch rückläufige abwärts gerichtete Bewegungen. Wegmann und Kranck haben das Gebiet um die Förde von Klockarfjärd in Sibbo östlich Helsingfors beschrieben. Wir hatten in den Jahren 1936—37 dort umfassende Detailuntersuchungen durchzuführen. Unsere Hauptergebnisse sind in dem Block der Abb. 4 wiedergegeben. Für die hier im Folgenden zu nennenden geographischen Bezeichnungen sei der Leser auf die Karten von Wegmann (1931) und Kranck (1937) hingewiesen. Die Gegend ist ein Teil des grossen Bogens um Simsalö-Norrkulla. Eine Serie von gebänderten Gneisen, die Wegmann als die Gneise von Skyttenskär bezeichnet hat, zieht im weiten Bogen gegen N. Diese Gneise bilden das, was wir als die Decke von Skyttenskär bezeichnen wollen (I.). An der Stirn dieser Decke findet sich ein Marmorhorizont, der auch über die ganze oberere Randzone der Decke verfolgt werden kann. Über dieser Decke von Skytten-

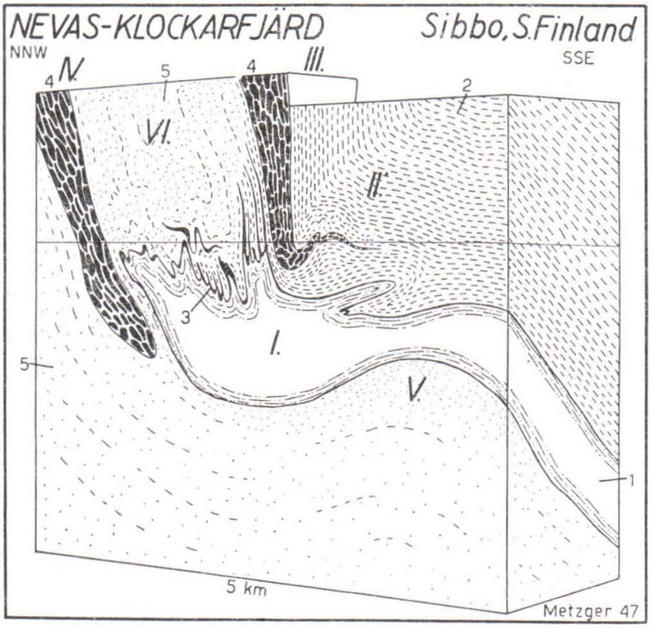

Abb. 4. Block aus der Gegend von Nevas-Klockarfjärd in Sibbo, S-Finnland. 1 = gebänderte Gneise von Skyttenskär, 2 = homogene Lutogenite, 3 = Marmor und Derivate, 4 = Pillow-Lava, 5 = Palingengranit und Migmatit. I. = Decke von Skyttenskär, II. = Decke von Löparö, III. = Tauchfalte von Djupsund, IV. = Tauchfalte von Bergarholm-Sandholm, V. = Antiklinale von Norrkulla, VI. = Migmatittauchfalten von Nevas-Majholmen. Orig. Zeichng.

skär erscheinen die Gneise von Löparö (II.). In diese dringt von Oben her eine Tauchfalte aus Pillow-Lava ein, die an der Stirn stark in die Gneise eingerollt worden ist. Dies ist die Tauchfalte von Djupsund (III.). Eine andere Tauchfalte aus Pillow-Lava erscheint weiter gegen N um die Inseln Bergarholm-Sandholm (IV.). Unter der Decke von Skyttenskär dringt die palingene Granitmasse von Norrkulla (V.) auf. Teils hat sie die Skyttenskärdecke angehoben. Die palingenen Ichors sind dann aber auch weiter nördlich um die Tauchfalte von Bergarholm herum gegen Oben gedrungen und haben hier das gesammte Gebiet migmatitisiert. Vor der Stirn der Skyttenskärdecke drangen die palingegen Massen wieder seitwärts gegen S bis sie auf die Tauchfalte von Djupsund stiessen. Hier war es anscheinend mit ihrer Kraft vorbei. Die hinter der Tauchfalte lagernden Löparö-Gneise sind weit weniger angegriffen. Bei der seitlich gerichteten Bewegung, die wahrscheinlich durch einen Druck aus dem nördlich anschliessenden Migmatitgebiet stark beeinflusst wurde, trat nun eine Rückfaltung und Zusammenschiebung der Stirn der Decke von Skyttenskär ein. Es entstanden eine ganze Reihe von Tauchfalten, zwischen denen der Marmor in langen schmalen und stark gepressten Antiklinalen angesammelt wurde. Auf diese Weise entstanden langestreckte Marmorlager, die nach E hin auskeilen. Ihre innere Tektonik zeigt aber ganz klaren antiklinalen Bau. Es sind dies die Marmorvorkommen von Nevas. Der Marmor ist hier passiv zwischen den durch den Palingengranit aktivierten Tauchfalten aufgeflossen. Seine grösste Ausdehnung ist in der Axenrichtung zu suchen. Wir sehen also, dass auch hier der Palingengranit recht aktiv in die serorogene Faltung eingegriffen hat, wobei wiederum Steilstellung der Elemente verbunden mit geringen Tiefgang ein wesentliches Ergebnis ist.

Diese Beispiele mögen vorerst genügen. Es wird offenbar, dass die serorogene Faltung zum grossen Teil unmittelbare Folge der Mobilität des aufdringenden palingenen Migmas ist. Es hat seitlichen Zusammenschub und damit Steilstellung vieler Gesteinskontakte bewirkt, wohingegen der Tiefgang der ursprünglich flachliegenden Elemente nur wenig vergrössert wurde. Infolgedessen ensteht das für den svekofennidischen Bau so charackteristische Bild flacher Tektonik mit steilen Grenzflächen. Tauchfalten sind ein besonders oft vorkommendes Stilelement. Der Marmor sucht sich ebenso wie der Palingengranit in die durch tektonisches Klaffen entstandenen Räume, wie Antiklinalen, Faltenschatten und Faltenstirnen. Sein Aufdringen ist aber mehr passiv, dauert aber länger an als die Bewegungen des Palingengranits. Der Palingengranit ist hingegen aktiv in der Raumschaffung und bildet gern Diapire, die, wenn in der Stirn gesprengt, zu stockwerksfremden, postkinematischen Granitmassiven anwachsen können.

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17.

ON THE GRADUAL DEVELOPMENT OF THE BASALTIC PARENT MAGMA DURING THE COURSE OF THE GEOLOGICAL PERIODS

BY

AHTI SIMONEN

ABSTRACT

The origin of the sial sphere from subalkalic basalt magma is presented and a hypothesis on the gradual development of the primary subalkalic parent magma into alkalibasaltic is put forward. From this hypothesis many petrological problems are discussed.

The opinions presented by different authors concerning the origin of granitic rocks and sial sphere in general show considerable deviations. In the following the present author will discuss only some points of the matter and set forth an idea on the gradual development of the basaltic parent magma.

Many geologists (Barth, 1939; Rittmann, 1939) have maintained that the basaltic parent magma shows an alkalibasaltic composition. The studies carried out in the oceanic regions underlain by volcanic rocks that are young geologically give support to this opinion. On crystallization the basalts of the oceanic regions and also the continental plateau basalts are not able to produce a residual magma of granitic composition. On the other hand, in the orogenic regions of the Earth the basalts do give a granitic residual magma. This can be understood by assuming that the alkalibasaltic magma has assimilated materials from the acidic sial sphere in the orogenic zones.

It is therefore very difficult to understand how the granitic rocks and the sial sphere in general could be differentiated from the alkalibasaltic parent magma, as this kind of magma does not give a granite as a late product of crystallization. Especially the great amount of free quartz in the sial sphere cannot be explained in that way.

Before the crystallization of the first earth crust all acid material of the present sial sphere was contained in the primary magma and it is probable that the composition of the parent magma was different from that of the present day. According to Daly's (1940) calculations, based upon the values of the thicknesses of the sima and the sial spheres, the sial sphere occupies 10 or 15 per cent of the lithosphere (sima + sial). It is probable that this amount of acid material was enough to give to the primary parent magma a composition from which the crystallization of a granitic residual magma became possible.

According to Barth (1938) very small differences in the basaltic magmas are able to cause, on their crystallization, great differences in the residual liquids. The enclosing of the acid sial sphere in the primary parent magma has most probably caused a composition nearly subalkalic, from which the granitic rocks were then able to originate through magmatic crystallization differentiation. In this way the great amount of free quartz in the sial is also easily understood.

During the earliest geological history of the globe, after the formation of a crystalline earth crust, the juvenile granites played a predominant role as material for the formation of sial. Especially during the orogenic periods the juvenile granitic pore magma of the subalkalic basalt sphere was squeezed upward (Eskola, 1932, 1933).

Some authors give a special significance to exogenous differentiation in the genesis of the granitic rocks. According to Wegmann (1938) the origin of the granites represents a granitization process of sediments. Similar opinions have been expressed by Backlund (1936, 1938) in his discussion of the granitic rocks of Fennoscandia as products of the granitization of sediments. In no one of the theories on the origin of the granites, based exclusively on the granitization or remelting of the sedimentary rocks, has the problem of the origin of these sediments been attacked. In order to explain the origin of such sediments available for granitization, one apparently must presuppose the existence of igneous rocks containing a great amount of granitic material.

According to the discussion presented above, the juvenile granites have formed the primary material of the sial and the amount of this juvenile material has increased until the primary subalkalic magma received an alkalibasaltic composition. It seems probable that the processes of differential anatexis and granitization began immediately after the origin of the first juvenile granites. In the orogenic movements palingenic pore magma can originate from the sediments formed by weathering of the primary granites. In the march of the geological periods differential anatexis and granitization probably play an increasingly predominant role, at the same time as the calc-alkalic basaltic magma is substituted by alkalibasaltic magma.

Also, when the basaltic parent magma has received an alkalibasaltic composition, there can originate under the sial sphere a subalkalic basaltic magma through the assimilation of the acid sial sphere in the orogenic regions. This basaltic magma gives, through the processes of magmatic crystallization and squeezing out, granites of a purely magmatic mode of occurrence.

During the remelting of the sial sphere there probably originates a dioritic or granodioritic magma, as in many orogenic zones magmatic granodiorites are predominant among the synkinematic intrusions. The great distribution of granodiorites in the Svecofennian territory has been emphasized especially by Wahl (1936). Kennedy and Anderson (1938) have presented a theory as to the origin of the plutonic rocks from a granodioritic parent magma. According to their opinion this magma has originated through the remelting of the sial sphere. Furthermore, according to van Bemmelen (1939), there forms under the sial a migmatic zone from which the magmas penetrating the earth's crust originate. It seems to the author, that in the origin of the plutonic rocks of the sial, magmatic crystallization from the magma, originated by pure remelting, has played a predominant role in the orogenic zones.

The above views concerning a possible gradual development of the parent magma have been based in many points on uncertain facts. The ideas presented, however, offer new possibilities for the explanation of many important petrological problems which so far have been very difficult to understand. In the following the writer will discuss briefly only a few more interesting problems.

Many geologists have pointed out the sharp contrast between the eruptivity of the early Pre-Cambrian and post-Cambrian times. Especially Daly (1933) has laid stress on the great eruptive activity during the Pre-Cambrian, represented for the most part by granitic rocks.

From the above explanation concerning the magmatic development of the globe this contrast is easily understood. During Pre-Cambrian times the composition of the parent magma was different from that of the later eras. Probably the sial was very thin and the greatest part of the present sial existed as material of the parent magma of subalkalic composition.

In a similar way the richness of the acid leptites in the oldest Svecofennian formation of Fennoscandia is easily explained.

The lack of dunitic rocks in the Svecofennian formation is also a characteristic feature. Probably in early Archaean times the composition of the parent magma was not basic enough for the crystallization of the dunites. However, already in the Karelidic orogeny there occur olivine rocks, showing that the composition of the parent magma had become more basic, caused by the enormous intrusions of the acid rocks into the sial during the early Archaean orogeny.

One of the plutonic rocks occurring mainly in the Pre-Cambrian formation, especially in late Pre-Cambrian, is according to Daly (1933) the anorthosite. During Pre-Cambrian times, especially in the late Pre-Cambrian, the composition of the parent magma must have been favourable for the crystallization of the anorthosites. The opinions as to the origin of the anorthositic rocks are varying, but it seems, however, that their crystallization has occurred from a nearly basaltic (subalkalic) magma.

In some cases dunites and anorthosites occur together in the same formation, showing that the crystallization of these rocks has occurred from the same kind of basaltic magma. According to our hypothesis of gradual development of the basaltic parent magma, the composition of the basaltic magma was subalkalic during the earliest geological periods, making possible a greater occurrence of anorthosites. When the basaltic parent magma became alkalibasaltic in composition, anorthositic rocks could originate only in the orogenic zones of the globe.

One rock group occurring only in the younger geological formations is according to Daly (1933) the group of alkaline rocks. Furthermore, Daly remarks that the distribution of the alkaline rocks in the earth's crust increases from older to newer times. The alkaline rocks are known already in the Paleozoic formations, but Daly notes that the greatest part of the alkaline rocks originated during the Cenozoic era. Supposing that the opinions of many authors on the origin of the alkaline rocks as crystallization products of alkalibasaltic magma are correct, the distribution of these in different geological formations is well understood from our hypothesis. During the earliest geological periods no alkaline rocks could originate, as the subalkalic basaltic parent magma could give rise to granitic residual liquids only. It seems to the present author that the first appearance of the alkaline rocks marks that point in the gradual development of the parent magma when it assumed an alkalibasaltic composition. This occurred probably in the Paleozoic era. The increase in the distribution of the alkaline rocks in the youngest geological formations shows that the composition of the parent magma continuously becomes more clearly alkalic.

It is a remarkable fact that the alkaline rocks occur in connection with great fault zones. We may say, in terms of the view we are following, that the alkaline rocks have originated from the residual liquids of the alkalibasaltic magma which have penetrated upwards into the earth's crust along the fault zones. The lack of alkaline rocks in the orogenic regions is, then, a consequence of the lack of alkalibasaltic magma.

In Fennoscandia the alkaline rocks are post-Jotnian. In the Jotnian formations there occurs also one rock which is met with only in this formation. This is the rapakivi granite, representing extremely anorogenic intrusions. Probably the rapakivi granites represent the composition of the residual liquids of the basaltic parent magma during the Jotnian period in a similar way as the alkaline rocks in the younger formations?

Daly (1933) points out the strong diversification of types among the diaschistic dikes after the Pre-Cambrian periods. In the Pre-Cambrian there occur predominantly aplites and pegmatites, while in younger formations a great number of lamprophyric rocks are also found. The difference in the dike rocks in the Pre-Cambrian and in the younger formations also points to the gradual magmatic development of the globe. In Pre-Cambrian times the basaltic parent magma was able to produce mainly granitic residual liquids, and aplitic and pegmatitic dikes, but in later periods there occurs a great diversity of types among the diaschistic dikes.

Several authors have earlier expressed ideas of a gradual development of the inorganic processes in the earth's history. Thus Lindgren (1933) has pointed out that in the western United States there is taking place — beginning in the Mesozoic and continuing up to the present day — a gradual magmatic development giving rise to the following rock groups: diorite, granodiorite, monzonite, quartzmonzonite and alkaline rocks. Diorite is the oldest member in this magmatic development and the alkaline rocks are the youngest. According to Lindgren the ore and mineral deposits stand in close connection with the magmatic development. The reason for the gradual magmatic development has been, according to Lindgren, the movement of the magmas eastward in the depths of the earth. The opinions presented by Lindgren give support to the gradual alteration of the basaltic parent magma during the course of the geological periods.

Rankama (1946) has set forth a theory of the geochemical differentiation of the globe. According to this theory so-called granitophile elements have become enriched in the granites in the course of geological times, thus showing that in the development of the globe the processes exhibit a tendency in a definite direction. The hypothesis of gradual development of the basaltic parent magma during the course of the geological times is in good harmony with the theory of geochemical differentiation. In orogenic periods granitic rocks always originate, but with the decrease of geological age differential anatexis and remelting become more significant with respect to the birth of granites, in which there then occurs an enrichment of the granitophile elements.

The discussion of the petrological problems presented above shows that the hypothesis of gradual change of the basaltic parent magma in the course of the geological periods is able to elucidate many geological phenomena. Unfortunately, the knowledge concerning the distribution of the different rock groups among the different geological formations is very deficient, and it is thus very difficult to follow step by step the alteration of the parent magma.

Especially the lack of alkaline rocks in the Pre-Cambrian formations shows that the composition of the parent magma has been subalkalic. When has the parent magma received an alkalibasaltic composition? This question is difficult to answer. Perhaps during the Paleozoic Era after the Kaledonidic orogeny? It seems most probable that the change has not occurred simultaneously in all regions. The hypothesis of the gradua change of the basaltic parent magma in the course of the geological periods receives its explanation from the growth of the sial sphere during the orogenic movements. In connection with the orogenic processes the granitic material of the primary parent magma has penetrated upwards from the depths into the earth crust, causing the alteration of the parent magma. The conditions on the earth's surface have been, however, throughout all geological periods very similar to those of the present day.

GEOLOGICAL SURVEY OF FINLAND, 1947.

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18.

ÜBER EINE ANTIMONREICHE PARAGENESE IN YLÖJÄRVI, SW-FINNLAND

VON

MARTTI SAKSELA

INHALT

EINLEITUNG

Das sog. Schiefergebiet von Tampere ist seit 1935 in lagerstättenkundlicher Hinsicht sehr interessant geworden. Das kleine Eisenerzvorkommen von Haveri im Kirchspiel Viljakkala, das schon im 18. Jahrhundert entdeckt worden war, erwies sich bei erneuerten Untersuchungen als so goldreich, das hier Bergbau schon einige Jahre betrieben worden ist. Hier wurde auch die erste reichere — obschon örtlich sehr begrenzte — Kobaltglanzkonzentration in Finnland angetroffen. Dieses Erzgebiet hat Stigzelius neulich ausführlich beschrieben (25).

Im Jahre 1937 sandte ein im Kirchspiel Ylöjärvi wohnender Bauer an die staatliche Geologische Forschungsanstalt ein Stück von reichem Pyriterz. Besonders wurde die Aufmerksamkeit des Verfassers darauf gerichtet, dass der Pyrit in Serizitschiefer vorkam. Offenbar stammte das Erzstück von genetisch gleichartigen Vorkommen her wie die Pyriterzvorkommen von Otravaara und Karhunsaari (Nordkarelien), die vom Verfasser früher untersucht worden waren (17, 18, 19). Bei dem Besuch an Ort und Stelle konnte festgestellt werden, dass das Probestück aus einem grösseren Moränenblock in der Nähe des Landgutes Järvenpää im Dorfe Takamaa (Ylöjärvi) losgebrochen war. In der nächsten Umgebung der Fundstelle wurden zwei neue pyritreiche Blöcke gefunden. Weiter fand man im anstehenden Fels Serizitschiefer, der von Pyrit imprägniert war. — Diese Observationen veranlassten weitere Untersuchungen vorzunehmen. Bei diesen ging es hervor, dass der in der Nähe von Järvenpää beobachtete Serizitschiefer ein ziemlich schmales, etwa 3 km langes einheitliches Gebiet bildet. Südlicher wurde noch eine andere Serizitschieferzone angetroffen (Ahdepääagebiet). Grössere Pyritanhäufungen wurden nicht entdeckt, trotzdem man neue und ziemlich reiche Pyritblöcke fand. — Das haupsächlichste Erzmineral in Serizitschiefer war also Pyrit. Nur in geringen Mengen kam Magnetkies vor.

Die Untersuchungen ergaben doch anderartige und ganz unerwartete Resultate. Ung. 5 km südlich von Järvenpää wurde, ausser pyrithaltigen Serizitschieferblöcken, mehrere Moränenblöcke entdeckt, die Kupferkies und Arsenkies enthielten. Der Verfasser konnte sofort feststellen, dass hier ein Erztypus vorlag, der bis jetzt nicht in Finnland bekannt war. Die Blöcke bestanden nämlich aus einer Breccie, worin ein basisches, vulkanogenes, dunkelgrünes Gestein die Bruchstücke bildet. Die Zwischenmasse war quarz- und turmalinreich. Schon in den ersten Blöcken konnte analytisch Edelmetallgehalte festgestellt werden.

Die grosse Zahl der Blöcke deutete darauf hin, dass die Abstammungskluft ziemlich nahe liegen musste. Noch im selben Sommer konnte man in der Gegend des kleinen Sees Paroisten Kaitajärvi (Paroinen), etwa 1800 m nordwestlich der zuerst gefundenen Blöcke, ein verhältnismässig kleines Gebiet begrenzen, worin sich das Anstehende aller Wahrscheinlichkeit nach versteckte. Im folgenden Sommer wurde das Erz schliesslich entblösst. Es liegt, wie die Diamantbohrungen gezeigt haben, teilweise unter dem obengenannten See. Seit 1943 hat man hier Abbau betrieben. — Es sei noch erwähnt, dass man in Paroinen-Erz später Wolfram in Form von Scheelit gefunden hat.

Im Sommer 1940 wurde die Aufmerksamkeit des Verfassers aufs Neue auf das Serizitschiefergebiet von Järvenpää gerichtet, denn nach meiner Meinung lagen triftige Gründe vor, den eventuellen Goldgehalt eingehender zu untersuchen. Die Untersuchungen führten zu einem positiven Resultat, wenngleich abbauwürdige Golderze bis jetzt nicht angetroffen worden sind. Etwa 300 m nach SO vom Gehöft Järvenpää fand man im Serizitschiefer quarzreiche Partien und im Zusammenhang mit denselben konnte vom Verfasser Mineralarten festgestellt werden, die man früher im Järvenpääagebiete nicht beobachtet hatte. Ausser Pyrit fand man namentlich Zinkblende, Bleiglanz, Kupferkies, Magnetkies, Arsenkies und ein stahlgraues, weiches Mineral, das makroskopisch nicht genauer bestimmt werden konnte. Es wurde vorläufig »Fahlerz» genannt. — Derartige Partien im Serizitschiefer erwiesen sich als bedeutend edelmetallhaltig. Den Analysen gemäss, die man aus den gesammelten Handstücken gemacht hat, können sich die Gehalte an Gold und Silber sogar auf den Betrag von 142, resp. 1373 g pro Tonne belaufen. Besonders gern schienen die Edelmetalle die Gesellschaft des »Fahlerzes» zu suchen.

In der Serizitschieferzone von Järvenpää konnte man später gleichartige Bildungen an mehreren Stellen finden. Auch im Ahdepäägebiete wurde im Serizitschiefer ungewöhnliche Minerale angetroffen, und zwar solche, die in erster Linie von hohen Gehalten an Wismut und Tellur charakterisiert sind.

Im folgenden Herbst fing der Verfasser an, das in Järvenpää- und Ahdepäägebieten zusammengebrachte Material erzmikroskopisch zu untersuchen. Es erwies sich bald, dass das »Fahlerz» nicht aus einer Mineralart bestand; wohl enthielt es Fahlerz, aber dazu noch eine ganze Reihe von anderen Mineralen, für welche besonders der Gehalt an Antimon charakteristisch war. Am reichlichsten kam irgendein Repräsentant der Boulangerit-Jamesonit-Falkmanitreihe vor.

Noch im selben Herbst hatte der Verfasser Gelegenheit, unter gütiger Anleitung seines Freundes, Dr. phil. Olof Ödman, in Stockholm weitere erzmikroskopische Untersuchungen über dieses Material vorzunehmen und es mit dem ziemlich analogen Material aus Boliden und anderen Gruben des Skelleftefeldes zu vergleichen.

Während des Krieges hat der Verfasser die Untersuchungen nur zeitweise fortsetzen können, weshalb es erst jetzt möglich gewesen ist, die Untersuchungsresultate an die Öffentlichkeit zu bringen. In der letzten Phase meiner Arbeit, bei den Kontrollbestimmungen der selteneren Erzminerale, ist Stud. J. Salemaa mir behilflich gewesen. Die spektrographischen Analysen sind von Mag. phil. O. Joensuu ausgeführt. Stud. V. Veltheim hat die Abbildungen gezeichnet und ist auch bei den photographischen Aufnahmen behilflich gewesen.

DIE SERIZITSCHIEFERFORMATION

von Järvenpää liegt am nördlichen Rande des Schiefergebiets von Tampere (siehe die Karten, Fig. 1 und 2), wo dieses an das grosse zentralfinnische Granitgebiet grenzt. In der Umgegend werden saure und basische Vulkanite samt sedimentogene Schiefer angetroffen. Die Untersuchungen haben gezeigt, dass der Serizitschiefer metasomatische Umwandlungsprodukte dieser Gesteine darstellt. Es ist kaum nötig, die Umwandlungsprozesse näher zu beschreiben. Gleichartige Bildungen sind vom Verfasser früher aus Otravaara- und Karhunsaarigebieten (17, 18, 19), sowie von Ödman

(7, 13), Sven Gavelin (5, 6) und Grip (7) aus dem Skelleftefelde geschildert worden. In vielen anderen Erzgebieten sind derartige Umwandlungen vorsichgegangen.

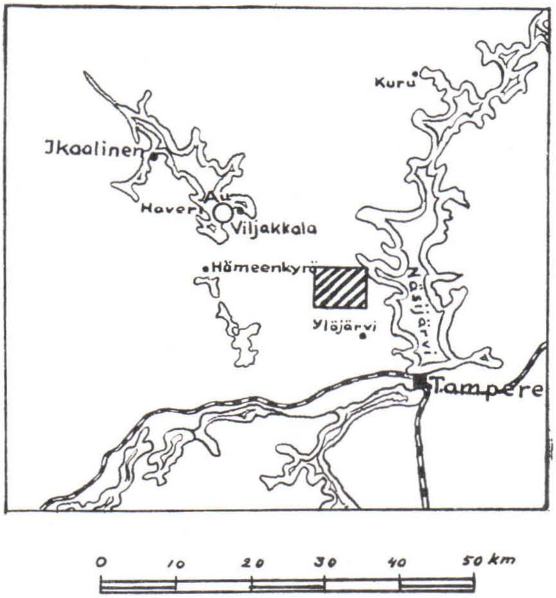

Fig. 1. Die geographische Lage des Gebietes in Fig. 2 (das linierte Rechteck).

Man kann verschiedene Umwandlungsstufen beobachten, die oft bandartig nebeneinander liegen. Das Endprodukt der Umwandlungsprozesse scheint ein grauer oder graugelber Schiefer zu sein, der hauptsächlich aus feinschuppigem Serizit und Quarz besteht. Es handelt sich vor allem um eine kräftige Kalium-Kieselsäure-Metasomatose (siehe auch die von Simonen und Neuvonen angeführten Analysen N:o 3, 4 und 5, S. 250 in diesem Bande (24). Der Serizitschiefer enthält hie und da ziemlich reichlich violetten Fluorit, der oft in haarfeine Streifen in der Richtung der Schieferung geordnet ist. Offenbar ist der Fluorit ein Produkt einer Fluor-Metasomatose und in den Reaktionen zwischen den von aussen hinzugekommenen fluorhaltigen Lösungen und den ursprünglichen kalziumreichen Mineralen entstanden.

Ein bedeutender Teil des Serizitschiefers ist von Pyrit imprägniert. Die Pyritkörner sind 0,1—0,5 mm im Durchmesser und ziemlich deutlich idiomorph. Kataklastische Strukturzüge fehlen. Der Pyrit ist beweislich Resultat einer Reaktion zwischen eisenhaltigen Silikaten der metasomatisch nicht veränderten Gesteine und von aussen hinzugekommenen Schwefelverbindungen. Das Gestein hat wahrscheinlich den Schwefel als eisenfreie Verbindungen, etwa in der Form von Schwefelwasserstoff, empfangen, denn wie aus den obengenannten Analysen von Simonen und Neuvonen, sowie aus den von Veltheim ausgeführten Bestimmungen (26) ersichtlich ist, ist bei der im Järvenpääagebiete erfolgten Metasomatose der Eisengehalt der Gesteine nicht grösser geworden.

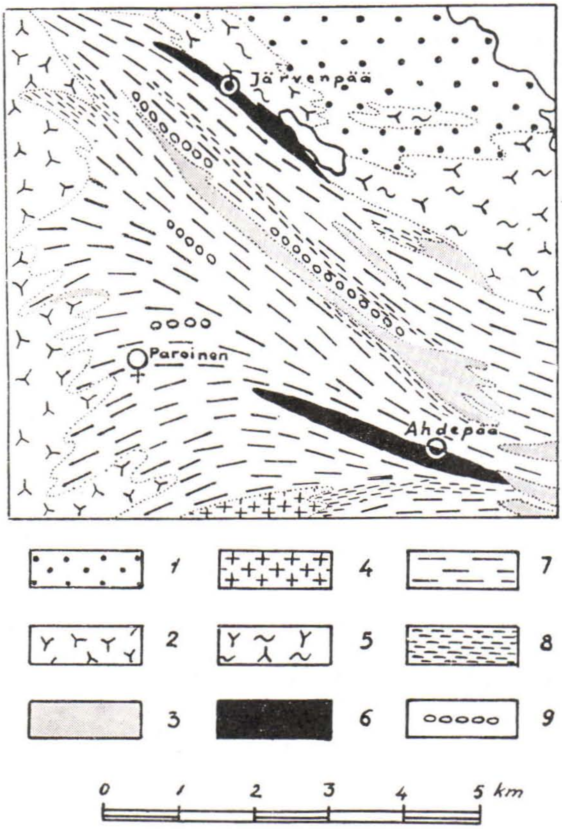

Fig. 2. Geologische Übersichtskarte des nördlichen Teils des Kirchspiels Ylöjärvi. Nach A. Simonen. 1. Mikroklingranit; 2. Granodiorit; 3. Glimmerschiefer und Phyllite; 4. Diorit; 5. Gneisartiger Granodiorit, teilweise migmatitisch; 6 Serizitschiefer; 7. Basische Vulkanite; 8. Saure Vulkanite; 9. Agglomerate und vulkanische Konglomerate.

Die als Resultat einer metasomatischen Verdrängung (replacement) entstandene Kiesimprägnation, die, wie schon erwähnt, zum kleinen Teil auch aus Magnetkies besteht, vertritt die erste Phase in der Bildung von Erzmineralen im Järvenpääagebiete (Pyrit I, Magnetkies I).

DIE GÄNGE VON KOMPAKTEM PYRIT IN SERIZITSCHIEFER

vertreten die folgende Phase. Diese Gänge folgen, soweit man hat beobachten können, genau dem Streichen des Serizitschiefers. Die Kontakte gegen das Nebengestein sind scharf. Die Gänge sind gewöhnlich schmal, etwa 1 cm breit. Die grösseren, 30—40 cm im Diameter messenden Blöcke von kompaktem Pyriterz, die man im Untersuchungsgebiete gefunden hat, zeigen jedoch, dass innerhalb der Serizitschieferzone grössere gleichartige Pyritkonzentrationen existieren müssen.

Der Pyrit ist grobkörnig (sogar 0,5 cm im Durchmesser) und die Körner sind oft kataklastisch zerbrochen. Wenn Pyrit an andere Erzminerale und Quarz grenzt, zeigt er grosse Neigung zu Idiomorphie. In reiner Pyritmasse hemmen sich die einzelnen Pyritkörner gegenseitig in ihrem Kristallwachstum und es entsteht deshalb eine pflasterartige Struktur, wo die Idiomorphie der Körner nicht hervortreten kann.

Ausser Pyrit kann man in den Gängen sehr wenig Kupferkies, Zinkblende und Magnetkies beobachten. Die zwei erstgenannten kommen gern in den schmalen Rissen und Spalten der Pyritkörner vor. Es entsteht hierbei ein Bild, das einer Verdrängungsstruktur gleicht. Wahrscheinlich handelt es sich nicht nur um Verdrängung, sondern Kupferkies und Zinkblende sind, um Schneiderhöhns Worte zu benutzen, zwischen die Pyritbruchstücke hineingeschmiert worden (22).

Es ist natürlich, dass dieses Kieserz sich nicht auf dieselbe Weise als der als Imprägnation im Nebengestein auftretende Kies hat bilden können, sondern hat sich als solches aus der Lösung ausscheiden müssen (vergl. 17, 19). Aller Wahrscheinlichkeit nach vertritt die Erzablagerung eine unmittelbar nach der erwähnten Schwefel-Metasomatose vorsichgegangene Phase der Erzmineralbildung. — Auf dieselbe Weise als Sven Gavelin (6) und Ödman (13) den Entstehungsmechanismus der Kieserze im Skelleftefelde gedeutet haben, möchte der Verfasser annehmen, dass die Pyritgänge im Järvenpäägebiete durch »displacement» (Verlagerung) gebildet worden sind. Ödman gibt darüber folgende Definition (13): ». . . the ores were brought into place by the intrusion of the ore solution into a more or less schisted rock complex in which the solution pressed the walls of suitable channelways apart thanks to its intrusive force.»

In der Altersreihe der Erzminerale des Järvenpäägebietes könnte man die Minerale der Kieserzgänge Pyrit II, Magnetkies II, Zinkblende I und Kupferkies I nennen.

DIE ANTIMONHALTIGEN QUARZPARTIEN IN SERIZITSCHIEFER

JÄRVENPÄÄGEBIET

Im Serizitschiefer wird hie und da Quarzpartien entweder als Gänge, grössere unregelmässig geformte Klumpen oder als kleine, 3—5 mm im Diameter messende rundliche Tröpfen (Fig. 3, 4 und 5) angetroffen. Bald

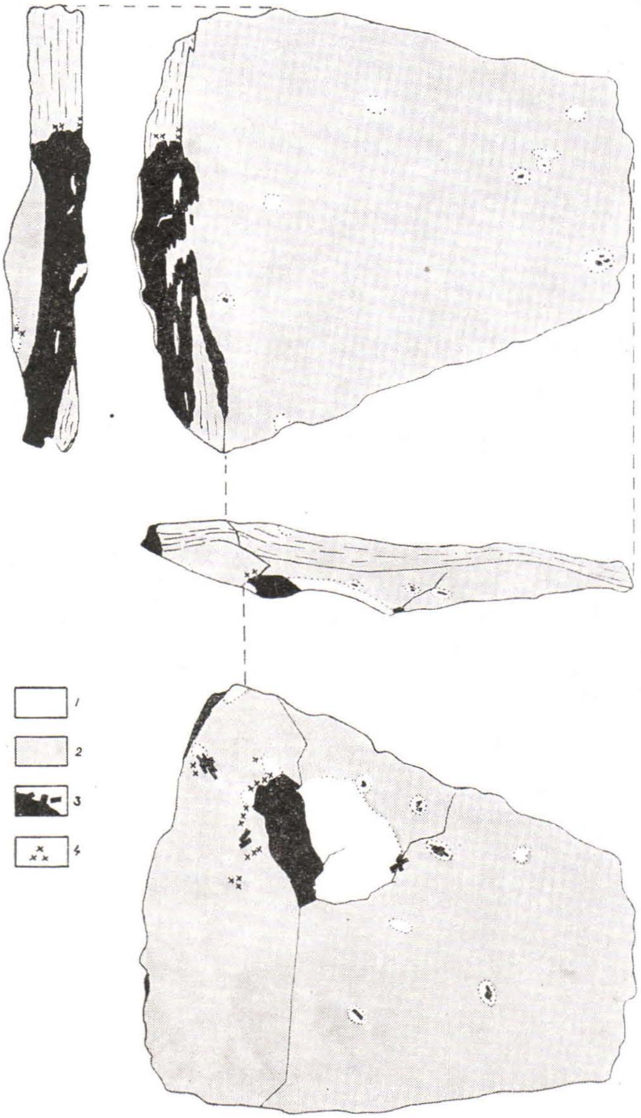

Fig. 3. Anhäufungen von Sb-haltigen Mineralen in Serizitschiefer. 1. Quarz; 2. Serizitschiefer; 3. Anhäufungen, reich an Sb-haltigen Mineralen; 4.Kupferkies. — Die schraffierten Flächen schneiden die Schieferung des Serizitschiefers. Ein Handstück von verschiedenen Seiten abgebildet. C:a ⅔ der nat. Grösse.

folgen die gangförmigen Partien dem Streichen des Schiefers, bald aber schneiden die Schieferung schroff ab, den Querspalten im Schiefer folgend. Oft sind diese Spalten nur teilweise von Quarz gefüllt. Besonders in denjenigen Quarzgängen, die parallel mit der Schieferung laufen, beobachtet man Fragmente von Serizitschiefer (Fig. 4 und 5). Oft sind die Kontakte derselben gegen die Quarzmasse sehr unscharf, was darauf beruht, dass die Schieferfragmente kräftig silifiziert sind. Bisweilen sind von ihnen nur »schattenförmige» Reste übrig. Auch das Nebengestein der Quarzpartien ist oft silifiziert. Mikroskopisch beobachtet man, das die Quarzpartien aus isometrischen Quarzkörnern aufgebaut sind, die nach Veltheim (26) keine einheitliche Gitterregelung aufweisen. Eine richtungslos massige Verwachsung ist vorherrschend. — Die anderen Gangarten fehlen, wenn man die hie und da vorkommenden kleinen Serizitschuppen, die Reste von silifizierten Nebengesteinsfragmenten darstellen, unbeachtet lässt.

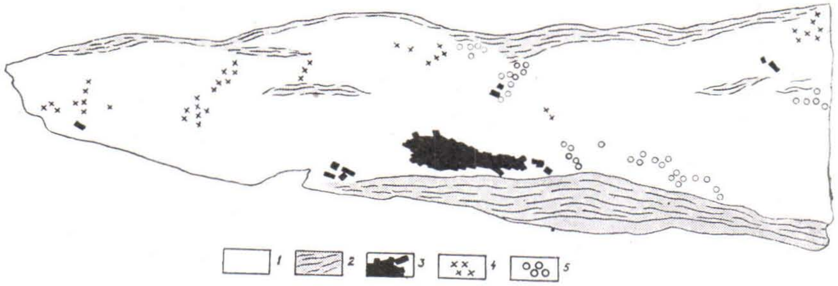

Fig. 4. Anhäufungen von Erzmineralen in einem Quarzgang in Serizitschiefer. 1. Quarz; 2. Serizitschiefer, teilweise silifiziert; 3. Partien, reich an Boulangerit und anderen „Sulfomineralen"; 4. Kupferkies; 5. Magnetkies. Ein Handstück, c:a $^2/_3$ der nat. Grösse.

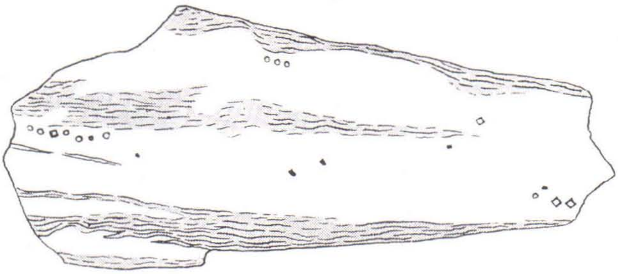

Fig. 5. Wie in Fig. 4. Die kleinen Vierecke sind Pyrit. Ein Handstück, ¾ det nat. Grösse.

In den obengeschilderten Quarzpartien sind mehrere gewöhnliche und seltene Erzminerale enthalten. Diese sind als kleine Einzelkörner in Quarz eingesprengt oder bilden sie grössere unregelmässig geformte Aggregate in der Quarzmasse (Fig. 4 und 5). Bisweilen kann eine streifige Anordnung der verschiedenen Erzminerale beobachtet werden (Fig. 6). Überhaupt kommen die Erzminerale im Verhältnis zum Quarz sehr spärlich vor. Die ältesten bilden die allgemeinbekannte Paragenese: Arsenkies, Pyrit, Magnetkies, Zinkblende und Kupferkies. An diese schliessen sich auch der Gudmundit und der Ullmannit an.

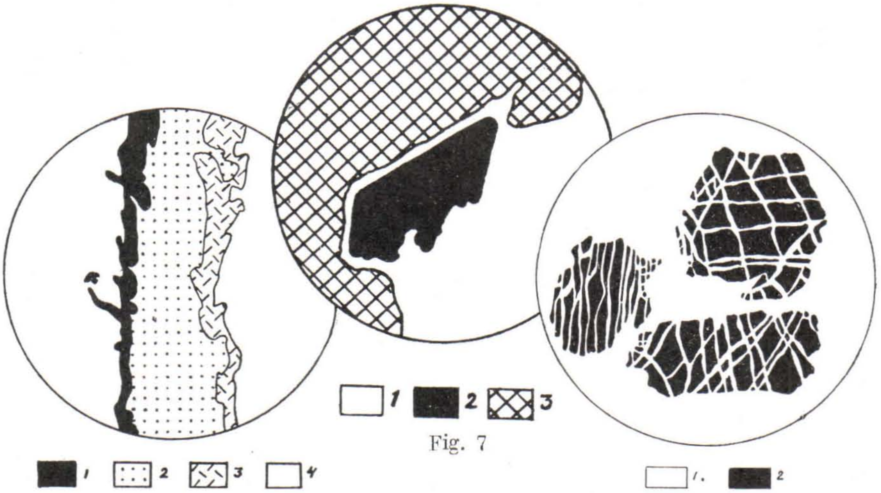

Fig. 7

Fig. 6

Fig. 8

Fig. 6. Streifige Anordnung der Erzminerale in Quarz. 1. Magnetkies; 2. Kupferkies; 3. Boulangerit und andere „Sulfominerale"; 4. Quarz. Vergr. 1.7 ×. Fig. 7. Boulangerit; 2. Arsenkies; 3. Pyrit. Vergr. 16 ×. Fig. 8. Zerbrochene Pyritkristalle in Quarz. 1. Quarz; 2. Pyrit. Ung. von nat. Grösse. (Nach Zeichnungen von Salemaa).

Arsenkies kommt bald von Quarz umgeben getrennt von den anderen Erzmineralen, bald zusammen mit diesen vor. Besonders im vorigen Falle bildet er schön idiomorphe Kristallindividuen, die bisweilen doch kataklastisch zerbrochen sein können. Im letzteren Falle sind die Arsenkieskörner allgemein und kräftig von den Nachbarmineralen, besonders von Fahlerz, Bleiglanz und Boulangerit korrodiert und verdrängt worden, sogar in dem Masse, dass von den ursprünglichen Körnern nur kleine Fetzen übrig sind (Fig. 2, Taf. II). Arsenkies kann jedoch auch in diesem Falle idiomorphe Kristalle bilden, auch wenn er an Pyrit grenzt oder innerhalb des letztgenannten vorkommt (Fig. 7). Arsenkies ist das einzige Mineral, das Kobalt enthält. Nach den Bestimmungen von Salemaa ist der Kobaltgehalt 0,9 %. Man dürfte also noch nicht von Danait sprechen können.

Pyrit gleicht in seinem Auftreten dem Arsenkies. Die Pyritkörner sind oft tektonisch beansprucht und von Sprüngen durchzogen (Fig. 8). Die Korrosions- und Verdrängungserscheinungen sind kräftig entwickelt (Fig. 2, Taf. II). Die einzelnen Körner können ziemlich gross, sogar 2—3 cm im Durchmesser sein. Die spektrographische Analyse hat gezeigt, dass der Pyrit ganz frei von Arsen und Kobalt ist. Der Gehalt an Nickel ist von der Grössenordnung 0,01 %.

In der Altersreihe von Erzmineralen haben wir hier mit Pyrit III zu tun.

Magnetkies bildet sehr oft gut entwickelte idiomorphe Kristallindividuen, besonders wenn er an Kupferkies oder Boulangerit grenzt (Fig. 2, Taf. I). Dieses ist ja geeignet in gewissem Masse Aufsehen zu erregen denn, wie bekannt, derartige Kristalle von Magnetkies sind selten. Wie Schneiderhöhn (20), Ramdohr (14) und Frebold (3) geschildert haben, sind die Kristalle tafelig || (0001). Im Järvenpäägebiete zeigen die Magnetkieskristalle im Anschliff einen langsäuligen Habitus. Offenbar dürften auch in diesem Falle die längsten in den Anschliffen sichtbaren Flächen für Basisflächen und die kleineren Endflächen für Pyramidenflächen gehalten werden müssen.

Der Magnetkies verdrängt kräftig den Pyrit und ist seinerseits von jüngeren Mineralen, speziell von Kupferkies, verdrängt und korrodiert worden. Der spektrographischen Analyse gemäss ist der Nickelgehalt des Magnetkieses von der Grössenordnung 0,1 %.

Eine sehr interessante Erscheinung ist die Umwandlung des Magnetkieses in eine Substanz, die der Verfasser hier *Wasserkies* genannt hat. Diese Substanz ist nicht homogen, denn schon die erzmikroskopische Untersuchung zeigt, dass sie zum Teil aus einem schwarzen Material, das sich gar nicht polieren lässt, zum Teil aus einem blassgelben Material besteht, das ein ziemlich starkes Reflexionsvermögen besitzt (Fig. 9). Die letztgenannte Komponente ist stark anisotrop. Die Farbeneffekte bei + Nicols sind jedoch nicht so stark wie bei Markasit, an welchen die Substanz vielleicht einigermassen erinnert, sondern viel schwächer: hellgrau — dunkelgrau. Auch ohne Nicols ist die Farbe nicht so deutlich gelb als bei Markasit und Pyrit (vrgl. Ödman, 12, S. 594). — Es entsteht oft ellipsoidisch-augenförmige Bildungen, worin die beiden mikroskopisch sichtbaren Komponenten konzentrisch angeordnet sind (Fig. 9). Das es sich wirk-

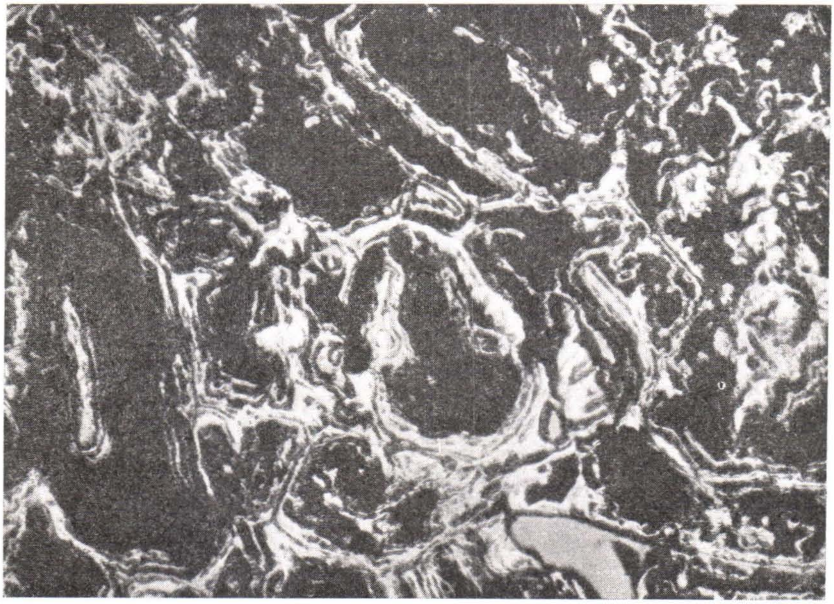

Fig. 9. Wasserkies. Vergr. 205 ×.

lich um eine Umwandlung von Magnetkies handelt, geht sowohl aus der Tatsache, dass man mitten im Wasserkies kleine Partien von unumgewandeltem Magnetkies finden kann, als besonders daraus hervor, dass die idiomorphen Magnetkiestafeln unter Aufbewahrung ihrer ursprünglichen Kristallform in Wasserkies umgewandelt sind (Fig. 10); wir haben sehr schöne Pseudomorphosen von Wasserkies nach Magnetkies. — Der spektrographischen Analyse gemäss ist der Gehalt an Ni in Wasserkies niedriger als in Magnetkies und zwar von der Grössenordnung 0,03 %.

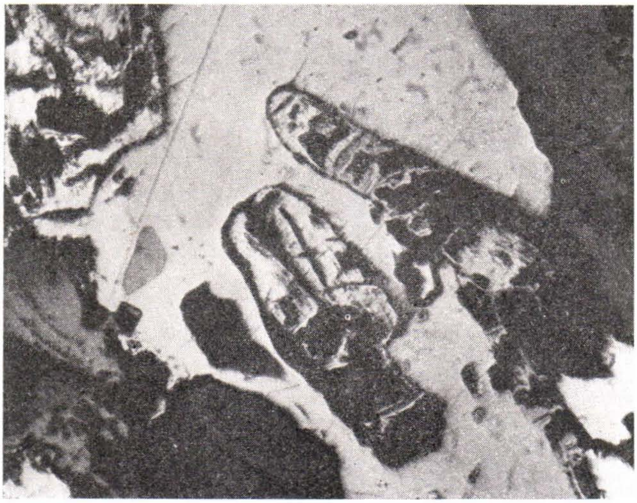

Fig. 10. Pseudomorphosen von Wasserkies nach Magnetkies in Kupferkies (hellgrau). Vergr. 205 ×.

Erzmikroskopisch ist ähnliche Substanz als der obenbeschriebene Wasserkies an mehreren Stellen in Finnland beobachtet worden. Er kommt z. B. im Kieserz von Otravaara vor (vrgl. Saksela, 19, Fig. 6, S. 41). Lokka hat über dieses Material interessante chemische Untersuchungen ausgeführt (9). Er hat gezeigt, dass hierbei ein wasserhaltiges Gemenge von verschiedenen Schwefelungsstufen des Eisens und Schwefel vorliegt. Der Wassergehalt beträgt 7,95 % und das Verhältnis S :Fe = 2,27.

Gudmundit bildet idiomorphe, im Anschliff rhombenförmige, 1—2 mm im Diameter messende Körner (Fig. 11). Er vermengt sich nicht gern mit den anderen Erzmineralen, sondern kommt entweder ganz von Quarz umgeben ausserhalb der von den letztgenannten gebildeten Anhäufungen, oder an den Rändern derselben vor. Unter dem Erzmikroskop zeigt Gudmundit sehr ähnliche Eigenschaften wie Arsenkies. Die einzige mehr deutliche Verschiedenheit findet man in der Schleifhärte: Gudmundit ist weicher und besitzt darum eine grössere Polierfähigkeit. Reflexionspleochroismus ist ein wenig stärker als bei Arsenkies. — Es gelang aus einem Anschliff unter gleichzeitiger Beobachtung unter dem Mikroskop etwa 2 mg ziemlich reines Gudmunditmaterial herauszubohren, welches dann spektralanalytisch untersucht wurde. Dabei gab Sb eine kräftige Reaktion, Co und As fehlten ganz.

Gemäss den obengenannten Bestimmungen dürfte der Gudmundit mit genügend grosser Genauigkeit identifiziert worden sein. Die Vertreter der Safflorit-Rammelsbergitreihe, wie auch Löllingit und Arsenkies, mit welchen der Gudmundit verwechselt werden könnte, kommen nicht mehr in Frage.

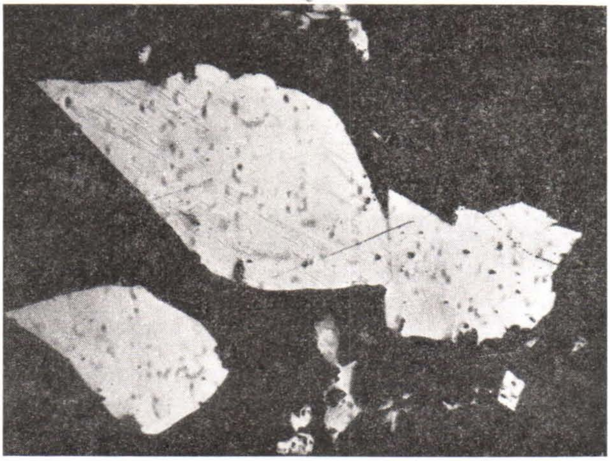

Fig. 11. Gudmunditkristalle (hellgrau) in Quarz.
Vergr. 80 ×.

Gudmundit ist nicht früher in Finnland gefunden worden. Sonst ist er eine ziemlich neue Mineralart, die zuerst von K. Johansson in Gudmundstorp b. Sala in Amphibolskarn entdeckt wurde (8). Buerger stellte fest (1), dass er strukturell dem Arsenkies sehr ähnlich ist. Später ist Gudmundit u. a. von Sven Gavelin (5 und 6) und Ödman (13) im Skelleftefelde, von Ramdohr aus der Grube Jakobsbakken bei Sulitelma (15) samt von Maucher aus der Grube Bayerland bei Waldsassen (11) und aus Turhal (Türkei, 10) gefunden worden.

Aus den Verdrängungsverhältnissen kann man schliessen, dass Gudmundit im Järvenpäägebiet früher als Kupferkies und Zinkblende kristallisiert ist. Die Verhältnisse zu Arsenkies, Pyrit und Magnetkies sind unsicherer.

Zinkblende II kommt in den Quarzgängen in ziemlich reichlichen Mengen vor. Sie ist völlig allotriomorph. Charakteristisch sind die kleinen tropfenförmigen oder mehr unregelmässig geformten Entmischungskörper von Kupferkies, die bisweilen regelmässig in der Zinkblende eigelagert sind. Zinkblende verdrängt sowohl Arsenkies als Pyrit und ist seinerseits von Kupferkies und Boulangerit verdrängt. — Der spektrographischen Analyse gemäss ist der Indiumgehalt der Zinkblende von der Grössenordnung 0,03 %. Ein derartiger Indiumgehalt dürfte nicht sehr ungewöhnlich für finnische Zinkblenden sein. Nach Erämetsä (2) enthalten die Zinkblenden von Paavola (Lohja), Träskböle (Perniö), Otravaara und Fiskars ung. gleich viel, die Zinkblenden von Pitkäranta und Lågnäs (Inkoo) ansehnlich mehr Indium.

Kupferkies II kommt in den quarzreichen Partien des Serizitschiefers in Verhältnis zu den den anderen Erzmineralen ziemlich reichlich vor. Er ist deutlich später kristallisiert als Arsenkies, Magnetkies, Pyrit und Gudmundit und ist von Boulangerit kräftig verdrängt worden.

Ullmannit ist eine Mineralart genannt worden, die erzmikroskopisch folgende Eigenschaften aufweist. Sie ist völlig isotrop, und der Reflexionspleochroismus fehlt gänzlich. Die Schleifhärte ist bedeutend höher als die des Kupferkieses, aber ein wenig niedriger als die des Magnetkieses. Sie poliert sich ziemlich gut und besitzt ein ansehnliches Reflexionsvermögen. Die Farbe ist weiss, mit schwachem Stich ins Rosa. — Ätzverhalten: mit konz. HNO_3 schnell ein schwarzer oder schwarzbrauner Beschlag. Verd. HNO_3 wirkt schwächer und langsamer. Nach einer Behandlung von 2 bis 3 Minuten bildet sich eine farbenreiche, irisierende Ätzfläche. Mit $HgCl_2$ und $FeCl_3$ tritt wenigstens dann und wann eine schwache irisierende Bräunung ein. HCl, KOH und KCN sind negativ.

Ullmannit kommt als deutlich idiomorphe Körner oder Korngruppen in Boulangerit vor (Fig. 1, Taf. II). An den Rändern der Ullmannitkörner, die im Anschliff oft als regelmässige Dreiecke erscheinen, findet man ziemlich allgemein Kupferkies.

Durch Bohren konnte aus einem Anschliff 1 mg Pulver losgemacht werden, das reichlich Ullmannit enthielt. Es wurde danach spektralanalytisch untersucht. Dabei erhielt man eine sehr kräftige Reaktion von Ni. Ebenso enthielt das Pulver Sb, Cu und Pb. — Co, Zn, Bi und Te fehlten gänzlich. Man dürfte gut annehmen können, dass Cu in Kupferkies samt Pb und ein Teil des Sb in Boulangerit enthalten sind. Unsere Mineralart wäre also in erster Linie von Ni und Sb charakterisiert. Wenn man noch diejenigen Resultate in Betracht zieht, die man bei Ätzversuchen an Vergleichsmaterial erhalten hat, dürfte mit ziemlich grosser Gewissheit behauptet können, dass hier wirklich Ullmannit vorliegt.

Ullmannit ist von Boulangerit und offenbar auch von Kupferkies verdrängt. Es sieht so aus, als ob der Ullmannit schon vor Ablauf der Kristallisation des Kupferkieses hätte zu kristallisieren begonnen.

Nach den obengeschilderten Mineralen ist eine Reihe von Mineralen gebildet worden, von welchen viele Sb-haltig sind. Es ist nicht möglich gewesen, die Kristallisationsfolge in allen Einzelheiten klarzulegen. In grossen Zügen dürfte sie folgende sein: Gold und Goldtellurid, Tellurwismut, Wismut, Bournonit, Boulangerit, Fahlerz und Bleiglanz.

Gold ist in der Regel im engsten Zusammenhang mit einem graublauen, schwach pleochroitischen, anisotropen Mineral angetroffen. Dieses ist ein wenig weicher als Gold und Boulangerit und stellt aller Wahrscheinlichkeit nach irgendeinen *Goldtellurid* dar. Diese zwei Minerale bilden oft myrmekitische Verwachsungen mit einander (Fig. 3, Taf. II); es handelt sich deutlich um Entmischung. Gold ist auch als schmale Gänge in Arsenkies gefunden worden (Fig. 4, Taf. I). Auch in diesem Falle kommt er im Zusammenhang mit Goldtellurid vor, und zwar auf die Weise, dass die beiden Minerale, wie aus derselben Fig. ersichtlich, im Gange nacheinander geordnet sind. In denselben Arsenkieskörnern findet man auch schmale Gänge, worin Kupferkies und Zinkblende auf dieselbe Weise nacheinander geordnet sind. — In den Spalten und Rissen der anderen Erzminerale ist Gold nicht angetroffen worden. Man hätte natürlich erwartet, Gold auch in den im Pyrit so allgemein vorkommenden Bruchspalten zu finden, denn, wie Schneiderhöhn bemerkt (21), dürfte die Ausfällungsfähigkeit beider Sulfide, Arsenkies und Pyrit, für Gold in grossem Durchschnitt wohl gleich sein. Es kann ja möglich sein, dass das zur Verfügung gestandene Material derart begrenzt gewesen ist, dass alle Erscheinungsweisen des Goldes nicht haben hervortreten können, aber ebenso möglich ist es, dass Gold im Järvenpäägebiete als Gänge in Pyrit gar nicht vorkommt. Schneiderhöhn bemerkt ja auch *(loc. cit.)*: »Im Einzelgang oder Einzelbezirk kann das eine oder andere auch einmal bevorzugt sein». — Es sei erwähnt, dass der Verfasser neulich im Erze des Nachbargebietes Haveri mikroskopisch Gold in unerwartet grossen Aggregaten gefunden hat. Das Gold kommt hier u. a. als schmale Gänge in Arsenkies vor. In Pyrit ist Gold bis jetzt nicht angetroffen worden.

Das mikroskopisch sichtbare Gold im Järvenpäägebiete ist tiefgelb; sein Gehalt an Silber ist offenbar sehr niedrig.

In denselben Anschliffen, wo die zwei obengeschilderten Minerale beobachtet worden sind, hat man auch cremgelblichen *Tellurwismut* gefunden. Er bildet oft längliche, scharfkantige Einschlüsse in Boulangerit (Fig. 1, Taf. I). Die Schleifhärte ist eine Ahnung höher als die des letztgenannten, aber niedriger als die des Bleiglanzes. Tellurwismut ist in reichlicheren Mengen im Ahdepäägebiete angetroffen und wird später näher besprochen.

Die Stellung der Telluride in der Altersreihe der Erzminerale ist im Järvenpäägebiete schwer genau zu fixieren. Wahrscheinlich sind sie früher als die im folgenden zu beschreibenden Sb-Minerale kristallisiert. Darauf deuten ja u. a. die erwähnten scharfkantigen Tellurwismuteinschlüsse hin.

Wismut hat man als grosse Seltenheit als kleine hellweisse und stark reflektierende tropfenförmige Einschlüsse in Boulangerit beobachtet. Die

Frage, ob die Wismuttröpfen ein Entmischungsprodukt darstellen (vrgl. z. B Ödman, 13, S. 111) oder auf irgendeine andere Weise entstanden sind, muss noch offen bleiben.

Bournonit kommt hauptsächlich in Grenzzonen zwischen Kupferkies und Boulangerit in derartiger Weise vor, dass man an seine Bildung als Produkt der Reaktionen zwischen boulangerithaltigen Lösungen und dem älteren Kupferkies nicht zweifeln kann. Wie aus der Fig. 3, Taf. I ersichtlich, hat sich ein typischer Reaktionssaum zwischen Boulangerit und Kupferkies gebildet (vrgl. auch Fig. 12). Als einzelne kleine Körner kann man Bournonit auch etwas weiter von den erwähnten Grenzlinien entfernt finden.

Bournonit unterscheidet sich ziemlich deutlich von seinen gewöhnlichen Nachbarmineralen. Der Farbton ist dunkler, die Anisotropieeffekte ansehnlich schwächer und die Schleifhärte niedriger als bei Boulangerit. Spaltbarkeit im Anschliff ist nicht beobachtet. Zwillingsbildung ist meistens in einer oder anderer Form zu sehen, obwohl sie bisweilen schwächer und nur in einer Richtung entwickelt ist.

Boulangerit bildet den grössten Teil des grauen und weichen Materials, das anfangs bei der Feldarbeit »Fahlerz» genannt wurde. Er kann verhältnismässig grosse, 1—2 cm im Diameter messende Kristallindividuen bilden. Diese sind oft säulig || c. Eine deutliche Spaltbarkeit in der Längsrichtung ist sowohl makroskopisch als bisweilen auch mikroskopisch zu beobachten (Fig. 3, Taf. I). Boulangerit ist ziemlich deutlich reflexionspleochroitisch. Zwillingsbildung ist nicht beobachtet.

Wie bekannt, sind die drei Bleispiessglanzerze, Boulangerit, Jamesonit und Falkmanit, unter dem Erzmikroskop einander sehr ähnlich. Auch was die chemische Zusammensetzung, das Ätzverhalten und die Paragenese anbelangt, gleichen sie sehr an einander (16). Für jeden einzelnen Fall ist eine sichere Identifizierung auch dem Kundigen unmöglich. — Durch Bohren unter dem Mikroskop wurde aus solch einen Stellen im Anschliffe, wo Boulangerit möglichst reine und grosse Stengel bildete, Pulver losgemacht. Eine röntgenographische Phasenanalyse wurde dann von Herrn A. Danielsson im Zentrallaboratorium der Grubenaktiengesellschaft Boliden in Stockholm ausgeführt. Der Analyse gemäss enthielt das Pulver Boulangerit. Offenbar besteht ja wenigstens der grösste Teil des Bleispiessglanzerzes in Järvenpää aus Boulangerit. In einigen Anschliffen kommt doch im Zusammenhang mit Boulangerit ein Mineral vor, auf welches die von Ramdohr für den Jamesonit festgestellten Eigenschaften einigermassen besser zu passen scheinen (16).

Bleiglanz und *Fahlerz* kommen hauptsächlich in myrmekitischer Verwachsung mit einander vor (Fig. 2, Taf. II), was offenbar auf gleichzeitige Kristallisation der beiden Minerale hindeutet. Teilweise bilden diese auch selbständige Körner und Kornaggregate. Das Fahlerz ist von schwach olivgrüner Farbe und völlig isotrop; es scheint sich um das gemeinste Fahlerz (CuSb-Fahlerz) zu handeln. — Die beiden Minerale verdrängen kräftig sowohl Pyrit als Arsenkies (Fig. 2, Taf. II). Die Verhältnisse zu Kupferkies und Boulangerit sind undeutlicher. In dem von Bleiglanz und Fahlerz gebildeten Myrmekit sieht man allgemein kleinere und grössere Kupferkiespartien, die vielleicht für Resorptionsreste gehalten könnten. — Wahrscheinlich sind Fahlerz und Bleiglanz nach der Bildung von Boulangerit und Bournonit kristallisiert. Man könnte eine Kristallisationsfolge Bournonit, Boulangerit → Fahlerz → Fahlerz + Bleiglanz → Bleiglanz denken (vrgl. Sven Gavelin, 5, S. 15).

Eigentliche *Silberminerale* sind in den ausgeführten Untersuchungen nicht gefunden worden, trotzdem der Silbergehalt in den analysierten Probestücken auf 1373 g pro Tonne ansteigen kann. Den spektrographischen Analysen gemäss ist der Silbergehalt einiger Minerale von folgender Grössenordnung[1] : in Arsenkies 0,03 %, in Tellurwismut 0,02 %, in Kupferkies 0,01 % in Gudmundit und Ullmannit wenigstens 0,1 %.

PARTIELLE UMLAGERUNG DES ERZMATERIALS

Eine interessante Erscheinung mag noch erwähnt werden. Auf den Spaltrissen des Boulangerits kann man stellenweise schmale, geradlinige »Gangstückchen» von Magnetkies, Kupferkies und mehr selten von Zinkblende beobachten (Fig. 3, Taf. I). Bisweilen findet man auf einem und demselben Risse alle drei Minerale nacheinander. Längs den Rändern der Kupferkies- und auch der Magnetkiesgängchen, auf beiden Seiten derselben, sieht man oft einen schmalen Saum von Bournonit (Fig. 13). Stellenweise bildet der letztgenannte auch »selbständige» Gangstücke, worin oft kleine Kupferkieskörner sitzen.

Wie soll man derartige Bildungen erklären? Es ist ja in gewissem Masse überraschend, dass z. B. Magnetkies, der, wie aus der obigen Schilderung hervorgehen dürfte, diejenigen Mineralbestandteile der Quarzgänge vertritt, die in ziemlich früher Phase auskristallisiert sind, jetzt wieder in Boulangerit, der verhältnismässig spät kristallisiert ist, hineindringt. Eine Erklärung wäre, dass nach der Kristallisation der »Sulfominerale» von aussen eine neue Portion von in erster Linie Cu- und Fe-reicher Erzsubstanz hinzugekommen ist. Diese hat sich dann leicht in den Spaltrissen des Boulangerit, die sich in den andauernden tektonischen Bewegungen geöffnet haben, absetzen können. Gleichzeitig sind Reaktionen unter Bildung von Reaktionssäumen von Bournonit stattgefunden. — Wahrschein-

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[1] Selbstverständlich ist es nicht möglich gewesen, reines Material durch Bohren unter dem Mikroskop zu erhalten. Man kann also in diesem Zusammenhang nur von kupferkies*reichem*, arsenkies*reichem* usw. Material sprechen

lich handelt es sich jedoch um eine Umlagerung des alten Materials, welche Umlagerung nicht nur durch plastische Verformung, sondern auch durch Lösungsumsatz erfolgte (Reaktionssäume!).

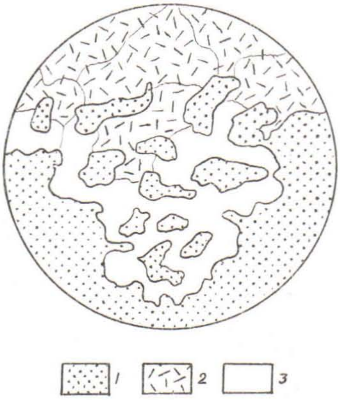

Fig. 12. Detailbild vom Reaktionssaum zwischen Kupferkies und Boulangerit. 1. Kupferkies; 2. Boulangerit; 3. Bournonit. Vergr. 230 ×.

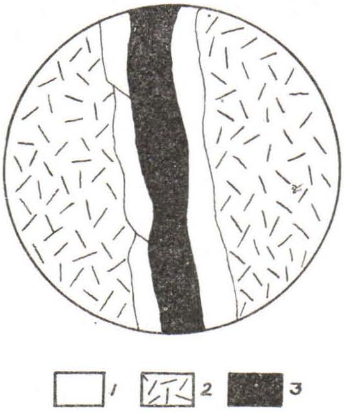

Fig. 13. Magnetkiesgang in Boulangerit. Zwischen diesen ein Reaktionssaum von Bournonit. 1. Bournonit; 2. Boulangerit; 3. Magnetkies. Vergr. 252 ×.

(Nach Zeichnungen von Salemaa).

Im

AHDEPÄÄGEBIET

wurde an ein paar Stellen in fluorithaltigem Serizitschiefer, in der nächsten Nähe von gangartigen Quarzanhäufungen, in verhältnismässig reichlichen Mengen ein graues, grafitähnliches Mineral angetroffen, das sich später als Tellurwismut erwies. Der Serizitschiefer, der an diesen Stellen quarzreich ist, enthält ausserdem Zinkblende, Kupferkies und Pyrit. Der erstgenannte kommt als gleichmässige Imprägnation vor.

Tellurwismut bildet sehr dünne, in der Richtung der Schieferung ausgezogene Kristallaggregate. Er poliert sich gut, Kratzer lassen sich allerdings schwer ganz vermeiden. Reflexionsvermögen ist hoch, Reflexionspleochroismus fehlt. Die Farbe ist weiss, mit deutlichem Stich ins Cremgelb. Die Anisotropieeffekte bei + N sind stark; die Farben wechseln von bräunlich grau bis bläulich grau. Nach Frondell (4) kann man den Tellurwismut von dem ihm sehr ähnlichen Tetradymit (und Joseit) durch Ätzen mit KOH unterscheiden. Nach ein paar Minuten tritt bei Tellurwismut, wie auch in unserem Falle, eine schwache Bräunung ein. Der Schwefelgehalt ist, gemäss einer Analyse von Salemaa, 1,2 %. Das anlysierte Material war doch nicht rein; Tellurwismut enthält Einschlüsse von Kupferkies, Pyrit und Zinkblende, die nicht beseitigt werden konnten. Der Bi-

Gehalt desselben Materials ist 44,8 %, also, wie erwartet, etwas niedriger, als die Formel Bi_2Te_3 voraussetzt. Der Tellurwismut von Ahdepää ist nach spektralanalytischen Bestimmungen Sb-haltig.

Fig. 14 stellt einen grösseren Kupferkieseinschluss in Tellurwismut dar Jener enthält seinerseits Einschlüsse von Pyrit, der von Zinkblende verdrängt ist. Die Kristallisationsfolge scheint in diesem Falle klar zu sein, nämlich: Pyrit, Zinkblende, Kupferkies, Tellurwismut. Wie oben (S. 212) angenommen wurde, sind die Telluride wahrscheinlich früher als z. B. Boulangerit kristallisiert. Wenn man also die Beobachtungen sowohl aus Järvenpää- als Ahdepäägebieten kombiniert, dürfte man mit ziemlicher Sicherheit annehmen können, das die Telluride in der Altersreihe zwischen Kupferkies und Boulangerit liegen.

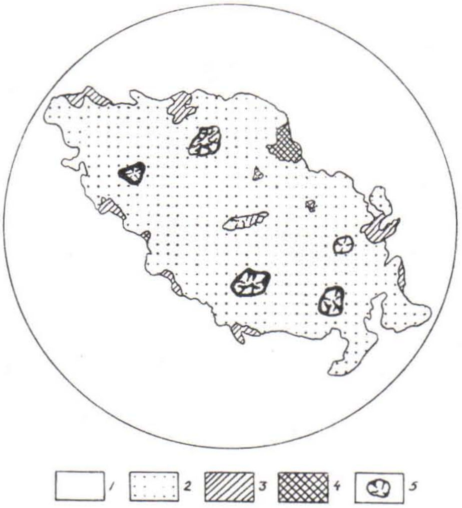

Fig. 14. Kupferkieseinschluss in Tellurwismut. 1. Tellurwismut; 2. Kupferkies; 3. Kupferglanz?; 4. Covellin; 5. Pyritkörner, die von Zinkblende verdrägt sind. Vergr. 87 ×. (Nach Zeichnung von Salemaa).

An den Rändern der Kupferkieskörner hat man stellenweise ein bläulich graues, anisotropes Mineral beobachtet, das wohl am ehesten *Kupferglanz* ist. Ung. an derselben Stelle ist auch *Covellin* gefunden worden (Fig. 14). Offenbar sind diese Minerale durch Einwirkung der deszendenten Lösungen zementativ auf Kupferkies entstanden.

Als Zusammenfassung mag folgendes Verzeichnis, woraus die verschiedenen Phasen der Mineralisation chronologisch hervorgehen, vorgelegt:

I. *Die metasomatische Bildung der Eisenkiesimprägnation in Serizitschiefer*

Pyrit I, Magnetkies I

II. *Die Bildung der Gänge von kompaktem Kieserz in Serizitschiefer*

Pyrit II
Magnetkies II
Zinkblende I, Kupferkies I

III. *Die Bildung der antimonhaltigen Quarzpartien in Serizitschiefer*

a. Die Hauptphase:

Arsenkies
Pyrit III, Gudmundit
Magnetkies III
Zinkblende II
Kupferkies II, Ullmannit
Gold und Goldtelluride, Tellurwismut, Wismut
Boulangerit, Jamesonit, Bournonit I, CuSb-Fahlerz, Bleiglanz

b. Die Umlagerungsphase:

Magnetkies IV, Kupferkies III, Zinkblende III, Bournonit II

IV. *Durch Einwirkung von deszendenten Lösungen verursachte Umwandlungen*

Kupferglanz?, Covellin
Wasserkies

DIE ENTSTEHUNGSVERHÄLTNISSE DER ANTIMONREICHEN PARAGENESE

Es dürfte kein Zweifel darüber bestehen, dass die Sb-haltigen Quarzpartien sich eng an die anderen Erzbildungen des Järvenpääegebietes anschliessen und mit diesen ein Ganzes bilden. Diese Auffassung findet weitere Unterstützung in dem Umstand, dass in neuerer Zeit ähnliche Sb-reiche Paragenesen im Zusammenhang mit genetisch offenbar sehr gleichartigen Kieserzen angetroffen worden sind. Es handelt sich um jüngere Gangfüllungen, mit mehr oder weniger Quarz als Gangart. Ramdohr bemerkt (15), dass eine Antimonvererzung dieser Art »ein mindestens recht verbreiteter Nachhall der Hauptvererzung in Kieserzlagerstätten ist». Er setzt fort: »Damit ist natürlich ein wichtiger Hinweis—natürlich noch lange kein Beweis—gegeben, dass auch die Hauptvererzung in diesen Fällen auf gleichartige oder sehr ähnliche und zwar hydrothermale Vorgänge zurückgeht.»

Wie schon klargelegt, ist die Bildung des Serizitschiefers mit Kiesimprägnationen sowie die der Pyritgänge durch hydrothermale Lösungen verursacht. Aller Wahrscheinlichkeit nach sind auch die Sb-haltigen Quarzpartien als hydrothermale Erzbildungen anzusehen. Aus welchen Magmen die erzhaltigen Lösungen emaniert sind, hat man nicht mit Sicherheit klarlegen können. Der Gedanke liegt aber nahe, dass sie von dem nördlich der Serizitschieferformation belegenen, synorogenen Gneisgranit herstammen.

Die Quarzgänge sind in ihrem Auftreten »weniger syntektonisch» als die Pyritgänge; sie füllen ja stellenweise Spalten, die schroff über die Schieferung schneiden. Offenbar sind sie auch jünger als die Pyritgänge. Die kräftige Silifizierung der Serizitschieferfragmente in den Quarzgängen und des Nebengesteins zeigt, dass die Bildung dieser Gänge ein späteres Ereignis ist als die Metasomatose, der der Serizitschiefer seine Entstehung verdankt.

In der orogenetischen Entwicklung des Gebietes gehört die Erzbildung im ganzen zu derselben orogenen Phase, den Endpunkt einer bestimmten synorogenen Intrusion darstellend.

Was die Bildungstemperatur der Sb-haltigen Quarzgänge betrifft, gehören sie hauptsächlich zur katathermalen Zone. Dafür spricht das verhältnismässig reichliche Vorkommen der hochthermalen Arsen- und Magnetkiese und die kräftige Silifizierung, die die Gangmasse verursacht hat. Die Temperatur ist jedoch während der Erzbildung gesunken, und die Kristallisation der letzten Bestandteile ist wahrscheinlich erst in der mesothermalen Zone eingetroffen. — Die Umlagerung des Erzmaterials könnte vielleicht auf eine zufällige Steigerung der Temperatur hindeuten.

Als selbständige Bildungen behandelt, kann man die antimon- und zugleich goldhaltigen Quarzanhäufungen zu derjenigen Gruppe der hydrothermalen Erzvorkommen rechnen, die Schneiderhöhn »hypoabyssische katathermale Goldquarzgänge» nennt und deren charakteristischen Züge er in seinem Lehrbuch (21, S. 318—324) geschildert hat. Derartige Gänge sind nach Schneiderhöhn »für die präkambrischen Schilde . . . . . charakteristisch, wie überhaupt für alle orogenetischen Zonen der Erdrinde, in denen grössere synorogene und postorogene dioritisch-granitische Intrusivkörper liegen.» Nach den sulfidischen Begleitmineralen sind die Goldquarzgänge von Schneiderhöhn folgendermassen gruppiert: 1. nur mit Pyrit; 2. mit Pyrit und Arsenkies; 3. mit Pyrit, Kupferkies, Zinkblende und Bleiglanz; 4. ebenso wie 3, aber noch mit Arsenkies; 5. mit Arsenkies und Antimonglanz; 6. mit Antimonglanz; 7. mit Tellurerzen; 8. mit Wismuterzen. — Über Magnetkies wird erwähnt, dass er »in durchgängigen und grösseren Mengen als Kennzeichen für pneumatolytische Gangteile angesprochen wird und in den eigentlich hydrothermalen Gängen zurücktritt». Neben Antimonglanz werden oft andere Sb-Minerale, wie Jamesonit, Boulangerit, Bournonit und Fahlerz, angetroffen.

Die Sb-Au-haltigen Quarzgänge des Järvenpääagebietes passen als solche in keine von obenerwähnten Untergruppen der Goldquarzgänge. Wir könnten aber die Altersfolge der Erzminerale ohne Schwierigkeit in Teilsukzessionen einteilen, von welchen dann jede ziemlich gut eine bestimmte von den Untergruppen Schneiderhöhns entspricht. Die Teilsukzession Arsenkies-Pyrit-Magnetkies-Zinkblende-Kupferkies entspricht die Untergruppe 4 (anstatt Bleiglanz haben wir also Magnetkies), die Telluride die Untergruppe 7 und die Serie Boulangerit-Bournonit-Fahlerz-Bleiglanz die Untergruppe 6.

Wie schon hervorgegangen, schliesst sich das mikroskopisch sichtbare Gold sehr eng an die Telluride und zwar besonders an die Goldtelluride an.

Im Ahdepäägebiete, wo die letztgenannten praktisch genommen fehlen und Tellurwismut das dominierende Erzmineral ist, ist der Goldgehalt anspruchslos. Nur in einem von den in diesem Gebiete genommenen Probestücken stieg der Goldgehalt bis 4,8 g pro Tonne. Gewöhnlich waren sie goldfrei.

Die Erzformationen von Järvenpää und Ahdepää schliessen sich an

DIE METALLOGENETISCHE PROVINZ DES SCHIEFERGEBIETES VON TAMPERE

an, zu welcher schon zwei grössere, abbauwürdige Erzkonzentrationen gehören. Die höchsten Bildungstemperaturen sind von dem Breccienerz von Paroinen vertreten. Die Beschaffenheit des Bindemittels zeigt deutlich, dass es hier, wenigstens zum grossen Teil, um eine pneumatolytische Erzformation handelt. Als Gangart kommt nämlich reichlich Quarz und Turmalin, als Erzart wieder Arsenkies, Kupferkies, Magnetkies und Pyrit vor. Stellenweise hat man verhältnismässig reichlich Scheelit angetroffen. Auch Molybdänglanz ist hie und da in geringen Mengen gefunden worden. Von Interesse ist ein kleiner Gehalt an Zinn, der sich bei Schwimmaufbereitung in das Kupferkieskonzentrat ansammelt. — Es sei noch erwähnt, dass die grösseren Bruchstücke in ihren Randpartien, die kleineren ganz und gar silifiziert sind.

Die lagerstättenkundliche Stellung des goldführenden Haveri-Erzes ist nicht ganz klar. Die Erze kommen hier in feinkörnigem Amphibolit vor und können in Eisenerze und Kieserze eingeteilt werden. Die erstgenannten stellen eine schwächere oder kräftigere Imprägnation von Magnetit in Amphibolit dar. Als sulfidische Erzarten in den Kieserzen kommen hauptsächlich Magnetkies, Pyrit, Kupferkies, Kobaltglanz und Arsenkies in sehr wechselnden Mengen vor. Dazu findet man hie und da ein wening Molybdänglanz, Zinkblende und Tellurwismut. Neulich ist auch Scheelit in Haveri-Erz angetroffen worden. Eine Breccienstruktur ist zu beobachten, obschon diese lange nicht so kräftig entwickelt ist als bei dem Paroinen-Erze. Das Nebengestein ist längs einer schmalen Kontaktzone metasomatisch umgewandelt, u. a. biotitisiert oder silifiziert worden.

Nach Stigzelius hätten sich die sulfidischen Erze des Haverigebietes unter Verhältnissen gebildet, die nahe der Grenze zwischen den pneumatolytischen und hydrothermalen Phasen liegen. Auch nach der Meinung des Verfassers muss es sich hier um hochtermale Erzbildungen handeln, worauf auch das Autfreten von Valleriit in Kupferkies hindeutet. Offenbar ist doch die Erzbildung in etwas niedrigerem Temperaturbereich vorsichgegangen als in Paroinen. — Am schwierigsten lässt sich die Genesis der magnetitischen Eisenerze deuten. Schon die epigenetische Natur derselben ist lange nicht bewiesen worden.

Es scheint, als ob wir in Järvenpää- und Ahdepäägebieten diejenigen Erze unserer metallogenetischen Provinz hätten, die sich in den niedrigsten Temperaturen gebildet haben. Am allerniedrigsten ist die Temperatur während der Bildung der Sb-haltigen Quarzpartien gewesen.

Nach Simonen und Neuvonen (24) gehören die metamorphen Gesteine des Ylöjärvigebietes zum grössten Teil zur Amphibolitfazies, die auch die primäre Fazies darstellt, in welcher alle Schiefer des Gebietes ursprünglich metamorphosiert worden sind. Längs verhältnismässig begrenzten Zonen findet man Mineralassoziationen, die die Gesteine der Niedertemperaturfazien (der Epidotamphibolit- und Grünschieferfazien) charakterisieren. Wir haben hier mit einer Temperaturerniedrigung zu tun, die im Zusammenhang mit dem späteren Vordringen der hydrothermalen Lösungen steht. An diese Zonen schliessen sich die Serizitformationen von Järvenpää und Ahdepää mit ihren Erzbildungen an. Die höhertemperierten Erze von Paroinen und Haveri sind von Gesteinen der Amphibolitfazies umgeben.

EINIGE VERGLEICHSPUNKTE AUS ANDEREN GEBIETEN

Boliden, Schweden. Nach Ödman (13) kommen die Erze in serizit- oder chloritreichen Schiefern vor, die metasomatisch umgewandelte Quarzporphyre, Keratophyre usw. darstellen. Die Erze lassen sich in drei Haupttypen einteilen: Arsenkieserze (ältest), Lamprophyrgänge mit Quarz,-Turmalin und sulfidische Erze samt Pyriterze. Die Aufmerksamkeit richtet sich in erster Linie auf die, in engem Zusammenhang mit den Arsenkieserzen vorkommenden Apophysen, die reichlich »Sulfominerale» und als Gangart bald mehr, bald weniger Quarz enthalten. Sie überqueren schroff die Schieferung (Fig. 15). Als zweifellose Erstausscheidungen in diesen Apophysen sind Arsenkies und Pyrit. In ziemlich frühem Stadium sind Magnetkies, Kupferkies und Zinkblende auskristallisiert. Nach diesen folgen Sb-Pb-reiche Minerale, wie Falkmanit, Bournonit, Jamesonit, Fahlerz, Gudmundit, Bleiglanz usw. Die Apophysen sind goldhaltig. — Nach Ödman sind die Apophysen Differentiate derselben Lösungen, aus welchen die Arsenkieserze selbst auskristallisiert sind.

Man beobachtet, dass in Boliden der Zusammenhang zwischen den Sb-reichen Bildungen und dem Hauptkieserz enger ist als in Järvenpää. Es dürfte doch keine prinzipiellen Verschiedenheiten vorliegen.

Jakobsbakken, Sulitelma. Ramdohr (15) hat aus dieser Grube Sb-reiche Paragenesen beschrieben, die jünger sind als das Hauptkieserz. Von den charakteristischen Mineralen mögen Gudmundit, Boulangerit, Bournonit, Jamesonit, Fahlerz, Bleiglanz und Geokronit genannt werden.

Die Grube Bayerland, Oberpfalz. Dieses Kieserzvorkommen, das von Maucher beschrieben worden ist (11), liegt konkordant in metamorphen Sedimentgesteinen. In verschiedenen Metamorphosen ist das ursprüngliche Pyritlager mit wenig Magnetkies teilweise in ein magnetkiesreiches Pyritlager übergegangen. Das Erz enthält ausserdem Zinkblende, Kupferkies und Bleiglanz. Die einzige Gangart ist Quarz. In Zusammenhang mit dem Kieserz sind quarzige Sb-Pb-reiche Gänge angetroffen, die auch edelmetallhaltig sind. Die Gänge kommen teilweise ausserhalb des eigentlichen Kieserzes, teilweise im Kieslager selbst vor, wo sie auf einige tektonisch stark gestörte Gebiete beschränkt sind. Im ersten Falle ist der als Nebengestein vorkommende Phyllit kräftig serizitisiert, eine Erscheinung, die in dem das Hauptkieserzlager umgebenden Phyllit nicht in dem Masse auftritt.

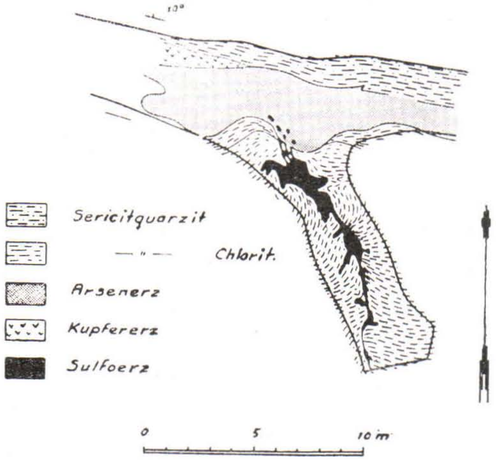

Fig. 15. Gang, von Arsenerz ausgehend. Boliden, 90 m — Sohle, Nach Ödman (16, S. 334).

Die Gänge sind nach Maucher deutlich jünger als das Kieserz. In der antimonreichen Paragenese sind, ausser Arsenkies, Pyrit, Magnetkies, Zinkblende und Kupferkies u. a. folgende Minerale enthalten: Gudmundit, Bournonit, Boulangerit, Jamesonit, Fahlerz und Bleiglanz.

Nach Maucher gibt es für die Genesis der Sb-reichen Gänge in der Grube Bayerland drei Deutungsmöglichkeiten. Eine Möglichkeit ist: »Aszendent hydrothermale Bildung in einem Bildungsgang mit der Kieslagerstätte, aber jünger als die Zufuhr der Kieserze.» Wenn man von dieser Annahme ausgeht, wäre nach Maucher auch die Kieserzlagerstätte als metasomatische Bildung angenommen.

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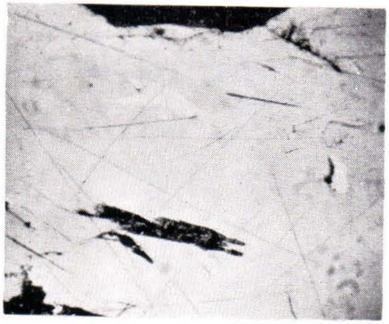

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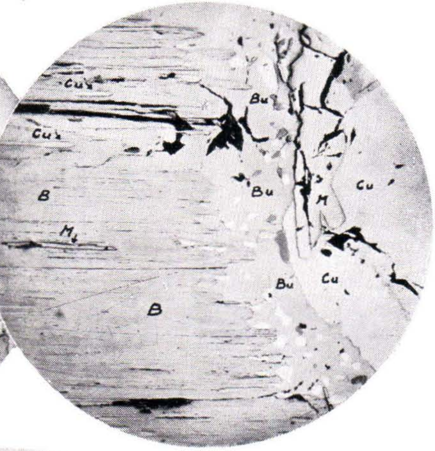

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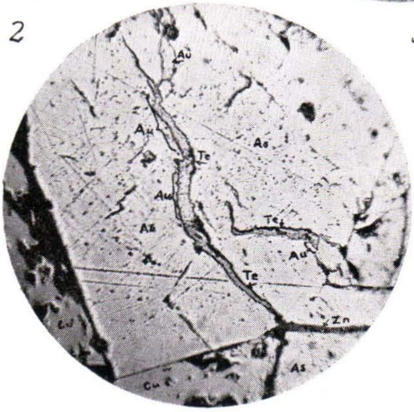

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Fig. 1. Scharfkantige Einschlüsse von Tellurwismut (weiss) in Boulangerit (hellgrau). Schwar = Gangart. Vergr. 82 ×.

Fig. 2. Magnetkieskristalle in Kupferkies (weiss). Die dunkelgrauen Flecke (links) sind Zinkblende. Vergr. 30 ×.

Fig. 3. Reaktionssaum von Bournonit (Bu) zwischen Kupferkies (Cu, hellgrau) und Boulangerit (B, ein wenig dunkler grau, mit Spaltrissen). In Kupferkies Magnetkieskristalle (M) und auf den Spaltrissen des Boulangerits schmale Gangstückchen von Magnetkies und Kupferkies. Im Reaktionssaum kleine Körner von Kupferkies und Zinkblende (dunkelgrau). Vergr. 30 ×.

Fig. 4. Gold-, Goldtellurid- und Zinkblendeadern in Arsenkies. Au = Gold, Te = Goldtellurid, As = Arsenkies, Cu = Kupferkies, Zn = Zinkblende. Vergr. 94 ×.

Martti Saksela: Über eine antimonreiche Paragenese in Ylöjärvi, SW-Finnland.

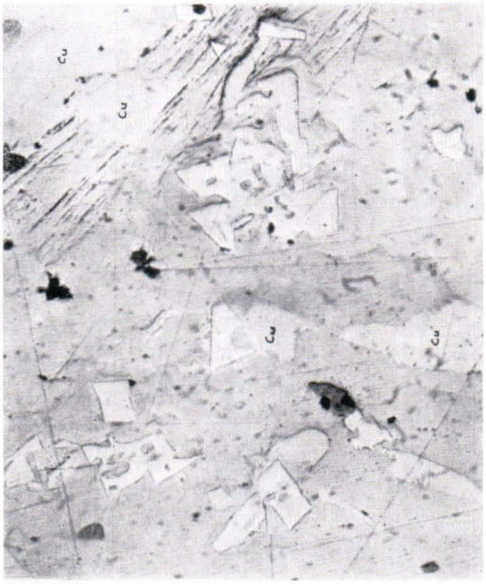

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Fig. 1. Idiomorphe Ullmannitkristalle (hellgrau) in Boulangerit. Cu = Kupferkies. Vergr. 161 ×.

Fig. 2. Myrmekitische Verwachsung von Bleiglanz (weiss) mit Fahlerz (grau). Stark resorbierte Pyrit- und Arsenkieskörner (As). Vergr. 94 ×.

Fig. 3. Myrmekitische Verwachsung von Gold (weiss) mit Goldtellurid (dunkelgrau bis schwarz). Vergr. 943 ×., Ölimmersion.

Martti Saksela: Über eine antimonreiche Paragenese in Ylöjärvi, SW-Finnland.

19.

NOTE SUR QUELQUES PROBLÈMES DE LA TECTONIQUE SUPERPOSÉE

PAR

EUGÈNE WEGMANN
(NEUCHÂTEL, SUISSE)

INTRODUCTION

On entend le plus souvent par tectonique superposée le jeu, dans une région, de deux sortes de structures attribuables à différentes époques tectogènes. La nomenclature en usage dans ce domaine n'est pas encore standardisée. Nous appellerons ici tectonique superposée l'intersection de structures métachrones, tandis que le terme d'interférence resterait réservé à l'interpénétration de structures synchrones.

Les structures superposées observées jusqu'ici ne sont que rarement du même style. Dans la plupart des cas, elles se distinguent par leurs caractères géométriques et par la nature des transformations. On a distingué un grand nombre de cas dans la nature et beaucoup d'autres combinaisons possibles peuvent encore être imaginées. Si l'on ajoute à cela que l'intensité des traces des deux actes de transformation peut varier jusqu'à la disparition de celles de l'un des deux groupes, on se rendra compte de la variabilité du phénomène. Beaucoup de ces combinaisons imaginées ne seront pourtant pas réalisables dans la nature.

L'étude de la tectonique superposée est l'un des très intéressants chapitres de la *tectonique comparée*. Ses résultats touchent aux questions fondamentales de la géologie des chaînes de montagne et de l'évolution de l'écorce terrestre. Ils touchent à l'antagonisme tant entre tectoniciens fixistes et mobilistes, (Argand, 1924, p. 289) qu' entre partisans des hypothèses magmatistes et transformistes (le migmatisme n'intéressant qu'une partie du domaine beaucoup plus étendu embrassé par cette controverse). L'étude de la tectonique superposée permettra de contrôler maintes affirmations émises par les différentes écoles géologiques et exigées par leurs principes d'interprétation. Elle favorise ainsi une critique constructive. Nous essaierons d'en esquisser quelques aspects dans cette note.

Pour discuter les phénomènes d'une façon plus rationnelle, il est nécessaire de distinguer tout d'abord deux éléments d'un segment orogénique:

a) le matériel sédimentaire nouveau,
b) le socle ancien.

Ce dernier est, dans la règle, formé par les restes d'un segment déformé plus ou moins pénéplainé. Les nouvelles formations transgressent sur cette pénéplaine. Le vieux socle est le plus souvent formé par les parties cristallines d'un ou de plusieurs cycles orogéniques antérieurs.

INFLUENCE DE LA STRUCTURE DU SOCLE SUR LES NOUVELLES SÉRIES SÉDIMENTAIRES

Un nouveau cycle orogénique commence par le dépôt d'une nouvelle série sédimentaire. La répartition des facies sera souvent influencée par les déformations »posthumes» du socle, qui sont par définition plus ou moins parallèles à la structure interne. Les exemples en sont fréquents et leur recherche a été à la mode à une certaine époque. Les idées d'Élie de Beaumont, d'Edouard Suess, dans les premiers volumes de l'ouvrage: »La face de la terre», et de Karpinski sont caractéristiques à ce sujet; mais, déjà dans le dernier volume de son ouvrage fondamental, Suess montra que les relations pouvaient être plus complexes. En effet, les études de paléogéographie montrent que c'est fréquemment la nouvelle direction structurale qui se faisait déjà sentir depuis le commencement d'une série sédimentaire. Si ce n'était pas le cas, les reconstructions d'embryotectonique ne seraient pas possibles, puisqu'elles tendent à montrer que les directions des zones de facies et celles des futurs plis et nappes sont plus ou moins parallèles. En ce qui concerne la zone helvétique des Alpes, par exemple, les auteurs sont d'accord pour rattacher la distribution des facies aux tendances positives des massifs hercyniens, pas nécessairement parallèles aux anciennes structures (cf. p. 230).

Les déformations qui règlent la distribution des facies pendant les premiers stades de l'évolution semblent presque toujours être des dômes et des bombements à grand rayon de courbure. La direction de leur axe est considérée, (à tort ou à raison) comme étant parallèle aux zones de sédimentation, elles-mêmes plus ou moins parallèles à l'ancien rivage. Des résultats de ce genre doivent donc, dans beaucoup de cas, être acceptés avec une certaine prudence. La forme des régions à tendance positive ressemblait probablement beaucoup plus à celle des »uplifts» de la région centrale des Etats-Unis qu'à une préformation des massifs actuels.

HIÉRARCHIE DES ÉLÉMENTS TECTONIQUES D'APRÈS LEUR ORDRE DE GRANDEUR

La discussion de l'influence des anciennes structures sera facilitée par une classification des éléments d'après leur ordre de grandeur.

On distinguera des unités de premier ordre comprenant un segment déformé qui sera composé des éléments d'une classe inférieure embrassant des *séries* de plis, de nappes, de blocs ou de noyaux granitiques, telles que les zones des Alpes (helvétique, pennique, austro-alpin) ou des Calédonides scandinaves ou d'autres segments zonés. Ces zones seront divisées en unités plus petites, faciles à déceler; ce qui permet de les individualiser c'est leur tendance de mouvement (nappes, plis, blocs faillés et intrusions), Certains de ces éléments ne montrent des traces de mouvements importants que sur leurs bords, comme les blocs limités par des failles et les nappes du style cassant. On les distinguera donc des unités avec traces de mouvements internes plus prononcés.

La division suivante est d'un ordre de grandeur s'étendant entre quelques mètres et quelques dizaines de mètres. Ce sont, par exemple, les plis visibles dans un affleurement, les blocs entre les grandes diaclases, les filons et autres phénomènes de cette grandeur. Cette division contient, comme la précédente, un certain nombre de formes, chez lesquelles le mouvement s'arrête aux limites de l'élément en question, en tout cas en ce qui concerne les mouvements synchrones. Le pas suivant nous mène à l'ensemble des structures et textures, à la trame des roches (Gefüge, petrofabric, Sander 1930). La limite des minéraux ou des fragments de minéraux (parfois aussi des fragments de roches) composant l'ensemble sépare cette catégorie de l'ordre inférieur. Nous laissons pour le moment de côté les traces des mouvements apparaissant à l'intérieur des minéraux.

Cette division n'est pas rigoureuse puisqu'on trouve toutes les transitions dans la nature. Ce sont donc plutôt des *types* de phénomènes choisis dans une série continue et fixés pour la commodité de l'exposé. Ils ne sont toutefois pas choisis au hasard; chaque ordre de grandeur se distingue des autres par des caractères dominants; les raisons du choix seront exposées ailleurs. Les phénomènes ressortant à des ordres de grandeur différents ne sont pas, au point de vue cinématique et dynamique, des répliques plus ou moins grandes, pour lesquelles les mêmes raisonnements seraient applicables sans autre. Pour les sédiments, par exemple, les galets et blocs de roches d'une part, et les grains de poussière ou particules colloïdales de l'autre, se comportent d'une façon très différente par rapport aux agents de transport. Les phénomènes tectoniques affectant différents ordres de grandeur varient tout autant depuis le haut vers le bas de l'échelle.

L'application des raisonnements per analogiam est limitée à certaines classes d'ordre de grandeur, aussi bien vers les catégories inférieures que supérieures. Certains auteurs ne procédent que depuis les étages supérieurs vers les étages inférieurs, ou inversément. Cette manière de faire est peut-être satisfaisante au point de vue philosophique ou à celui de la logique pure, puisqu'on obtient une simplification étonnante. Elle ne donne toutefois pas toujours de bons résultats dans la pratique des recherches, où il est bon d'avoir certaines possibilités de contrôle. Une comparaison permettra de saisir de quoi il s'agit: admettons des êtres plus petits que nous, obligés de copier une mosaïque représentant une image d'une certaine complexité. Ils peuvent procéder de deux manières: a) ils décrivent et mesurent pierre par pierre, en notant la couleur, la forme et les distances au prochain élément. En travaillant très exactement, ils obtiendront une bonne copie; il ne faut toutefois pas oublier que les petites fautes ne se compenseront pas, dans la règle; b) ils essayeront d'obtenir une vue d'ensemble et de dessiner les grandes surfaces, contenant des pierres rouges, vertes ou bleues, et de placer les petits éléments dans les surfaces ainsi délimitées. La forme générale du dessin ne sera pas influencée par les petites fautes commises en passant d'un élément à l'autre.

Les exemples de la nature contiennent beaucoup plus de catégories d'ordres de grandeur. En divisant les phénomènes en classes de grandeur, on peut les décrire sous deux points de vue susceptibles de se contrôler mutuellement: ils sont considérés à chaque étage, d'une part, comme continus, et, d'autre part, comme des ensembles composés par les éléments discontinus de l'étage inférieur. Les méthodes tectoniques s'inspirant des deux manières de voir (phénomènes continus et discontinus) permettent une coordination rationnelle des observations.

L'ANISOTROPIE DES MATÉRIAUX NOUVEAUX

L'étude des facies et de leur distribution peut nous renseigner sur les influences que pourraient manifester les structures du premier et du deuxième ordre de grandeur. Mais les mouvements des socles se font aussi sentir dans les phénomènes des étages inférieurs, les failles et les diaclases. Si le socle se divise en compartiments se mouvant le long de failles, les séries sédimentaires suivent ce mouvement. La différence de niveau des blocs du socle est souvent amortie dans les couches sédimentaires et se traduit en flexure. Le tracé des failles peut correspondre à des surfaces d'inhomogénéité du socle, parfois à d'anciennes failles ou charriages. Le dessin en échelon de beaucoup de ces accidents montre qu'il y a interférence entre l'anisotropie du socle, marquant l'ancienne structure et les nouvelles tendances de mouvement indiquées par la direction du train général des échelons. La représentation de la répartition des facies ne marquera les échelons que s'ils sont assez grands, ou si les recherches ont été assez détaillées pour les déceler (cf. Groenland oriental, Vischer 1944).

Les exemples de ce genre sont très nombreux dans la littérature géologique ancienne et récente, aussi bien dans les pays tabulaires que dans les régions plissées. Les pays tabulaires faillés entourant les vieux massifs hercyniens de l'Europe occidentale et centrale offrent un grand nombre de bons exemples.

Les régions plissées, comme le Jura, en fournissent d'autres. Les travaux de Glangeaud (1942) et d'Aubert (1945) en montrent toute l'importance. Glangeaud a décrit »de petits accidents aberrants» qu'il appelle des »pincées». Il les définit de la manière suivante: »Dans une série régulière et continue de calcaires subhorizontaux, apparaît brusquement une solution de continuité, occupée par des sédiments fortement plissés, appartenant à un niveau supérieur. La largeur de l'accident est de 300 m environ.» La plus grande extension longitudinale constatée jusqu'à maintenant est de 40 km.

Des exemples encore plus frappants ont été découverts dans la série des Alpes Helvétiques par Guenzler (1941, 1944). Ces failles d'âge mésozoïque, correspondant à la phase cimmérienne, ont joué à plusieurs reprises. Elles gardent ainsi l'empreinte d'un socle sur lequel elles ne reposent plus, puisque toute la série helvétique est décollée et a glissé vers le Nord. L'anisotropie provoquée par les failles a, en grande partie, déterminé la localisation et la direction des replis de la nappe du Wildhorn.

Un autre exemple bien étudié est la région côtière du Groenland oriental. La chaîne calédonienne forme le socle par rapport aux formations dévoniennes et post-dévoniennes. Avant le dépôt du Dévonien, (Old Red Sandstone) le socle a été profondément érodé. Les dépôts clastiques remplissent, grosso modo, un synclinal calédonien, sans toutefois en être une réplique exacte, ce qui est aussi le cas en Norvège occidentale. Les déformations du Dévonien et du Carbonifère (Buetler 1935, 1939) recoupent les structures calédoniennes. Les séries post-carbonifères allant du Permien au Crétacé supérieur sont coupées en blocs d'une largeur moyenne de 15 à 20 km. Les failles prenant leur origine dans les socles ont rejoué à diverses époques et ont chaque fois rejeté la couverture sédimentaire. Elles sont inclinées vers l'Est. Les blocs ont effectué un mouvement de bascule, dont les étapes se répartissent dans le mésozoïque. La partie orientale des compartiments est montée, tandis que la partie occidentale s'est abaissée. Ces mouvements ont déterminé, comme Vischer (1943) l'a montré, la nature de la topographie mésozoïque (archipel avec des îles allongées formées par les bords remontés des blocs) et la distribution des facies (séries de conglomérats au pied des facettes de faille). En comparant les coupes transversales et longitudinales, on remarque que le jeu des blocs semble avoir changé au cours de l'évolution. Les accidents formant la côte actuelle, s'étendant environ entre 73° 15′ et 70° 15′, du N au S, ont une autre direction que les failles actives jusqu'à la fin du mésozoïque, et, probablement, au tertiaire.

Le fait que ces failles recoupent le socle calédonien n'implique pas qu'elles soient déterminées par sa structure. On ne voit pas bien pourquoi les directions de leurs mouvements auraient changé. Il semble plutôt que leur direction et le sens de leurs mouvements indiquent une nouvelle tendance dans l'évolution de la marge continentale. L'évolution du Groenland méridional semble donner des résultats analogues. (Wegmann 1938, 1939).

Citons un exemple encore plus évolué: le Tessin méridional (Luganais et Mendrisiotto) en Suisse. La région se subdivise en plusieurs blocs basculés. Le socle est formé par du cristallin recouvert d'une série de porphyrites et de porphyres (d'âge permien), sur lesquels se sont déposés des sédiments allant du Trias au Tertiaire avec un certain nombre de lacunes. Le mouvement de bascule a commencé à se faire sentir après la pénéplénation, puisque les canaux nourriciers de la série des porphyres ont rempli les espaces de discontinuité entre les blocs. Le mouvement continue pendant une partie du mésozoïque. Il est particulièrement sensible au Lias (brocatello d'Arzo) Les compartiments penchent de plus en plus vers l'Est comme une rangée de livres qu'on inclinerait de côté. Les plissements alpins ont influencé cette structure de plusieurs manières: les terrains de couverture sont en partie plissés; l'ensemble a subi une rotation à axe méridien sensiblement horizontal; des failles et des flexures de direction équatoriale recoupent la structure; cette dernière est renforcée. Les anciennes failles fonctionnent à plusieurs reprises d'une manière différente.

Il est possible d'ajouter à cette série des types encore plus évolués. Pour le but que poursuit cette note, il suffit d'avoir esquissé cette lignée. Une grande partie de ce que Stille désigne sous le nom de »tectonique germanotype» trouve sa place dans cette catégorie.

Les formations tabulaires montrent le plus souvent des systèmes de partage correspondant à différents ordres de grandeur, compartiments délimités par des failles et des diaclases. Les failles proviennent, dans la plupart des cas, de semblables mouvements du socle. Il en est autrement d'une grande partie des diaclases. Rappelons d'emblée qu'il y a beaucoup de termes de passage entre les failles et les diaclases.

Le phénomène des diaclases, déjà étudié par Daubrée (1880), dans le bassin de Paris, a fait l'objet de beaucoup de travaux sur les formations tabulaires mésozoïques de l'Europe centrale, (voir bibliographie dans Philipp 1935, Thiele 1935 et Sonder 1938) et du paléozoïque ancien de la région baltique (Kaufmann 1931, Teichert 1927). Les interprétations étonnent parfois ou paraissent contradictoires. Il semble toutefois s'en dégager quelques points principaux: dans certains cas, l'on peut distinguer entre diaclases et failles. La plupart des auteurs sont d'accord pour attribuer cette différence à la nature dissemblable des causes, il est aussi possible de l'attribuer à une différence d'âge, peut-être aux deux raisons à la fois.

Les diaclases peuvent changer de direction dans une série sédimentaire contenant des zones lithologiques de nature différente. Ceci correspond à l'allure disharmonique des séries de ce genre dans les pays plissés. L'âge de la fissuration dans les parties supérieure et inférieure d'une grande série peut être différent. Le phénomène est souvent impressionnant à la base d'une série: sur un socle fissuré repose une couche de conglomérats de base parfois épaisse qui ne montre généralement que des diaclases espacées, tandis que des calcaires, par exemple, situés plus haut dans la série pourront montrer un réseau plus dense et de direction différente.

Les diaclases sont importantes parce qu'elles marquent l'anisotropie des sédiments par rapport aux directions parallèles aux couches. Elles peuvent être l'effet de mouvements du socle, de gauchissements par exemple et ne sont donc qu'*indirectement dépendantes* de la structure du socle. Même dans le cas où les directions des diaclases du socle et de la couverture sont plus ou moins parallèles, cette coïncidence n'est pas la preuve d'un rapport direct de causalité. Les diaclases des séries nouvelles ne sont pas la continuation de celles du socle.

TECTONIQUE SUPERPOSÉE DANS LE SOCLE

Le rapport entre l'ancienne structure et la nouvelle déformation doit être étudié dans le socle. Dans ce but, il est nécessaire de diviser le socle en un certain nombre de niveaux; dans le cadre de cette note, il suffit d'en distinguer deux:

a) la partie haute, caractérisée dans ses niveaux supérieurs par une anisotropie déterminée par les diaclases, failles, plans de charriage et autres discontinuités de ce genre. Les unités qui la composent sont les blocs tels qu'on les détache des carrières.

b) La partie inférieure, caractérisée par des déformations continues, (en tout cas à l'échelle macroscopique). Cette zone est le siège de regroupements de phases et de migrations chimiques, dont nous ne nous occupons pas pour le moment. L'anisotropie est marquée par la limite des roches à composition et structure différentes.

Tout pays, dont la surface topographique est formée par du cristallin ancien, est montée de la zone inférieure dans la zone supérieure. Pendant leur montée, les matériaux ont ajouté à l'anisotropie marquée par la différence des roches, celle du réseau des fissurations. Ces surfaces de partage, à l'exception des grandes failles, montrent des directions variables dans un même district, si celui-ci n'est pas formé par une masse rocheuse homogène. Dans le sens de la hauteur, les systèmes changent généralement d'étage en étage. On peut s'en convaincre dans les coupes naturelles d'une certaine hauteur, par exemple au Groenland méridional (Wegmann 1938), au Groenland oriental (Wegmann 1935) et en Norvège. La mosaïque parfois compliquée des diaclases dans les pays à relief peu prononcé, comme la Fennoscandie ou le bouclier canadien, s'explique facilement par le fait que la surface topographique recoupe la structure à des niveaux très différents. Les joints transversaux des plis du type pennique (aussi bien dans les Alpes qu'ailleurs), semblent être les systèmes les plus étendus, au point de vue de la longueur de chaque train de fissures. Les cartes des fissures marquées par la topographie, publiées pour la Fennoscandie (Sederholm 1913, Hausen 1942, Ljungner 1930) et pour le bouclier canadien (Quirke 1936), montrent un dessin assez chaotique. Ce fait est dû à plusieurs facteurs: l'inégalité de la documentation due à la morphologie, la différence des matériaux, et surtout aussi la différence du niveau d'érosion variant d'un district à l'autre. Les directions principales de l'anisotropie de fissuration acquise lors de la montée changent donc horizontalement et verticalement. Nous ne nous occuperons pas ici des causes de la fissuration (cf. Cloos 1936).

On distinguera deux types principaux de tectonique superposée:

1) Un socle cristallin ayant acquis une première fissuration est déformé dans la zone supérieure (cf. p. 229);

2) un socle cristallin replonge dans la zone inférieure ou sous-socle.

1) Différents exemples ont été étudiés. Nous citerons en premier lieu les travaux minutieux de N. Oulianoff (1937, 1944, voir bibliographie). Celui-ci a étudié la tectonique superposée dans les massifs du Mont-Blanc, des Aiguilles Rouges et de l'Aar. Les études de N. Oulianoff montrent tous les termes de passage entre les districts à structure de prévalence hercynienne et des régions, où la tectonique alpine prédomine.

La première a pris naissance à la limite du niveau de la granitisation pour monter à la surface pendant une période d'érosion profonde antéstéphanienne (Lugeon 1930). Le Permien et le Trias déposés en discordance marquent des déformations de moindre intensité. Le cristallin a de nouveau été enseveli sous les sédiments mésozoïques et tertiaires et sous les masses charriées de la zone helvétique. L'enfoncement n'a toutefois pas atteint les profondeurs indiquées par les conditions hercyniennes, de sorte que, comparée à la première transformation, la déformation alpine a eu lieu à un niveau supérieur, caractérisé par la présence de la chlorite.

Les anciennes structures influencent les déformations alpines, l'ancienne anisotropie décompose les nouveaux mouvements dans les premiers stades; dans les cas avancés, elle tend à disparaître. La déformation alpine, bien que revêtant souvent un style cataclastique dans le détail, montre, dans les ordres de grandeur supérieur, beaucoup de formes à courbure continue. Pour les détails, nous renvoyons aux bonnes descriptions de N. Oulianoff.

2) Le socle replonge dans la zone inférieure (cf. p. 229.). Les exemples sont assez fréquents; mais, en général, ils ne sont pas étudiés à ce point de vue. Nombre de districts du vieux cristallin présentent cette particularité qui ne se révèle qu'à une analyse tectonique approfondie. Nous en avons décrit en Finlande (Wegmann 1929) et au Groenland (Wegmann 1935, 1938). Beaucoup de régions caractérisées par des gneiss à hypersthène et des séries semblables appartiennent à cette catégorie. Nous ne mentionnerons que l'aspect structural qui nous intéresse ici directement.

Les études montrent que les diaclases, les failles et les anciennes zones de mylonites cessent de jouer un rôle au point de vue de l'anisotropie. Les traces restent parfois visibles, soulignées par des traînées de feldspaths, d'amphiboles, de biotites, de grenats, ou d'autres minéraux de ce genre. Les nouveaux plans d'écoulement passent à travers ces configurations qui ne représentent de ce fait qu'un dessin de contrôle. L'anisotropie est marquée par la géométrie des espaces remplis par les différentes roches qui font partie de l'édifice.

La structure antérieure peut disparaître, aussi bien en ce qui concerne les grandes lignes que la trame des roches. Les structures relictiques et les traces dans la symétrie de la trame (Sander 1930), ainsi que d'autres indices de ce genre, permettent parfois de reconstituer l'état antérieur en plein milieu de la zone transformée. La limite de ce changement de structure est souvent si marquée qu'elle peut être dessinée, voire photographiée dans les affleurements favorables (Wegmann 1935). Elle a passé à travers les espaces inférieurs, entraînant une série de transformations minéralogiques et géochimiques. C'est le train des *fronts de transformation endogène*. Doris Reynolds (1946) vient d'en montrer les règles géochimiques qui seront de la plus grande importance pour l'analyse tectonique des sous-socles. Nous reviendrons ailleurs sur les principes de l'analyse structurale de ces régions.

La coupe verticale d'un segment orogénique montre donc deux espaces où les nouvelles structures prévalent: les terrains de couverture et le sous-socle. Le haut du socle est dans ce cas le seul espace qui aura conservé les anciennes structures et anisotropies. Cette partie intermédiaire peut disparaître quand les fronts de transformation endogène montent dans les terrains de couverture. Ce sera le cas si les nouveaux matériaux sont très épais, ce qui peut être dû à une sédimentation abondante dans un géosynclinal, à des plissements ou charriages, ou à ces deux causes à la fois.

Il est donc nécessaire de faire une distinction dans un segment orogénique, non seulement entre le socle et les matériaux nouveaux exogènes, mais aussi entre le socle à structure ancienne et les matériaux renouvelés, produits de la transformation endogène. Ces matériaux, devenus très mobiles, peuvent s'introduire dans les diaclases et autres fissurations de la partie haute du socle (rhéomorphisme). Ils remplissent ainsi des moules provenant du cycle orogénique antérieur et peuvent passer de là dans les matériaux nouveaux.

Au cours des époques orogéniques, l'écorce terrestre se voit partagée par des segments entièrement renouvelés depuis le haut et depuis le bas, comparables à des *soudures*. Branchés sur ces soudures ou à une certaine distance de celles-ci, s'étendent des segments incomplètement renouvelés. Ce sont les nombreux plis de fond, de types en partie très différents (cf. Huang 1945). Cette manière de voir permet de saisir quelques aspects du problème de l'orogénèse, moins bien visibles à partir d'autres points de vue. Le renouvellement des socles est un problème commun à la pétrogénèse, à la géochimie et à la tectonique.

Les soudures peuvent se recouper; les points d'intersection sont particulièrement intéressants.

LE ROLE CINÉMATIQUE DES FISSURATIONS

On se demandera peut-être pourquoi nous avons insisté sur les anisotropies qui correspondent aux fissurations? Aussi devons-nous dire quelques mots de leur signification cinématique.

Pour les besoins de la discussion, les mouvements tectoniques peuvent être décomposés de différentes façons. Dans le cas d'une plaque, (couche, série, filon etc.) les mouvements parallèles à la plaque constituent, soit un élargissement ou étalement, soit un raccourcissement ou rétrécissement. Deux points de ces espaces, alignés dans le sens du mouvement, se rapprochent ou s'écartent au cours de la déformation. L'étalement ou le rétrécissement d'une plaque constituée de matière mobile augmentera ou diminuera son épaisseur. Une plaque plus ou moins solide, fissurée, réagira d'une autre façon: l'étalement ne sera pas continu, mais discontinu, c'est à dire que les fissures préexistantes s'y ouvriront. Nous avons traité la cinématique d'un cas de ce genre en décrivant la tectonique comparée du phénomène du boudinage (Wegmann 1932).

Dans le cas du boudinage, la plaque qui subit l'élargissement est comprise entre deux couches de matériaux plus mobiles. Dans le cas qui nous occupe, il n'y a que le substratum qui est mobile. Les glaciers arctiques fournissent des modèles intéressants de ce genre de mouvements. Les couches supérieures ayant une température très basse ne sont que peu mobiles, tandis que les parties profondes sont douées de la plasticité caractéristique de la glace des glaciers. Les jeux des crevasses, tels qu'ils ont été décrits, et que les photographies aériennes modernes les révèlent, sont du plus haut intérêt pour en donner la compréhension.

Les pincées du Jura (Glangeaud 1942) s'expliquent par une phase d'étalement et une phase de rétrécissement. Les failles cimmériennes

(Guenzler 1941, 1944) ainsi que les accidents semblables du Groenland oriental et du Tessin sont le résultat de phases d'étalement précédant ou alternant avec des phases de rétrécissement. L'évolution est dans chaque cas quelque peu différente, ce qui permet de définir plusieurs types cinématiques.

Les élargissements peuvent être mis en relation avec les accidents des ordres de grandeur supérieur. Ce sont les bombements, les dômes, gauchissements et autres phénomènes de ce genre. Ces relations ont été mises en évidence par Cloos (1939). Il a souligné les analogies de plusieurs exemples de socles soulevés en dôme avec les résultats d'expériences intéressantes. La ressemblance des modèles obtenus par la déformation de grands gâteaux d'argile aux structures observées dans la nature est frappante, ce qui est d'autant plus curieux que les matières argileuses nécessaires pour ces expériences ont une résistance à la traction que les roches solides fissurées ne possèdent pas dans la règle.

Les failles d'une phase antérieure peuvent jouer un rôle important, mais assez variable d'un cas à l'autre, lors du *plissement*. Nous ne rappellerons que les exemples déjà cités des pincées, (Glangeaud 1943) où les petits segments sont serrés comme dans un étau, et des failles cimmériennes ayant déterminé en grande partie la structuration de la nappe du Wildhorn (Guenzler 1941, 1944). Leur rôle a été différent, mais, néanmoins, fondamental dans l'exemple du Tessin. Le plissement est donc *précédé* dans ces cas par une phase d'étalement qui accompagne probablement un bombement. Dans le cas du Jura, le soulèvement en dôme était probablement d'âge oligocène; pour la zone helvétique, il date du Jurassique et du Crétacé, tandis qu'il est activé à plusieurs reprises au Tessin. L'exemple du Groenland oriental est probablement unique par la persistance des mêmes tendances depuis le Permien jusqu'au Tertiaire.

REMPLISSAGE DES FISSURATIONS

Les fissures qui tendent à s'ouvrir servent de moules pour les masses mobiles endogènes. Les réseaux de roches basiques, surtout des basaltes, conservent les images cinématiques d'un certain stade de l'évolution structurale. Traversant le socle et les terrains de couverture, les filons basiques offrent, d'autre part, des marques chronologiques, une espèce de signal horaire géologique, qui peut être pris à tous les étages d'un segment orogénique, depuis le sous-socle jusqu'à la surface. Ces filons peuvent ainsi servir à la *synchronisation* des phénomènes de profondeur et de surface. Ils peuvent être suivis sur de longues distances même en plein terrain granitisé.

Les marques de la structure antérieure étant souvent peu visibles en arrière des fronts de transformation endogène, on s'adressera en premier lieu au témoignage des anciens filons basiques qui conservent, dans la règle, le mieux la marque des anciennes anisotropies. Leur dessin permet, d'une part, de reconstruire une partie de l'ancienne structure, et, d'autre part, de suivre les changements structuraux survenus lors du passage des fronts de transformation endogène. Ils peuvent ainsi servir de dessin de contrôle.

Le réseau des filons et des intrusions basiques permet dans beaucoup de cas de distinguer dans le vieux socle des phases d'élargissement et de rétrécissement, parfois même dans des sous-socles renouvelés d'une façon assez intense. Ce sont probablement les meilleurs guides dans l'étude de la tectonique superposée des socles cristallins. Leur rôle chronologique a été souligné depuis longtemps par J. J. Sederholm.

Les fentes du socle, tant soit peu ouvertes pendant un plissement de fond, gardent souvent les traces de cet épisode sous forme de revêtements de chlorite, d'hématite ou d'autres minéraux de ce genre. Ces revêtements sont parfois écrasés, ou transformés en miroirs avec stries pendant une phase subséquente. Les remplissages peuvent être conservés dans certains cas même lors d'un mouvement descendant du socle.

RELATIONS AVEC D'AUTRES RECHERCHES ET HYPOTHÉSES

Les anciennes fissurations se font surtout sentir dans les vieux socles soulevés, dans les plis de fond d'Argand (1924). Nous employons ce terme pour décrire la structure, sans y inclure un jugement sur les causes qui lui ont donné naissance. Il ne nous semble pas qu'il faille admettre implicitement que les plis de fond aient été uniquement formés par des poussées latérales (cf. Argand 1924, p. 176). Les anisotropies de l'écorce terrestre déterminent dans les régions à plis de fond surtout des phénomènes d'extension. La morphologie de ces pays est très caractéristique par le dessin des vallées, et ne peut pas être confondue avec d'autres formes. Rappelons les paysages de la Fennoscandie (Sederholm 1913, Hausen 1942, Ljungner 1930), du bouclier canadien (Quircke 1936), du Brésil, du Sinaï, du Groenland et aussi ceux de la Forêt Noire et des Vosges, du Plateau Central de France, et de l'Écosse. Les traits des paysages de ce genre peuvent parfois être mis en relation avec plusieurs générations de structures antérieures. Ce n'est pas seulement la forme des vallées, résultant des facteurs de transport et des changements du niveau de base, mais aussi, le tracé du chevelu hydrographique qui mérite une attention spéciale, car il peut révéler des mouvements récents ou vivants, ouvrant les anciennes fissurations. Les caractères morphologiques et les séries de couvertures avec leurs anisotropies des différents ordres de grandeur peuvent être considérés comme phénomènes enregistreurs des plis de fond. Puisque leurs fissurations préexistantes s'ouvrent dans les bombements, l'étude consistera en grande partie dans l'analyse géométrique et cinématique de ces discontinuités. Elle ne peut pourtant pas se limiter aux ordres inférieurs, par exemple aux diaclases, mais doit procéder à la coordination de tous les étages de grandeur, depuis les diaclases jusqu' aux plis de fond. La technique de l'analyse diffère, par conséquent, de celle des plis géosynclinaux.

Les travaux d'Émile Argand (1924) et de Huang (1945) montrent l'importance des plis de fond et l'immense répartition des régions à tectonique superposée.

Les segments orogéniques à matériaux principalement renouvelés (cf. p. 231) montreront de nouvelles structures. Celles-ci peuvent êt replus ou moins parallèles aux anciennes; mais il n'est tout de même pas prudent d'affirmer pour cette raison: mêmes directions, mêmes causes! Dans d'autres cas les diaclases et failles transversales d'un ancien segment sont souvent parallèles aux structures principales de la soudure consécutive.

On a voulu voir dans les chaînes postcambriennes des répliques plus ou moins exactes de la chaîne alpine s'ajoutant autour d'un noyan formé par les plissements antérieurs. Un examen plus attentif du tracé des chaînes anciennes depuis les Svécofennides jusqu'aux Calédonides et aux Altaïdes montre que les segments à matériaux renouvelés peuvent aussi bien se recouper que s'ajouter en zones concentriques autour d'un noyau. L'étude approfondie de la zone méditerranéenne indique des relations entre les déformations des socles et les plissements géosynclinaux passablement plus compliquées que ne l'admet le schéma classique. Les travaux de Fallot (1941) sur la coordination des mouvements des chaînes comprises entre l'Anti-Atlas et la Cordillère Bétique montrent que les plis de fond ne peuvent pas être le contrecoup des plissements géosynclinaux de cette région. D'autre part, dans une note qui mérite l'attention des tectoniciens, Fallot (1944) appelle à la prudence: »Au lieu de vouloir chercher dans les Altaïdes une réplique, ou mieux, une préfiguration des Alpes, le tectonicien objectif doit, au contraire, constater que ces deux édifices diffèrent par leur essence même.» Une telle différence n'existe pas seulement entre les Alpes et les Altaïdes, mais aussi entre les chaînes plus anciennes. Chaque segment étudié un peu plus en détail révèle des traits caractéristiques créant ainsi une physionomie individuelle.

Durant assez longtemps on a voulu créer une image standardisée de ce qu'un segment orogénique devrait être et on a ainsi uniformisé l'aspect structural des chaînes de montagnes. Ces schémas ne sont dans la plupart des cas que des images de première approximation; ils peuvent servir d'hypothèses de travail destinées à être dépassées par les recherches. Les traits individuels des chaînes qui sortent du cadre de ces schémas commencent de plus en plus à se dessiner. Ils nécessitent une nouvelle classification, qui ne sera pas uniformiste, mais tiendra compte des différences.

Le critère des transformations endogènes du socle y jouera un rôle important. Les schémas provenant de la géologie alpine n'en tiennent compte que d'une façon accessoire. Le phénomène spectaculaire des nappes de recouvrement constitua longtemps le foyer de l'intérêt de la géologie alpine, tandis que les manifestations des transformations endogènes attiraient l'attention des chercheurs dans les vieilles chaînes, où l'analyse géométrique et cinématique occupait un des derniers rangs. Les manifestations endogènes ne sont nullement négligeables dans les Alpes, bien qu'elles n'apparaissent pas dans les coupes synthétiques, nécessairement virtuelles, construites en projetant les données d'un niveau supérieur vers un niveau inférieur. Cette manière de faire était légitime aussi longtemps qu'on ne connaissait pas les styles de profondeur et elle pouvait représenter une première approximation. En essayant de reconstruire les parties inférieures des coupes alpines d'après les expériences des vieilles chaînes, on obtiendrait des représentations quelque peu différentes des vues officielles.

Une autre lignée de théories se rattache aux noms d'Elie de Beaumont, de R. T. Chamberlin et de Hobbs. Le principe de base est l'existence de certaines directions privilégiées de l'écorce terrestre préfigurées depuis les premiers temps de la consolidation. En combinant ce principe avec certaines données géophysiques du sous-sol des continents et des océans, Sonder (1938, 1939) semble obtenir les détails en ordre descendant, par voie de *déduction.*

Bien que cette théorie soit susceptible de grouper un certain nombre de connaissances d'une façon intéressante en découvrant de nouvelles relations, elle n'est pas encore suffisamment développée pour qu'on puisse en tirer un questionnaire pouvant être contrôlé dans la nature.

Dans cette note, nous avons voulu suivre l'un des nombreux fils qui relient la tectonique des vieux pays cristallins à celle des chaînes plus jeunes. L'évolution des étages supérieurs doit nécessairement être suivie en même temps que celle des parties profondes. Il ne semble guère prudent d'échafauder des hypothèses tectoniques basées uniquement sur les déformations superficielles (même si on les extrapole et les projette en profondeur), et il est peu satisfaisant d'étudier les régions profondes en n'ayant que le ciel au-dessus de soi-même, c'est à dire : sans savoir à quels phénomènes de surface il faut relier les structures constatées en profondeur. Une série de méthodes permettant de relier les deux domaines ont été développées pendant cette dernière vingtaine d'années. Les résultats de leur application seront à la base des synthèses de demain. Des segments orogééiques coupés par un niveau d'èrosion, faisant voir en même temps les ntages supérieurs et le sous-socle, seront appelés à jouer un très grand rôle. Les chaînes calédoniennes du Groenland oriental et de Scandinavie représentent quelques régions-clefs pour l'étude de ces questions.

La chaîne des Carélides, le premier segment précambrien analysé avec les méthodes de la tectonique moderne, a révélé la nécessité de pouvoir relier les structures du sous-socle aux étages supérieurs (1928). Les relations entre les deux espaces sont mieux connues aujourd'hui, mais le champ est vaste et il reste beaucoup à explorer. Les méthodes qui combinent l'analyse tectonique avec les techniques de la petrographie sont parfois méconnues autant par les géologues que par les pétrographes. Faudrait-il se laisser décourager par les recommandations de quelques autorités? Faudrait-il toujours rester dans une ornière?

Qu'on permette à l'auteur de cette note d'attacher un remerciement à ces lignes présentées en hommage à M. Pentti Eskola: c'est le jubilaire d'aujourd'hui, qui, en août 1927, a montré à l'auteur les merveilleux paysages de Carélie et l'a encouragé dans ses recherches sur la tectonique des chaînes précambriennes. Que ces lignes soient une expression de reconnaissance envers lui et son Pays!

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20.

ON THE FABRIC HABIT OF FIBROUS CELESTITE

BY

OTTO MELLIS

1. THE TEXTURE OF FIBROUS CELESTITE

The peculiar properties of the fibrous celestite of Dornburg near Jena, Germany, have long aroused the interest of investigators and this celestite is that most described in mineralogic literature.

Already in the year 1899 O. Mügge came to the conclusion that the seemingly parallel-fibrous aggregate is not formed of fibers, but of coarse about 2—10 mm great individuals with well developed cleavage. The boundaries of these individuals do not coincide with fiber faces and their orientation to fiber axis is various. This made Mügge think that the aggregate must be a pseudomorph after an unknown fibrous mineral (perhaps gypsum). Fibrous celestite samples of Frankstown, Pennsylvania have — after Mügge — the same properties.

Linck (1926) and Linck & Noll (1928) examined the fibrous celestite of Jena again and found the cone-in-cone texture. Both investigators tried to explain cone-in-cone texture from Mügge's point of view — as pseudomorph after some original structure. They agreed with Mügge as to the celestite being pseudo-fibrous, but they did not fully deny the presence of a certain fibrous character. It seemed to them that the fibers are — in a section cut perpendicular to the direction of fibers — allotriomorphic, marble like, sometimes idiomorphic.

In connection with my studies on the fibrous gypsum ten years ago, I investigated the cone-in-cone texture of fibrous celestite from Latvia, which occurs together with fibrous gypsum (Mellis, 1937 [1]).

Already a slight examination of thin sections cut parallel or perpendicular to fiber axis shows clearly that the fibers of celestite whose cone-in-cone formations are to be found in fibrous gypsum of Latvia are allotriomorphic. The diameter of fibers varies considerably from about 0,05—0,2 mm. By closer examination under the microscope fibers are usually seen to be made up of still smaller fibers having optically nearly the same

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[1] In this paper must be: in cone-in-cone formations cone-side and cone-basis have an angle of 60—70° (but not of 30—40° as it is misprinted).

orientation (fiber bundles). This is to be seen because of the undulatory extinction, which appears best in thin sections cut parallel to fiber axis.

It is peculiar to observe that besides the undulatory extinction in each bundle of fibers the side-by-side fibers sometimes have continually decreasing or increasing extinction angles. In turning the object stage extinction position my be seen like waves moving over great areas of thin section. Now and then quite large parts of fibrous celestite are changed into single-crystals which have a simultaneous extinction. In such cases the extinction has seemingly no connection with the primary structure (f. i. formely bent fibers have straight extinction).

Microscopic examination of fibrous celestite of Jena reveals generally the same features, except that of the undulatory extinction, which is not so easily observed. The observations of Mügge and Linck may be therefore explained more precisely: actually, the celestite of Jena shows often both fibrous and pseudofibrous (granular) development as is evident from Fig. 1. In the inside of cone-in-cone formations (the upper and lower part in Fig. 1) are celestite fibers. The part in the midst contains celestite grains. It is notable, that although such a texture may be seen in thin sections, in a fracture only fibers are to be seen even in the middle of celestite layer. In a thin section cut perpendicular to the direction of fibers from the fibrous part of celestite (inside of cone-in-cone) the fibers show allotriomorphic boundaries and a seemingly irregular optical orientation to each other (Fig. 2).


Fig. 1. Celestite of Dornburg near Jena. Cut parallel to fiber axis. Well developed cone-in-cone structure is to be seen and once existing central parting (upper part of phot.). Magnif. 5, 5 ×, only polarizer.

2. ORIENTATION OF CELESTITE FIBERS

As the described celestite is parallelfibrous and belongs to the so called cross-fiber typy, everything here mentioned about orientation relates to the celestite fabric habit.

Mügge's (1899) conception that fiber axes of celestite are orientated irregularly against the crystallographic directions was assumed by Linck and

Noll. Further, they observed that extinction direction is sometimes parallel to cone sides. Linck has the opinion that (001) cleavage plane shows an angle at about 52° to layer walls, which fact allowed him to make some conclusions about the replacing of gypsum substance by $SrSO_4$.

In 1933 Jansen carried out X-ray investigations of fibrous celestite of Jena. On examining the rotating crystal X-ray photographs of fiber bundles he found that no layer lines are present and therefore no special orientation could be ascertained. This might be due to the coarseness of fibers taken into consideration. He continued to find out the possible orientation of fiber bundles and suggested at least that each bundle had a special photograph, different from all others. This made him think: »Damit war schon die vollständige kristallographische Unorientierung des vorliegenden Fasercölestins bewiesen» (Jansen, 1933). Single fiber bundles were located: ⊥ (001), ⊥ (002), ⊥ (210) or ⊥ (303) parallel to fiber axis.

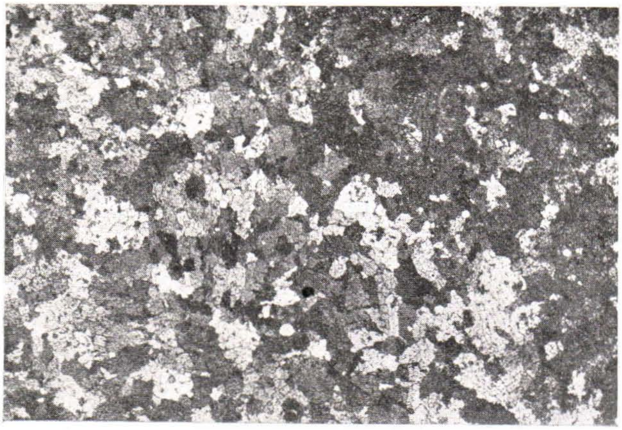

Fig. 2. Celestite of Dornburg near Jena.
Cut perpendicular to fiber axis. Magnif. 24 ×, crossed nicols.

One celestite sample of Girgenti, Sicily, examined by Jansen showed well developed orientation of fiber axis parallel to a-axis.

In 1937 the present writer published a preliminary report on the optical examinations of fibrous celestite from Navessala, Latvia (Mellis, 1937). He observed that the fiber axis of celestite has a tendency to orientate parallel to γ respectively to a-axis. Later he had the opportunity to determine the orientation of celestite fibers in some other samples from Latvia. At the same time he also proved the orientation of fibrous celestite from Dornburg, Jena.

The measurements were carried out by means of five-axis U-stage (after Emmons).

Celestite belongs to the orthorombic system and the crystallographic orientation is clearly defined by determining the position of indicatrix, because $\alpha = c =$ [001], $\beta = b =$ [010] and $\gamma = a =$ [100].

Fig. 3 shows that the cone-in-cone formations of Latvian celestite contain fibers which have a tendency to orientate with (001) plane parallel to the direction of fibers, for most of the fiber axes lie in a broad zone about 20° (40°) perpendicular to α respectively to c-axis. The concentration of poles is the greatest at [100]. More delicate accumulations of poles at ⊥ (210) and [010]. Single poles lie near ⊥ (101).

The results concerning the fibrous celestite of Jena (Fig. 4) are:

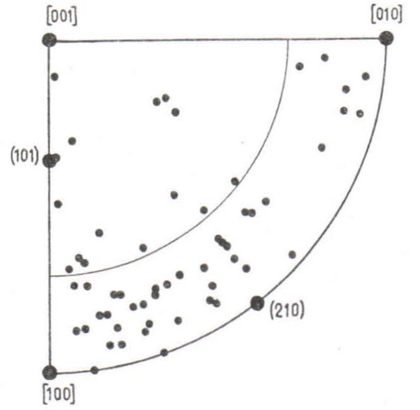

Fig. 3. Celestite of Latvia. Orientation of fiber axes according to fixed optical indicatrix. Stereographic projection.

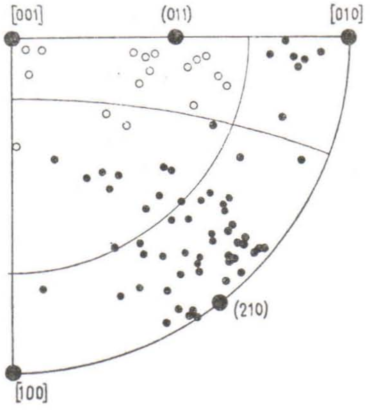

Fig. 4. Celestite of Dornburg, Jena. Orientation of fiber axes according to fixed optical indicatrix. Filled circles — fibers, open circles — coarse crystals. Stereographic projection.

a. *Fibers* (in the cone-in-cone formations). Fiber axes show a remarkable tendency to orientate parallel to (001). Most of the fiber axes are placed in a zone about 20° (40°) perpendicular to c-axis. Distinct accumulation of poles is to be seen at ⊥ (210). Single poles are at [010] but none at [100].

b. *Coarse crystal*-grains (region between cone-in-cone formations in the midst of layer). These grains are so coarse that a normal thin section contains only few of them; therefore it was possible to measure only 15 such. But even then the tendency to locate with (100) plane parallel to the original fiber axis is to be seen. Most of the poles are placed at ⊥ (011), single poles are at [001] but none at [010].

These results show clearly that in spite of the conception of Mügge, Linck, Noll and Jansen the fibrous celestite has an evident tendency to orientate in certain crystallographic directions. This tendency is here even better evolved than in fibrous gypsum.

The directions of fibers mentioned by Jansen are all to be found on the diagrams reprinted here; this, too, shows that the directions determined by the present writer are not fortuitious.

3. THE ORIGIN OF THE FIBROUS CELESTITE

The fibrous celestite of Jena is, according to the interpretation of all previous investigators, a pseudomorph after fibrous gypsum. Linck (1926) tries to explain the cone-in-cone formations from this point of view, but the conclusions drawn by him are not clearly defined and satisfactory. Not to carry the discussion too far I will simply add that the conception of celestite being a pseudomorph after gypsum (not to mention the origin of cone-in-cone texture which problem I hope to be able to discuss in the nearest future) does not explain some important realities.

For instance, how can we imagine the permeating of Sr-bearing solutions in gypsum layers from a chemical point of view? It seems to be natural that these solutions, if they were present at all, would by touching calcium sulphate have precipitated celestite. Or if Sr-bearing solutions have been present, how did the cone-in-cone formation take place? It is characteristic for the celestite cone-in-cones of Latvian fibrous gypsum that they lie spread abroad and have no connection with one another. The Sr-solutions could have formed a continuous pseudomorph celestite layer. And — one characteristic reality — the cone-in-cones never go over the central parting. The central parting which is to be seen in a fracture as a delicate coloured line and distinguished from the other parts of gypsum layer because of different orientation of gypsum fibers, parallel to [010], this line could not be an obstacle for pseudomorphs.

During my studies on fibrous gypsum I came to the perception that the crystallization of fibrous gypsum is beginning at the central parting, crystals growing on both sides in the direction from central parting to wall (see also Taber (1918) and Richardson (1920)), The growth can be defined as bilateral-symmetric. The celestite cone-in-cones lie symmetrical to the whole process of growth and therefore they must have been formed syngenetical with fibrous gypsum. Even the quantity of $SrSO_4$ in Latvian fibrous gypsum (Mellis, 1937) is adequate to that which could be evolved by changing anhydrite into gypsum and be precipitated as celestite.

The peculiar orientation of celestite fibers may be due to recrystallization without changing the chemical composition of the original structure of growth, which is to-day not preserved or only partly seen. Such a suggestion seems to be more plausible and simple. No doubt, the original structure of growth is here changed, as may be seen from the investigations of Mügge, Linck, Noll and the previous writer. But the pseudomorphous character of fibrous celestite is not established.

Linck (1926 p. 486) has made attempts to show how the pseudomorphs of fibrous gypsum have occurred: the (001) plane of celestite space lattice coincided with (101) plane of gypsum lattice, whereby the a-axis of celestite was covered with b-axis of gypsum. Let us admit with Linck (Linck, 1926, p. 484) that the fibrous gypsum is orientated with fiber axis parallel to c-axis. In such a case b-axis of gypsum fibers lies perpendicular to fiber axis. If Linck's suggestion were correct, a-axis of the new-formed celestite fibers would have to lie perpendicular to fiber axis. Further, the fiber axis would have to lie in the (100) plane of celestite, but the investigations made until now do not establish this presumption. Only the celestite grains have such a position. Celestite fibers, as seen from Figs. 3 and 4, do not lie in (100) plane but in (001).

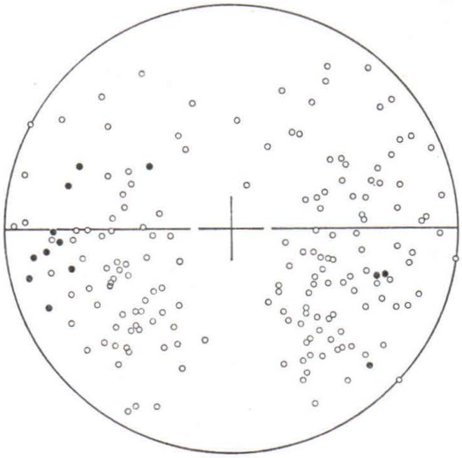

Fig. 5. Orientation of b-axes of gypsum fibers (open circles) and a-axes of celestite fibers (filled circles). Fiber axes lie parallel to EW direction. Sample from Saurieši, Latvia. Stereographic projection.

A thin section was made of the Latvian fibrous gypsum which included fibrous celestite parallel to the direction of fibers in order to examine the orientation of fiber axes for both minerals. As Fig. 5 shows, gypsum fibers have a tendency to orientate parallel to their b-axis. Celestite fibers, although only few in this thin section are approachable for measurements are inclined to lie with fiber axis parallel to a-axis.[1] This is the case which Linck assumed — b-axis of gypsum to be parallel to a-axis of celestite — except that both axes are lying parallel and not perpendicular to fiber axis.

The seemingly regular orientation of gypsum and celestite fibers to each other may be due to various obstacles (f. i. single directions being

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[1] The coincidence for both axes is reality more complicated. The results of measurements of gypsum fibers are published in an other paper (Mellis, 1947) and show maximums for this section at $\perp (\bar{1}31)$, $\perp (1\bar{2}1)$, $\perp (120)$ and $\perp (130)$. Single celestite fibers are placed nearly $\perp (210)$.

preferred during growth) and has not necessarily to be explained as a consequence of pseudomorphs.

The fact that a well orientated fibrous aggregate may be changed into a comparatively irregular one with coarse fibers and grains might be called in question. But the measurements of Beckenkamp (1890) show how the growth of great porphyroblastic gypsum crystals in fine fibrous gypsum is taking place — their c-axis showing angles at 30°, 40° and 70° to fiber axis (these data are obtained by means of graphical construction from Fig. 5a — 7a by Beckenkamp). The recrystallization process during which a fine aggregate is changed into a coarser one is not necessary to form well orientated aggregates. The new crystals due to recrystallization are arranged in such manner that their orientation can be observed only as statistical phenomenon.

Several times I had the opportunity to observe how a finely fibrous aggregate is changed into a coarse-grained one showing several stages of the recrystallization process. First aggregates with nearly equal individuals are formed. During the recrystallization new and coarse-grained individuals are formed, the single individuals having still better orientation till at last one single crystal arises of all. This is well to be seen in the celestite from Latvia (fiber bundles with wave-like extinction and coarse single crystals).

The diagrams reprinted in this paper greatly resemble recrystallization diagrams for metals (Burgers, 1941).

There are types of fibrous gypsum to be found in their outward appearance and structure very like the one of Jena, which reveal in fracture the fibrous structure and are really recrystallized to a coarse-grained crystal aggregate, but there is no evidence to assume a pseudomorph.

The change of original structure by means of recrystallization is a reality to be taken more into consideration than has been done as yet. It is possible that most of the fibrous aggregates reveal a changed original structure of growth and that their orientation does not correspond with the primary one. Experimentally this was shown with fibrous silver and copper by R. Schenk and his cooperators (R. Schenk, R. Fricke & G. Brinkmann, 1928).

This brief study I dedicate to Professor Dr. Pentti Eskola on his 65th birthday as a special expression of esteem and gratitude.

UPPSALA, MINERALOGISK-GEOLOGISKA INSTITUTIONEN, FEBRUARY 1947.

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21.

ON THE METAMORPHISM OF THE SCHISTS IN THE YLÖJÄRVI AREA

BY

AHTI SIMONEN and K. J. NEUVONEN

CONTENTS

INTRODUCTION

The Ylöjärvi area on the western side of Lake Näsijärvi belongs to the classical Bothnian schist areas in the Tampere field described thoroughly by Sederholm (1897). In the year 1945 the Geological Survey of Finland began a new, more detailed geological mapping in the Tampere schist area. The reason for the new survey was stated to be the ore occurrences met with in the region. In the summer of 1945 the present authors mapped the Ylöjärvi area, in which they were assisted by Mr. Erkki Viluksela, to whom the authors express their most sincere thanks.

During the years 1945—46 the present authors carried out laboratory studies of the rocks in the Ylöjärvi area and in this paper they will give only some main characteristics of the metamorphism of these schists. Especially the principles of mineral facies classification presented by Eskola (1915, 1921 and 1939) will be applied.

Concerning the petrographical characteristics of the rocks in the Ylöjärvi area the present authors refer to the investigation by Sederholm (1897). Sederholm applied succesfully the actualistic method to the Archaean metamorphic schists in the Tampere field. In the work he describes many beautiful relic textures in the rocks in which the primary horizontal layers show today a vertical position.

BASIC SCHISTS

Basic tuffitic rocks (porphyrotoids according to Sederholm) are widely distributed, especially in the northern part of the Ylöjärvi schist area (Map I). Basaltic lavas represented by uralite and plagioclase porphyrites occur in considerably less degree. In the work mentioned above (pp. 64—81) Sederholm gives a voluminous description of the relic textures of the basic volcanics in the Tampere area, showing the blastoporphyritic texture of the originally basaltic lavas and the inhomogeneous texture of the tuffitic rocks, in which can be observed the banding of the different layers. The material of the tuffitic schists has been predominantly volcanic ash, which has often been influenced by flowing water. In many cases it has been intimately mixed with a clayey weathering material. Also there occur in the Ylöjärvi area some calcareous schists as narrow bands lying in between the layers of the other supercrustal rocks.

The most common mineral association of the basic volcanics in the Ylöjärvi area is:

(1) hornblende-biotite-plagioclase (An_{30})

corresponding to the amphibolite facies of Eskola. In some cases biotite occurs as an alteration product of hornblende, especially in schists which have been influenced strongly by tectonic movements. Chlorite occurs only in minute part as an alteration product of hornblende.

In the basic schists of the Ylöjärvi area it can often be observed that the anorthite component of plagioclase, stable in the amphibolite facies, has begun to alter into epidote. The plagioclase thus contains inclusions of epidote and the hornblende has altered in minute part into chlorite and epidote. We thus have the mineral composition:

(2) hornblende-(biotite)-(chlorite)-epidote-plagioclase (An_{15-20}).

This kind of mineral composition represents a transitional form between mineral parageneses typical of the amphibolite and epidote-amphibolite facies.

In some cases basic plagioclase has altered into epidote and albite, causing the mineral assemblage:

(3) hornblende-chlorite-epidote-albite.

This mineral association is characteristic of the epidote-amphibolite facies. In some thin sections biotite can be observed as an instable relic.

In the Ylöjärvi area there also occur basic schists which do not contain hornblende, it having altered into chlorite, giving rise to the mineral paragenesis:

(4) chlorite-epidote-albite.

This mineral association is characteristic of the greenschist facies. The alteration to this mineral paragenesis has occurred in connection with the extraction of CaO from silicate minerals. The occurrence of calcite is very common in this type. Metasomatic processes have also occurred, pyrite always being present in the chlorite schists.

In some cases metasomatic processes at a low temperature have played a predominant role, causing sericitization of basic volcanogeneous rocks. A gradual transition from chlorite schists into sericitic rocks can be observed. Sericitization stands in close connection with the sulphide occurrences met with in the Ylöjärvi area. In this present volume Saksela (1947) describes these ore minerals in greater detail.

Among the basic schists in the Ylöjärvi area there also occur some calcium-rich varieties in which can be observed the mineral association:

(5) diopside-hornblende-plagioclase (An_{35}).

The most characteristic feature of these rocks is a well developed granoblastic texture, lacking all signs of its primary mode of occurrence. The high CaO-content makes it probable that these rocks have originally been calcium-rich sediments. These diopside bearing rocks are met with only in the boundary zones near the granodiorites.

Furthermore there occur, as small beds in other supercrustal rocks, calcareous schists showing the mineral paragenesis:

(6) epidote-biotite-plagioclase (An_{10-20}).

Muscovite occurs in some cases and chlorite is observed as an alteration product of biotite. The above-mentioned mineral association shows a similarity to the epidote-biotite schist facies of Vogt (1927) in the Sulitelma region corresponding to the greenschist facies.

The distribution of the above-mentioned mineral assemblages is presented in Map II, showing the predominant occurrence of the basic schists belonging to the amphibolite facies. The basic schists represented by the mineral parageneses corresponding to those of the epidote-amphibolite and greenschist facies occur only in a limited area. The occurrence of epidote-biotite schists in an area in which there occur predominantly rocks showing a mineral association characteristic of the greenschist and epidote-amphibolite facies gives one the idea that in calcareous schists the combination epidote-biotite is stable in the PT-conditions of the low-temperature facies.

Table I. Basic metamorphic schists in the Ylöjärvi area and their metasomatic derivates.

	1		2		3		4		5	
	%	Mol. prop.	%	Mol. prop.	%	Mol. prop.	%	Mol. prop.	%	Mol. prop.
SiO_2	50.98	8454	50.10	8308	48.81	8094	57.44	9526	74.79	12403
TiO_2	0.59	74	2.18	272	1.28	160	0.68	85	0.68	85
Al_2O_3	13.56	1327	18.50	1810	17.39	1702	13.40	1311	13.69	1339
Fe_2O_3	1.85	116	4.69	294	4.30	269	7.08	443	1.24	78
FeO	10.58	1472	7.30	1016	6.93	964	1.26	175	0.11	15
MnO	0.24	34	0.26	37	0.21	30	0.14	20	0.01	1
MgO	7.42	1840	2.30	570	4.36	1081	2.05	508	0.31	77
CaO	9.20	1640	8.08	1440	8.43	1503	5.27	940	2.25	401
Na_2O	2.56	413	2.32	374	1.43	231	1.43	231	0.52	84
K_2O	0.84	89	1.90	202	0.52	55	2.38	253	3.30	350
P_2O_5	0.33	23	0.28	20	0.28	20	0.44	31	0.26	18
CO_2	n. d.		n. d.		3.02	686	2.27	516	n. d.	
S	n. d.		n. d.		n. d.		4.82	1510	0.55	172
H_2O+	2.17		2.38		2.65		1.54		1.97	
H_2O-	0.05		0.09		0.10		0.05		0.02	
	100.37		100.38		99.71		100.25		99.70	
									—O = S_2 = 0.14	
									99.56	

Mineral associations:

Hornblende	—	—	—	—
Chlorite	Chlorite	Chlorite	Chlorite	—
Epidote	Epidote	Epidote	Epidote	—
Albite	Albite	Albite	Albite	—
Quartz	Quartz	Quartz	Quartz	Quartz
—	—	—	Sericite	Sericite
—	—	Pyrite	Pyrite	Pyrite
—	—	—	—	Fluorite
—	—	Calcite	Calcite	—

1. Uralite porphyrite. N. of Keijärvi, Ylöjärvi. Anal. H. B. Wiik.
2. Basic tuffitic schist. Parosten Kaitajärvi, Ylöjärvi. Anal. H. B. Wiik.
3. Chlorite schist. S. of Hirvijärvi, Ylöjärvi. Anal. Pentti Ojanperä.
4. Chlorite-sericite schist. Ahdepää, Ylöjärvi. Anal. H. B. Wiik.
5. Sericite schist. Ahdepää, Ylöjärvi. Anal. H. B. Wiik.

In Table I there are presented some chemical analyses of the basic schists in the Ylöjärvi area and their metasomatic derivates.

Anal. 1 was made from a uralite porphyrite in which the primary basic plagioclase had altered into epidote + albite and primary pyroxene into actinolitic hornblende + epidote + chlorite. The chemical analysis clearly shows a basaltic composition similar to the plateau basalts.

Anal. 2 shows the chemical composition of a basic tuffitic rock. This kind of rock is very common in the Ylöjärvi area. The analysed specimen was rich in aggregates of epidote and chlorite which were probably pseudomorphs after hornblende.

Anal. 3 was made from a chlorite schist which was rich in calcite.

Anal. 4 represents the metasomatic derivate of the chlorite schist of anal. 3 originated through sericitization.

Anal. 5 was made from the last product of the process of sericitization. Anal. 3—5 show the alteration of primary composition through metasomatic sericitization, as has occurred in the schists NNW. of the church village, Ylöjärvi, near Ahdepää.

ACID VOLCANICS

A new characteristic of the Ylöjärvi area is the rich occurrence of acid volcanics represented by pyroclastic and quartz-porphyritic types. Texturally these rocks show in many cases great similarity to the volcanogeneous leptites of Southwestern Finland described earlier, especially by Eskola (1914) in connection with the Orijärvi region.

Fig. 1. Vitrophyric texture in acid volcanogeneous rock. Lielahti, Ylöjärvi.

More coarse-grained types of acid volcanogeneous rocks are observed in the boundary zones near the granodiorites. Generally the acid volcanics are fine-grained, types similar to the so called »hälleflinta» also occurring. Also one vitrophyric texture relic is observed, characterized by occurrence of concentric zones (Fig. 1). This kind of texture originates according to Rosenbusch (1910) through crystallization of acid volcanic glass. Banding of different layers is characteristic of the pyroclastic types. Blastoporphyritic textures containing phenocrysts of quartz and feldspar occur only in minute part.

Table II. Acid volcanogeneous rocks in the Ylöjärvi area.

	6		7		8		9		10	
	%	Mol. prop.	%	Mol. prop.	%	Mol. prop.	%	Mol. prop.	%	Mol. prop.
SiO_2	69.08	11456	66.95	11103	72.29	11988	75.02	12441	57.18	9483
TiO_2	0.38	47	0.67	84	0.18	22	0.15	19	0.72	90
Al_2O_3	13.99	1369	15.40	1507	14.31	1400	12.84	1256	18.96	1855
Fe_2O_3	1.88	118	0.00		1.32	83	0.45	28	2.16	135
FeO	1.48	206	3.42	476	0.97	135	1.48	206	3.10	431
MnO	0.13	18	0.06	8	0.03	4	0.04	6	0.08	11
MgO	0.39	97	0.76	188	trace		0.42	104	1.80	446
CaO	2.27	405	1.13	201	0.92	164	1.20	214	4.76	849
Na_2O	3.48	561	3.40	548	7.72	1245	6.00	968	4.74	764
K_2O	4.28	454	7.20	764	0.82	87	1.20	127	3.94	418
P_2O_5	0.56	39	0.33	23	0.30	21	0.23	16	0.40	28
CO_2	1.23	280	n. d.		n. d.		n. d.		n. d.	
H_2O+	0.65		0.67		0.75		0.73		1.57	
H_2O-	0.02		0.07		0.03		0.26		0.13	
	99.82		100.06		99.64		100.02		99.54	

Main mineral associations:

Quartz	Quartz	Quartz	Quartz	Quartz
Anorthite	Anorthite	—	Anorthite	—
Albite	Albite	Albite	Albite	Albite
Biotite	Biotite	—	Biotite	—
Muscovite	—	Muscovite	—	—
—	—	Chlorite	—	Chlorite
—	—	—	—	Epidote
Microcline	Microcline	—	—	Microcline
—	—	Calcite	—	—

6. Quartz porphyry. Farm Härkilepo, Ylöjärvi. Anal. H. B. Wiik.
7. Potash-rich porphyry. Lielahti, Ylöjärvi. Anal. H. B. Wiik.
8. Soda-rich »hälleflinta». Lepomäki, Ylöjärvi. Anal. H. B. Wiik.
9. Fine-grained acid volcanic rock. Kiviniemenlahti, Ylöjärvi. Anal. H. B. Wiik.
10. Keratophyre. E. of Mastosjärvi, Ylöjärvi. Anal. H. B. Wiik.

Chemically the acid volcanics show generally a rhyolitic composition (anal. 6 in Table II). It is remarkable, however, that there occur also soda and potash extreme types (anal. 7—9 in Table II), soda extreme types being predominant. Furthermore, there occur in the Ylöjärvi area lava beds showing a keratophyric bulk composition (anal. 10 in Table II).

In the metamorphic schists, which have been originally acid volcanics, the occurrence of mineral parageneses corresponding to different mineral facies is not as easy to observe as in the basic schists. The mineral association:

(1) biotite-(muscovite)-plagioclase (An_{10-30})

is most common among the acid volcanics corresponding to the PT-conditions of the amphibolite facies.

As a transitional form between the amphibolite and epidote-amphibolite facies, there occur, among the acid volcanic schists, types in which chlorite occurs as an alteration product of biotite and in which also the anorthite component of plagioclase has to some extent altered into epidote.

In some cases biotite has completely altered into chlorite and plagioclase has altered into albite + epidote, giving rise to the mineral paragenesis:

(2) epidote-chlorite-(muscovite)-albite.

According to Eskola (1939) this kind of mineral paragenesis is possible both in the epidote-amphibolite and the greenschist facies.

In acid volcanogeneous rocks there has also occurred metasomatic sericitization, accompanied by increase in sulphides. Chlorite has altered into sericite and the feldspars have become instable. The amount of free quartz has increased in many cases during the process of sericitization.

The distribution of the above-mentioned mineral parageneses of the acid volcanics in the Ylöjärvi area is presented in Map II. The greatest distribution is shown by the mineral association biotite-plagioclase. Mineral parageneses corresponding to PT-conditions similar to those of the epidote-amphibolite or greenschist facies occur generally in the same limited areas as the basic epidote amphibolites and chlorite schists.

PHYLLITES

The weathering sediments represented by phyllites commonly occur in the southern part of the Ylöjärvi area. Also phyllites often occur as narrow bands in the volcanogeneous rocks. The most characteristic feature of the phyllites is a well developed varved texture. Also mica schists in which the primary varved texture is not clearly observable can be met with in the Ylöjärvi area.

Table III. Chemical composition of the rocks occurring in the phyllite formation of the Ylöjärvi area.

	11		12		13	
	%	Mol. prop.	%	Mol. prop.	%	Mol. prop.
SiO_2	58.89	9766	77.00	12769	67.87	11255
TiO_2	0.86	107	0.42	52	0.58	72
Al_2O_3	20.39	1995	11.32	1108	14.99	1467
Fe_2O_3	1.16	73	0.88	55	0.56	35
FeO	6.01	836	2.02	281	3.53	491
MnO	0.06	8	0.04	6	0.04	6
MgO	2.04	506	0.68	169	1.25	310
CaO	1.20	214	1.52	271	2.88	513
Na_2O	2.00	323	2.84	458	4.36	703
K_2O	3.92	416	1.68	178	2.12	225
P_2O_5	0.19	13	0.14	10	0.17	12
H_2O+	3.31		1.05		1.25	
H_2O-	0.05		0.13		0.06	
	100.08		99.72		99.66	

Main mineral associations:

Quartz	Quartz	Quartz
Plagioclase	Plagioclase	Plagioclase
Biotite	Biotite	Biotite
Muscovite	Muscovite	—

11. Pseudoporphyroblastic varved phyllite. S. of Valkeekivi, Ylöjärvi. Anal. H. B. Wiik.
12. Coarse-grained part in a thick varve of phyllite. Niemen saha, Ylöjärvi. Anal. H. B. Wiik.
13. Fine-grained phyllite-like schist. Valkeekivi, Ylöjärvi. Anal. H. B. Wiik.

Great variations can be observed chemically in the phyllite formation. Unweathered tuffitic material is intermixed with the phyllites in many cases. In Table III there are presented 3 analyses giving an idea of the variations in the chemical composition of the rocks occurring in the phyllite formation.

Anal. 11 was made from a phyllite in which the thickness of the varves was only a few centimeters. This analysis shows a very typical clayey composition.

Anal. 12 was made from a coarse-grained part of a thick varve (thickness of varve about 1 m.). The great amount of free quartz is worthy of notice.

Anal. 13 was made from a dark phyllite-like schist which does not, however, show the characteristics of a typical clayey composition. Al_2O_3-excess is only very small and the amount of Na_2O is higher than that of K_2O. It seems probable that this rock consists mainly of a chemically unweathered tuffitic material.

In the boundary zones near infracrustal rocks there occur mica schists containing aggregates of sericite and chlorite. Apparently these aggregates are pseudomorphs after Al-rich minerals In these mica schists small relics of cordierite, andalusite and almandite, which have been greatly altered, can be observed only as rarities. The occurrence of the pseudoporphyroblastic mica schists along the contact zones near the intrusive rocks shows clearly that their origin stands in close connection with the thermic metamorphic action of the intrusions (Map I).

In the southernmost part of the mapped area the primary varved texture can only be seen very weakly. South of Lake Vihnusjärvi the phyllites gradually change into mica schists and mica gneisses. According to Sederholm (1897) the mica schists south of the phyllites are older than the Bothnian schists of the Tampere field. This opinion of Sederholm does not receive support from the new observations, which show only that recrystallization has been stronger in the southern part of the area, the ages of the schists being similar.

The porphyritic granodiorite north of Nokia is rich in inclusions of schists which have originally been clayey sediments. These schists are coarse-grained mica schists, but in some cases the primary stratification is weakly observable as a texture relic. According to microscopical studies one is able to see that these schists have at some time been porphyroblastic mica schists containing cordierite and almandite porphyroblasts. Cordierite has altered into aggregates consisting of chlorite and muscovite. In some cases almandite is present as small corroded grains in the rounded or six-angled aggregates which consist chiefly of chlorite, biotite and muscovite.

The greatest part of the phyllites and mica schists in the Ylöjärvi area show the mineral association:

(1) biotite-muscovite-plagioclase $(\mathrm{An}_{>10})$

which belongs to the amphibolite facies. Muscovite is lacking in some varieties.

As a transitional form between the above-mentioned mineral assemblage and that corresponding to the association characteristic of the epidote-amphibolite facies, there are types in which chlorite occurs as an alteration product of biotite and epidote as an alteration product of anorthite.

The disappearance of biotite indicates PT-conditions corresponding to that of the epidote-amphibolite facies. There then occurs the mineral paragenesis:

(2) muscovite-chlorite-albite.

This mineral combination is also stable in the conditions corresponding to those of the greenschist facies. Especially some sericite phyllites occurring in close connection with metasomatic sericite schists indicate a low-temperature facies.

The distribution of the above-mentioned mineral parageneses is presented in Map II, showing that the phyllites, represented by the mineral association muscovite-chlorite-albite, occur in the same limited area as the low-temperature facies of other supercrustal rocks.

CONCLUDING REMARKS

In the metamorphic schists of the Ylöjärvi area there can be observed mineral associations corresponding to the different mineral facies conditions. Equilibriums of the amphibolite, epidote-amphibolite and greenschist facies are met with.

The points representing the chemical analyses given above are plotted in the ACF-diagrams (Fig. 2) presented by Eskola (1939). The points plotted in the diagrams constitute a survey of the mineral parageneses met with in the Ylöjärvi area.

The distribution of different mineral parageneses in the Ylöjärvi area is seen in Map II, which shows that associations corresponding to the PT-conditions of the amphibolite facies are most predominant. The rocks represented by the mineral assemblages typical in the epidote-amphibolite and greenschist facies occur only in limited areas. Schists showing the mineral parageneses of the low-temperature facies occur most abundantly in a zone going through the schist area from the northern side of the copper mine at Paroinen to Lake Näsijärvi. The Paroinen copper mine is situated near the boundary zone between the low-temperature and amphibolite facies. The mineral associations most commonly met with in the mine correspond to those of the amphibolite facies.

In the Ylöjärvi area one is able to trace a gradual transition from the mineral associations of the amphibolite facies into that of the low-temperature facies. The instable relics met with in the rocks of the low-temperature facies clearly point out that at some time these metamorphic schists have also had mineral associations corresponding to the PT-conditions of the amphibolite facies.

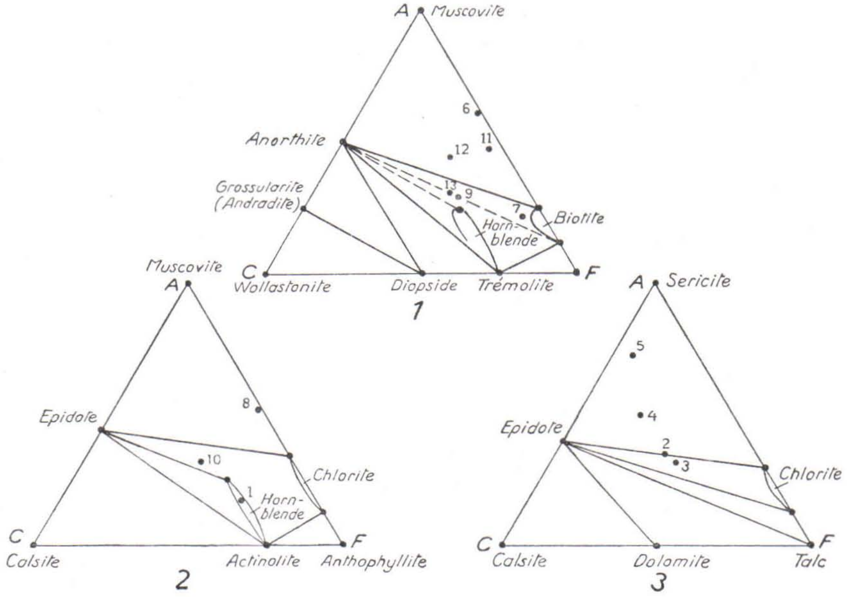

Fig. 2. ACF-diagrams. 1. amphibolite facies, 2. epidote-amphibolite facies and 3. greenschist facies. The numbers of the points refer to the chemical analyses presented above. The chemically studied rocks do not, however, represent all mineral assemblages met with in the Ylöjärvi area. The rocks showing the following mineral parageneses of the amphibolite facies have not been studied chemically: hornblende-biotite-anorthite and diopside-hornblende-anorthite. In plotting the points of the acid schists showing the association muscovite-chlorite-epidote, it has been difficult to decide whether they belong to the epidote-amphibolite or greenschist facies. From these rocks the authors have plotted only the sericitic schists occurring in connection with the chlorite schists in the ACF-diagram of the greenschist facies.

The most primary metamorphic facies of all schists in the Ylöjärvi area has been the amphibolite facies, but later a secondary decrease in temperature has taken place in some zones. It seems to the authors that this decrease in temperature stands in close connection with the metasomatic processes taking place at low temperature. Generally the schists represented by mineral parageneses of the low-temperature facies occur only in a narrow zone around the metasomatic sericite schists. The sericite schist at Ahdepää is surrounded by low-temperature facies covering a large area. At Ahdepää sericitization has been most effective and also among the sulphide minerals connected with metasomatism there occur Bi-tellurides, representing the low-temperature parageneses of ore minerals met with in the Ylöjärvi area (Saksela, 1947). It is worthy of note that in the Ylöjärvi area, based on the occurrences of silicate rocks of the low-temperature facies, it is possible to limit the areas in which the occurrences of low-temperature parageneses of ore minerals are most probable.

Some pseudoporphyroblastic mica schists can be observed in the Ylöjärvi area as relics of an earlier stage of metamorphism. This metamorphism is due to intrusions of granodiorites and diabases, showing that these infracrustal rocks are older than later phases of metamorphism. Apparently the above-mentioned intrusions are synkinematic intrusions of orogeny connected nowadays to the Svecofennian orogeny.

GEOLOGICAL SURVEY OF FINLAND, 1947.

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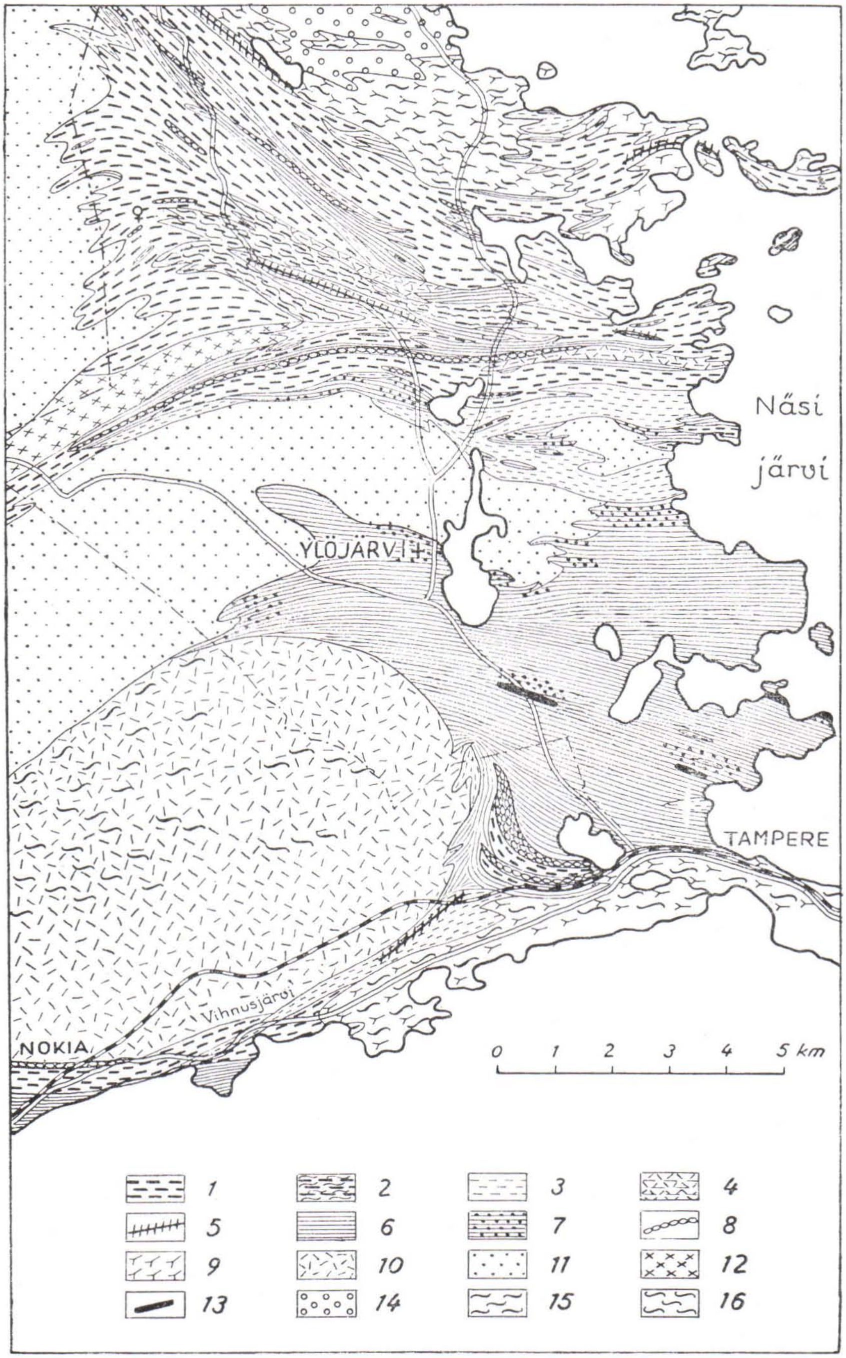
Map I.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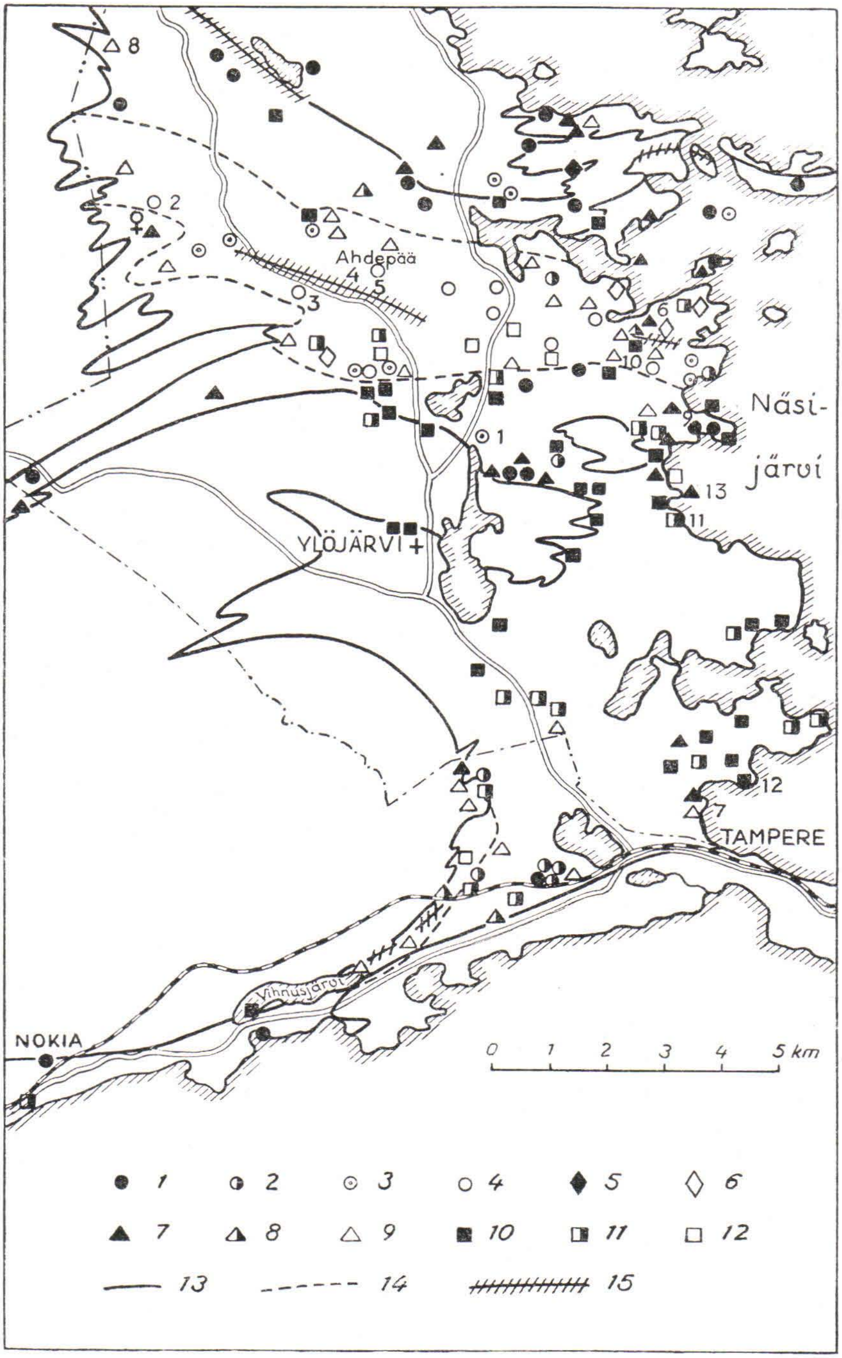
Map II.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EXPLANATIONS TO THE MAPS

Map I

Geological map of the Ylöjärvi area. 1 = basic schists, predominantly tuffites. 2 = chlorite schists. 3 = acid volcanogeneous rocks. 4 = keratophyres. 5 = sericite schists. 6 = phyllites and mica schists. 7 = pseudoporphyroblastic mica schists. 8 = conglomerates and agglomerates. 9 = gneissose granodiorites. 10 = porphyritic granodiorites. 11 = even-grained granodiorites. 12 = diorites. 13 = diabases. 14 = porphyritic microcline granites. 15 = rocks granitized by the influence of microcline granite. 16 = schist inclusions in the infracrustal rocks.

Map II

Distribution of different mineral associations in the schists of the Ylöjärvi area. Points plotted in map were determined by the microscopical studies of thin sections. Numbers refer to the chemical analyses given in Tables I—III.

1—6 mineral parageneses of basic schists
1 hornblende-biotite-plagioclase (An_{30})
2 transitional form between the amphibolite and epidote-amphibolite facies
3 hornblende-chlorite-epidote-albite
4 chlorite-epidote-albite
5 diopside-hornblende-plagioclase (An_{35})
6 epidote-biotite-plagioclase (An_{10-20})
7—9 mineral parageneses of the acid volcanogeneous rocks
7 biotite-muscovite-plagioclase (An_{10-30})
8 transitional form between the amphibolite and epidote-amphibolite facies
9 epidote-chlorite-muscovite-albite
10—12 mineral parageneses of phyllites
10 biotite-muscovite-plagioclase ($An >_{10}$)
11 transitional form between the amphibolite and epidote-amphibolite facies
12 muscovite-chlorite-albite
13 boundaries of the schist area
14 isogrades showing distribution of low-temperature facies.
15 sericite schists

22.

THE VÄHÄJOKI IRON ORE IN TERVOLA, NORTHERN FINLAND

BY

AIMO MIKKOLA

CONTENTS

INTRODUCTION

In the Summer of 1943 the Geological Survey carried out magnetic investigations in Vähäjoki district, Koivu village, Tervola parish. These were occasioned by reports to the Survey about strong compass disturbances in the said district. The investigations performed resulted late in the Summer of the same year in the discovery of a considerable area of magnetic disturbance. The cause of the disturbances was a magnetic iron ore, which seemed to be so promising that investigations were continued, both by drilling, and by thorough magnetical measurements, until the following autumn, when they had to be stopped, owing to force of circumstances. In the Summer of 1945 the author carried out the geological mapping of the area, where the field researches also were concluded. As the final result it was stated, however, that no practical ore exists on the area. Nevertheless, being situated in the Karelian formation the ore, though of little economic significance, is interesting from the geological point of view.

The area of magnetic disturbance is located about 8 km SE. of Koivu station in the NW. part of Vähäjoki settlement. The extent of the whole area is about 1 × 2 km, where there are eight separate disturbance maxima.

The ground of the disturbance area is low and watery swamp, where exposed rocks are scarce. This fact has made both the actual ore investigations and the geological mapping more difficult. The same features are common to the whole North-Bothnian schistose area NW. of Kivalovaarat. The ground is either low swamp, or moraine heaths, from which some lonely spur of rock sticks out. Exposures are further found only in beds of small rivers.

GENERAL GEOLOGICAL FEATURES

The Karelian schists of North-Bothnia form a consistent area between the Kemi and Tornio rivers. To the S. and SE. it is limited by pre-Karelian gneissose granite and to the N. by post-Karelian granite. Geologically the area has been very little investigated, and even the investigations performed are already mostly obsolete. (Hackman 1918, Mäkinen 1915). Later Hausen has treated the schist formation chiefly from stratigraphical and tectonical points of view (1936), while Väyrynen has dealt with the tectonics of the area in his unpublished paper (1945).

On the southern boundary the quartzites of the schist area lie directly on the old gneissose granite without bottom formations. Mäkinen, however, regards as such a weathering breccia in the neighbourhood of Kemi. The boundary is to be considered as exclusively tectonic. In their character the quartzites are mostly clastic, though there are even noticeable differences from the sericite-carbonate-cemented types down to the glassy ones, the colour changing at the same time, largely from white to very dark grey. The differences, however, are in the first place caused by the primary composition, and the nearness of penetrating igneous rocks and post-Karelian granite.

Above the quartzites there is a thick phyllite layer, which varies even more than the quartzites. The most unchanged phyllite is somewhat fine-grained dark grey bedded rock, the predominating minerals of which are quartz and biotite. There is astonishingly little feldspar, but a remarkable plenitude of magnetite. A carbon content is also common, although not as a general feature. The current bedding and transverse schistosity characterize these types. With growing metamorphism the phyllites are converted into mica- schists, where there sometimes are found contact minerals, for instance andalusite and garnet. On the N. border of the area the schists are turned into veinous gneisses under the influence of the post-Karelian granite.

In the phyllite there are often thinnish strata of very fine-grained amphibole schists, which differ from normal schists of erosion origin not only in appearance, but also in their mineral composition. Hausen calls them »green schists of phyllite formation». He has pointed out, that they presumably are originally pyroclastic sediments, and have been brought forth by subaqueous eruptions. The author, too, thinks this to be likely, as is shown later on (p. 267). From these pure amphibole schists we have to distinguish actinolite-micaschists, very similar to them, which have been originated by metamorphism from marl sediments, or as reaction skarns between phyllite and dolomite.

The changing of phyllite into dolomite happens through marl schists, or even so, that there appear thin dolomite layers, the number and thickness of which grow in stratigraphical series succession upwards. Wholly pure dolomites are generally infrequent. They always contain rather much quartz, and even actual layers of quartzite. In addition, there are mica minerals and talc in varying quantities. Dolomite contains an amount of skarn minerals. The commonest of which is tremolite-actinolite. Diopside is seldom to be met with and then only in E. and W. parts of the schist area. According to the stage of metamorphism the North-Bothnian calcareous rocks limestones belong chiefly to the first and the second dolomite groups of Eskola, viz. the quartz dolomites and the tremolite dolomites (Eskola 1927).

The sedimentary series mentioned above is penetrated by basic eruptives (metabasites of Hackman, lava rocks and ophiolites of Hausen). To these belongs a series of rocks of greally varying composition and structure, which have been metamorphosed chiefly in green stone facies. The primary composition varies from peridotitic to dioritic. Distinct plutonic rocks, diabases, feldsparporphyres, amygdaloidal rocks, lavas, and agglomerates can be told apart structurally.

Younger than all rocks mentioned above is the red microcline granite (post-Kalevian granite of Hackman). N. and E. of the schistose area it lies as a large coherent massif, forming migmatites on the border of sediments. S. of Kivalovaarat there is a smaller massif, which migmatitizes the pre-Karelian gneissose granite. Only pegmatitic veins of this granite are so far known on the actual schistose area.

The stratigraphical succession of sediments within the schistose area is revealed from above. The strike of the stratification of the quartzites on the bottom follows fairly exactly the contact of gneissose granite. The general strike is SW. — NE. and the dip on the bottom about 50° NW. Northward the position of strata is fairly steep. The quartzites nearest the bottom are foliated into thin, easily cleavable schists, the strike of which follows the bottom contact even more accurately than the stratification (Fig. 1). Northward the schistosity keeps the direction SW. — NE., at the same time as the strike of the bedding varies, resulting in the transverse schistosity characteristic of phyllites.

The metamorphism has generally been weak, and no big overthrust planes are met with. Instead of this small faults and joints are common.

In dolomite this appears as brecciation, which when weakest causes mosaic cracks, but can occur even as a distinct pressure breccia (Fig. 5). The strike of the folding axis, so far as it has been possible to measure on a slightly sloping folded area, is generally W. and SW., and the dip about 30°—60°. But also northern and north-western folding axes are of fairly common occurrence. The dip of these is generally steeper, about 60°—80°.

Fig. 1. Quartzitic schist. Strike N. 40 E. Dip 50° NW. Runkausvaara Tervola. Photo by P. Haapala.

MAGNETIC DISTURBANCES

The magnetic disturbances form a distinct zone parallel to the schistosity, but the individual disturbance maxima are, however, so far separated from each other, that they cannot be caused by the same source of disturbance (Fig. 2). In addition to this, there are on both sides of the main zone — though at the same time parallel to it — noticeable disturbances, which in no way are connected with same. Wholly different in its character is the maximum of disturbance (R, in Fig. 2), that lies SW. to the first-mentioned ones. Its length axis is parallel to the bedding, or N. 60° W. Even its form is not broken, on which grounds one could think it to be originated by a coherent ore body. I am sorry to state that no drillings have been carried out on this area, and no exposures exist, so that there is no knowledge as to the character of the cause of disturbance. The drillings performed on the area A (Fig. 2) verify the incoherence, which is to be expected on the grounds of the magnetic maps. The ore is situated in the contact zone of the dolomite and phyllite layers as small irregular accumulations, the continuity and position of which are impossible to conjecture, even with the aid of several drill holes.

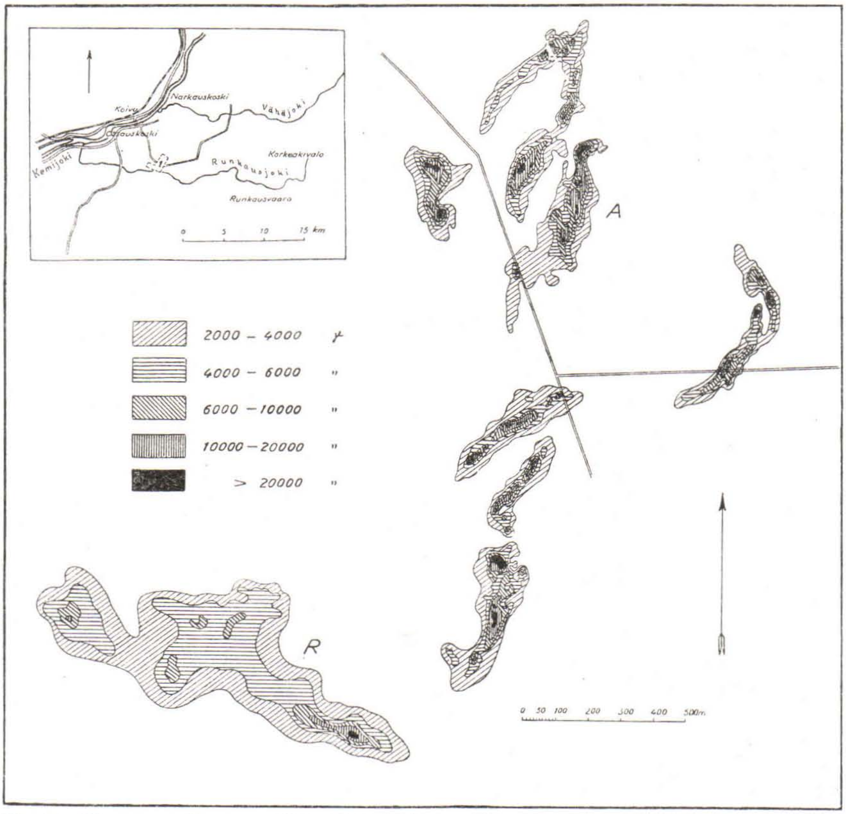


Fig. 2. The magnetic map of the Vähäjoki area. Strengths of the vertical components in gammas. In the small map 1 denotes the disturbance area (schematically).

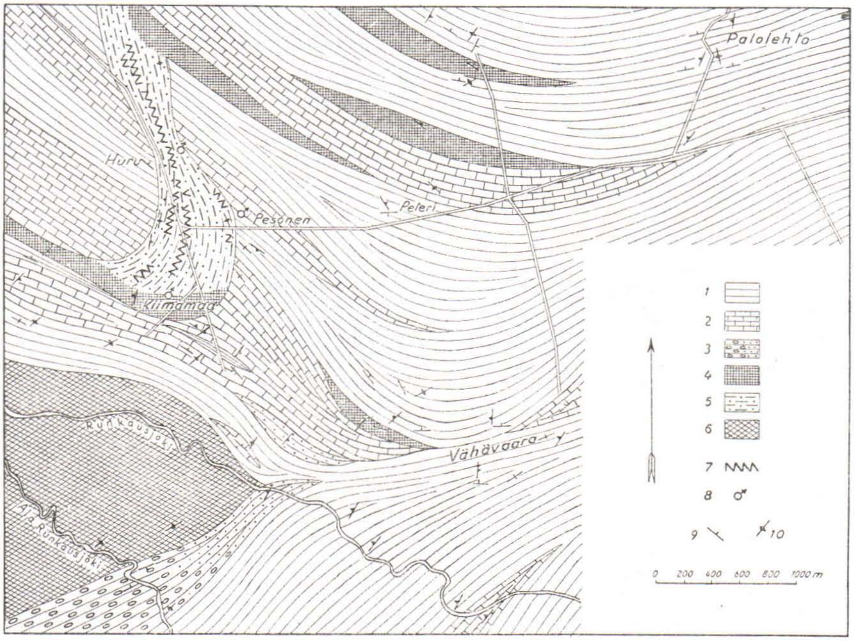


Fig. 3. The geological map of the Vähäjoki area. 1. phyllite, 2. dolomite, 3. quartzite and conglomerate, 4. skarn, 5. actinolite-micaschist, 6. basic eruptive rocks, 7. breccia, 8. ore, 9. stratification, 10. schistosity.

ROCKS

Sedimentary rocks

Dolomite. The orebuilding zone belongs to the dolomite horizon, where there is lively interbedding with phyllite. The unchanged dolomite, when pure, is pale yellow and very fine-grained. In addition to carbonate minerals it contains a little quartz and chloritisized biotite. Talc and tremolite are secondarily found on the cleft surfaces. Within the ore area the dolomite is never homogeneous; paler and coarser parts abound as fissures or patches. These are carbonate minerals, born secondarily from solutions having circulated in the dolomite. Among them we may mention calcite and, rarely, the pale pink rhodochrosite. Besides the inhomogeneity mentioned above, brecciation is characteristic of the dolomites of Vähäjoki. When weakest it is thin cracking, which gives a mosaic appearance to the surface. Chlorite and talc have filled up the cracks (Fig. 4). The following stage is a distinct press breccia (Fig. 5), which is met with at the E. end of Vähävaara about 2,5 km S. from the ore. The fragments in it are wholly unchanged fine-grained dolomite. In the fissures and the cement itself, nearest to dolomite is fine-grained magnetite and pyrite impregnate, which

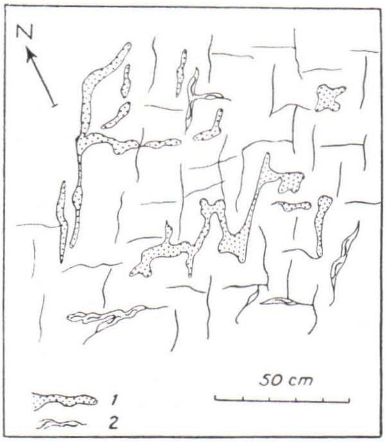
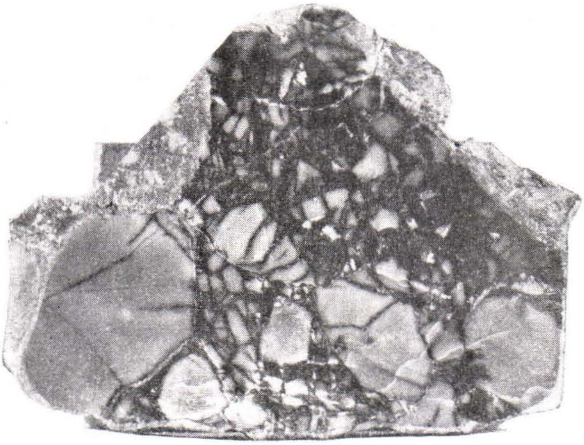


Fig. 4. Dolomite with mosaic cracks. 1. Quartz cracks, 2. Chloritic racks. Vähäjoki, Laitakangas.



Eig. 5.' Press breccia. The fragments dolomits, the cement carbonate, impregnated by magnetite. Vähäjoki, E. end of Vähävaara. $^4/_5$.

is the cause of the black colour. After this follows a calcite frame around the fragments and a quartz frame farthest out. The simultaneous occurrance of quartz and calcite shows in how low a temperature this brecciation has taken place. The biggest part of the dark matter, however, is fine-pulverized, unchanged dolomite, impregnated by magnetite. In the ore zone the dolomite fragments have been lengthened and flattened (Fig. 6). At the same time contact-metasomatical changes have taken place between dolomite and phyllite. As a result of these actinolite-micaschist has been formed, and this together with magnetite composes the cement. The original matter of the breccia has been a rock formed by thin phyllite and dolomite layers. It is often possible to follow along the original dolomite layer several meters by means of the fragments.

Anal. 1 in Table I represents the chemical composition of unchanged dolomites. It corresponds to 50,5 % $CaCO_3$ and 40.9 % $MgCO_3$. Analyses of fairly analogous dolomites appear in the explanation of the map by Hackman (1918).

Phyllite. Purely primary phyllite is very sparsely found in connection with the ore formation. All contain more or less secondary amphibole minerals. Immediately S. of the ore the dolomite shows interbedded phyllite, the thickness of layers varying from 2 to 50 cm. The mineral composition in the phyllite layers is as follows: biotite, plagioclase, quartz, and carbonate.

Fig. 6. Schistose dolomite breccia. Pebbles grey yellowish dolomite. Cement actinolite-micaschist with veinous quartz (white). — After photograph drawn by Mrs. Toini Mikkola. Vähäjoki, Laitakangas.

About 0,5 km E. from the ore, near Peteri, occurs stratification between pale green and dark green layers in schists belonging to phyllites. It refers to the tuffite layers mentioned before (p. 263). The pale green layers are in the field found to be distinctly banded, and the microscopical investigation points to erosion sediments. There is seen in the schist very fine-grained magnetite in rings, these being possibly of original cement. The primary mineral composition, certainly, has wholly disappeared. The schist contains plenty of amphibole, which according to the optical properties, determined by Mrs. Toini Mikkola, MA., corresponds to about 50 % ferrotremolite in the Table of Winchell. In addition, there is some quartz, carbonate, and tourmaline. The marl sediment should approximately corresponds to the original composition of the schist. In the thinner dark layers magnetite and also pyrite on the other hand appear as big corroded grains, or as fine-grained aggregates, lengthened parallel to schistosity, having, however, still clear crystal forms of magnetite. In these layers no sort of bandedness is found in the field. The rock is fairly pure, very fine amphibole mass, which according to its optical properties is 80 % ferro-tremolite.

The chemical composition of the phyllites and of the amphibole schists is represented in Table I, analyses 2 and 3. Of these, 2 is a banded, fairly unchanged erosion sediment. The SiO_2-content is unusually low for phyllites, and the amount of Fe high, but Al_2O_3 and K_2O on the other hand are typically high for erosion sediments. When counting normatively the surplus of Al_2O_3 is 5.53 %.

Table I

	1		2		3	
	%	Mol. prop.	%	Mol. prop.	%	Mol. prop.
SiO_2	—	—	48.40	0.8026	59.88	0.9930
TiO_2	—	—	3.41	0.0426	0.42	0.0052
Al_2O_3	0.30	—	15.30	0.1497	13.97	0.1367
Fe_2O_3		—	8.91	0.0558	2.24	0.0140
FeO	1.78	0.0248	7.04	0.0980	5.35	0.0745
MnO	0.03	0.0004	0.06	0.0008	0.16	0.0023
MgO	19.65	0.4873	5.38	0.1334	5.14	0.1275
CaO	28.32	0.5049	2.77	0.0494	4.29	0.0765
Na_2O	—	—	1.62	0.0261	6.75	0.1089
K_2O	—	—	4.66	0.0495	0.62	0.0066
P_2O_5	—	—	0.26	0.0018	0.26	0.0018
H_2O +	0.15	—	1.13	—	0.83	—
H_2O —	—	—	0.10	—	0.08	—
CO_2	43.58	0.9905	1.03	0.0234	—	—
insol	6.54	—	—	—	—	—
	100.35	—	100.07	—	99.99	—

1. Dolomite-rock. Huru, Vähäjoki, Tervola. Analyst. Pentti Ojanperä.
2. Banded phyllite. Palolehto, Vähäjoki, Tervola. Analyst. Pentti Ojanperä.
3. Amphibole-schist. Narkauskoski, Kemijoki, Tervola. Analyst. Pentti Ojanperä.

Anal. 3 shows amphibole schist interbedded in the original marl sediment. It has been taken from Narkauskoski, Kemi river. Its composition differs noticeably from the usual kind of phyllite. The most striking difference is in the amount of alkalies. Further, the amount of CaO is remarkable, and no excess of Al_2O_3 is found. The difference of composition when compared to the amphibole-micaschists born by reaction skarnbuilding is in the first place owing to the amounts of the alkalies and Al_2O_3. (Table III, anal. 4 and 5). Thus we can regard as definitely indicated by their chemical composition, that the dark and compact amphibole schist layers belong to volcanic sediments ,bedded simultaneously with the phyllites.

Quartzite. Quartzite is not found at all within the ore area, but 1 km S. of same there are a couple of small exposures. The rock in these is slightly reddish or white clastic quartzite, the cement being sericite and iron pigment. The conglomerate of Ukonköngäs, Ala-Runkausjoki river, described by Hackman, belongs already to the same horizon. He considered it to be a Jatulian bottom formation, but in fact it is an interformation conglomerate in the same series of sediments.

Basic eruptive rocks

NW. from the ore there is a wide area of basic eruptive rocks of different types. Nearest to the ore in the Runkausjoki river is found a slighly schistose amygdaloidal rock, the amygdaloids being of calcite, epidote, or chlorite. The rock is of the same type also in Ossauskoski of Kemijoki river, where it has a distinct schistosity N. 10 E., and dip about 70° W. In Ala-Runkausjoki river, N. of the conglomerate mentioned before, the rock has a distinct intrusive structure. Its original composition should in the first place correspond to diorite, but has been changed in the low metamorphism. NE. of the ore area, at a distance of about 4 km, there is the nearest exposure belonging to this group. The rock is here somewhat coarse uralite gabbro, or here and there uralite peridotite. Also in this range of exposures variations of amygdaloidal rocks are found nearer to Kemijoki river. — The basic eruptive rocks contain plenty of epidote, and sulphide grains in some quantity. The schistosity of these rocks is slight, but so far as it has been noticed, it usually is similar to that of the surrounding sediments.

Pegmatite veins

Of the post-Karelian granite there are met within the ore area only some pegmatite veins, penetrating sediments and basic eruptive rocks. The veins contain chiefly perthitic microcline, albite and quartz. The pure quartz veins of the same age are somewhat more common than the former ones. The nearest known post-Karelian granite massif is situated on Korkeakivalo, about 15 km E. from the ore.

Reaction skarn

Tremolite skarn. At Vähäjoki skarn-building is often met with in the contacts of dolomite and phyllite layers, forming sometimes remarkable layers of skarn. They are skarns, born in regional metamorphism between limestone and silicatic wall rock before actual migmatitization, as Magnusson has defined (1930). Later Ramberg, when descriping the metamorphic rocks of the Norwegian west-coast, has given more exact information about the petrology of reaction skarns. The mineral composition depends on pressure and temperature conditions, and on the mineral composition of wall rocks. All typical pneumatolytic minerals are lacking. This applies also to the Vähäjoki skarn. Almost pure Mg-tremolite appears there as predominant mineral (10—15 % Fe-trem.). Secondarily it has to a great degree been turned into serpentine and talc. Some amounts of epidote, titanite, chlorite and quartz occur in addition to tremolite. In pure skarns ore minerals are scace. Epidote is born in a low temperature as a reaction result between carbonate and the clay material of phyllite. According to Ramberg the reaction of kaolin: $3\,Al_2Si_2O_5(OH)_4 + 4\,CaCO_3 = 2\,Ca_2Al_3Si_3O_{12}OH + 4\,CO_2$ takes place already before muscovite and chlorite begin to react with calcite. In some skarn zones there are several big grains of plagioclase, the anorthite bearing of which is 60 % according to the determination of Mrs. Toini Mikkola. Ramberg has shown that in calcium-metasomatose plagioclase can not only turn into epidote, but even become richer in An. Such is the case, if in addition to the physical conditions favourable to the birth of basic plagioclase, also the pore solutions are rich in Al_2O_3. The high Al_2O_3 content of the phyllites makes it understandable, that this has been the case at Vähäjoki. Garnets, on the other hand, are entirely lacking in the Vähäjoki. The author has met with only one garnet crystal, 1 mm in diameter.

The chemical composition of reaction skarn comes out in Table II, anal. 7.

Actinolite-micaschist. The turning of amphibole skarn or dolomite into phyllite takes place through schist rich in mica. The mineral composition of such a schist differs entirely both from the phyllites and from the skarn. As relics of phyllite in less changed parts quartz and feldspar grains are found. Actinolite porphyroblasts on the other hand resemble skarn. Other minerals are biotite, or chlorite, epidote, quartz, calcite, and apatite, and plenty of ore minerals. Secondary changes have happened in actinolite in such manner, that the borders have been turned into talc, or the grains are here and there broken and corroded, leaving only torn remnants. Epidote appears, as well as actinolite, in well crystallized porphyroblasts. The predominant mineral of these schists, however, is biotite-chlorite. The unchanged biotite is very dark brown, showing high iron content. The main part of it is chloritized. All stages of transition between biotite and chlorite are now met with. The turning into talc of actinolite grains is connected with the same phase of low metamorphism as the chloritization mentioned above.

Actinolite-micaschist forms the matter cementing dolomite and phyllite fragments in the breccia of the ore zone (Fig. 6). In some cases reaction seam can be seen around the dolomite fragments, in the same way as Hackman described in the case of a loose block (1928). Nearest to the unchanged dolomite is a frame of coarse-grained calcite-actinolite, the grain size decreasing outwards.

Magnusson has called »sööls» the changing results of leptites rich in mica and connected with reaction skarns, and »söölskarn» the transition between sööl and skarn (1930). In Långban the söölbuilding is a metasomatic changing between on the one hand leptite, rich in potassium, and on the other hand skarn, dolomite or ore. The final result depends on the primary components. The building of Vähäjoki actinolite-micaschist can be compared to this process, although the conditions and original matter were different. When comparing the composition of the unchanged phyllites to chlorite schist and to actinolite-chloriteschist, regularity is to be found (Table II). The analyses are given in the Table II according to the transition stages. Anal. 1 (the same as 2 in Table I) represents unchanged phyllite outside the ore area, 2 and 3 different stages of chloritization types of phyllite in the same phyllite layers, in the wall rock of ore, and 4, 5, and 6 actinolite-micaschists in the ore building zone. The latter analyses contain magnetite and some sulphides. When comparing in the first place the proportions of phyllite and of chlorite schist, the amount of alkalies ($Na_2O + K_2O$) and Al_2O_3 is found to be smaller in the chloritized schist than in the unchanged phyllite. Their mutual proportion (molecularly

Table II

	1		2		3		4		5		6		7	
	%	Mol. prop.	%	Mol. prop.	%	Mol. prop.	%	Mol. prop.	%	Mol. prop.	%	Mol. prop.	%	Mol. prop.
SiO_2	48.40	0.8026	42.52	0.7051	44.72	0.7416	31.92	0.5293	27.92	0.3630	15.66	0.2597	47.00	0.7794
TiO_2	3.41	0.0426	1.36	0.0170	1.10	0.0137	0.83	0.0104	0.70	0.0087	0.24	0.0030	1.19	0.0149
Al_2O_3	15.30	0.1497	16.92	0.1656	12.52	0.1225	11.45	0.1120	9.14	0.0894	1.96	0.0192	12.21	0.1195
Fe_2O_3	8.91	0.0558	3.27	0.0205	6.17	0.0386	13.84	0.0867	22.96	0.1438	17.70	0.1108	2.14	0.0134
FeO	7.04	0.0980	13.05	0.1816	14.30	0.1990	14.62	0.2035	19.37	0.2696	8.33	0.1159	12.93	0.1800
MnO	0.06	0.0008	0.06	0.0008	0.06	0.0008	0.09	0.0013	0.04	0.0006	0.17	0.0024	0.28	0.0039
MgO	5.38	0.1334	10.52	0.2609	12.86	0.3189	13.13	0.3256	10.92	0.2708	8.69	0.2155	7.64	0.1895
CaO	2.77	0.0494	0.42	0.0075	0.06	0.0011	1.32	0.0235	0.38	0.0068	22.56	0.4028	8.08	0.1440
Na_2O	1.62	0.0261	0.49	0.0079	0.27	0.0043	0.36	0.0058	0.21	0.0034	0.48	0.0077	1.41	0.0227
K_2O	4.66	0.0495	4.98	0.0529	0.49	0.0052	5.73	0.0608	2.36	0.0250	0.08	0.0008	0.19	0.0020
P_2O_5	0.26	0.0018	0.71	0.0050	0.77	0.0054	0.25	0.0018	0.16	0.0011	0.43	0.0030	0.23	0.0016
H_2O+ ...	1.13	—	5.32	—	6.88	—	3.03	—	5.25	—	2.04	—	2.61	—
H_2O- ...	0.10	—	0.07	—	0.04	—	0.09	—	0.19	—		—	0.09	—
CO_2	1.03	0.0234	—	—	—	—	—	—	—	—	16.48	0.3746	4.07	0.0925
CuO	—	—	—	—	—	—	0.41	0.0052	—	—	—	—	—	—
S_2	—	—	—	—	—	—	2.89	0.0451	0.82	0.0128	7.40	0.1158	—	—
$-O=S_2$.	—	—	—	—	—	—	0.72	—	0.20	—	1.85	—	—	—
	100.07	—	99.69	—	100.24	—	99.24	—	—	—	100.37	—	100.07	—

1. Banded phyllite, Palolehto, Vähäjoki, Tervola. Analyst. Pentti Ojanperä.
2. Biotite-schist. Drillhole 9, » » » » »
3. Chlorite-schist. » 9, » » » » »
4. Actinolite-biotiteschist. Drillhole 5 Vähäjoki, Tervola. Analyst. H. B. Wiik.
5. Actinolite-chloriteschist. » 2 » » » » »
6. Actinolite-schist. » 5 » » » Pentti Ojanperä.
7. Tremolite-skarn. Huru » » » » »

counted) has changed for the benefit of Al_2O_3, and this fact shows the decreasing of alkalies to have been stronger than that of Al_2O_3. A similar transport of the material has, according to Magnusson, taken place in the söölbuilding zone, to which in the first place corresponds the chlorite-schist. When comparing chlorite-schist to actinolite-chloriteschist (anal. 5), which is a transition when passing on towards the dolomite, the amount of MgO is found to be reduced in the latter, while the sum of the alkalies has grown. There is a smaller amount of Al_2O_3 than in the chlorite-schist, but proportionately also here there is more Al_2O_3 than in the phyllites. Thus the transport has also in actinolite-micaschist taken place to the benefit of Al_2O_3. In the total amount of iron a continous enrichment is to be found when going from phyllite to actinolite-chloriteschist. The transport of iron has thus occurred from the phyllite into the more changed matter, or actinolite-chloriteschist. The conditions there have been favourable to the birth of magnetite, and iron has stopped here as magnetite. At the same time as there is only a fair half of same in the form of silicatic iron. The case is just the reverse in chlorite-schist and biotite-schist, there being manifold amounts of silicatic iron. In skarn the iron appears almost entirely as silicate. The transport of potassium, on the other hand, has taken place down from the dolomite, but it has stopped chiefly in skarn. The actinolite and chlorite part contains but little of same SiO_2 decreases continuously when passing from the unchanged phyllite into actinolite-chloriteschist, where the decrease has been strongest. SiO_2 has passed into the skarn in order to form silicates.

We can conclude from the above, that the transport of materiel by reaction skarn building of phyllite and dolomite has taken place so, that into the transition zone MgO and CaO have come from the side of carbonate, and Fe alkalies, and Al_2O_3 from the side of phyllite. SiO_2, necessary to the birth of silicates, can be traced to the phyllites, although plenty of it exists also in the dolomites.

ORE TYPES

At Vähäjoki magnetite occurs as almost the sole ore mineral. Sulphides are found in such small quantities that they have no significance, although at place there is considerable in impregnation of iron pyrite. This is the case especially in actinolite-chloriteschist and in phyllite. On the other hand chalcopyrite is quite occasional. The different types of ore cannot be designated according to their ore minerals, but according to, which of the rocks contains magnetite.

Breccia ore. In the border parts of the ore-building and the transition zone the magnetite has intruded along the cracks of dolomite, as has been pointed out in connection with dolomite. Breccia of this kind can be originated even by dissolving, as Tiberg mentions when treating Långban (1931). He calls it pseudo-breccia. But at Vähäjoki this is not in question, the brecciation is here distinct breaking-brecciation. The intrusion of magnetite has occurred in low temperature, because no contact effect between it and dolomite is to be seen (Fig. 7).

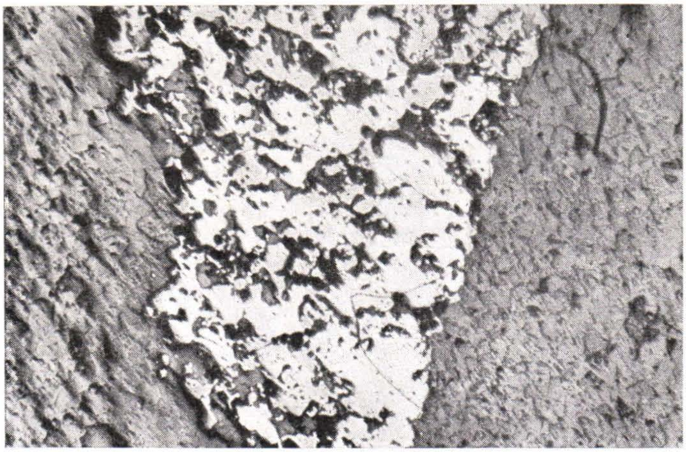


Fig. 7. The contact of magnetite (pale) agains dolomite in the breccia ore. Pol. sec. ord. light., 40 ×.

On the edges of the dolomite fragment a narrow ring of magnetite is often to be found, this being the same kind of impregnate as the interstitial magnetite. This type of breccia contains only very little sulphides. Pyrite appears in large, well developed crystals, which have been broken in some degree. In the cracks either small lamellae of chalcopyrite or magnetite grains are found. The latter are rather fine-grained and well-grown, showing the crystallization to have occurred in quiet conditions. After the crystallization, however, slight crushing has taken place in magnetite grains, at which stage calcite or talc has crystallized in the cracks.

Actinolite-micaschist and tremolite-skarn ore. At Vähäjoki the ore-building forms its main type in zones most strongly changed, viz. in actinolite-micaschist. There are remnants of the changed dolomite fragments, the matrix of which is formed by the schist mentioned above. Magnetite follows the schistosity surrounding the dolomite fragments already mentioned. Only seldom has it impregnated them. There is found distinct zonarity around the fragments, in such manner that large idiomorphic actinolite crystals are nearest to the dolomite. Around these and in the cracks appear pyrite and chalcopyrite. The former has beautiful crystal forms, which are entirely lacking in the latter. After this comes magnetite, and in the same way as dolomite breccia, also this shows a considerable quantity of crystal forms.

Magnetite occurs in actinolite-schist, too, as a very fine powderlike impregnation in the chlorite part, forming a sort of bedding together with actinolite (Fig. 8). The impregnation is found either evenly through the chlorite part, or often in small aggregates, which have been lengthened in the direction of the schistosity. At the same time magnetite in the actinolite part occurs in distinct, well developed grains, crystallized after actinolite, and fairly unaffected by the movements. This is due to the fact, that there are two kinds of magnetite. The fine-grained one has been in the original phyllite layers. As is pointed out before (p. 272) the coarse magnetite has at the same time come from the chlorite zone.

Fig. 8. The contact of fine-grained chlorite-schist and course actinolite-schist, with magnetites of different ages. Pol. sec., ord. light. 40 ×.

At the south edge of the ore area is seen a tremolite fels as an exposure, containing strong magnetite impregnate. At places there is to be seen feeble parallel structure, which could be primary bedding. But usually the impregnation has gone evenly through the whole rock. Magnetite occurs also here as ore mineral, appearing well developed in the same way as in actinolite-micaschist. Hematite is found as a new mineral in tremolite supplanting it.

Phyllite-impregnate ore. Magnetite occurs in phyllite, outside the transition zone, as an even, fine-grained impregnate, which entirely follows the original bedding (Fig. 9). Phyllite is compact, and rather unchanged. Secondary silicates are wholly lacking. Biotite is partly chloritized, and feldspar, so far it occurs, is sericitized. Magnetite is fine-grained, in diameter about 0.01—0.02 mm, and it has even more developed forms than actinolite-micaschist. Tectonic movements are shown by microfolding and by small joints, which intersect the magnetite layers. The cracks, born at the same time, have been filled with quartz. The impregnate ore is insignificant, as according to drillings it is restricted to the phyllite nearest to the transition zone.

Minor magnetic disturbances of the environment. Minor magnetic disturbances are often seen in the environs of the ore area. They are all located in phyllite. The longitudinal direction of the disturbances is exclusively the same as the strike of bedding. Magnetic indications are caused by the slight magnetite content of the phyllite, contemporary to the sedimentation, as has been mentioned in connection with phyllite.

Fig. 9. Fine-grained magnetite impregnate, parallel to the original bedding in phyllite. Pol. sec., ord. light. 75 ×.

CHEMICAL COMPOSITION

The iron content (soluble in HCl) of the Vähäjoki ore varies in the actinolite-micaschist type 25—35 %. In the dolomite part magnetite in places is also very compact. The specimen, taken from the exposure dug out near to Pesonen, contains the following percentages:

Fe (HCl) 57.5 %, Fe (total) 60.2 %, Ti 0.26 %, V 0.00 %, P 0,075 %.

The magnetite content of the phyllite impregnate ore is even feebler than that of the actinolite-micaschist type. The Fe-amount (HCl) is about 25—28 %. Magnetite can, however, even here be concentrated into layers of various thickness, the Fe-amount rising remarkably. An example of this is the following analysis from the drill core, the sample of which has been taken from a magnetite layer 3 m in thickness. Fe (HCl) 41.9 %, S 0.59 % P 0.036 %. Specific gravity 3.3. The averages of sulphur and phosphor in all analyses are S 1.04 %, and P 0.031 %. The Mn- and Cu-amounts are very insignificant. The average of Cu in places rich in sulphides is found to be about 0.1 %. From the ore of the dolomite type two total analyses have been performed (Table III). When counting normatively the magnetite amount in the former is 66.35 %, and in the latter 74.22 %. In the latter hematite is found 5.35 %. The TiO_2 amount is remarkably smaller than that of the sediments in the environment, but P_2O_3 is of the same order.

Table III

	1		2	
	%	Mol. prop.	%	Mol. prop.
SiO_2	22.18	0.3678	13.74	0.2278
TiO_2	0.21	0.0026	0.14	0.0017
Al_2O_3	0.71	0.0069	0.42	0.0041
Fe_2O_3	45.83	0.2870	56.54	0.3535
FeO	25.82	0.3594	23.13	0.3219
MnO	0.05	0.0007	0.02	0.0003
MgO	3.09	0.0766	3.48	0.0863
CaO	0.96	0.0171	0.66	0.0118
Na_2O	0.14	0.0023	0.18	0.0029
K_2O	0.15	0.0016	0.12	0.0013
P_2O_5	0.18	0.0013	0.32	0.0023
H_2O+	0.56	—	0.88	—
H_2O-	0.07	—		—
	99.95	—	99.64	—

1. Magnetite ore in tremoliteskarn, Kiimamaa. Vähäjoki, Tervola. Analyst. Pentti Ojanperä.
2. Magnetite ore in dolomite, Pesonen. Vähäjoki, Tervola. Analyst. Pentti Ojanperä.

Mr. Oiva Joensuu, M. A., has determined spectrographically from the different ore types described before the Mn_2O_3- CrO_2-and V_2O_5-content (Table IV). The determinations have been made with a Zeiss Three Prism Glass Spectrograph. In magnetite, enriched by an electric magnet, the error can be about 20 % of the amount of the material to be determined. The determination has been made by measuring the length of the line.

Table IV

	Magnetiteore			Magnetite			
	Mn_2O_3	CrO_2	V_2O_5	Mn_2O_3	CrO_2	V_2O_5	
1.	0.05	0.001	0.005	0.032	0.0035	0.030	1—2 Compact magnetite ore
2.	0.03	0.005	0.01	—	—	—	
3.	0.5	0.005	0.005	—	—	—	3—6 Dolomite-breccia ore
4.	0.5	0.003	0.005	0.1	0.010	0.035	
5.	0.3	0.001	0.005	—	—	—	7—16 Aktinolite-chlorite-schist ore
6.	0.3	0.005	0.005	—	—	—	
7.	0.3	0.003	0.03	—	—	—	17—18 Tremolite-skarn ore
8.	0.5	0.01	0.03	—	—	—	
9.	0.3	0.01	0.03	—	—	—	19 Phyllite ore
10.	0.1	0.01	0.03	—	—	—	
11.	0.03	0.03	0.1	—	—	—	
12.	0.03	0.03	0.1	—	—	—	
13.	0.03	0.03	0.03	0.2	0.04	0.08	
14.	0.5	0.01	0.03	—	—	—	
15.	0.03	0.005	0.01	—	—	—	
16.	0.03	0.01	0.01	—	—	—	
17.	0.1	0.003	0.01	—	—	—	
18.	0.1	0.003	0.01	0.2	0.0045	0.01	
19.	0.1	0.03	0.03	0.06	0.06	0.07	

Mn_2O_3, CrO_2 - and V_2O_5 - content spectrographicelly determined by Mr. O. Joensuu M. A., with a Zeiss Three Prism Glass Spectrograph.

The amounts of unenriched ore specimens have been determined by visually comparing the strength of the lines (Rankama, Joensuu 1946). The results can be at most either three times too great, or too small. The occurrance of Cr and V in separate ore types or rather in the sedimentary wall rock of these, corresponds entirely to their geochemical character in weathering and bedding. Cr and V are present most abundantly in hydrolysate sediment and least in carbonates. The richness in Mn of carbonates is caused by rhodochrosite born in metamorphism, which has been found to exist there. According to Landergren (1943) these low values of ferrides (Ti, V, Cr, Mn) are peculiar to non-magmatic or palingenetic iron enrichments.

GENESIS OF THE ORE

The Vähäjoki iron ore is located in the contact zone between the phyllite of sediment series and dolomite, where remarkable contact-metasomatic changes have taken place. Rocks belonging to the basic eruptives in the nearest vicinity are to be met with as magmatic rocks. The pegmatite and quartz veins of post-Karelian granite have so insignificant a size and are so scarce, that they cannot be considered as the carriers of ore.

When examining the chemical composition of the ore we see, that it does not fit into any group of magmatic oxide iron ores, viz. titan iron ores, chrome iron ores, ore magnetite ores rich in phosphor. The geochemical character and the lack of suitable intrusives point to a non-magmatic or palingenetic origin. Bearing at the same time in mind that neither in the transition zone, nor in skarn, actinolite-chlorite, and chlorite-schist are minerals met with, showing any metasomatic changes to have taken place in a pneumatolytic or hydrothermalic way, nothing supports the supposition of magmatic ore. On the other hand many facts speak on behalf of the origin of the sedimentary ore. In the first place the unchanged phyllites of the area are very rich in iron. Magnetite occurs in them as layers several millimeters thick, in conformity with the original bedding of phyllites. These cause noticeable magnetic indications, as has been seen. Remarkable enrichment of iron has happened during the sedimentation. The iron must have been sedimentated in a chemical way, because magnetite appears in the original clay sediment. The occurrence of the heavy magnetite as a mechanical weathering product in the very finest sediments bedded in water ought not to have been possible. Magnetite is even in the present loose sediments met with only in coast sands. The whole amount of iron can hardly be derived only from the pre-Karelian rock ground. Iron has surely come forth in connection with the volcanic activity contemporary with the sedimentation. On the other hand, we have been able to show with analyses, that in the reaction skarn building between phyllite and dolomite, the iron in phyllite has become mobile, and has been moved farther from it (p. 272). The conditions have been favourable for the birth of magnetite in a transition zone of actinolite-chloriteschist, and the iron has stopped there. We cannot consider the Vähäjoki iron ore to have been born palingenetically, because the metamorphism prevailing in the area is so weak, that it cannot have brought the iron back to the magma-stage.

The only intrusive massif know with in the ore area or in its immediate vicinity, is the diorite between the Runkausjoki rivers. It belongs to the same series of basic eruptives as the amygdaloidal rocks nearest to the ore. The brecciation is connected with this eruption, as is common for instance in the Karelides of Kainuu. The possibility of tectonic breccia is present, too, because strong differences in the directions of the folding axis are stated on both sides of the brecciation zones. The ore-building in the transition zone has taken place in the stage of the Karelian folding, because the crystallization of magnetite has occurred in peaceful conditions, possibly in connection with the intrusion of the post-Karelian granite.

SUMMARY

The North-Bothnian schists belong to the Karelides. They are bordered on the S. side by pre-Karelian gneissose granite, and on the N. and E. sides by post-Karelian microcline granite. The continuation of the schists to W. is unknown so far, because the geological map of the Swedish side

is not available. Both weathering and volcanic sediments are present in the North-Bothnian area. The first-mentioned form a continuous series from quartzite to dolomite; the bottom conglomerate, however, is missing. Basic eruptive rocks are younger than the sediments, to which belong both deep-seated and superficial rocks. All of these have been metamorphosed in the green stone facies.

In the Vähäjoki reaction-skarn building and contact metasomatic changing between phyllite and dolomite, have taken place and this has caused transportation of material in the transition zone. Ca and Mg have come from the side of the dolomite; transportation of the alkalies, Al, Fe and SiO_2 has occurred on the side of the phyllite. The result of this is the gradual changing of the phyllite into chlorite schist, and actinolite-chlorite-schist. In the last mentioned zone the conditions have been favourable for the birth of magnetite, and the iron has stopped there, chiefly being bound to ore minerals.

The situation of the ore in the reaction skarn zone, its geochemical character and geological environment show that the ore is not of magmatic origin. Primarily the iron has precipitated into the phyllites, when they have been bedding as a chemical sediment. At the final stage of folding the iron has been crystallized as magnetite in the special conditions prevailing in the ore-building zone. Magnetite occurs in unchanged phyllites as layers of varying thickness parallel to primary bedding, and is causing remarkable magnetic indications.

COMPARISON TO MAGNETITE ORES OF KARELIDES

In the Karelian formation only a few iron ores are known, and all of these, — being insignificant as regards size and content — have remained without more detailed investigation. Some known hematite occurrences need not be dealt with at all in this connection. In Finnish Karelia two magnetite deposits occur, in one of which, at Kirjokallio, Ilomantsi parish, magnetite appears as layers of mechanical sediment in quartzite — hornblende schist. At Havukkavaara, Kontiolahti parish, a small magnetite ore is to be met with, also this in quartz schist. According to Väyrynen the ore is situated in the contact of intrusive amphibolite, and he considers it to be one of the striped iron ores.

The only known more remarkable iron ore in Karelian formation is at Tuomivaara, Sotkamo parish. It was investigated in 1920 but no consistent treatise is available about some. The ore appears in garnet »Strahlstein» schist, being connected with quartzites. The ore appears to be somewhat striped, and contains phosphor, in abundance differing thereley from the Vähäjoki ore. The high garnet content shows a metamorphism thronger than that at Vähäjoki, but the geological features, however, are the same. As most similar to Vähäjoki is to be considered the magnetite occurrence of Kivesvaara, Paltamo, which is situated in folded and brecciated quartzite. In its mode occurrence it is as irregular as the Vähäjoki ore. — There are in Kainuu several other magnetic disturbance areas, which have been stated to be caused by magnetite. Magnetite is mostly met with in layers in the striped quartzite, which alternates with amphibole schists.

As is apparent from the examples mentioned above, the sedimentary magnetic ores in the Karelian formation are fairly common. All of them are, it is true, so small, that they have not lead to detailed investigations or mining activity; and so they are as yet not sufficiently thoroughly known.

I desire to express my best thanks to Mrs. Toini Mikkola, M. A., who has performed the mineral determination with U-stage, and has together with Mrs. Helvi Vasara, M. A., translated the manuscript, and also to Mr. Oiva Joensuu, M. A., for the spectral analyses.

GEOLOGICAL SURVEY, APRIL 1947.

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23.

PETALITE, A MINERAL NEW TO FINLAND

BY

TOINI MIKKOLA and H. B. WIIK

DISCOVERY AND OPTIC PROPERTIES

(TOINI MIKKOLA)

For more than a hundred years the pegmatite occurrences in Tammela parish have been known for their rare minerals. The foundation of Åvik glass works in 1748 caused closer investigations of pegmatites in the neighbourhood to be undertaken. Many finds were made and they could supply the new industry with raw material. These pegmatites, their geology and richness in rare minerals have been comprehensively treated by E. Mäkinen (1).

In August 1945 Mr. L. Suvenmaa sent a piece of white mineral to be examined in Helsinki. The person to whom it was sent, who was not a geologist, supposed it to be plagioclase. The specimen was forwarded to the present writer for the determination of its An-content. The low refractive indices soon revealed the fact that some new substance had been immersed in the liquids.

$$\alpha = 1.504 \qquad \beta = 1.510 \qquad \gamma = 1.516 \qquad 2V\gamma = 83^\circ$$

The colour of the Li- flame was to be seen after dissolving pulverized material in Na_2CO_3. These properties showed the material to be petalite, not earlier met with in Finland.

For the sake of comparison, the same determinations were made with Varuträsk petalite from Sweden, described by P. Quensel (2). The specimen for this was kindly lent by Prof. A. Laitakari. The optic properties were much the same.

$$\alpha = 1.504 \qquad \beta = 1.509 \qquad \gamma = 1.516 \qquad 2V\gamma = 84^\circ$$

While examining the mineral I made an interesting discovery (Fig. 1). One grain showed clear lamellar structure very similar to that of the plagioclases. The values of α' and γ' were very low and $\gamma'—\alpha'$ small, $\alpha' \parallel$ lamellae. This made me think that the slice represents a 010 section where α and β are horizontal. The twinning plane might be 001, thus agreeing both with the Manebach and the Albite twinning of plagioclases.

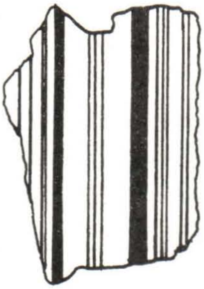
Fig. 1. Twinning lamellae in petalite.

The chemical composition and the conclusions made by means of analyses are treated by Mr. Wiik.

In the summer of 1946, I had the opportunity to visit the place where petalite occurs, Luolamäki hill, Koivula village, Somero parish, together with Prof. Laitakari. The occurrence was on the whole very small, only a few big crystals, at its highest 30—40 cm in diameter. They were in an old quarry in disuse for


Fig. 2. The old quarry at Luolamäki. Standing Mr. L. Suvenmaa, the sender of the first specimen.


Fig. 3. Clevelandite and **muscovite. $^2/_3$ of natural size.**

more than a hundred years. The entrance was so low that we had to creep into it. (Fig. 2). Close to the petalite crystals were quartz, clevelandite (Fig. 3), microcline, muscovite, biotite, beryl, serpentine and tourmaline. The biggest single crystal of petalite is represented in fig. 4.

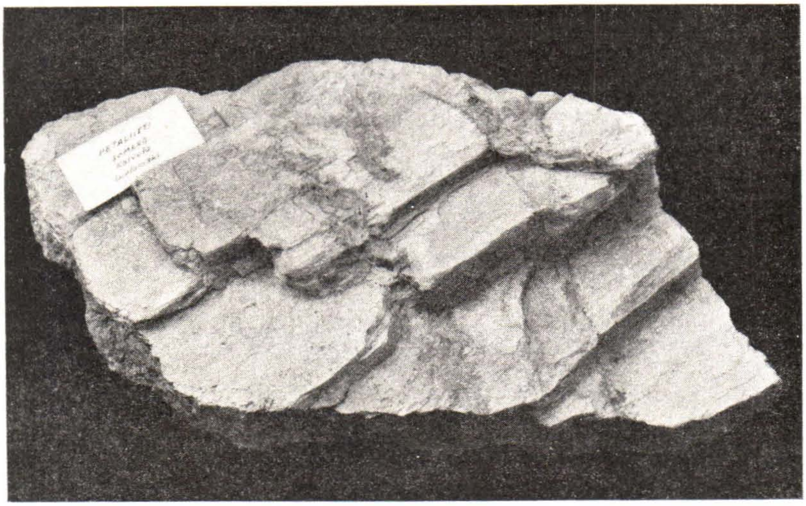

Fig. 4. The biggest petalite crystal.

CHEMICAL PROPERTIES

(H. B. Wiik)

The petalite from Somero in S. W. Finland has been analysed and the result is to be seen in the table below. For the sake of comparison a specimen of the petalite from Varuträsk, Sweden, was analysed too. These two new analyses and all analyses of petalites hitherto published (except a couple of old inaccurate specimens) are given in the table.

	1.	2.	3.	4.	5.	6.	7.	8.	9.	10.
SiO_2	77.47	77.52	77.18	76.91	78.68	76.19	77.87	77.95	77.90	77.29
TiO_2	0.00	0.00	0.00	—	—	—	—	—	—	—
Al_2O_3	17.12	16.88	17.05	16.85	16.62	16.48	17.53	16.63	15.85	16.95
Fe_2O_3	0.24	0.20	0.60	—	0.09	0.21	—	0.62	0.51	tr.
MnO	tr.	tr.	0.00	—	0.004	tr.	—	—	—	tr.
MgO	0.00	tr.	0.00	—	0.00	0.54	—	0.21	0.26	—
CaO	0.13	0.12	0.14	0.27	—	—	—	—	—	—
Li_2O	3.95	4.14	4.11	4.15	4.13	3.72	2.77	3.74	3.52	2.62
Na_2O	0.78	0.34	0.61	0.73	0.08	0.36	1.04	0.48	0.53	2.39
K_2O	0.30	0.24	0.17	—	0.00	0.18	0.43	—	—	tr.
P_2O_5	tr.	tr.	0.00	0.31	—	—	—	—	—	—
F	—	—	0.03	—	—	—	—	—	—	—
H_2O+	0.25	0.21	0.44	0.84	0.01	1.04	0.34	0.60	0.70	1.03
H_2O-	0.03	0.06	0.10		0.00	1.22				
	100.27	99.71	100.43$^{1)}$	100.06	99.614	99.94	100.00	100.23	99.27	100.28
Sp. gr.	2.412	2.404	2.418	—	2.41	2.38	—	—	—	—

$^{1)}$ in original 100.03

1. Petalite, Somero, Finland. Anal. H. B. Wiik.
2. Petalite, Varuträsk, Sweden. Anal. H. B. Wiik.
3. Petalite, Varuträsk, Sweden. Anal. Thelma Berggren (2).
4. Petalite, Utö, Sweden. Anal. K. Sondén Geol. För. Förh. 6. 39. 1882 (from Doelter).
5. Petalite, Londonderry, Australia. Anal. H. P. Rowledge (3).
6. Petalite, Ibid. Anal. C. R. Le Mesurier (3).
7. Petalite, Elba. Anal. C. F. Rammelsberg. Monatsberg. Akad. Berlin 1878. (From Doelter).
8. Petalite, Bolton, Massachussets. Anal. L. Smith and G. J. Brush. Ann. J. Sci. 1.373. 1853, (From Doelter).
9. Petalite. Ibid.
10. Petalite, Peru, Maine, U. S. A. Anal. F. W. Clarke. Bull. geol. Surv. U. S. A. 149. p. 257, 1910. (From Doelter).

As will be seen, the petalites from Sweden and Finland are very similar in composition. One could almost say that they form a class of their own of definite composition. This is furthermore to be seen from the following tables, where the ratios R_2O:R_2O_3:SiO_2 are calculated for them and the other analyses.

1. Petalite, Somero.

SiO_2	77.47	1.2899	12899		
Al_2O_3	17.12	0.1679	3358	3398	R_2O:R_2O_3:SiO_2 0.877 : 1 : 3.797
Fe_2O_3	0.24	0.0015	30		
CaO	0.13	0.0023	23	2983	
Li_2O	3.95	0.1322	2644		
Na_2O	0.78	0.0126	252		
K_2O	0.30	0.0032	64		

and so on

1. Somero, Finland	0.877 : 1 : 3.797	0.892 : 1 : 3.823
2. Varuträsk, Sweden	0.886 : 1 : 3.867	
3. Varuträsk, Sweden	0.879 : 1 : 3.755	
4. Utö, Sweden	0.926 : 1 : 3.873	
5. Australia	0.850 : 1 : 4.000	
6. Australia	0.890 : 1 : 3.890	
7. Elba	0.663 : 1 : 3.768	
8. Mass., U. S. A.	0.795 : 1 : 3.885	
9. Mass., U. S. A.	0.796 : 1 : 4.086	
10. Maine, U. S. A	0.759 : 1 : 3.869	

An 0.8 % error is made if we set the medium ratio of the four first analyses (the Swedish and Finnish analyses) = 3.8 : 1 : 0.9. The petalite formula in common use, $LiAlSi_4O_{10}$, was published in 1882 by P. Groth (4) and requires a ratio 1 : 1 : 4. This formula was founded on evidences from the

Utö-analysis. As may be seen, Groth made a 7.4 % error in setting 0.926 = 1.000, but if the analyses 1 to 3 were to be made the basis of the formula $LiAlSi_4O_{10}$, there would be an error of 12 % (0.88 = 1.00). A better formula for these analyses would be (if the ratio 3.8 : 1 : 09 be made the basis) $9\ Li_2O \cdot 10\ Al_2O_3 \cdot 76\ SiO_2$; a very inconvenient formula indeed. — Better results are obtained if we eliminate the minor Na, K and Ca- content as felspars. According to W. Kunitz (5) Li is not replaced by Na and K. The differences in ion radius are too great. After the elimination of Na, K and Ca the above-mentioned rations would be:

1. 0.898 : 1 : 4.002
2. 0.883 : 1 : 3.948
3. 0.875 : 1 : 3.855
4. 0.946 : 1 : 4.035
5. 0.851 : 1 : 4.011
6. 0.801 : 1 : 3.935
7. 0.615 : 1 : 3.878
8. 0.785 : 1 : 3.928
9. 0.784 : 1 : 4.148
10. 0.686 : 1 : 4.141

8.124 : 10 : 39.882

The medium is: 0.812 : 1 : 3.988, from this the formula should be:

$Li_8\ Al_{10}Si_{40}O_{99}$ The factor 8 is, however, in many analyses near 9, but never near 10, as the old formula of Groth demands. Compare the formula of Groth $5\ Li_2O \cdot 5\ Al_2O_3 \cdot 40\ SiO_2$ with that now suggested: $4\ Li_2O \cdot 5\ Al_2O_3 \cdot 40\ SiO_2$.

The pale pink decomposition products mentioned by Quensel (2) are also readily to be seen on the surfaces and along the cleavage planes of the, otherwise very clear and unaltered Somero-petalite. The altered sheet is, however, very thin. P. Quensel has determined this pink to yellow claylike mineral to be montmorillonite $Al_2O_3 \cdot 3SiO_2 \cdot 6H_2O$.

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Geological Survey, April 1947.

24.

A COMPOSITE LAVA FLOW FROM LOUNATKORKIA, HOGLAND

BY

WALTER WAHL

1. INTRODUCTION

Only few inhomogeneous lava flows have been more closely studied and described. The first reference to a differentiated lava flow, of which the present author is aware, is the description by Leopold von Buch (1) in 1825 of an obsidian flow from Teneriffe containing felspar phenocrysts, where the basal layers are crowded with such phenocrysts, while the upper parts of the flow contain phenocrysts but sparsely. During his visit to the oceanic islands Charles Darwin observed a similar instance (2). He found that the lower part of a basalt flow from James Island in the Galapagos group was crowded with plagioclase crystals.

From the period of microscopic description of rocks we have the description by Lane (1897) of certain basaltic effusives from the Keweenaw Point region (3) and by Iddings (1899) of a sheet S. E. of Electric Peak, Montana (4) [1].

Later Lane and Queneau (6) and Lane (7) gave further descriptions of such lava flows from Michigan and from New Jersey, and analyses were published of the different components of the Watchung sheets near Springfield, New Jersey. The last paper on the subject of »Magmatic Differentiation in Effusive Rocks» by Lane was published in collaboration with Powers in 1916 (8). In this paper a number of lava flows from the neighbourhood of Cape d'Or, Nova Scotia, were described and a great number of quantitative microscopic data and chemical analyses of the lava from different levels of the flows were given.

In his great Skye memoir Harker (1904) describes some basaltic rocks which he considered as composite dikes (9). Later it has been shown by Kennedy (10) that these occurrences in reality are composite sheets. From one of these sheets, that of Druin na Criche, Skye, analyses of both the rock varieties are given.

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[1] Note: This occurrence is, however, by Kennedy (5) considered to be a differentiated sill, and not a lava flow.

Also the lava stream of the 1910 eruption of Etna seems, according to the statements of Washington, Aurousseau and Keyes (1926), to be of a composite character (11), but so far only one analysis has been published.

In 1931 Kennedy published an interesting paper on certain composite lava sheets from Renfrewshire, near the shore of the Firth of Clyde, together with a review of the whole problem of composite lavas. The lavas described by Kennedy (12) consist of a lower fine-grained »Muguerite» (olivine-andesine-andesite) layer without phenocrysts and an upper basaltic layer containing large amounts of porphyritic phenocrysts of basic plagioclase, up to 1,5 cms in length. »No chilled contact or intermediate layer of slag separates the two types, and the lower non-porphyritic rock is in perfect continuity with the ground-mass of the upper.»

According to the analyses published the composite lavas from Michigan, Nova Scotia, Skye and Renfrewshire show no very pronounced difference in the chemical composition between the two members of these composite lava streams, the principal difference being the occurrence of a more or less considerable surplus of basic plagioclase in one of the rock members.

It is therefore of much interest that the great porphyry lava flow of the island of Hogland in the center of the Gulf of Finland shows much greater variation in chemical and mineralogical composition than the composite lavas mentioned above. In this paper it is the intention of the author to describe in some detail only the most extreme case of differentiation of these porphyries of Hogland, the lavas of the most southerly and highest of the three big »humps» of Hogland, that of Lounatkorkia. Before entering into a description of this composite part of the lava mass, it seems well to give a short description of the principal rocks of the island and the general occurrence of the porphyries, in part quoting the excellent short description by Kranck (13), who has carried out the most extensive investigation on the geology and petrology of Hogland in recent years, although only two short reviews of the principal geological problems have so far been published.

2. SOME GENERAL REMARKS WITH REGARD TO THE YOUNGER IGNEOUS ROCKS OF HOGLAND

»The oldest formation of Hogland is an assemblage of gneissic, amphibolitic and leptitic rocks invaded, and partly assimilated by granite.» »Intense granitization has in most places changed the rocks into migmatites, crossed by numerous veins and dikes of granite and pegmatite. The main direction of strike within the highly folded and schistose formation is about NW-SE, the dip almost vertical.» »The northernmost and southernmost parts of the Archean terrain consist of fairly pure microcline granite.» In its middle part the island is crossed by a broad belt of gabbro, diorite and ultrabasic rocks. There are two granites, one grey, slightly gneissic, of granodioritic type and one red microcline granite, often accompanied by coarse, red pegmatite. »On the low rounded hills Majakallio, Suur Somerikonkallio and Mustakallio on the northern part of the island below the steep western slope of the quartz-porphyry mountains there is preserved a quartzite boulder conglomerate unconformably resting on the basement of the Archean rocks». »This conglomerate shows no traces of having been influenced by later orogenic movements, and the layers have evidently been preserved almost in their original position. They dip slightly eastward, but are nowhere folded or deformed by stress.» »The quartzite conglomerate formation was deposited on an uneven surface, as is clearly shown by the stereographic map.» »The bottom of the conglomerate follows the slopes of the valleys.» »The thickness of the remaining strata of conglomerate and quartzite nowhere exceeds 5 m.» »At Vähäkorkia the conglomerate immediately underlies the quartz-porphyry which is intruded between the conglomerate pebbles.»

»The youngest formation of Hogland, the porphyry rocks, occupy the eastern part of the island and compose its highest hills. There is every probability that they are symmagmatic with the rapakivi granites of eastern Finland (Ramsay 1890 (14), Wahl 1925 (15). As Ramsay has pointed out, the greatest part of these rocks forms a single bed of a thickness exceeding 100 m. and covering the rather uneven surface of the old bedrock. Before this magma was erupted, smaller quantities of it flew out in the form of thin lava streams. Contemporaneously tuffs and tuff agglomerates were deposited. The first outpouring lava had a composition varying from andesitic to basaltic. The dark porphyritic rocks belonging to this phase of the eruption occur in 1 to 5 m. thick beds along the western border of the porphyry area underlying the younger quartz-porphyries of Pohjoiskorkea, Majakallio, Vähäkallio and Lounatkorkia. They frequently show well developed amygdaloidal and block lava structures. At Välikallio there is under the lava a layer of tuff agglomerate containing pebbles of Archean rocks.» »In the later outpouring big lava mass the most widespread type is an acid quartz-porphyry, generally more coarse-grained than the rocks described above (often holocrystalline). It contains phenocrysts of potash felspar and quartz, often measuring up to 1 or 2 cm. in a dark grey aphanitic ground-mass.»

It may here be pointed out that in several localities the ground-mass when fresh is very dark, almost black. When the ground-mass is slightly decomposed (late magmatic autometamorphosis) the color changes to brown. In certain varieties (Hirsikallio) dark, almost black 2 to 3 mm. large quartz-grains occur in great abundance in a greenish black ground-mass, orthoclase phenocrysts being scarce and of small size or lacking.

Already Lemberg (16) and Lagorio (17) have described fairly basic plagioclase-porphyrites as forming part of the porphyry mass of Lounatkorkia and also Ramsay (14) mentions that plagioclase phenocrysts occur in the porphyry of this hill. Kranck in his »Stereogram of Suursaari» (Hogland) (13), in the »Explanation of colors» says: »1. Quartz-porphyry including basic differentiation products belonging to the same magma mass.» In the text we find the following statement: »A more basic variety (of the porphyry) occurs near the hill Lounatkorkia. It contains phenocrysts of plagioclase and hornblende and in part resembles very much the porphyritic (older) lavas earlier described. The basic lavas gradually grade into acidic porphyries without any sharp contact, showing that they are differentiated from the same magma.» In his very interesting paper on the geology of Hogland, published a year later, Kranck (18) once more refers to the occurrence of plagioclase- and hornblende-carrying varieties of the younger porphyry and to the occurrence of such in the southern part of the island, to some extent already at Tervamäki, but especially in the surroundings of Lounatkorkia, »where the potashfelspar nearly wholly is replaced by plagioclase phenocrysts.» Kranck seems, however, not to make any definite distinction between these plagioclase-rich differentiation products of the porphyry and the porphyrites erupted earlier, although he states that the »differentiation tendency here is quite clear.» Also the present author has earlier referred to the occurrence of these plagioclase-bearing porphyrites at or near Lounatkorkia (15) and pointed out that among the porphyries they probably occupy the same position as, and correspond to, the basic plagioclase- and hornblende-rich varieties of the rapakivi of the Aspö islands (19) farther to the north.

Since the older publications give no clear description of the occurrence of these plagioclase-bearing porphyries at Lounatkorkia, the present author at the end of August in the summer of 1939 visited Hogland in order to study more closely the basic varieties of the porphyry. He was on that occasion accompanied by Magister Åke Stenius, who rendered valuable assistance.

3. THE PORPHYRIES OF LOUNATKORKIA

Ascending the north-eastern slopes of the highest hill of Hogland, the Lounatkorkia, one finds that the rock of the uppermost portion consists of a dark green plagioclase-porphyrite which in appearance differs considerably from the quartz-orthoclaseporphyries of which the other porphyry hills of Hogland are composed.

Here, on the top of this hill, we also meet with another feature not observed in other parts of the big porphyry mass of Hogland: The porphyrite contains numerous xenolithes of the older country rocks, mostly such of a coarse reddish granite-pegmatite, but also such of the greyish gneiss-granites. The xenolithes are seldom larger than one foot, are sometimes cracked and fritted, and often also to a certain extent digested by the lava.

forming big lumps in it. By some of the older writers these xenolithes have been mistaken for huge »porphyric orthoclase felspars», »porphyric micropegmatite», or concretions of felspar and quartz. A careful study in the field of a great number of them, together with microscopical investigation, shows beyond doubt that we here have to do with xenolithes from the country rocks, which the lava has carried with it and fritted and digested to a smaller or larger extent.

It is, however, remarkable that these xenolithes only seem to occur as a fairly limited horizontal band or streak in the lava, indicating that we here have to do with the remainder of a lateral part of a lava mass passing a channel or crack in the country rocks and sweeping with it chips from the wall-rock. We are thus here at Lounatkorkia probably not far from the orifice or crack in the country rock through which the Hogland porphyry lava has escaped.

The western-southwestern slope of the hill is in places very steep, but where it is accessible one finds that the plagioclase-porphyrite downwards passes over into quartz-porphyry of a light greyish-brown colour. Farther down the slope, but still at a great height above the sea-level and the shore plateau of Archean rocks, we encounter the contact between the porphyry, which is here black, and the underlying gneiss-granite. At this locality there is no tuff between the porphyry and the old weathered surface of the country rock. The gneiss-granite does not seem to have been altered by the lava mass covering it, but the porphyry is somewhat denser close to the granite contact and thus shows a chilled contact phase. On passing up and down from the porphyry level to that of the plagioclase-porphyry, one is not able to locate any border line between the two rocks. They merge everywhere gradually into each other, and, the weathering of the two varieties being very similar, it is difficult to tell where the one rock ends and the other begins without using the hammer.

The surface of the old countryrock has, as Kranck (18) has shown beyond doubt, been very uneven, and towards the southern slope of Lounatkorkia the granite-gneiss basement slopes sharply downward, the quartz-porphyry here occurring at a much lower level than the above described contact. Kranck has published a sketch map of the valley between Lounatkorkia and Vähäkorkia (Fig. 9, 1929). According to this map and the text there occurs under the southern steep wall of the Lounatkorkia porphyry a porphyrite bed, the porphyrite of which has an amygdaloidal character. A corresponding porphyrite layer is found at the foot of Lounatkorkia, on the northern side of the hill. Here there is a thin layer of a tuff breccia between the porphyrite and the covering porphyry. The lower amygdaloidal porphyry thus belongs to the earlier series of porphyry eruptions, which preceded the outpouring of the big porphyry masses. The lower part of the valley between Lounatkorkia and Vähäkorkia is occupied by gneiss-granite, at three places capped by plagioclase-porphyrite. This occurrence of porphyrite was visited by the members of the International Geological Excursion to Finland in the summer of 1933.

THE LABRADOR-QUARTZSYENITEPORPHYRY OF LOUNATKORKIA

The »plagioclase-porphyry» is a dense, greenish black rock with abundant phenocrysts of greenish-white plagioclase. In those varieties of the rock from the upper part of the hill, which the present author has studied, the larger phenocrysts are mostly from one to two centimeters in length, about one centimeter wide and about half as thick. But Lemberg states that phenocrysts up to a size of four centimeters occur. Among the smaller plagioclase phenocrysts lath-shaped ones are not infrequent. Rounded, black lustrous quartz grains are seen here and there, but are not numerous. No dark minerals occur as phenocrysts visible to the eye. Besides these phenocrysts mentioned above, the porphyrite contains occasional grains of quartz, of sizes from one centimeter and downwards, which seem to be derived from the surrounding quartzite or pegmatites, since they consist of clusters of corroded and cracked quartz-grains.

Lemberg has published three analyses of the plagioclase phenocrysts (1868 p. 341) according to which the average composition is that of a labradorite $Ab_{50}An_{50}$.

In thin slices the ground-mass of the »plagioclase-porphyrite» reveals itself as holocrystalline; it is, however, of a rather extraordinary composition. Roughly estimated, the dark minerals constitute some 20 % of the whole, and of these about four fifths is a green to bluish-green »ordinary hornblende.» This hornblende forms only to a small extent separate idiomorphic individuals; mostly quite small grains of the hornblende are clustered together in lumps, containing great numbers of ground-mass plagioclase, orthoclase and quartz grains between the hornblende grains, the whole forming a kind of micropoikilitic structure. The dark mica, which is greenish-brown and not of as pronounced a pleochroism as one is accustomed to meet with in dark mica, also forms similar clusters of small grains as does the hornblende, often together with the latter in the same cluster. It is further characteristic of this rock, that when we meet with a foreign small xenolithe of for instance quartz-grains, this is surrounded by a layer of small hornblende grains, which separate the xenolithe from the surrounding ground-mass and also fill the cracks in the xenolithe.

The principal part of the ground-mass, occupying the interstices between the phenocrysts of labrador and quartz and the hornblende and groundmass clusters described above, is constituted by a kind of irregular micropegmatitic growth of orthoclase and quartz which in this rock, however, are peculiar in this respect that they are penetrated in all directions by great numbers of long plagioclase laths fine as needles.

Iron ore is not abundant and may accordingly be placed with the accessory constituents. It consists probably of ilmenite and at certain points in the slides small rounded clusters of grains of iron ore and titanite are found. Idiomorphic titanite grains not associated with iron ore have not been observed. Apatite is fairly plentiful, occurring in stout prisms, often having a hollow channel along part of the prism center.

It is rather remarkable for a rock of effusive character like this, having been consolidated as a big lava mass and containing a comparatively anorthite-rich plagioclase, that neither in the big plagioclase phenocrysts nor in the smaller laths of later crystallization is any zoning to be found.

In Table I several analyses of the plagioclase porphyrite are given. No 1 is a new analysis made by Dr. N. Sahlbom of Stockholm. The specimen was collected by the author from the top of the hill. Nos 2, 3, 4, 5 are older analyses by Lemberg (from the western part of the hill). No 6 is an analysis by Lemberg of the groundmass of No 5. A, B, and C in Table I are analyses by Lemberg of the labradorite phenocrysts of the rock.

Table I.

	1	2	3	4	5	6	A	B	C
SiO_2	65,60	62,75	61,72	60,54	60,48	60,79	55,28	54,87	54,76
TiO_2	0,60	—	—	—	—	—	—	—	—
Al_2O_3	14,60	17,11	15,83	16,50	16,93	14,68	26,15	26,54	26,29
Fe_2O_3	1,58	2,20	8,22	8,60	7,83	9,87	1,87	2,18	2,36
FeO	5,03	5,61							
MnO	0,20	—	—	—	—	—	—	—	—
MgO	0,56	0,77	1,50	1,30	1,04	1,81	0,40	0,35	0,23
CaO	3,70	4,57	4,01	4,84	4,98	2,71	9,79	9,22	9,31
Na_2O	2,74	2,57	1,95	2,71	2,74	1,37	5,15	5,19	5,37
K_2O	4,23	4,41	5,40	4,02	4,02	6,93	1,14	1,20	1,04
P_2O_5	0,38	—	—	—	—	—	—	—	—
H_2O +	0,50	0,50	1,16	1,06	1,14	0,95	0,22	0,45	0,64
H_2O —	0,05								
	99,77	100,49	99,79	99,58	99,16	99,13	100	100	100
Sp. Grav.	—	—	—	2,762	2,764	—	—	—	—

1. Labrador-quartzsyeniteporphyry, Top of Lounatkorkia hill, Hogland. N. Sahlbom analyst.

2—5. » Western part of Lounatkorkia, Hogland. Joh. Lemberg analyst.

6. Ground-mass of Labrador-quartzsyeniteporphyry No 3. Joh. Lemberg analyst.

A, B, C, Labradorite phenocrysts from same rock Joh. Lemberg analyst.

The mineralogical as well as the chemical composition reveals that the labrador containing dark differentiates of the porphyries of Lounatkorkia, which so far, according to their macroscopic properties, have been called »plagioclase-porphyrites», in reality are quartz-syeniteporphyries containing labrador phenocrysts. It seems therefore proper to use the designation Labrador-quartzsyeniteporphyry for these rocks.

In Table II some analyses of porphyrites of the older series of eruptions are given. No 1 is a new analysis by Dr. Sahlbom of the labrador-porphyry from Sommarö to the NE of Hogland, originally published by the author (1938) (19). No 2 is an analysis by Lemberg of the labrador-porphyrite with slender labradorite phenocrysts from Majakallio. No 6 is the ground-mass of the same rock. Nos 2, 3, 4 are »porphyrites» lacking phenocrysts from the same locality. A is an analysis by Lemberg of the labradorite phenocrysts from No 2.

Table II.

	1	2	3	4	5	6	A
SiO_2	51,05	48,49	43,65	37,46	44,14	46,70	53,36
TiO_2	2,00	—	—	—	—	—	—
Al_2O_3	18,55	17,18	16,92	18,98	19,73	15,06	25,42
Fe_2O_3	6,08	19,32	20,31	21,12	17,82	22,59	3,43
FeO	5,38						
MnO	0,11	—	—	—	—	—	—
MgO	2,62	3,58	3,83	4,47	3,37	3,93	1,00
CaO	7,34	5,16	10,60	14,03	6,52	4,83	8,78
Na_2O	3,59	2,96	1,52	0,10	1,21	2,69	4,06
K_2O	2,16	2,48	2,09	0,24	5,54	2,39	2,34
P_2O_5	0,35	—	—	—	—	—	—
$H_2O +$	0,44	0,83	0,91	3,32	1,77	0,50	1,61
$H_2O -$	0,13						
	99,80	100	99,92	99,72	100,10	98,69	100

1. Labrador-porphyry from Sommarö E-NE from Hogland. N. Sahlbom analyst.
2. Labrador-porphyry with slender phenocrysts of labradorites, Majakallio, Hogland. J. Lemberg analyst.
3. Porphyrite without phenocrysts from same locality as 2. J. Lemberg analyst.
4. Porphyrite » » » » » Epidotized Majakallio, Hogland. J. Lemberg analyst.
5. »Porphyrite», same locality as 2—4. Hogland. J. Lemberg analyst.
6. Ground-mass of the porphyrite No 2. » » »
A. Labradorite phenocrysts from No 2.

As shown by the analyses in Table II, when compared with those of Table I, the labrador-porphyries of Sommarö and Majakallio with labradorphenocrysts are much more basic rocks than the labrador-quartzsyeniteporphyries of Lounatkorkia. They contain less potash than soda and may be classed as ordinary labrador-porphyrites. The porphyrites Nos 3 and 4 without phenocrysts are still more basic and of a basaltic character. The porphyrite, an. No 5, is again of quite a different and decidedly »lamprophyric» character.

THE QUARTZ-PORPHYRIES OF THE LOWER PORTION OF LOUNATKORKIA

Near the contact towards the gneissgranite the quartz-porphyry is black and dense with numerous round grains of black quartz. But at several places along the western cliff front there occurs a light greenish grey to yellowish grey quartz-porphyry which forms »schlieren» in the dark rock. In thin slides the two varieties are very different and therefore will be separately described.

The dark quartz-porphyry contains rounded lustrous black quartz-grains in a black dense ground-mass. The orthoclase phenocrysts are seldom more than 5 mm in size and are not numerous. In thin slides the ground-mass is a very fine-grained mass of small crystals of quartz and felspar and looks as if it were a devitrification product of an originally glassy ground-mass. There is a certain amount of small lath-formed grains, which probably have been plagioclase, but which now are much altered. Small clusters of biotite grains occur here and there. No hornblende has been observed, nor any apatite, but some small grains of magnetite are found here and there. The orthoclase phenocrysts are much altered and seem to be kaolinized and epidotized. The quartz grains are much larger than in the plagioclase-porphyry and much more numerous. They are strongly corroded and in ordinary light one sees a border rim round each quartz phenocryst, which is lighter than the surrounding ground-mass and of different appearance on account of the lack of darker particles which impregnate the rest of the ground-mass. This lighter border rim probably consists of the marginal solution and reaction products of the quartz-grains resulting from the corrosion of the quartz, which here still remain in the place where they were originally formed. This indicates that there has been no movement of the magma after the corrosion took place and that this corrosion thus must have immediately preceded the final solidification.

The light quartz-porphyry is a quite different rock. It contains the same corroded black quartz grains as the dark rock, but about two thirds of the ground-mass consist of a fairly even-grained microcrystalline assemblage of quartz and orthoclase grains and about one third of large clusters of epidote grains. This epidote is slightly yellowish green in ordinary light, has a high refractive index and shows strong interference colours between crossed nicols. There is neither plagioclase nor hornblende, nor any other dark mineral to be seen in the slides.

We have thus here to do with a »helsinkitic» or »unakitic» facies of the quartz-porphyry. The epidote grains are well developed and transparent, and some of them have often grown to stout idiomorphic prisms somewhat larger than the surrounding grains of epidote. These epidote clusters enclose or are interwoven with the microcrystalline felsparquartz ground-mass, but seem at least in part to have crystallized before the crystallization of this ground-mass had become complete. All these circumstances taken together point towards the epidote of this unakite-porphyry being a primary crystallization product of the magma. The analysis shows that this porphyry contains much lime as compared with sodium, and, in portions of the magma rich in watervapour and gases, the crystallization of the lava seems to have been retarded and has later taken place at such a low temperature that under these conditions epidote instead of hornblende, iron ore and plagioclase, has been the stable product of mineral formation and thus an unakitic facies of the lava has been formed.

Epidotization of the quartz-porphyries of Hogland is by no means uncommon. Already Lemberg (16) and Lagorio (17) have described epidotization of both phenocrysts and ground-mass of the porphyries, but in these cases it has mostly been the question of secondary alteration of rocks originally consolidated as quartz-porphyries, whereas we here have to do with a primary crystallization of the lava as an unakite. In his interesting work on the granophyres from the bottom of the Bothnian Gulf, which are found in boulders all along the western coasts of Finland, Eskola (20) has been the first to observe and describe such unakites among effusive quartz-porphyries. There is evidently a strong similarity between these rocks described by Eskola and the unakite-porphyry from Lounatkorkia.

In Table III an analysis of the quartz-porphyry from the neighbourhood of the gneiss-granite contact at Lounatkorkia is given and for comparison an analysis of a quartz-porphyry from Central Hogland and one of the quartz-porphyry of Sommarö. All these analyses are by Dr. N. Sahlbom in Stockholm.

Table III.

	1	2	3	4	5	6	7	8	9	10	11	12
SiO_2	70,74	72,64	74,57	74,21	71,52	71,95	73,94	68,93	68,94	67,62	64,29	74,0
TiO_2	0,20	0,60	0,32	—	—	—	—	—	—	—	—	—
Al_2O_3	14,56	11,60	12,67	12,51	12,74	12,18	12,07	14,41	14,31	16,25	17,00	10.53
Fe_2O_3	0,48	2,38	2,02	0,78	1,78	1,96	4,45	4,36	2,29	4,92	—	5,01
FeO	3,60	3,59	2,29	2,94	1,81	1,81			2,75		4,30	
MnO	0,04	0,10	0,05	—	—	—	—	—	—	—	—	—
MgO	0,40	0,21	0,49	0,20	0,30	0,18	0,13	0,33	0,47	0,34	0,76	0,28
CaO	3,22	1,72	0,86	1,82	1,10	1,43	0,35	1,01	2,25	1,74	1,91	0,99
Na_2O	0,93	1,12	0,75	0,54	0,72	0,78	0,83	0,44	1,13	0,35	1,95	0,60
K_2O	5,25	5,00	4,94	6,67	7,70	7,53	6,68	9,62	7,32	6,92	7,00	7,57
P_2O_5	—	0,35	—	—	—	—	—	—	—	—	—	—
H_2O+	0,78	0,50	0,78	0,50	0,39	0,63	0,60	0,59	0,46	1,42	0,98	0,86
H_2O-	0,16		0,20									
Spec.	100,36	99,81		100,17	98,06	98,45	98,06	99,69	99,98	99,56	99,19	99.84
Grav.	—	—	—	2,702	2,698	2,681	—	2,667	—	—	2,657	—

1. Quartz-porphyry from near border line to gneiss-granite, Lower part of Lounatkorkia, Hogland. N. Sahlbom analyst.
2. Quartz-porphyry, northern part, of top of Tervamäki, Hogland. E. H. Kranck. leg., N. Sahlbom analyst.
3. Quartz-porphyry, Sommarö NE-E from Hogland. N. Sahlbom analyst.
4. Black quartz-porphyry with dark orthoclase, south from Lapinlaks, Hogland. Lemberg analyst.
5. Black quartz-porphyry with red orthoclase, N from Lounatkylä near seasshore, Hogland. Lemberg analyst.
6. Dark violet quartz-porphyry with red orthoclase phenocr. Hirsikallio, Hogland. Lemberg analyst.

7. Black quartz-porphyry with red orthoclase, Purjeniemi, Hogland. Lemberg analyst.
8. Black quartz-porphyry with red orthoclase, Pohjoiskorkia, Hogland. Lemberg analyst.
9. Black quartz-porphyry with red orthoclase, Pohjoiskorkia, Hogland. Lemberg analyst.
10. Black quartz-porphyry from E shore opposite Suurhälli, Hogland. Lemberg analyst.
11. Black quartz-porphyry, Lounatkorkia, Hogland. Lemberg analyst.
12. Ground-mass of the rock No 6.

The analyses Nos 4—12 are old analyses by Lemberg and are here cited because they give information about the variations in composition of the porphyry body of Hogland. Interesting is that the MgO content, generally low as it is, may sink to as small a value as 0,13 % (No 7), CaO to 0,35 % (No 7) and Na_2O to 0,35 % (No 10).

No 12 is an analysis of the ground-mass of the porphyry No 6. As seen from these analyses, SiO_2 and ironoxides are higher in the ground-mass and Al_2O_3, CaO and Na_2O lower. But the differences are not large.

The black porphyry No 11 from Lounatkorkia, containing labradorite phenocrysts, seems to be a variety forming a link between the quartz-porphyries and the labrador-quartzsyeniteporphyries of this locality.

4. THE CHEMICAL CHARACTER OF THE PORPHYRIES OF LOUNATKORKIA

The mineralogical composition of the porphyry from Lounatkorkia containing labradorphenocrysts, as well as its chemical composition, is rather extraordinary. The labrador-porphyries in the usual sense of the name are porphyrites of the augite-andesitic and basaltic rockfamilies containing phenocrysts of labradorite but the porphyry of Lounatkorkia does not belong to this group of rocks. Mineralogically it is a quartz-syeniteporphyry containing large numbers of phenocrysts of labradorite and a small amount of quartz phenocrysts. Chemically it is characterised by a much larger content of potash than of soda and also contains a fairly large amount of lime.

The present author has earlier pointed out (15) that the rapakivigranites belong to the potash clan of rocks in the sense of Niggli and not to the alkali-calcic clan. They represent the silica-rich end members of the potash clan of series of rocks, which often have, however, not been dealt with as belonging to a separate clan, but have been thrown together with the rocks of the alkali-calcic series. When intermediate and basic rocks are split off by the rapakivi magmas either intermediate lime-potash rocks or gabbroic-anorthositic rocks are frequently formed. The formation of gabbroic-anorthositic differentiation products is analogous to the splitting off of gabbroic, pyroxenitic, peridotitic, hornblenditic and anorthositic rocks from the granites of the alkali-lime series. But instead of the dioritic-andesitic intermediate differentiation products of the alkali-lime series the rapakivi-granites of the potash-series often give quartz-syenitic differentiation products of a peculiar »lamprophyric character», combining high potash with high lime content. So far as these products have been investigated, they are in the rapakivi series poor in magnesia but very rich in iron. They are perhaps to be considered as the lamprophyric portion in the formation of »diashist» differentiation products of the rapakivi magmas.

Table IV.

	1	A	B	2	3	4	5	6	7	8	9	10
SiO_2	65,60	64,69	61,06	61,37	58,05	47,73	44,14	65,05	59,41	58,21	55,44	51,68
TiO_2	0,60	0,68	1,04	—	—	1,91	—	0,44	1,00		0,16	1,08
Al_2O_3	14,60	16,02	17,12	16,59	17,89	15,86	19,73	16,22	19,06	19,19	18,60	14,07
Fe_2O_3	1,58	0,62	1,06	8,27	4,31	4,31	17,82	1,01	1,87	4,11	2,09	4,71
FeO	5,0	4,64	6,35			5,89		2,48	3,42	0,87	4,48	4,57
MnO	0,20	0,02	0,10	—	—	0,18	—	—	—	—	—	—
MgO	0,56	0,63	0,68	1,15	1,53	5,36	3,37	1,46	2,05	0,98	4,75	7,72
CaO	3,70	3,62	4,40	4,60	4,99	7,56	6,52	3,19	4,01	3,58	6,76	6,65
Na_2O	2,74	2,34	2,79	2,49	0,91	2,06	1,21	2,62	2,58	2,57	1,79	2,45
K_2O	4,23	5,54	4,99	4,46	10,55	5,10	5,54	5,52	5,29	9,18	6,63	4,16
P_2O_5	0,32	0,38	0,48	—	—	1,35	—		0,29	—	—	0,72
H_2O	0,55	0,60	0,20	0,96	1,56	0,68	1,77	1,60	1,55	0,74	0,25	2,09
	99,77	99,80	100,27	99,89	99,78	99,44	100,10	100,33	100,61	100,14	100,95	100,03

1. Labrador-quartzsyeniteporphyry, Lounatkorkia, Hogland. N. Sahlbom analyst.
A. Syenitic Rapakivi containing Labradorite phenocrysts, Luppi, Aspö Islands. N. Sahlbom analyst.
B. Syenitic Rapakivi, more basic than A. Harbour of Aspö. N. Sahlbom analyst.
2. Average of four older analyses 2—5, Tab. I of the Labrador-quartzsyenitheporphyry of Lounatkorkia by Lemberg.
3. Felsitic rock. Large xenolithe in 2. Lounatkorkia, Hogland. Lemberg analyst.
4. »Lamprophyre» belonging to the Åva granite (Rapakivi?), Brändö, Åland.
5. »Porphyrite» of the older porphyrite series (?). Majakallio, Hogland. Lemberg analyst.
6. Labrador trachyte (toscanite) mean of 7 analyses. Monte Amiata, Italy.
7. Biotite latite, La Cava, Viterbo, Italy. Washington analyst.
8. Vulsinite, Bolsena, Italy. Washington analyst.
9. Ciminite, Fontana Fiesole, Viterbo, Italy. Washington analyst.
10. Absarokite, Two Ocean Pass, Yellowstone Park. Whitefield analyst.

The rocks corresponding to the labrador-quartzsyeniteporphyries of Lounatkorkia among abyssal rocks are the hornblende-rich basic rapakivis of Aspö described by the author (19). Some analyses of »lamprophyric» differentiation products of rapakivi and similar magmas have been brought together in Table IV — together with some analyses of effusive rocks of the potash series from some other districts — in order to facilitate a comparison and the discussion of the chemical relations of these rocks. The analyses of these Aspö rocks, A and B of Table IV, are almost identical with those of the Lounatkorkia rocks. The analytic figures of all these rocks correspond also in a remarkable degree to those of the »toscanites» (labrador-bearing trachytes), No 6, Table IV among the effusive rocks of the potash series in Italy. Lemberg has described and analysed a large xenolithe, which he mentions as occurring in the labrador-quartzsyenitic facies of the porphyries of Lounatkorkia (an. 3, Tab. IV), which is extremely potash-rich and of a chemical composition very similar to the »vulsinite», an. No 8. In his last posthumous paper on the rocks of the archipelago east of the Åland Islands Sederholm (21) briefly describes a granite from Åva of similar composition as the rapakivis, which is accompanied by numerous »lamprophyre» dikes. The analysis No 4, Table IV, shows that we here have a rock related to the Lounatkorkia labrador-porphyry although more basic and chemically corresponding to the vulsinites and ciminites (Washington), an. 9, Table IV, of the Italian potash province of rocks.

Also among the analyses by Lemberg of the older porphyrite series of Hogland, which mostly are of basaltic and anorthositic character, although extremely iron-rich, there is one analysis, No 5, Table IV, which resembles the rock from Åva and the Italian ciminites and the still more basic banakites, shoshonites and absarokites, an. No 10, Table IV, from the Yellowstone Park.

The quartz-porphyry an. No 1, Table III of Lounatkorkia has a higher content of lime than any of the other Hogland porphyries, and also here we thus meet with a lime-potash rock, although it by its high silica content clearly belongs to the quartz-porphyries of the potash series.

5. THE DIFFERENTIATION OF THE PORPHYRY MAGMA OF LOUNATKORKIA

In comparing the figures of the new analyses of the labrador-quartz-syeniteporphyry No 1 of Table I with those of the quartz-porphyry N:o 1 of Table III we find that the syenite-porphyry contains about the same amounts of Al_2O_3 and MgO as the quartz-porphyry, but less SiO_2 and K_2O and more TiO_2, Fe_2O_3, FeO, MnO, CaO, Na_2O and P_2O_5. The differences in the SiO_2, total of iron oxides and Na_2O are the biggest. If we consider that the quartz-porphyry contains no apatite and the syenite-porphyry considerable amounts, the difference in the lime amount of the two rocks which enters into the anorthite molecule can be said to be negligible.

However, the other analyses of the syenite-porphyry contain less SiO_2 and more CaO than the analysis No 1, from 4—5 %, and the remainder of the analyses of quartz-porphyries from Hogland only from 0,35—2,25 % of CaO. There is thus generally a very considerable rise of CaO in the syenite-porphyries of Lounatkorkia as compared with the bulk of the quartz-porphyries of Hogland. A closer comparision shows that it is principally the components of anorthite and albite together with the femic components forming hornblende which have been enriched in the syenite-porphyry as a result of the magmatic differentiation.

If we compare the figures for the ground-mass for the two types of rocks which according to Lemberg's work is given in Table V we find similar conditions.

TABLE V.

	SiO_2,	Al_2O_3,	Fe_2O_3,	MgO,	CaO,	Na_2O,	K_2O
Labrador-syeniteporphyry, ground-mass	60,79	14,66	9,89	1,81	2,71	1,37	6,93
Quartz-porphyry, ground-mass........	74,00	10,53	5,01	0,28	0,99	0,60	7,57

These figures show that also in the ground-masses of the two kinds of porphyry it is the components of anorthite and albite together with those femic components of which the hornblende is built up which have been enriched in the labrador-syeniteporphyry, whereas the silica and orthoclase components have been enriched in the quartz-porphyries. *During differentiation the components of orthoclase and of albite of the parent magma thus move in opposite directions.* Table V shows also beyond all question that the differentiation of the porphyry magma has not taken place simply by an accumulation of certain kinds of phenocrysts in a limited portion of the magma, but that, since also the ground-masses of the different porphyries differ in such a high degree, the whole bulk of the rock magma has been subject to a differentiation process.

In the case of the earlier described composite lava flows the principal difference between the components has chemically not been large and it is possible to explain the difference by assuming that an accumulation of plagioclase crystals in the lower parts of the lava streams has taken place. In the Nova Scotia rocks investigated by Lane and Powers the explanation given by these authors is that a concentration of the leucocratic components, lighter than the bulk of the magma, has taken place towards the top of the flow, while the melanocratic constituents, heavier than the average of the magma, have been concentrated towards the base of the flow.

In the case of a difference in composition of also the ground-masses of the different varieties of the lava, especially if these are glassy, such an explanation is not satisfactory. We would then have to assume that gravitative separation of the crystallized mineral-components had first taken place, and then had been followed by a sudden remelting of the crystals before final crystallization into a porphyric lava occurred. It is difficult to see how such a complicated process could have taken place in a fairly rapidly solidifying lava flow, although such assumptions may well account for many differentiation processes in deep-seated magma basins.

In the lava flow, part of which now forms the Lounatkorkia hill, we have the labrador-syeniteporphyry portion of the lava located above the quartz-porphyry portion. The spec. gravity of the labrador-syenite is, according to the determinations of Lemberg in different specimens, 2,762 and 2,764, that of the quartz-porphyry 2,657, 2,681, 2,698 and 2,702. The specific gravity of the labradorite phenocrysts ought, according to their chemical composition, to be 2,69. Here we thus have a case where the heavier portion of the lava forms the upper part of the flow. Of course the present surface does not correspond to the original surface portion of the lava, this having been carried away by erosion, since the holocrystalline ground-mass of the labrador-syenite-porphyry could not have been formed very close to the upper surface of the lava flow. But even if there had been still another variety of lava above the plagioclase porphyry, the fact remains that this heavier rock now forms the upper part of the flow and the lighter quartz-porphyry the lower portion. The hypothesis of crystallization-differentiation therefore does not apply to the differentiation of the Lounatkorkia lava flow.

The whole occurrence of the big porphyry lava at Hogland points towards it being an »areal eruption» of the Wiborg rapakivi magma emanating from one or several fissures. The differentiation of the lava has most probably taken place already before the eruption of the lava. The almost identical chemical character, of the basic rock varieties among the rapakivis of the Aspö Islands north of Hogland with the labrador-syenite-porphyry shows that similar differentiates as those in the lava at Hogland have occurred in the rapakivi-magma of the huge Wiborg rapakivi-mass, solidifying here as coarse-grained quartz-syenites of the potash series. The occurrence at Lounatkorkia may therefore be regarded as a place where a portion of such an already differentiated magma has been poured out through a fissure and formed a composite lava flow.

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Les publications de la Société seront éditées dans la série *Bulletin de la Commission géologique de Finlande* et seront distribuées

1. aux membres de la Société,
2. aux personnes ayant reçues la présente série,
3. aux institutions et aux associations désirant entrer en échange des publications.

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