GEOLOGINEN TUTKIMUSLAITOS

BULLETIN
DE LA
COMMISSION GÉOLOGIQUE
DE FINLANDE

N:o 215

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

XXXVI

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WALTER WAHL
ZUM 85-JÄHRIGEN GEBURTSTAG.

Am 9. Sebtember beging unser Ehrenmitglied Professor emeritus Dr. Walter Wahl seinen 85-jährigen Geburtstag. An diesem Tage konnte er auf ein Leben mannigfältiger wissenschaftlicher Tätigkeit zurückblicken.

Für Walter Wahl begann die wissenschaftliche Laufbahn 1897, als er die Universität zu Helsinki bezog, damals noch Alexanders Universität. 1902—03 ging er doch nach Heidelberg, um dann 1903 das FK-examen in Helsinki zu bestehen. Sein Hauptfach war Chemie, aber schon damals hatte Wahl grosses Interesse für Mineralogie und Petrographie. In den nächsten Jahren begab er sich nach Ostkarelien, hierdurch einem allgemeinen Bestreben finnischer Naturwissenschaftler zur Erforschung dieses von Finnen bewohnten Landstriches folgend. Es waren die postjotnischen Quarzdiabase des Svirgebietes, die Svirdiabase, die Wahls Interesse besonders erweckten. Aus diesen Studien kristallisierte seine Doktorsarbeit über die Enstatitaugite aus, die er 1906 öffentlich verteidigte. Damit war für Wahl der Weg zur Mineralogie und Petrographie gefunden, ein Weg, den er nur selten verlassen hat. In den nächsten Jahren finden wir ihn wieder im Auslande, in Göttingen 1909—10 und in London am Davy-Faraday Laboratorium 1911—13. Er besuchte auch den Internationalen Geologenkongress in Mexico sowie die Vereinigten Staaten.

Im Jahre 1918 wurde er an die Åbo Akademi als Professor der Chemie berufen, eine Stellung die er bis 1924 innehatte. Unter dieser Zeit war er auch zwei Jahre Dekan der chemisch-technischen Fakultät. Unter dieser Zeitspanne entstand Wahls Arbeit über die Gesteine des Wiborger Rapakivigebietes, die er 1925 veröffentlichte. 1924 verliess er seine Stellung an der Akademie und zog nach Helsinki zurück. Er lebte nun ganz seiner Forschung, wurde aber 1937 auf eine persönliche Professur für Chemie an die Universität Helsinki berufen, eine Stellung, die er bis 1946 innehatte. Wahl hatte mit eigenen Mitteln den ersten Massenspektrographen nach Finnland geschafft, der nun in seinem Laboratorium aufgestellt wurde. Der leidige Krieg hatte jedoch für Wahl, wie auch alle anderen Wissenschaftler, eine hemmende Wirkung. Ferner aber war er auch gezwungen, nach dem Tode seines Bruders 1940 die Leitung des Familienunternehmens Wahl & Co zu übernehmen, eine Aufgabe, die ihn bis 1961 festhielt.

Trotz der vielen anderweitigen Verpflichtungen, hat Wahl sich immer wieder der Wissenschaft hingegeben. Seine umfassenden regionalen Arbeiten über das Onega Gebiet und das Wiborger Rapakivi Gebiet sind von grundlegender Bedeutung. Sie zeigen uns auch, dass der Chemiker ein umsichtiger Feldgeologe war, der dem Beobachtungsmaterial mit grosser Schärfe eine Deutung zukommen liess. Für die Auffassung von dem geologischen Geschehen innerhalb der Svekofenniden wurde dann Wahls Deutung der Genesis der Granite in diesem Gebiet. Durch diese Arbeit gab Wahl der Graniteinteilung durch Sederholm erst eine genetische Deutung, die von grösster historisch-geologischer Bedeutung ist. Diese Gedanken von Wahl sind im Übrigen auch auf andere Orogene anzuwenden. Ein anderes Gebiet, welches das Interesse von Wahl erweckte, waren die Meteorite und Tektite. Man kann gut sagen, dass Walter Wahl einer der anerkannt besten Kenner dieser Bildungen ist.

Die wissenschaftlichen Arbeiten von Wahl zeigen vor Allem, dass ihm jegliches Schematisieren fern war. Ein Zug der ja heute leicht bei vielen Verfassern zu finden ist. Wahl sieht in den durch die Natur dargebotenen Dingen stets das jeweils Spezifische und Individuelle. Dadurch werden seine Deutungen auch den mannigfaltigen Entwicklungen in der Natur gerecht. So muss das Werk den Meister loben. Und so kam es auch, dass er, der Chemiker, 1956 zum Ehrenmitglied unserer Gesellschaft ernannt wurde. Wir betrachten ihn als einen der unseren. Später 1963 verlieh ihm die Åbo Akademi den Dr. phil. honoris causa, eine Ehrung über die sich alle mit ihm freuten.

Eine Passion hat Walter Wahl, nämlich das Reisen. Auch heute noch folgt der Rüstige nach dem Ruf in die Ferne. Er will Sehen und Erleben, und aus dem Kaleidoskop der Beobachtungen findet er immer noch wertvollen Stoff zur Weiterarbeit. Wir wünschen unserem geehrten Ehrenmitglied noch viele Jahre voller Gesundheit und Wirkens. Glück Auf Walter Wahl!

A. A. T. M.

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ON THE STRUCTURE AND PETROGRAPHY OF THE IPERNAT DOME, WESTERN GREENLAND

BY

RAIMO LAUERMA

Geological Survey of Finland, Otaniemi

CONTENTS

ABSTRACT

A well exposed area with a Precambrian gneiss dome was investigated. Photogeological methods were used to a great extent; the aerial photographs are reproduced and a geological map, scale 1:75 000, is presented.

The folding appears to be rather gentle. For the greatest part the rocks crystallized under the conditions of hornblende-granulite subfacies. Subsequently, a large proportion of them metamorphosed retrogressively.

Chemical analyses of three granodioritic and quartz dioritic rocks and one garnet are presented. Refraction indices and unit cell dimensions of three garnets, optical properties of some pyroxenes and triclinicity values of some potash feldspars are given. Twenty determinations of the sodium and potassium content of representative samples indicate that potassium is not concentrated in the gneiss core of the dome.

ACKNOWLEDGMENTS

My first information about the gneiss dome that is the subject of this paper came from Mr. K. Ellitsgaard-Rasmussen, who is at present Director of the Geological Survey of Greenland. It was while I was a member of an expedition that carried out geological investigations under his direction in the Precambrian area of West Greenland in the summer of 1952. I asked to be assigned the geological investigation of the dome and its surroundings, and I received the assignment the following year. The executive staff and other personnel of the Geological Survey of Greenland have also endeavored to arrange the best possible facilities for field work and follow-up research. I wish to extend particular thanks to Mr. K. Ellitsgaard-Rasmussen and Prof. A. Noe-Nygaard for their encouraging attitude toward this study.

The field investigations were carried out in the summers of 1953, 1954 and 1958. During the first two summers the work was done in conjunction with the reconnaissance mapping of the Precambrian area of West Greenland. In 1953 the expedition was headed by Mr. H. Sørensen and in 1954 by Mr. A. Berthelsen (both now professors). It pleases me to acknowledge their help in, for example, organizing my field investigations.

During the course of the field work, I was assisted by a number of people: in June 1953 by Mr. N. Henriksen and in July of the same year by Mr. E. Bondesen, who lent me a hand the following summer as well; then, for a short time, in August 1954 by Mr. Adam Dahl, a Greenlander from Itivleq. During the entire five-weeks' expedition in the summer of 1958, I had the assistance of a pair of Greenland seminar students, Ilanguaq Jensen from Thule and Samuel Kleinschmidt from Godthaab. To all the persons mentioned I wish to express my heartfelt gratitude for their valuable support and good companionship under conditions that were sometimes not quite enjoyable.

The laboratory work was done for the most part in 1955—56 at the Geological Institute of the University of Helsinki and in 1959—63 at the Geological Survey of Finland. For placing the facilities of these establishments at my disposal I wish to thank Professor M. Saksela, head of the former institute; Professor A. Laitakari, ex-director, and Professor V. Marmo, present director of the Geological Survey; Professor A. Kahma, chief of the Ore Department, and Professor A. Simonen, chief of the Petrological Department of the Survey.

When I presented a portion of my research results at the University of Helsinki for my degrees of Mag. Phil. and Lic. Phil. in 1955 and 1956, Professors M. Saksela and J. Seitsaari read my papers and offered valuable criticism. Dr. M. Härme read the manuscript of the present study and Dr. M. Okko scrutinized it, mainly from the formal point of view; both of them made valuable suggestions.

Dr. H. B. Wiik made the chemical analyses. Dr. K. Hytönen carried out many of the mineralogical determinations. Mr. L. Hyvärinen, Mag. Phil., made the ore-microscopical investigations. Mr. I. Rajala and Mr. A. Nurmi, Mag. Phil., prepared the mosaic of aerial photographs. Mr. E. Halme took the photomicrographs and Miss Karin Dahl drew the appended maps. Mr. Paul Sjöblom, M. A., translated the greatest part of the manuscript from the Finnish. Mr. T. C. R. Pulvertaft made some linguistic corrections. It gives me pleasure to acknowledge my appreciation to all these persons.

The Geodetic Institute, Copenhagen, kindly gave its permission to publish the appended aerial photographs.

At various stages of my work, I received financial aid from the University of Helsinki, the Outokumpu Foundation (Outokumpu Oy:n Säätiö) and the Cultural Foundation of Finland (Suomen Kulttuurirahasto), and for this I wish to express my gratitude.

Finally, I want to record my particular debt to Professor E. Wegmann, who during my visits — totalling three months — in 1957 and 1961 — to Neuchâtel, Switzerland, spared neither time nor effort in teaching me his methods of tectonic analysis.

Helsinki, April, 1964.

Raimo Lauerma

INTRODUCTION

When the Grønlands Geologiske Undersøgelse (Geological Survey of Greenland, abbr. GGU) started its investigations in western Greenland in 1946, one of its goals was to carry out in ten years a preliminary survey, mapping included, of the exposed coastal area situated between latitudes 61° and 69°N (Fig. 1). Roughly 1 000 km long and between 50 km and 150 km broad, this area is bare of ice and snow in the summertime. It is mountainous and very difficult to travel in, but it is broken up by scores of great fjords, which with their numerous branches and bays offer the investigator revealing geological profiles. The longest of the fjords are from 100 km to 150 km long. In view of the time reserved for the preliminary investigations, the limited possibilities of the expeditions, made up as they were of only a few geologists and students of geology, and the vast extent of the region as well as the difficult terrain, it was necessary to concentrate the investigations during the ten-year period almost wholly on the shores of the fjords and islands. It was only by photogeological methods in the main that data could be obtained from the areas, many of them stretching from 20 km to 40 km in breadth, which lie between the fjords. Any geological map drawn on the basis of such observation material can give at best only a very limited picture of the geology of such an area — predominantly comprising in this case migmatic rocks. Consequently, in addition to the reconnaissance mapping, a few suitable, smallish areas were investigated in fair detail in order to obtain a better view of the geology of the region as a whole. One sample area of this kind selected by GGU was the vicinity of the gneiss dome called Ipernat dome in the present paper.

This dome structure was first observed by H. Sørensen in 1950 while examining aerial photographs of the Godthaab district for geological reconnaissance mapping. These photographs, which are reproduced in Plates IV and V, clearly reveal a ring shaped structure. The area of the dome itself had most probably not been visited by any geologist before the present author started his investigations there in the spring of 1953. On the shore of Godthaabsfjord, near the dome, however, some geologists had done research work in connection with geological reconnaissance mapping.

The name »Ipernat dome» cropped up spontaneously during the field work carried out in the summer of 1953, when it was observed that it is situated in a boggy and sheltered tundra area, which breeds mosquitoes in extraordinary abundance (mosquito in Greenlandic = »ipernaq», plural »ipernat»). The name was originally intended only to serve temporarily in the field, but since not a single place name is known to exist in the area of the dome, it has remained in regular use. Ipernat dome refers only to the geological structure in question and thus does not represent any actual geographical place name.

The structure and genesis of different types of domes have long been the subject of numerous studies and discussions. The bedrock in large areas of Greenland is exceedingly well exposed. And when the present investigation was started, it was therefore known that this circumstance would make it possible to map and investigate the structure of this gneiss dome with considerable accuracy. Only in the light of detailed investigations of many examples can a reliable picture of the mode of origin of different kinds of domes be formed.

SITUATION AND OUTLINE OF GEOLOGY

Geographically, the situation of the area investigated is between Fiskefjord and the northernmost branch of the Godthaabsfjord, Qugssug, or between latitudes 64° 35'N and 64° 55'N and longitudes 51° 15'W and 51° 35'W. The site lies some 60 km northeast of the capital of Greenland, Godthaab, and roughly 80 km west of the western margin of the Inland ice. The area surveyed and mapped covers some 300 sq. km and its location is marked in Fig. 1.

When the GGU started its investigation in West Greenland in 1946, scarcely anything more was known about the geology of the region between the 61st and 69th parallels — where the Ipernat dome is situated — than that the rocks were principally migmatic, and probably belonged to the Precambrian. The southernmost part of Greenland — the part south of latitude 61°N — was geologically much better known. Wegmann (1938), for example, had published a masterly study on the geological structure of southern Greenland, with special attention given to the age relations of the rocks. To the Precambrian orogeny described in this work, he gave the name »Ketilidian».

During the early stages of the field explorations carried out by GGU in 1946—47, conclusive evidence was discovered in the vicinity of Søndre Strømfjord, situated between latitudes 66° and 67° N, of the existence of two Precambrian orogenies of different ages (Ramberg 1948a, Noe-Nygaard 1952). The younger orogeny, extending from around Søndre Strømfjord northwards at least as far as Jacobshavn (latitude 69°N), was named by Ramberg (1948a) the »Nagssugtoqidian».

The region between the 61st and 66th latitudes, or between the area of the Ketilides studied by Wegmann (1938) and the area of the Nagssugtoqides, is referred to by Noe-Nygaard and Ramberg (1960) simply as the pre-Nagssugtoqidian fold belt. The region is generally regarded as belonging to the Ketilides, but this view has not yet been proved.

The pre-Nagssugtoqidian rocks have been divided into several complexes, each showing specific structural characteristics. (Berthelsen, 1957, 1961; Noe-Nygaard and Ramberg, 1960). The area of the present study is situated in the eastern part of the Nordland Complex, near the boundary of the Godthaab Complex. Berthelsen (1957, p. 176) has drawn the Nordland Complex as covering an area of approximately 5 000 sq. km. However, the eastern and northern boundaries of the complex are very imperfectly known.

It is only in the last few years that reliable age determinations based on radioactive disintegration have been published from the Precambrian rocks of West Greenland. According to Moorbath, Webster and Morgan (1960) the Rb-Sr date for the biotite contained in the Julianehaab granite is 1590×10^6 years and the K-Ar date for the same biotite 1597×10^6 years. The emplacement of the Julianehaab granite occurred during the late stages of the Ketilidian orogeny (Wegmann, 1938), and these age determinations thus give a fairly reliable minimum age for the orogeny. On geological grounds Wegmann (op.cit. p. 132) expounded the conception that the Ketilides probably belong to a middle Precambrian cycle.

Recently, Armstrong (1963) has published K-Ar dates for two biotites and one feldspar from the Precambrian of West Greenland. Biotite from the granodioritic gneiss at the northeast end of Søndre Strømfjord gave 1650×10^6 years as a minimum age for the Nagssugtoqidian fold belt. A date of 2700×10^6 years was obtained for the biotite from a granodioritic gneiss at Godthaab in the pre-Nagssugtoqidian fold belt. The place where the latter sample was taken belongs to the Godthaab Complex and it is situated about 60 km southwest of the area of the present investigation.

According to the studies so far made, the part of the western coast of Greenland between the 61st and 69th parallels N lat. consists mainly of migmatic rocks. The most common rocks are granodioritic to quartz dioritic gneisses, hypersthene gneisses, various amphibolites and hornblende gneisses. To a considerably lesser extent there occur sillimanite-, garnet-, cordierite-, graphite- or pyrite-bearing gneisses and schists as well as skarn-bearing rocks. Ultrabasic rocks are not uncommon and generally there is quite an abundance of basic dikes. The intensity of the regional metamorphism varies from granulite facies to amphibolite and epidote-amphibolite facies, according to the mineral facies classification of Eskola (1939).

A geological reconnaissance map on a scale of 1: 500 000 of the region between the latitudes 69°N and 63° 45′N, or between Jacobshavn Isfjord in the north and Buksefjord in the south, was published by Noe-Nygaard and Ramberg (1960). This map covers the area of the Nagssugtoqidian fold belt studied so far and the northern part of the pre-Nagssugtoqidian fold belt. Together with its explanatory text this map gives a fairly good idea of what is nowadays known of the geology of this region, in which the area dealt with in the present study is situated.

The Ipernat dome consists of the central gneiss and several concentric rings. The two most conspicuous rings both in the field and in the aerial

Fig. 1. Location of area investigated in western Greenland.

photographs consist in the main of amphibolite. The outer ring extends across a relatively broad area in the northern, western and southwestern parts of the dome, reaching a breadth of as much as a couple of kilometers, wheareas in the eastern part it is only between 200 and 400 meters wide. The inner amphibolite ring, which varies from 200 to 300 meters in width, conformably surrounds the homogeneous gneiss massif comprising the entire center of the dome. The area between these two amphibolite rings, measuring between one and two km in width, varies petrographically: it contains, for example, granodioritic and quartz dioritic gneisses, hypersthene gneisses and minor amphibolite zones. The area covered by the dome structure is some 9 km long and 6 km wide. The longest diameter runs approximately N—S. The shape of the outer amphibolite ring may be described as a deformed ellipse, the southeastern part of which has been flattened.

The center of the dome consists of a granodioritic to quartz dioritic gneiss massif, the length of which on the present erosion surface is about 2 km and the maximum width 1.2 km. The form of this central gneiss and the inner amphibolite ring enclosing it is a fairly regular ellipse.

The rocks of the surrounding country are largely the same as those of the area of the dome itself, namely, quartz dioritic and granodioritic gneisses, hypersthene gneisses, pyroxene amphibolites, amphibolites and hornblende gneisses. Present are also ultrabasic rocks as well as diabase and pegmatite dikes. In one place north of the dome, there occurs a layer of garnet-cordierite gneiss about 50 meters thick.

The surroundings of the dome, with the exception of the area to the east, are noticeably more complex in structure than the dome itself. Investigation has brought to light numerous large folds, some of which are between 0.5 km and 2 km in breadth. Many of them can be clearly distinguished in aerial photographs even without a stereoscope. Their situation, form and size are best observed from the geological map and the appended aerial photographs.

PHYSIOGRAPHY

Topographically, the area of the dome and its immediate extension eastward and northeastward is exceptionally level and low-lying compared to the surrounding country and to western Greenland as a whole. The highest points in the area of the dome are in its northeastern part, rising to altitudes of between 100 and 190 meters above sea level. Particularly the southern half of the dome, with the exception of the amphibolite zones, is very level (Fig. 2).

At distances of ten to twenty kilometers toward the southwest, west, north and northeast of the dome, the terrain generally rises to elevations of 300 to 700 meters; but even the highest summits are rounded. The tract situated to the southeast of the area of the investigation, on the southern and southeastern side of the northernmost branch of Godthaabsfjord, differs sharply from the topography just described. On the islands of Godthaabsfjord and to the southeast of them, there rise steep-walled, in many cases pinnacled mountains, which generally reach altitudes of between 1 000 and 1 600 meters; and the district is one of the most elevated in all of western Greenland. This sharp topographical difference between two neighboring tracts has long commanded the interest of researchers (Bøggild 1917, pp. 2 and 25).

The topography of the area investigated reflects amazingly faithfully the local geological structures and also changes in the composition of the rock. The resistance of different rocks to erosion has differed, and in many cases it has been nearly as selective as, for instance, the polishing hardness utilized in ore microscopy as a method of identification. The amphibolite zones occur as ridges, which rise distinctly higher than the granodioritic and quartz dioritic gneisses surrounding them. The height of these ridges above the surrounding terrain varies considerably and seems to depend on, among other things, the breadth of the amphibolite zone. In most cases it is in the order of one to twenty meters but rises in some cases to a height of about thirty meters in otherwise level stretches.

A good example of the foregoing is the eastern and southeastern portion of the dome's outer amphibolite ring. In the eastern part of the ring the

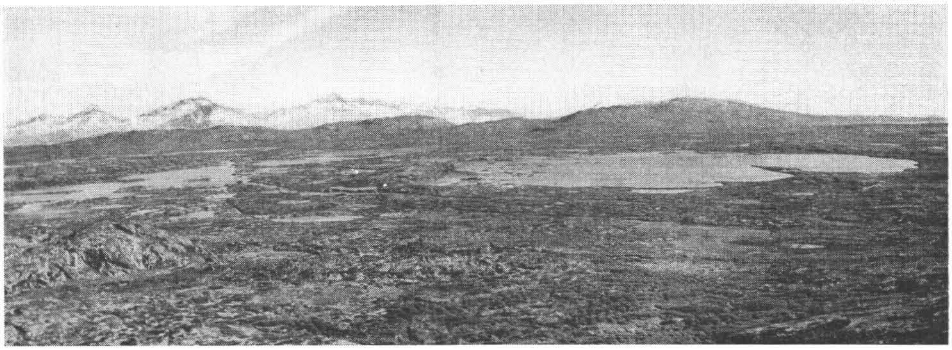

Fig. 2. View over southern part of dome towards south. Rounded hills in background are situated between dome and Godthaabsfjord and snow-covered mountains on islands of Godthaabsfjord.

rock consists of pyroxene amphibolite and in this area the zone forms a ridge about 250 meters wide which rises some 20 to 30 meters above the rather flat surroundings of migmatic rock. The slopes of the ridge follow the contact between the amphibolite and granodioritic or quartz dioritic gneiss mostly within 20 to 40 m. As the amphibolite zone proceeds southward, more siliceous portions appear in it in increasing measure. No sharp contacts between them and the amphibolite have been observed; in lieu of them there are gradual transitional zones. As the composition of the zone becomes more quartz dioritic, the height of the ridge also diminishes. In the southeastern part of the ring, where the composition of the zone and the surrounding rock is almost the same, the zone scarcely rises above the rest of the terrain.

In odd contrast, as it first seems, to the topographic mode of occurrence of the amphibolite zones is their very strong tendency to disintegrate under present conditions. In many places the amphibolites have disintegrated into debris to depths of 20 cm to 50 cm or perhaps more (Fig. 3), which makes the obtaining of fresh rock specimens at times troublesome. The granodioritic and quartz dioritic gneisses situated in the immediate proximity of the strongly weathered amphibolites have, on the other hand, scarcely undergone similar disintegration. However, in some instances, especially in low-lying and boggy places, these gneisses have broken up into large sharp-edged blocks.

The reasons for the varying mode of topographic occurrence of the rocks in the area just described have not been throughly clarified in connection with the present study. The observations with reference to it were made principally in an area where the elevation above sea level is less than 200 meters and where the relief usually amounts to only a few dozen meters.

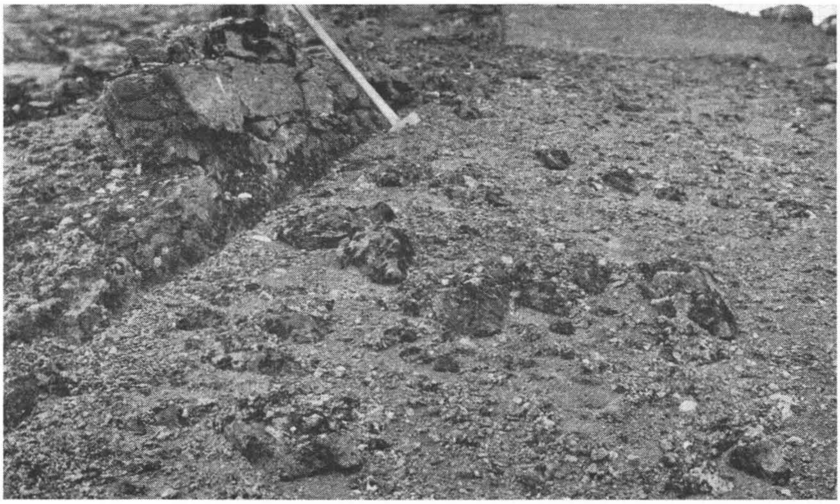

Fig. 3. Debris in situ, formed by disintegration of amphibolite layer. More resistant bed at hammer. Southern part of Quagssugtarssuaq. Altitude about 250 m.

The most plausible explanation is to be found in the difference between the amphibolites and the granodioritic and quartz dioritic gneisses with respect to their resistance to glacial erosion. The granodioritic and quartz dioritic rocks split and crack fairly easily into large, sharp-edged blocks, and these the glacier was able to break loose and carry away much easier than in the case of the tougher amphibolite.

The dome is situated in a fairly sheltered area in the interior and most of it and its surroundings are less than 100 meters above sea level. Accordingly, the local vegetation is rather abundant compared to the coast and higher regions. Lichen, moss and shrubs are a common occurrence in the area, while bushes between 0.5 m and 1 m in height also grow in the most protected places. Compared to the ideal conditions prevailing on the shores of the fjords and islands as well as in the regions situated at altitudes of more than 500 or 600 meters, the quality of the exposures is much inferior and less suitable for detailed petrological investigations.

On the other hand, the density of exposures in the area under investigation is quite sufficient for much more detailed geological mapping than it has until now been feasible to carry out. Large areas are wholly exposed and the distances between outcrops are at most a few dozen meters. There are hardly any glacial deposits in the area and the debris is also predominantly in situ. In applying photogeological methods one will observe that the local vegetation tends to clarify rather than conceal the lithological variations of the rocks. In aerial photographs taken during the summer

(Plate IV) the rock zones consisting of dark amphibolites appear distinctly paler than the light granodioritic and quartz dioritic gneisses surrounding them. This phenomenon is due to the various lichen and moss vegetation, following as it does with astonishing exactness the different rocks. The vegetation boundaries visible in the aerial photographs follow the contacts of the amphibolite zones in many cases within 5 to 10 meters.

Considering the local circumstances, it is not difficult to propose plausible explanations as to the reasons for this correlation between geology and vegetation. The amphibolites rise topographically higher, weather much more easily and afford plant life different moisture and pH conditions and chemically a different kind of ground to grow on than granodioritic and quartz dioritic gneisses.

The correlation between the rocks and the vegetation has been perceived and it has been utilized in the geological mapping only in the area of the dome itself and on its eastern and southeastern sides, or, in other words, where the vegetation is relatively rich and where aerial photographs have been taken during the summer.

MAPPING METHODS

The map material available for the present study has included the topographic map of the Danish Geodetic Institute on a scale of 1: 200 000, contact prints of aerial photographs on a scale of 1: 40 000, and a sketch map on a scale of 1: 40 000 drawn according to the aerial photographs. In 1954 topographic maps on a scale of 1: 50 000, which had just been completed, were also obtained.

In simplifying the latter maps for the geological map, over 200 smaller lakes and ponds have been left out of the map for sake of clarity. During field investigations they have been of great value in the precise location of observation sites.

The field work was conducted from camps located in different parts of the area in the summers of 1953, 1954, and 1958. The field observations in the largest part of the region were made by the present author. E. Bondesen in the summer of 1954 mapped a part of the area north of the dome. Some details of the geological map in the vicinity of the small inlet Kikiagdlit on the shore of Godthaabsfjord are drawn according to a sketch map made by H. Sørensen.

Two different methods have chiefly been used in investigating the geological structure of the dome and its surroundings. The one is the photogeological method and the other is based on tracing suitable rock zones in the field step by step.

The dome was discovered by examining aerial photographs, which have also greatly facilitated structural investigations. The geological structure of the region was analyzed in a preliminary manner before field investigations were undertaken by means of a stereoscopic study of the aerial photographs, and these pictures have been used at various subsequent stages of mapping. During the field work, moreover, they have proved exceedingly useful. Structural features, such as the strike of the bedding or foliation, the position of contacts, folds and faults, are often revealed by aerial photographs in considerable detail, even without a stereoscope, and it has been possible the whole time to compare them with the results obtained through field observations. It has thereby been possible to concentrate the field investigations directly on the structural keypoints.

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The strike of the rock zones generally appears in aerial photographs as a banded pattern. The reason for this pattern can usually be traced to many factors of secondary character, the analysis of which is not always by any means easy.

One cause is the occurrence of two distinct joint sets in the stratified and foliated rocks. One joint set runs parallel to the bedding or foliation. The strike of the other joint set is parallel to the strike of the bedding or foliation, but its dip is roughly perpendicular to the dip of the bedding or foliation. The snow and moisture collecting in the stepped furrows produced by these two joint sets, the frequently richer vegetation in them, their tonal and textural relations, and other such factors enhance the banded pattern running parallel to the bedding or foliation of the rock. The joint sets cutting across the stratification perpendicularly or obliquely do not have the same effect in this respect because they are less regular and they lack the corresponding continuity.

The morphological features caused by selective erosion have been described already including the occurrence of amphibolite zones as ridges rising mostly from one to thirty meters above the surrounding terrain. Since it is possible with a stereoscope to distinguish without difficulty on the 1: 40 000-scale aerial photographs available altitude differences of about one meter between two adjacent spots, the differences in altitude dealt with in the foregoing have been more than sufficient to enable one to trace the zones.

The correlation between the rocks and the vegetation which has similarly been elucidated previously, greatly facilitated the geological mapping of the dome itself by photogeological means.

Faults, mylonitic zones and joints are to be seen best in aerial photographs taken early in the spring (Plate V). Erosion has usually worn such places deeper than the surrounding terrain, and in thus produced, commonly rectilinear depressions and ravines the snow lies longer and accentuates their appearance in the pictures. Otherwise, pictures taken at this season serve the purposes of photogeological research much less effectively than aerial photographs taken during the summer.

Four different series of aerial photographs taken of the mapped area have been available, all of them contact prints on a scale of approximately 1: 40 000. Only one series was taken in the summertime and it starts from the northern end of the dome, then extends over it toward the south. In the photogeological sense the first pictures in this series must be judged excellent (Plate IV).

Another method in the investigation of the geological structures has been that of following suitable rock zones in the field step by step as far as possible. Usually the same zone has been followed for several kilometers, sometimes for distances of as much as 10 km to 20 km. The majority of the rock zones followed have been amphibolites or related rocks, which occur as ridges rising above the surrounding ground. This procedure of tracing of a single zone furthermore makes it possible to study systematically any evidences of petrographic changes occurring along the strike of the zone. In most cases, the tracing of rock zones in the field has amounted to checking and verifying the results achieved by photogeological means.

The lack of suitable beds with sufficiently distinctive characteristics to serve as reliable marker beds has proved a major hindrance in carrying out structural investigations. Of the three chief rock types in the region, granodioritic and quartz dioritic gneisses, hypersthene gneisses and amphibolites, none suits this purpose, for they occur in extensive areas much in the same way and seem, in addition, to grade into each other along the strike of the layers. Garnet-cordierite gneisses, which probably would have proved suitable for the purpose, were met during the course of the investigations in only one place. Sporadically, there occur in the amphibolites narrow rusty layers. They might have been of considerable help in this respect except that far more detailed field work and topographic maps would have been required.

One difficulty, too, has been the fact that no primary features have been found that could be used for the determination of the top of beds, such as cross-bedding, graded bedding or pillow structures.

PETROGRAPHIC DESCRIPTION

The following laboratory methods were used in the petrographic study of the rock specimens.

The optical properties of the pyroxenes, the amphiboles, and some of the plagioclases reported were determined by means of the Leitz universal stage.

If not otherwise stated, the composition of the plagioclase was determined by comparing some refractive index of the plagioclase with that of Canada balsam and/or determining the maximum extinction angle of albite twins in sections normal to 010.

The triclinicity values of the potash feldspar were determined by X-ray methods, described by Goldschmith and Laves (1954). The potash feldspar was separated for this purpose by means of heavy liquids. The determinations were carried out by K. Hytönen.

In most cases the modal analysis was carried out by means of the Leitz integration stage, but in some cases the point counter method was used. Each analysis is based on a single thin section. Accordingly, the results must be regarded only as semi-quantitative, particularly with respect to the rocks of coarser grain.

The specific gravity of the rocks was determined by weighing the rock samples in the air and in the water with a balance constructed specially for this purpose.

PYROXENE AMPHIBOLITES, AMPHIBOLITES AND HORNBLENDE GNEISSES

This category of rocks includes numerous fairly different rocks that have been marked on the geological map with the same green color. These rocks often occur in the same zones and also grade into one another. In the structural descriptions they have simply been lumped together as amphibolites.

It has been difficult to find an accurate and generally approved term for the rocks here designated as pyroxene amphibolites. In the light of microscopic studies, their typical mineral assemblage in the mapped area is plagioclase, green hornblende, diopside and hypersthene in varying proportions. Rocks of more or less the same description have long been the object of active investigation in many countries and they have been given many different names, such as »basic division of the charnockite series», »basic charnockites», »basic granulites», »basic pyroxene granulites», »norite granulites», and »pyroxene amphibolites». Recently, Berthelsen (1960, pp. 20—21) suggested yet another nomenclature for these rocks, designating the pyroxene-hornblende-plagioclase rocks pyribolites and the pyroxene-plagioclase rocks pyriclasites.

The term pyroxene amphibolites has been used in the present study because they grade in the region investigated into normal amphibolites, and hornblende seems to belong as an essential part of the mineral assemblage, even in rocks mainly composed of pyroxenes and plagioclase. The term is, however, less appropriate when applied to rocks with very little hornblende.

In the field the pyroxene amphibolites, amphibolites and hornblende gneisses are more or less schistose. The pyroxene amphibolites, in particular, are generally distinctly layered, with dark and light layers alternating. The thickness of the layers varies from a few millimeters to some meters, and the impression is most often gained in the field that the layering is due to the original stratification. On the other hand, in some cases, at least, the possibility of segregation banding of metamorphic origin (Turner, 1941, pp. 1—16; Billings, 1954, pp. 344—345) has to be taken into consideration. This is true notably when the thickness of the layers is to be measured only in millimeters or centimeters.

The pyroxene amphibolites often occur in layers alternating with hypersthene gneisses, and in some places they appear to grade into each other along the strike of the layers.

Lenses of ultrabasic rocks generally occur in the pyroxene amphibolites, amphibolites and hornblende gneisses, and along their contacts, especially against granodioritic and quartz dioritic rocks. They are described more in detail on page 56. Furthermore, the pyroxene amphibolites and amphibolites sporadically contain narrow rusty intercalations, which vary in thickness from a few centimeters to several meters. The rustiness on the weathered surface seems generally to be due to the presence of very slight amounts of megascopically hardly perceptible disseminated sulphide minerals. On the other hand, the weathered hypersthene-bearing rocks are also generally rusty, and this circumstance reduces the value of rusty layers as marker beds, except when they are traced step by step.

In a few places, such as on the eastern shore of the small inlet of Kikiagdlit, small calcite lenses between five and thirty centimeters in diameter have been found in the amphibolite. Skarn lenses of the same size are a fairly common occurrence in the amphibolites.

Of the typical rock samples collected during the field investigations, some fifty were classified as various amphibolites and hornblende gneisses. Nearly forty thin sections were prepared from these samples, a few of which were seen under the microscope to be hypersthene gneiss. After a study of the thin sections, those specimens best representing the petrography of the amphibolites and hornblende gneisses of region as a whole were selected, and they are described in the following. The refractive indices, optic axial angles and extinction angles ($c \wedge \gamma$) reported here were determined by K. Hytönen and the ore microscopical investigations were carried out by L. Hyvärinen. The specimen numbers refer to the collection of the Geological Survey of Greenland.

Pyroxene amphibolite. Specimen No. 14 511. Locality 1 (Plate I), eastern part of the outer amphibolite ring.

The rock is distinctly layered. Dark and light layers a few dozen centimeters thick alternate in it. The specimen here described represents a dark layer. The color of the rock is a dark gray with a faint greenish tinge. Both megascopically and microscopically the rock is clearly schistose. The grain size varies from 0.5 mm to 2 mm. The texture is nematoblastic, almost granoblastic (Fig. 4). The specific gravity is 3.07.

The chief constituents are plagioclase, green hornblende, clinopyroxene and orthopyroxene. In addition, there is a small amount of opaque material, which appears to be for the greatest part limonite. Under the ore microscope a few very small grains of pyrite and pyrrhotite were discovered. The modal analysis is presented in Table I on p. 28.

The plagioclase is mostly polysynthetically twinned. It is clear and lacking in alteration products. It is somewhat zoned in that the centers of the grains are more basic than the marginal portions. Its α varies 1.563 — 1.567 and its β 1.568 — 1.571, which corresponds to An_{68}—An_{75} (Deer et al. Vol. 4, p. 131)

The hornblende is pleochroic from yellowish green to green. Its $2V\alpha = 76°$ and $c \wedge \gamma = 16°$.

The clinopyroxene is almost colorless, being faintly greenish. Its $2V\gamma = 58°$, $c \wedge \gamma = 45° — 47°$, $\alpha \sim 1.690$ and $\gamma \sim 1.720$. Refractive indices were not determined from the same grains as other optical properties. If a mineral of the pure diopside-hedenbergite series is involved, the refractive indices correspond to about 35—40 mol. % hedenbergite (Deer et al. Vol. 2, p. 62).

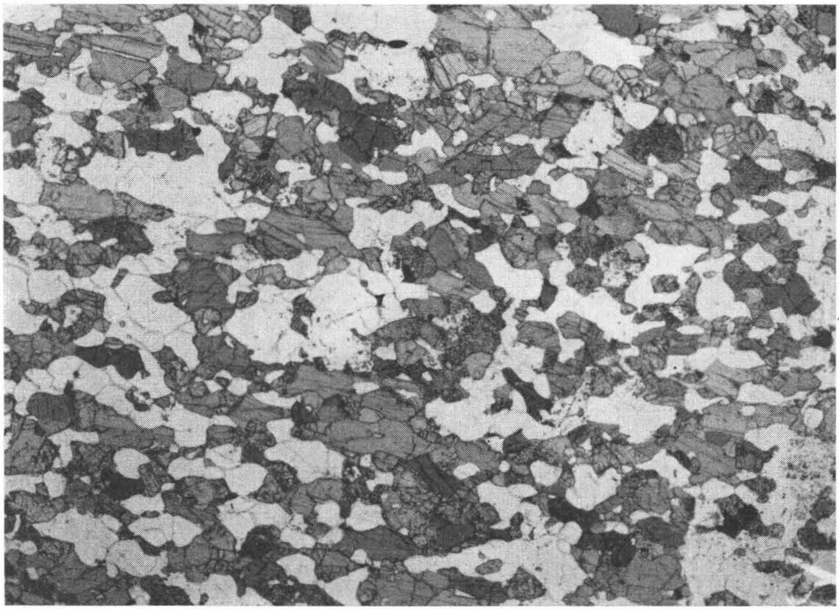

Fig. 4. Texture of pyroxene amphibolite. Specimen No. 14 511. One nicol. Magn. 9 x. White grains are plagioclase and darker grains are hornblende and pyroxenes. Photo: E. Halme.

The orthopyroxene is distinctly pleochroic, α red, β yellow and γ grayish green. Its $2V\alpha = 59°$, $\alpha \sim 1.705$ and $\gamma \sim 1.719$. Refractive indices were determined from other grains than the optic axial angle. The γ refractive index, which according to Deer et al. (Vol. 2, p. 29) is the most accurate optical method of determining the En:Fs ratio of an orthopyroxene, corresponds to about 40—45 mol. % ferrosilite, i.e., to hypersthene.

The thin section has a couple of spots where the hornblende at least seems to contain diopside as an inclusion, and several diopside grains possess a narrow green rim, i.e., evidence of alteration into hornblende. These are, however, exceptional instances. By far the greatest portion of mineral grains are wholly lacking in inclusions and alteration products. The mineral assemblage thus appears to be in a fairly steady state of equilibrium.

Pyroxene amphibolite. Specimen No. 14 512. Locality 1 (Plate I), eastern part of the outer amphibolite ring. About 0.5 m across the strike from the spot where the preceding specimen was taken.

This specimen represents a lighter, rusty layer. Megascopically the rock is conspicuously lighter than the previous specimen and brownish gray in color. Both megascopically and microscopically it is distinctly schistose.

The grain size is 1—2 mm. The texture is nematoblastic, almost granoblastic. The specific gravity is 3.04.

The principal constituents are plagioclase, clinopyroxene, orthopyroxene, titaniferous magnetite and green hornblende. In addition, the thin section contains a trace of biotite and apatite as well as, evidently on the surfaces of other mineral grains, limonite. Under the ore microscope there were detected, in addition to ilmenite and magnetite, a few very small grains of pyrrhotite. The modal analysis is presented in Table I on p. 28.

The plagioclase is generally polysynthetically twinned. It is clear and lacking in alteration products. Its $\alpha = 1.552$ and $\gamma = 1.561$, corresponding to An_{46}—An_{48} (Deer et al. Vol. 4, p. 131).

The clinopyroxene's $2V\gamma = 55° - 58°$, $c \wedge \gamma = 43°$, $\alpha \sim 1.695$ and $\gamma \sim 1.725$. Its color is a very pale green. The refractive indices were not determined from the same grains as other optical properties. If the mineral belongs to the pure diopside-hedenbergite series, the refractive indices correspond to about 45—50 mol. % hedenbergite (Deer et al., Vol. 2, p. 62).

The orthopyroxene's $2V\alpha = 52°$, $\alpha \sim 1.715$ and $\gamma \sim 1.730$. Refractive indices were determined from other grains than the optic axial angle. The γ refractive index corresponds to 50—55 mol. % ferrosilite i.e., to ferrohypersthene (Deer et al., Vol. 2, p. 28). It is distinctly pleochroic, α red, β yellowish green and γ greenish.

The hornblende is pleochroic from yellowish green to bluish green. Its $c \wedge \gamma = 12°$. Hornblende occurs in most cases as grains measuring 0.1—0.3 mm in diameter. Some of them occur as separate independent grains, some, again, as inclusions in hypersthene or diopside. In certain cases, the hornblende is clearly enclosed by a rim of diopside, which thus appears to be younger than the hornblende. In certain other grains, on the other hand, the diopside appears to alter into hornblende.

The very slight biotite content of the rock appears to be quite secondary and consequently younger than the other minerals. It occurs in many cases in the immediate proximity of oxide ore grains.

Pyroxene amphibolite. Specimen No. 14 515. Locality 2 (Plate I), northwestern part of the outer amphibolite ring.

The rock is megascopically distinctly schistose and its color is greenish dark gray. The mafic minerals, notably hornblende, can be distinguished as grains 1—6 mm long in a somewhat finer-grained matrix containing chiefly plagioclase. The diameter of the plagioclase grains is 0.5—1 mm. The texture is nematoblastic, almost granoblastic. The specific gravity is 3.12.

The chief constituents are plagioclase, green hornblende and hypersthene. Under the ore microscope only a few, quite tiny iron sulphide grains, evidently pyrite, were found. The modal analysis is presented in Table I on p. 28. On account of the coarse grain and broken character of the specimen, this analysis is not altogether reliable and the true proportion of plagioclase is likely to exceed the result obtained.

Polysynthetic twinning is to be seen in the plagioclase faintly or not at all. Weak zoning appears in some of the grains. The refraction index $\beta = 1.573$, corresponding to An_{80} (Deer et al., Vol. 4, p. 131).

The hornblende is pleochroic from pale green to dark green. Its $2V\gamma = 88° - 90°$ and $c \wedge \gamma = 23°$.

The orthopyroxene is markedly pleochroic, α red and γ grayish green. Its $2V\alpha = 78° - 81°$, $\alpha \sim 1.692$ and $\gamma \sim 1.702$. Refractive indices were determined from other grains than the optic axial angle. The γ refractive index corresponds to hypersthene containing about 30 mol. % ferrosilite (Deer et al., Vol. 2, p. 28).

Under the microscope, the plagioclase, hornblende and hypersthene are clear and nearly totally lacking in alteration products. Only in certain slightly broken grains is a weak alteration to be detected. The rock's mineral assemblage thus appears to be wholly in equilibrium.

Pyroxene amphibolite. Specimen No. 14 524. Locality 3 (Plate I), northwestern part of the outer amphibolite ring.

Megascopically the rock appears brownish dark gray. The weathered surface is rusty brown. The schistosity is weak, hardly detectable. The grain size is 0.5—1.5 mm. The texture is nematoblastic, almost granoblastic (Fig. 5) The specific gravity is 3.14.

The principal constituents are plagioclase, clinopyroxene, orthopyroxene, titaniferous magnetite and green hornblende. In addition, there is a small amount of apatite as well as, on the surfaces of the other mineral grains, limonite. Under the ore microscope a trifle of chalcopyrite and pyrrhotite were detected. The modal analysis is presented in Table I on p. 28.

The plagioclase is polysynthetically twinned. Its $\beta = 1.559$ and $\gamma = 1.563$, corresponding to An_{52} (Deer et al., Vol. 4, p. 131).

The clinopyroxene is very slightly pleochroic from pale green to somewhat darker green. It is slightly zoned. $2V\gamma$ varies from 52° to 64°. Its $c \wedge \gamma = 42°$ and $\gamma \sim 1.720$. In the pure diopside-hedenbergite series this refractive index corresponds to about 40 mol. % hedenbergite (Deer et al., Vol. 2, p. 62).

The orthopyroxene is weakly pleochroic, α red and γ grayish green. Its $2V\alpha = 55°$, $\alpha \sim 1.715$ and $\gamma \sim 1.730$. The latter refractive index corresponds to ferrohypersthene with 50—55 mol. % ferrosilite (Deer et al., Vol. 2, p. 28). The optic axial angle was determined from other grains than the refractive indices.

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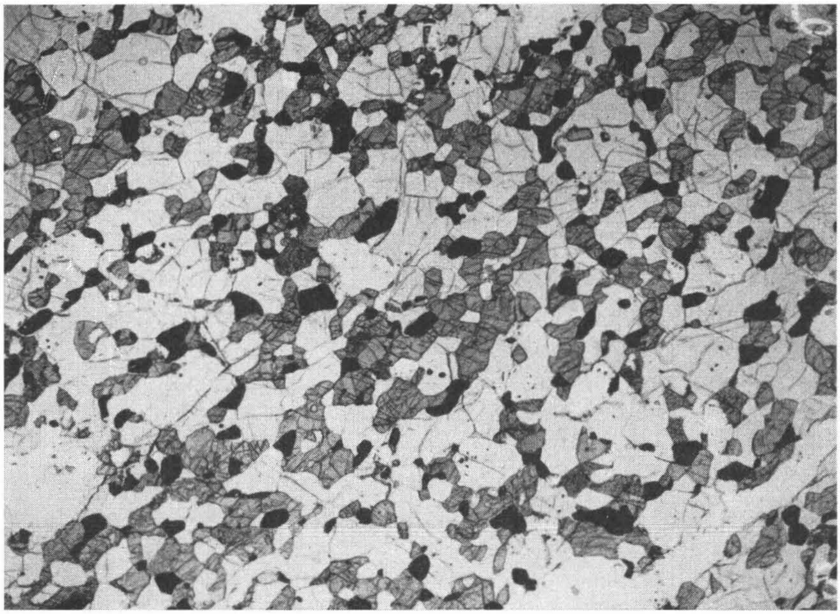

Fig. 5. Texture of pyroxene amphibolite. Specimen No. 14 524. One nicol. Magn. 8 x. White grains are plagioclase, gray grains are mainly diopside and hypersthene and black grains are magnetite. Photo: E. Halme.

The opaque ore minerals generally occur as independant grains, of which the largest are the same size as the pyroxene grains, or 0.5 mm in diameter. According to the ore-microscopic examination, they consists for the most part of pure magnetite. Many of the magnetite grains contain ilmenite lamellae. Furthermore, there are ilmenite grains with magnetite and/or spinel lamellae. In addition to the oxide ore minerals, the polished section was observed to contain several tiny grains of sulphide minerals less than 0.1 mm in diameter. They are principally chalcopyrite, but there is a very slight amount of pyrrhotite, too.

Green hornblende occurs only sporadically as small grains. Its properties are by and large the same as in the three rock samples described in the foregoing. All the silicate minerals are clear and lacking in alteration products. The mineral assemblage thus seems to be quite in equilibrium.

Amphibolite. Specimen No. 14 505. Locality 4 (Plate I), eastern part of the dome, between the inner and outer amphibolite rings.

Megascopically the rock is distinctly banded and also under the microscope it is schistose. Dark and light bands a few millimeters thick alternate in it. The cause of the banding seems mainly to be the layered occurrence of hornblende. The grain size is 1—2 mm, but the hornblende occurs also as larger grains, which reach a length of as much as 1—2 cm. The texture is nematoblastic. The specific gravity is 2.96.

The principal constituents are green hornblende, plagioclase (ca. An_{35}) and biotite. In addition, there are small amounts of chlorite, apatite, quartz and sericite as well as, according to the ore microscopical examination, ilmenite and a few tiny grains of pyrite and chalcopyrite. The ilmenite occurs mainly along the fissures of hornblende. The modal analysis is presented in Table I on p. 28.

The green hornblende is clearly in a state of alteration and in many places it is bordered by biotite. A few large hornblende grains have evidently been crushed and in them the hornblende appears to be in a state of alteration into chlorite.

In some of the plagioclase grains, polysynthetic twinning is clearly to be seen, in others faintly or not at all. The plagioclase contains a trace of sericite, but otherwise it is clear and unaltered in appearance.

Hornblende gneiss. Specimen No. 14 550. Locality 5 (Plate I), southern part of the outer amphibolite ring.

Megascopically the rock is dark gray and its schistosity is rather weak. The grain size is 1—3 mm and the texture is nematoblastic. The specific gravity is 2.81.

The principal constituents are plagioclase (ca. An_{40}), hornblende, chlorite, biotite and quartz. Accessories are the opaque minerals, apatite, zircon, sericite and epidote. The modal analysis is presented in Table I on p. 28.

Polysynthetic twinning is clearly visible in majority of the plagioclase grains. The plagioclase contains a certain amount of sericite.

Quartz occurs as xenomorphic grains and it has a distinctly undulating extinction.

The hornblende is pleochroic from yellowish green to dark green. In addition to the green hornblende, there are also amphibole varieties with a bluish green tinge that usually occur in green hornblende grains. Their birefringence is in places conspicuously low and the variety presumably represents hornblende in process of chloritization. The hornblende is obviously in a state of alteration, and the alteration product is principally biotite, though chlorite also occurs.

The mineralogical compositions of pyroxene amphibolites, amphibolites and hornblende gneisses are best grasped from Table I. Their petrographic differences are due chiefly to the fact that great portions of them have undergone strong retrogressive metamorphism. Moreover, some of the amphibolites and, in particular, hornblende gneisses seem to have also

Table I. Modal analyses of pyroxene amphibolites, amphibolites and hornblende gneisses (vol. %).

	1.	2.	3.	4.	5.	6.
Plagioclase	42.2	61.3	23.9	56.5	32.4	60.8
Hornblende	34.4	1.2	49.5	0.1	52.1	23.5
Hypersthene	8.2	11.9	26.6	16.9	0.0	0.0
Diopside	15.0	19.4	0.0	15.9	9.7	0.0
Oxide ore	0.2	6.2	present	10.6	0.2	present
Biotite	0.0	present	0.0	0.0	1.7	6.1
Quartz	0.0	0.0	0.0	0.0	1.7	5.6
Chlorite	0.0	0.0	0.0	0.0	2.2	3.3
Others	< 0.1	< 0.1	< 0.1	< 0.1	< 0.1	0.7
Total	100.0	100.0	100.0	100.0	100.0	100.0

1. Pyroxene amphibolite. Specimen No. 14 511. Description on p. 22.
2. Pyroxene amphibolite. Specimen No. 14 512. Description on p. 23.
3. Pyroxene amphibolite. Specimen No. 14 515. Description on p. 24.
4. Pyroxene amphibolite. Specimen No. 14 524. Description on p. 25.
5. Amphibolite. Specimen No. 14 505. Description on p. 26.
6. Hornblende gneiss. Specimen No. 14 550. Description on p. 27.

changed metasomatically. The mineral facies conditions are discussed in more detail on pp. 59—62.

The data available do not suffice to draw reliable conclusions concerning the origin of these amphibolitic rocks. The most plausible explanation is that they are of volcanic origin. The calcite lenses and skarn occurrences described on p. 22 suggest, however, that at least part of the amphibolitic rocks in the region may originally have been calcareous sediments.

HYPERSTHENE GNEISSES

Hypersthene gneisses often occur in the region investigated as intercalations in pyroxene amphibolites, but to some extent they form rather large homogeneous areas. Their quantitative share of the rocks marked in the geological map as pyroxene amphibolites, amphibolites and hornblende gneisses is rather substantial and in more detailed mapping a part of them could be drawn separately. In places the hypersthene gneisses have been observed to grade into granodioritic gneisses and, according to some field observations, they appear to grade into pyroxene amphibolites along the strike of the layers.

The most extensive uniform hypersthene gneiss area met with in the region investigated is situated between the inner and the outer amphibolite ring in the western part of the dome, and it covers roughly ten square kilometers. In its typical occurrence the hypersthene gneiss is here very homogeneous. Its foliation is weak and can hardly be detected megascopically. The color of the weathered surface of the hypersthene gneisses is a pale reddish brown. This color has proved not only in the western Greenland but also in other countries, to be a reliable indication of hypersthene gneisses. In the vicinity of the dome this color has been observed to indicate the presence of hypersthene, especially in gneisses, very reliably. In a few localities it was surmised in the field, mainly on the basis of the color nuance of the weathered surface, that certain homogeneous gneisses represent a transitional type between hypersthene gneisses and granodioritic gneisses. Microscopic examinations supported this view.

In the western part of the dome near the inner amphibolite ring, the hypersthene gneiss grades into a dark gray gneiss (Specimen No. 14 516, p. 54) resembling the central gneiss. In the area northwest of the central gneiss, there appear in the hypersthene gneiss numerous inclusions and layers of amphibolite. The hypersthene gneiss area also contains smallish areas of garnet-bearing granodioritic rocks. The contacts between them and the hypersthene gneiss seem to be gradual rather than sharp.

In the following two typical hypersthene gneiss samples and one transitional type will be described.

Hypersthene gneiss, Specimen No. 14 514. Locality 6 (Plate I), western part of the dome between the inner and outer amphibolite rings.

Megascopically the rock is brownish gray and the color of its weathered surface is reddish brown. The rock is homogeneous and its foliation is very weak. The grain size is 1—2 mm and the texture is granoblastic. The specific gravity is 2.63.

The principal constituents are plagioclase (ca. An_{25}), microcline, quartz, biotite, hypersthene, serpentine and opaque ore minerals. In addition, there are traces of chlorite, sericite and zircon. The modal analysis is presented in Table II on p. 31. According to the chemical analysis, the rock contains 4.38 % Na_2O and 2.86 % K_2O by weight.

Polysynthetic twinning is visible in the plagioclase only faintly or not at all. The plagioclase contains very small amounts of antiperthite and sericite as inclusions. The microcline is for the most part cross-hatched, and it occurs as xenomorphic grains 0.1—0.3 mm in diameter. It commonly contains tiny plagioclase inclusions. The triclinicity of the potash feldspar is 0.9. The quartz to a certain extent exhibits an undulating extinction.

The hypersthene occurs in the thin section as grains 0.2—0.8 mm long. No crystal forms are visible and the hyperstene is obviously in a state of alteration. The principal alteration product is a brownish, fibrous substance, which for the most part appears to be serpentine. In addition, there frequently occur oxide ore, biotite and a slight amount of chlorite in the proximity of the hypersthene grains.

Hypersthene gneiss. Specimen No. 14 548. Locality 7 (Plate I), southwestern part of the dome between the inner and outer amphibolite rings.

The color of the rock is a dark, brownish gray. The weathered surface is reddish brown. The rock is megascopically even-grained and homogeneous. The foliation is very weak, hardly perceptible. The grain size is 1—2 mm and the texture is granoblastic. The specific gravity is 2.66.

The principal constituents are plagioclase (ca. An_{25}), quartz, biotite, hypersthene, potash feldspar and oxide ore. In addition, serpentine and zircon are present in slight amounts. The modal analysis is presented in Table II.

Polysynthetic twinning is clearly visible in the plagioclase. The slight amount of potash feldspar in the rock occurs wholly as antiperthite in the plagioclase. The diameter of the hypersthene grains is 0.3—1 mm, and the hypersthene is in a state of alteration in the way described in the foregoing.

Transitional type between hypersthene gneiss and granodioritic gneiss. Specimen No. 14 549. Locality 8 (Plate I), southern part of the dome between the inner and outer amphibolite rings.

In the locality where the sample was taken, the extensive hypersthene gneiss area marked on the geological map grades according to field observations, into granodioritic and quartz dioritic gneiss. In the field this grading is to be seen by the gradual change in the color of the rock's weathered surface from reddish brown into light gray.

On the site where the sample was taken, the rock is light gray with a faint brownish tinge. The color of the weathered surface has a faint suggestion of reddish brown. The rock is very slightly foliated, even-grained and homogeneous. The grain size is 1—2 mm and the texture is granoblastic. The specific gravity is 2.61.

The constituents of the rock are plagioclase (ca. An_{20}), microcline, quartz, biotite, oxide ore, sericite and zircon. In addition, the thin section shows a slight amount of a mineral resembling serpentine and extremely tiny particles of hypersthene 0.01—0.02 mm long. The modal analysis is presented in Table II.

Polysynthetic twinning is plainly to be seen in the plagioclase. The plagioclase contains a slight amount of potash feldspar as antiperthite. The microcline occurs for the most part as xenomorphic grains 0.3—1 mm in diameter, which in most cases are clearly cross-hatched and contain plagioclase as perthite. The triclinicity of the potash feldspar is 0.9—1.0.

Table II. Modal analyses of hypersthene gneisses (vol. %).

	1.	2.	3.
Plagioclase	51.8	79.6	54.7
Potash feldspar	17.0	1.1	20.4
Quartz	28.5	15.3	23.0
Biotite	0.9	1.8	1.0
Hypersthene	0.8	1.4	trace
Ore	0.3	0.8	0.3
Others	0.7	< 0.1	0.6
Total	100.0	100.0	100.0

1. Hypersthene gneiss. Specimen No. 14 514. Description on p. 29.
2. Hypersthene gneiss. Specimen No. 14 548. Description on p. 30.
3. Transitional type between hypersthene gneiss and granodioritic gneiss. Specimen No. 14 549. Description on p. 30.

The quartz has a faintly undulating extinction. The biotite is pleochroic from light brown to dark brown. In addition, the thin section reveals the presence of a slight amount of green biotite, which appears to represent a variety at the incipient stage of chloritization. Under the microscope it is green in color, but its birefringence is almost the same than that of the brown biotite.

In several spots in the thin section between the grains and as small aggregates, there occurs a brownish, fibrous mineral, which appears to be serpentine. Sericite also often occurs in association with it. In the middle of one such aggregate, which probably contains serpentine and sericite, tiny grains 0.01—0.02 mm long could be observed. Their refraction indices, birefringence and straight extinction correspond to those of hypersthene. All these grains have a simultaneous extinction. The evidence indicates that they are the remains of an almost totally altered hypersthene grain.

As the foregoing descriptions and Table II make clear, these hypersthene gneisses are very poor in dark minerals. In composition they vary from quartz dioritic to granodioritic. In all the cases investigated the hypersthene is more or less in a state of alteration owing to retrogressive metamorphism.

GARNET-CORDIERITE GNEISS

Garnet-cordierite gneisses have been met with in the mapped region in only one spot, which is situated at the northern edge of the geological map. Here the garnet-cordierite gneiss forms a continuous layer, the visible portion of which is approximately 1 km long. At both ends it terminates in a lake, so the true length of the layer could not be determined.

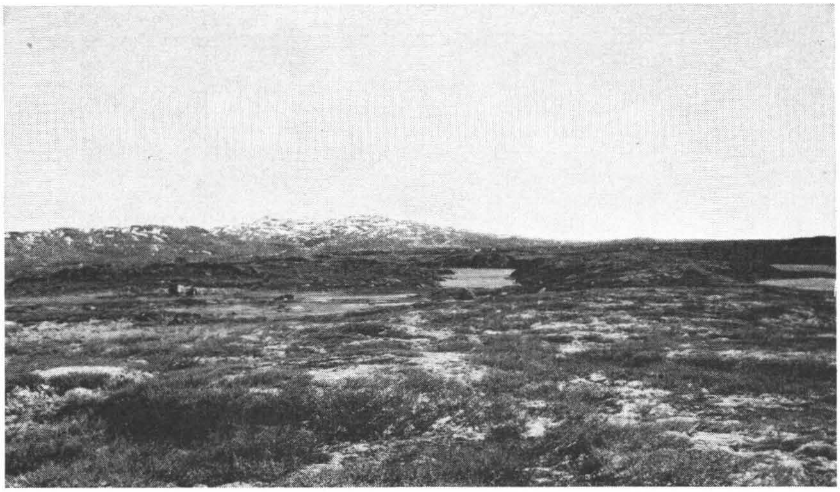

Fig. 6. View towards south over western part of inner amphibolite ring and central gneiss. Amphibolite in foreground and right of lake, central gneiss at left.

The layer is a few dozen meters thick, being about 50 m at its thickest. It has a vertical dip and on both sides there occurs amphibolite. In color the rock is brownish gray and the weathered surface is a pale rusty brown. The southern end of the layer contains very little or no garnet, whereas the northern end has an abundance of it. The occurrence of cordierite was not detected until the microscopic examination was carried out, and the few specimens do not suffice for the drawing of any significant conclusions concerning its distribution or abundance throughout the layer.

In the following a sample is described which was taken from a spot containing considerable garnet.

Garnet-cordierite gneiss. Specimen No. 14 559. From the northern part of the layer described above. Locality 9 (Plate I).

Megascopically the rock is pale rusty brown and distinctly foliated. The grain size is generally 1—2 mm. The diameter of the garnet grains varies between 1 mm and 5 mm. The texture is granoblastic. The specific gravity of the rock is 2.94.

The principal constituents are quartz, garnet, biotite, plagioclase (ca. An_{40}), cordierite and oxide ore. In addition, there is a slight amount of zircon and some micaceous mineral as an alteration product of cordierite and garnet.

Polysynthetic twinning is clearly visible in the plagioclase. The quartz has a weakly undulating extinction. The biotite is pleochroic from very

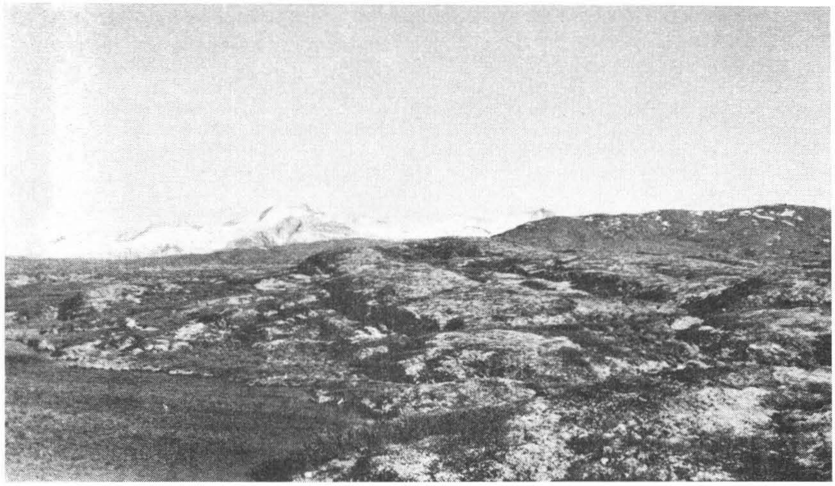

Fig. 7. View towards southeast over southwestern part of central gneiss and inner amphibolite ring. Amphibolite in foreground forming ridge about 5 m high; central gneiss at left behind lake.

pale brown to reddish brown, and there are distinct pleochroic haloes around the zircon grains contained in it.

The garnet forms xenoblasts with quartz and oxide ore inclusions. Under high magnifications it is also possible to distinguish a micaceous mineral, which is evidently a product of an incipient alteration of garnet.

The garnet's refraction index $n = 1.778$, the cell edge $a_0 = 11.518$ Å and the specific gravity 3.99 (Determinations by K. Hytönen).

Calculated according to the chemical analysis (Table IV, p. 42) the garnet contains about 40—45 mol. % pyrope, 49—54.5 mol. % almandine, 0—3 mol. % grossular, 1.5—5 mol. % andradite and 1—1.5 mol. % spessartite. This calculation of the percentages of the garnet's end-members is, however, somewhat uncertain, because the chemical analysis does not closely agree with the theoretical ratio of 3:2:3 for the $Si:R^{+3}:R^{+2}$ of the garnet group. The FeO seems to be low and the Fe_2O_3 correspondingly too high. The determination of FeO was for this reason repeated several times, but without significant differences. The Pratt method was used in all these determinations. The garnet's incipient stage of alteration described above is one possible explanation for this discrepancy.

The cordierite similarly exhibits incipient alteration, with a micaceous mineral as a product. Around the tiny zircon grains contained in the cordierite there are distinct pleochroic haloes.

This garnet-cordierite gneiss is the only rock found in the mapped region which might be interpreted as an argillaceous sediment by origin.

Fig. 8. Northernmost part of central gneiss. Contact of central gneiss and amphibolite is situated on upper slope. View towards northwest.

THE CENTRAL GNEISS

The center of the dome structure as a whole comprises a homogeneous granodioritic and quartz dioritic gneiss massif. It is approximately 2 km long and 1.2 km broad, and its shape is a fairly regular ellipse. The foliation even in the marginal parts of the central gneiss is so weak that its strike and dip are difficult to determine in places. With the exception of the jointing, no reliable tectonic observations have been obtained of the center of the gneiss. This weakness of the foliation is due, at least in part, to the paucity of mafic minerals. The structure of the central gneiss and its surroundings will be described in more detail on pp. 76—78. Figures 6 and 7 on pp. 32—33 show general views from the central gneiss.

There are relatively few exposures in the area of the central gneiss and their quality is not the best possible, either. The middle parts of the central gneiss, especially, are largely covered with vegetation and shallow lakes. The best exposed is the northernmost part of the central gneiss, which is practically a single outcrop (Fig. 8).

No inclusion or schlieren have been found in the central gneiss any more than granitic, pegmatitic or other such veins. The field observations have brought to light very little variety in the central gneiss.

In the following, typical samples taken from different parts of the central gneiss will be described.

Granodioritic rock. Specimen No. 14 501. Locality 10 (Plate I), middle of the central gneiss.

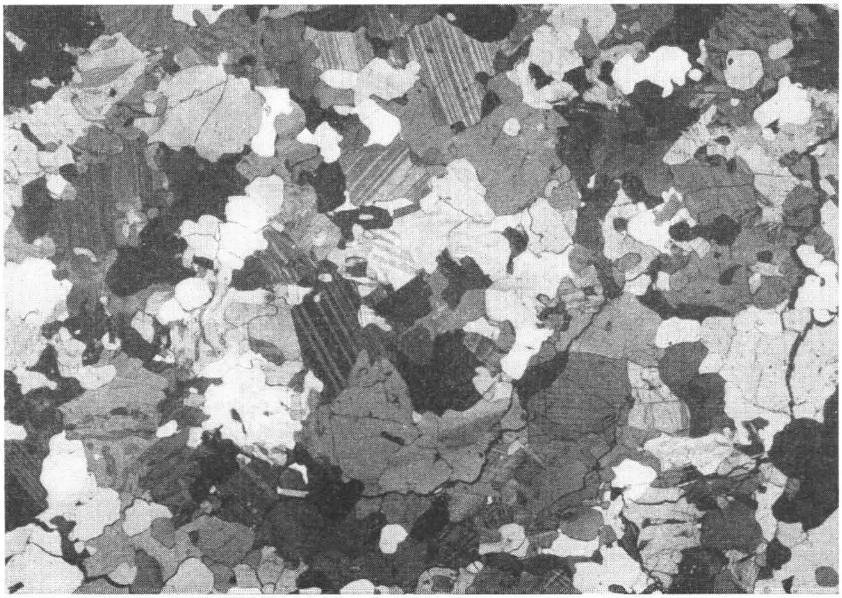

Fig. 9. Texture of granodioritic variety of central gneiss. Specimen No. 14 501. Nicols +. Magn. 11 x. Photo: E. Halme.

In the field the rock is rather homogeneous. No distinct foliation is to be seen. The rock is dark gray with a tinge of blue. The grain size is 1—2 mm. The biotite forms larger flakes and flake aggregates, the diameter of which rises in some cases as high as 5—10 mm. The texture is granoblastic, though in spots there are features that could, perhaps, be interpreted as blastohypidiomorphic (Fig. 9). The specific gravity is 2.64.

The constituents are plagioclase (ca. An_{20}), microcline, quartz, biotite and chlorite as well as oxide ore and small amounts of sericite and zircon. The modal analysis is presented in Table III on p. 40.

According to the chemical analysis, the rock contains 4.83 % Na_2O and 2.41 % K_2O by weight.

Polysynthetic twinning is clearly visible in the majority of the plagioclase grains. Many plagioclase grains contain microcline up to as much as 50 %, forming stripes. In other plagioclase grains, on the other hand, no potash feldspar is in evidence.

For the most part the microcline is cross-hatched. Partly it occurs as stripes in the plagioclase and partly as interstitial grains 0.2—0.4 mm in diameter. The interstitial microcline grains also contain tiny plagioclase inclusions. The triclinicity of the potash feldspar is 0.9.

The quartz has a distinct undulating extinction. Some of the biotite is clearly pleochroic from light brown to dark brown. A number of the biotite flakes are greenish or green and are evidently in a state of alteration into clorite.

A certain mineral aggregate appearing in the thin section is composed chiefly of chlorite and ore. Judging by the form of the aggregate, what is involved is a chloritized amphibole grain.

Quartz dioritic rock. Specimen No. 14 502. Locality 11 (Plate I), middle of the central gneiss.

According to field observations, the rock is fairly homogeneous. Its foliation is extremely weak or none is apparent. The rock is light gray with a faint violet tinge. The grain size is 1—2 mm and the texture is granoblastic (Fig. 10). The specific gravity is 2.62.

The principal constituents are plagioclase (ca. An_{20}), quartz, microline, biotite and sericite. The accessories are apatite, ore, chlorite, zircon and epidote. The modal analysis is presented in Table III on p. 40, and the chemical analysis in Table IV on p. 42.

Polysynthetic twinning is visible faintly or not at all in the plagioclase, which contains conspicuous amounts of sericite and some microcline as antiperthite.

The microcline is cross-hatched and occurs partly as antiperthite in plagioclase and partly as interstitial grains measuring 0.1—0.4 mm in diameter. The triclinicity of the potash feldspar is 0.95.

The biotite is pleochroic from light brown to dark brown. Neither greenish nor green biotite varieties are present in the thin section.

The thin section has one spot in which there is an epidote grain in the middle and chlorite around it. Possibly a chloritized amphibole grain is in question.

A number of the plagioclase grains and biotite flakes are clearly bent, suggesting movements that took place after the crystallization of these minerals.

Quartz dioritic gneiss. Specimen No. 14 506. Locality 12 (Plate I), northern part of central gneiss, about 100 m from the contact.

The color of the rock is dark gray. The foliation is so weak that its strike and dip are difficult to measure in the field. The grain size is 1—3 mm. The biotite in places forms larger flakes and flake aggregates measuring some 4—5 mm. The texture is granoblastic. The specific gravity is 2.64.

The principal constituents are plagioclase (ca. An_{20}), quartz, biotite, potash feldspar and sericite. The accessories are sphene, apatite, epidote, ore and zircon. The modal analysis is presented in Table III on p. 40.

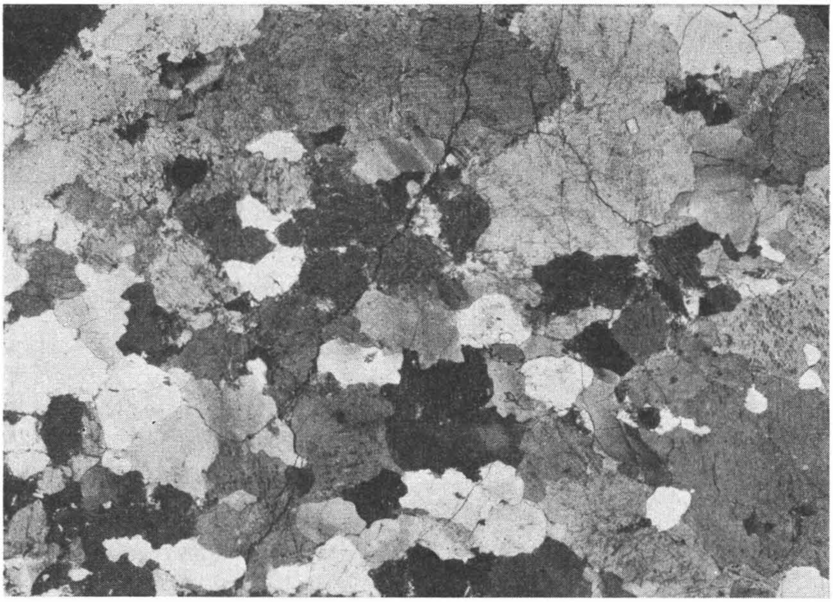

Fig. 10. Texture of quartz dioritic variety of central gneiss. Specimen No. 14 502. Nicols +. Magn. 11 x. Photo: E. Halme.

According to the chemical analysis, the rock contains 5.50 % Na_2O and 1.15 % K_2O by weight.

For the most part, polysynthetic twinning is only faintly visible in the plagioclase or not at all. The plagioclase is conspicuously sericitized and it contains a certain amount of antiperthite.

The potash feldspar occurs partly as antiperthite in the plagioclase, partly as xenomorphic grains around 0.3—0.5 mm in diameter.

Most of the biotite is pleochroic from light brown to dark brown. Some of the biotite is greenish and evidently in a state of alteration into chlorite. The quartz exhibits distinct undulating extinction.

Quartz dioritic gneiss. Specimen No. 14 507. Locality 13 (Plate I), western part of central gneiss, about 100 m from contact.

Megascopically, the rock is dark gray, and its foliation is rather weak. The grain size is 1—2 mm and the texture is granoblastic. The specific gravity is 2.64.

The principal constituents are plagioclase (ca. An_{20}), quartz, biotite, chlorite and potash feldspar. The accessory constituents are sericite, apatite, ore, sphene and epidote. The modal analysis is given in Table III on p. 40.

According to the chemical analysis, the rock contains 5.52 % Na_2O and 0.90 % K_2O by weight.

Polysynthetic twinning is visible in the plagioclase only faintly or not at all. The plagioclase is conspicuously sericitized.

The potash feldspar occurs partly as antiperthite in the plagioclase, partly as interstitial grains 0.1—0.2 mm in diameter.

The quartz exhibits a strong undulating extinction. The biotite occurs in the thin section mostly in small aggregates. Furthermore, it is present around the ore grains. A portion of the biotite is brown and another portion greenish.

Quartz dioritic gneiss. Specimen No. 14 508. Locality 14 (Plate I), eastern part of central gneiss, about 20 m from contact.

Megascopically the rock is dark gray. Its foliation is rather weak. The grain size is 1—2 mm. The biotite occurs as slightly larger flakes and flake aggregates 4—5 mm in diameter. The texture is granoblastic. The specific gravity is 2.63.

The principal constituents are plagioclase (ca. An_{20}), quartz, potash feldspar, biotite, sericite and green hornblende. The accessories are ore, sphene, apattie and epidote. The modal analysis is given in Table III on p. 40.

According to the chemical analysis, the rock contains 5.83 % Na_2O and 0.85 % K_2O by weight.

Polysynthetic twinning is visible in the plagioclase only faintly or not at all. The plagioclase is conspicuously sericitized.

The potash feldspar occurs for the most part as antiperthite in the plagioclase. In the entire thin section there are only a few interstitial potash feldspar grains and they are 0.1—0.2 mm in diameter. The triclinicity of the potash feldspar is 0.75.

The hornblende seems to be in a state of alteration and around it there occur biotite and some epidote.

Part of the biotite is brown and part greenish. Around a few ore grains there occur leucoxene and biotite.

Granodioritic gneiss. Specimen No. 14 510. Locality 15 (Plate I), southern part of central gneiss, about 50 m from contact.

Megascopically this rock differs from the other varieties of rock included in the central gneiss and described in the foregoing in that, for instance, light, slightly reddish layers containing chiefly quartz and feldspar alternate with darker layers richer in biotite. For this reason the foliation of the rock is more distinct than in the specimens from the central gneiss described previously. The grain size is 1—2 mm and the texture is granoblastic. The specific gravity is 2.62.

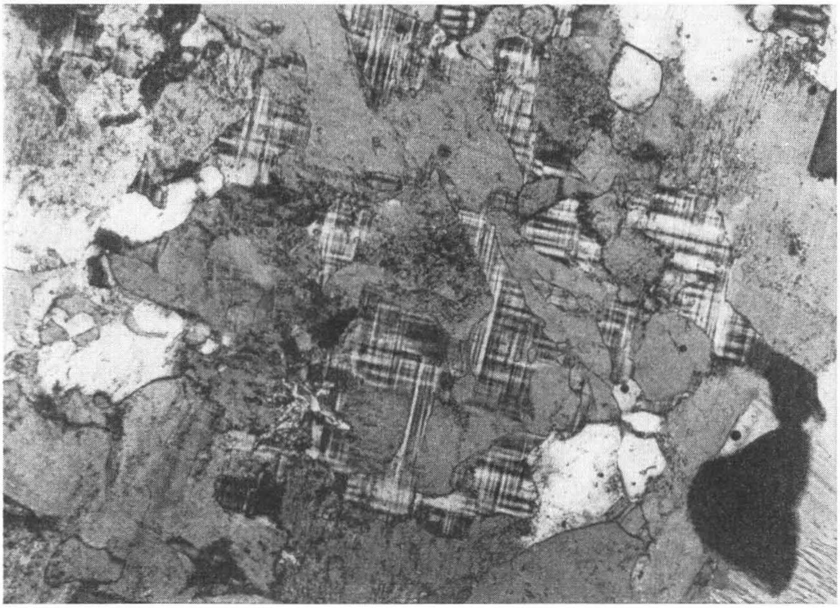

Fig. 11. Textural mode of occurrence of plagioclase and potash feldspar in granodioritic variety of central gneiss. Specimen No. 14 510. Nicols +. Magn. 55 x. Photo: E. Halme.

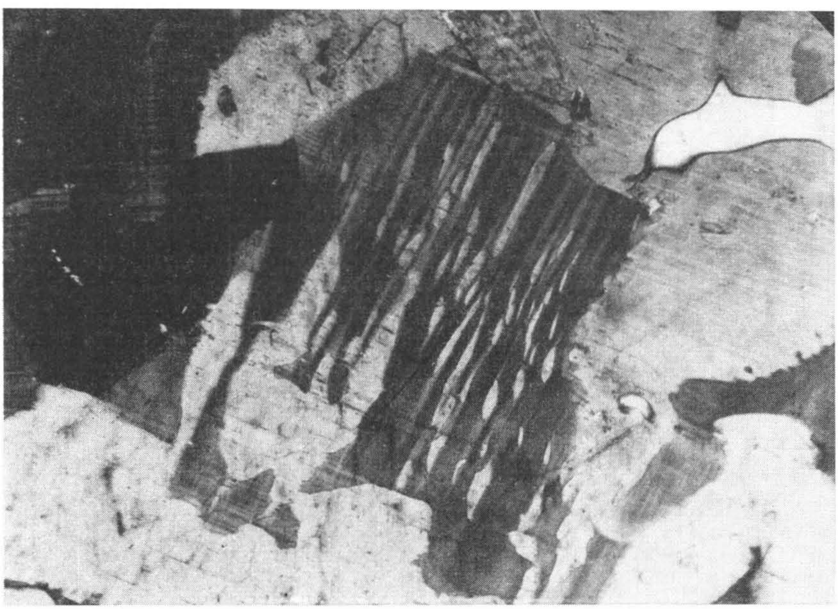

Fig. 12. Textural mode of occurence of plagioclase and potash feldspar. Same thin section as in Fig. 11. Nicols +. Magn. 130 x. Photo: E. Halme.

Table III. Modal analyses of specimens from central gneiss (vol. %).

	1.	2.	3.	4.	5.	6.
Plagioclase	61.6	65.2	63.9	77.0	67.3	43.6
Potash feldspar	19.7	4.9	3.9	1.3	3.1	17.5
Quartz	16.8	25.0	23.8	18.3	24.4	31.5
Biotite and Chlorite	1.7	3.3	5.8	2.1	2.7	6.3
Hornblende	0.0	0.0	0.0	0.0	0.9	0.0
Sericite	present	1.4	2.6	1.1	1.5	1.1
Others	0.2	0.2	< 0.1	0.2	0.1	< 0.1
Total	100.0	100.0	100.0	100.0	100.0	100.0

1. Granodioritic rock. Specimen No. 14 501. Center of central gneiss. Description on p. 35.
2. Quartz dioritic rock. Specimen No. 14 502. Center of central gneiss. Description on p. 36. Chemical analysis in Table IV.
3. Quartz dioritic gneiss. Specimen No. 14 506. Northern part of central gneiss. Description on p. 36.
4. Quartz dioritic gneiss. Specimen No. 14 507. Western part of central gneiss. Description on p. 37.
5. Quartz dioritic gneiss. Specimen No. 14 508. Eastern part of central gneiss. Description on p. 38.
6. Granodioritic gneiss. Specimen No. 14 510. Southern part of central gneiss. Description on p. 38. Chemical analysis in Table IV.

The principal constituents are plagioclase (ca. An_{20}), quartz, potash feldspar, biotite and sericite. Accessory constituents are ore, zircon and sphene. The modal analysis is presented in Table III and the chemical analysis in Table IV.

Polysynthetic twinning is clearly visible in some of the plagioclase grains, while in others it is faint or not apparent at all. Part of the plagioclase is sericitized.

The mode of occurrence of the potash feldspar and the relations between the potash feldspar and the plagioclase vary a good deal. There occurs very little potash feldspar in plagioclase as antiperthite. A large part of the potash feldspar appears as xenomorphic grains 0.1—0.3 mm in diameter and often containing small plagioclase inclusions. In a few cases the plagioclase is present as separate »islets», the interstices of which are filled with cross-hatched microline (Figs. 11 and 12 on p. 39). The parts of each mineral, which seems to be situated separately, have a simultaneous extinction.

The triclinicity of the potash feldspar varies from 0.4 to 0.8. The greatest portion of it seems to have the latter value.

In the light of the foregoing descriptions and Table III, the composition of the central gneiss varies from quartz dioritic to granodioritic. The amount of dark minerals is only about two to six vol. %. The triclinicity of the potash feldspar varies from 0.4 to 0.95. According to these determinations, it is microcline. The content of potassium and sodium in quartz dioritic and granodioritic rocks of the mapped region as a whole will be discussed on pp. 63—66.

OTHER AREAS OF HOMOGENEOUS GNEISSES

In addition to the central gneiss there occur in the mapped region several other areas of homogeneous granodioritic and quartz dioritic rocks. In other places these rocks form migmatites in association with other varieties of rocks.

Immediately north and northeast of the dome, there is situated a homogeneous gneiss area. In shape it is an elongated triangle, the sides of which are concave. Its length is some 3.5 km and greatest breadth about 2 km. Megascopically the rock closely resembles the central gneiss. Its foliation is very weak even in the marginal parts, though it conforms to the amphibolite layers bordering the gneiss. Megascopically apparent foliation is almost totally missing from the middle portions of the gneiss.

In the following are descriptions of one typical sample from the middle of this gneiss and several other samples from other homogeneous gneiss areas in the region investigated.

Quartz dioritic rock. Specimen No. 14 554. Locality 16 (Plate I), north of the dome.

Megascopically the rock is generally light gray, but portions richer in biotite form darker splotches in places. Foliation can scarcely be detected. The grain size tends to be 1—2 mm. The diameter of some of the quartz grains is as much as 3 mm and the biotite also forms larger flakes and flake aggregates measuring some 3—5 mm in diameter. The texture is granoblastic. The specific gravity is 2.65.

The principal constituents are plagioclase (ca. An_{20}), quartz, biotite and sericite. Accessory constituents are epidote, ore, apatite, zircon, potash feldspar and sphene. The modal analysis is presented in Table V on p. 49.

According to the chemical analysis, the rock contains 5.46 % Na_2O and 1.00 % K_2O by weight.

Polysynthetic twinning is visible in the plagioclase only faintly or not at all. The plagioclase contains an abundance of sericite and some epidote. The very small amount of potash feldspar contained in the rock (less than 0.1 vol. %) is situated almost wholly in the plagioclase as antiperthite.

The quartz has a strong undulating extinction. The biotite is pleochroic from nearly colorless or very light brown to dark brown. In the middle and around many of the biotite flakes there are epidote grains and in places the biotite is bordered by muscovite. Around the ore grains leucoxene often occurs.

Granodioritic rock. Specimen No. 14 561. Locality 17 (Plate I), northwest of the dome.

Table IV. Chemical analyses of granodioritic and quartz dioritic rocks and of garnet. (weight %). Analyst, H. B. Wiik.

	1.	2.	3.	4.
SiO_2	73.73	73.38	70.68	37.18
TiO_2	0.06	0.25	0.11	0.10
Al_2O_3	15.03	14.43	17.68	23.50
Fe_2O_3	0.89	0.67	0.60	6.62
FeO	0.28	0.93	0.36	20.04
MnO	0.01	0.02	0.00	0.59
MgO	0.34	0.93	0.54	10.41
CaO	2.15	1.31	3.38	1.64
Na_2O	4.86	4.50	5.05	—
K_2O	1.50	3.38	1.11	—
P_2O_5	0.05	0.01	0.04	—
H_2O+	0.29	0.43	0.27	0.22
H_2O-	0.07	0.06	0.03	0.02
CO_2	0.00	0.00	0.00	—
Total	99.26	100.30	99.85	100.32

1. Quartz dioritic rock. Specimen No. 14 502. Center of central gneiss. Description on p. 36. Modal analysis in Table III, p. 40.
2. Granodioritic gneiss. Specimen No. 14 510. Southern part of central gneiss. Description on p. 38. Modal analysis in Table III, p. 40.
3. Quartz dioritic gneiss layer. Specimen No. 14 536. Northwestern part of dome. Description on p. 46. Modal analysis in Table V, p. 49.
4. Garnet from garnet-cordierite gneiss. Rock specimen No. 14 559. Description on p. 32.

Megascopically, the rock is light gray with a tinge of yellowish brown. The biotite content is quite small. The foliation is weak, scarcely perceptible. The grain size is 1—2 mm and the texture is granoblastic. The specific gravity is 2.61.

According to the chemical analysis, the rock contains 5.77 % Na_2O and 1.29 % K_2O by weight.

The principal constituents are plagioclase (ca. An_{20}), quartz, biotite and cross-hatched microcline. In addition, there are small amounts of oxide ore, sericite, apatite, sphene and zircon.

The plagioclase contains some potash feldspar as antiperthite.

Granodioritic gneiss. Specimen No. 14 568. Locality 18 (Plate I), northwestern shore of the lake Quagssup Taserssua, southwest of the dome.

Megascopically, the rock is dark gray and contains biotite in relative abundance. The foliation is fairly distinct. The grain size is 1—3 mm and the texture is granoblastic. The specific gravity is 2.63.

According to the chemical analysis, the rock contains 4.57 % Na_2O and 2.50 % K_2O by weight.

The principal constituents of the rock are quartz, plagioclase (ca. An_{20}), biotite and cross-hatched microcline. The accessories are oxide ore, apatite, sericite and zircon.

The plagioclase is on the whole noticeably sericitized.

Granodioritic gneiss. Specimen No. 14 576. Locality 19 (Plate I), northwestern shore of Godthaabsfjord, about 500 m northeast of Nûa.

Megascopically, the rock is light gray and contains biotite in relative abundance. The foliation is fairly distinct. The grain size is 1—3 mm and the specific gravity is 2.66.

According to the chemical analysis, the rock contains 3.43 % Na_2O and 4.29 % K_2O by weight.

The principal constituents of the rock are plagioclase (ca. An_{20}), quartz, cross-hatched microcline and biotite. In addition, small amounts of sphene, oxide ore, muscovite, apatite, zircon and epidote are found.

The plagioclase contains some potash feldspar as antiperthite and the microcline some plagioclase as perthite. The plagioclase is somewhat sericitized.

Granodioritic rocks. Specimen No. 14 587. Locality 20 (Plate I), northeast of the northeastern corner of the dome.

Megascopically, the rock is light gray with a pale reddish tinge. The biotite content is rather slight. The foliation is weak, scarcely perceptible. The grain size is 1—2 mm and the specific gravity is 2.62.

According to the chemical analysis, the rock contains 5.09 % Na_2O and 3.21 % K_2O by weight.

The principal constituents of the rock are plagioclase (ca. An_{20}), quartz, cross-hatched microcline and biotite. The accessory constituents are sericite, sphene, apatite and zircon.

The plagioclase exhibits distinct polysynthetic twinning. It contains a very small amount of potash feldspar as antiperthite and a bit of sericite. The microcline contains a noticeable quantity of plagioclase as perthite.

Quartz dioritic gneiss. Specimen No. 14 597. Locality 21 (Plate I), between the southeastern part of the dome and Niaqornatsiaq.

Megascopically, the rock is light gray. The biotite content is rather slight. The foliation is only faintly in evidence. The grain size is 1—2 mm and the specific gravity 2.66.

According to the chemical analysis, the rock contains 5.63 % Na_2O and 0.76 % K_2O by weight.

The principal constituents of the rock are quartz, plagioclase (ca. An_{20}), biotite and sericite. The accessory constituents are epidote, potash feldspar, sphene, oxide ore and zircon.

Fig. 13. Nearly horizontal bedding in northwestern part of dome. View towards northeast. Quartz dioritic gneiss layer in lowest part of wall in right margin of figure. Upper part of wall is mainly amphibolite. Height of wall is about 15 m.

The slight amount of potash feldspar in the rock is wholly situated in the plagioclase as antiperthite. The plagioclase is noticeably sericitized.

Granodioritic gneiss. Specimen No. 14 613. Locality 22 (Plate I), about 1 km east of the northeastern corner of the dome.

Mecascopically, the rock is light gray. Its biotite content is rather small. The foliation is distinctly perceptible. The grain size is 1—2 mm and the specific gravity is 2.67.

According to the chemical analysis, the rock contains 5.08 % Na_2O and 2.41 % K_2O by weight.

The principal constituents of the rock are plagioclase (ca. An_{20}), quartz, cross-hatched microcline, biotite and sericite. The accessory constituents are epidote, apatite, sphene, oxide ore and zircon.

The plagioclase exhibits distinct polysynthetic twinning. In places it contains some sericite but scarcely any potash feldspar as antiperthite. The microcline contains a bit of plagioclase as perthite.

Granodioritic gneiss. Specimen No. 14 573. Locality 23 (Plate I), southeastern part of the dome between the inner and outer amphibolite rings.

Megascopically, the rock is light gray. Its biotite content is rather small. The foliation is weak but none the less distinguishable. The grains size is 1—2 mm and the specific gravity is 2.63.

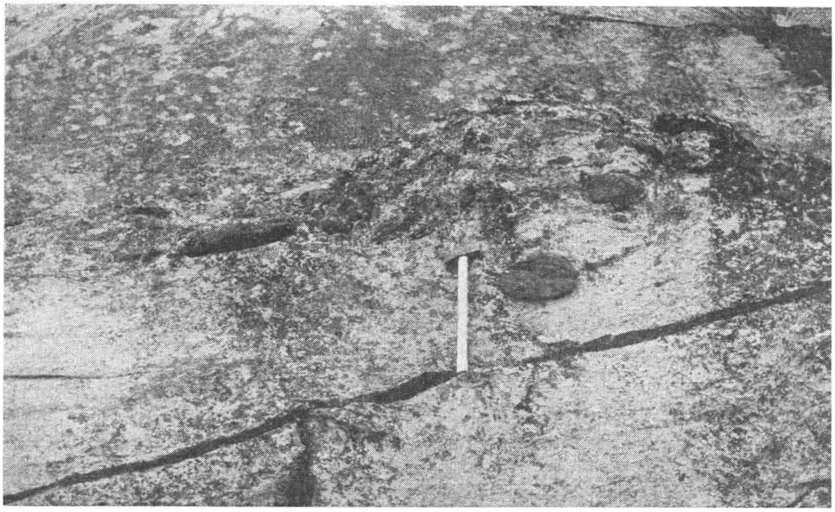

Fig. 14. Skarn lenses, mainly diopside, in quartz dioritic gneiss layer. Same layer as in Fig. 13.

According to the chemical analysis, the rock contains 4.90 % Na_2O and contains 3.36 % K_2O by weight.

The principal constituents of the rock are plagioclase (ca. An_{20}), quartz, cross-hatched microcline and biotite. The accessory constituents are sphene, muscovite, apatite and zircon.

In the light of the descriptions in the foregoing, these homogeneous gneisses vary in composition from quartz dioritic to granodioritic. No essential petrographic difference between these rocks and the central gneiss has been noticed.

QUARTZ DIORITIC INTERCALATIONS IN AMPHIBOLITES

At various places in the region investigated, the amphibolites contain intercalations of quartz dioritic gneisses. The contacts between the gneiss layers and the amphibolites are usually sharp and in many cases the contact is a shear surface. No apophyses intruding from the gneiss layers into the surrounding amphibolites have been found. The thickness of the gneiss layers is generally about 5—15 meters. Fig. 13 represents the mode of occurrence of such a gneiss layer.

In the gneiss layer shown in this figure, a horizon has been found which includes an abundance of skarn lenses containing chiefly diopside (Fig. 14).

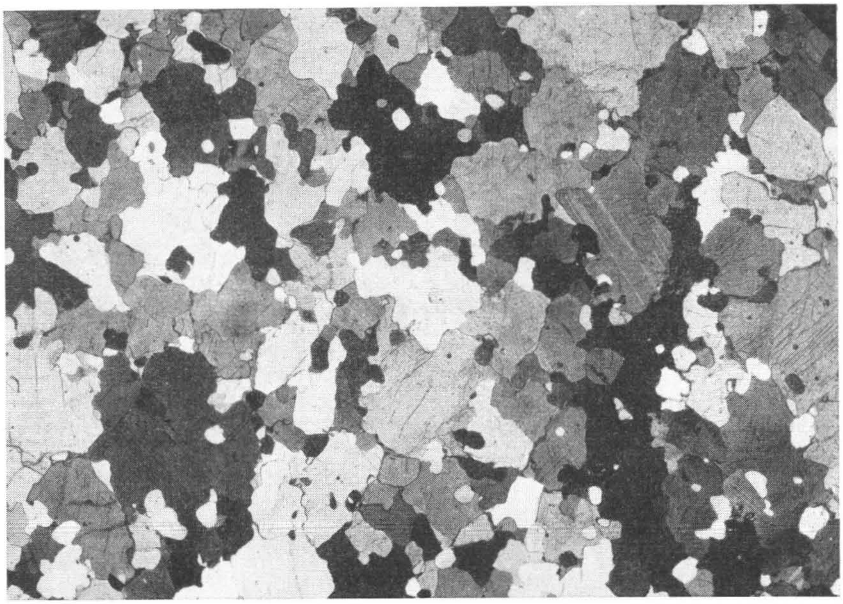

Fig. 15. Texture of quartz dioritic gneiss layer. Specimen No. 14 536. Same gneiss layer as in Figs. 13 and 14. Nicols +. Magn. 11 x. Photo: E. Halme.

This horizon of skarn lenses is exposed for a distance of more than 50 meters.

The quartz dioritic intercalations in the amphibolites occur most plainly in the western and northern parts of the dome, where the layers dip very gently or lie almost horizontal. A couple of these gneiss layers have been followed partly in the field and partly by means of aerial photographs, for a distance of some 5 km.

In the following a description will be given of two of typical specimens taken from such a gneiss layer.

Quartz dioritic gneiss layer. Specimen No. 14 536. Locality 24 (Plate I), northwestern part of the dome. The layer is the same as that shown in Fig. 14.

Megascopically the rock is dark gray. Its foliation is weak, scarcely perceptible. The grain size is 1—2 mm and the texture is granoblastic (Fig. 15). The specific gravity is 2.64.

The principal constituents are plagioclase (ca. An_{20}), quartz and potash feldspar. In addition, there is a small amount of biotite and the accessory constituents are sericite, ore, chlorite and zircon. The modal analysis is presented in Table V on p. 49 and the chemical analysis in Table IV on p. 42.

Polysynthetic twinning is clearly to be seen in most of the plagioclase grains. In places the plagioclase contains considerably potash feldspar as antiperthite. Moreover, the potash feldspar occurs as interstitial grains 0.1—0.2 mm in diameter. The quartz has a weak undulating extinction.

The biotite is pleochroic from light brown to dark brown. Sporadically the thin section reveals the presence of a very fine-grained material, which seems to consist principally of chlorite and sericite.

Quartz dioritic gneiss layer. Specimen No. 14 569. Locality 25 (Plate I), southwestern part of the dome.

Megascopically the rock is very much like the specimen No. 14 536, just described, except that it is a trifle lighter in shade. The grain size is 1—2 mm and the specific gravity is 2.62.

Also microscopically the rock is very much the same in appearance as the preceding sample, except that its plagioclase reveals a substantial content of sericite.

According to the chemical analysis, the rock contains 5.31 % Na_2O and 0.55 % K_2O by weight.

According to the descriptions in the foregoing, these gneiss intercalations in the amphibolites are quartz dioritic in composition and very poor in mafic minerals. Under the microscope they are rather similar to quartz dioritic rocks forming larger homogeneous areas.

GARNET-BEARING GRANODIORITIC ROCKS

Garnet-bearing rocks are rather rare in the mapped region. With the exception of the garnet-cordierite gneiss layer described on pp. 31—33, they have been met with in noteworthy amounts only in the area of the dome itself, between the inner and outer amphibolite rings, especially to the west, north and northeast of the central gneiss. Here garnets occur in the granodioritic rocks in smallish areas, which are mostly a few dozen meters in diameter. According to the field observations, the garnet-bearing granodioritic varieties of rocks grade without any sharp boundaries into garnet-free rocks. The garnet content is rather small, generally around one vol. % or less.

In the following, descriptions are given of four samples of garnet-bearing granodioritic rocks.

Garnet-bearing granodioritic gneiss. Specimen No. 14 518. Locality 26 (Plate I), north of the central gneiss, between the inner and outer amphibolite rings.

Megascopically, the rock is light gray, slightly reddish. The foliation is rather weak. The rock contains red garnet grains 0.5—1 mm in diameter. The garnet grains are partly arranged in bands and generally occur in the same bands together with biotite. This banded mode of occurrence is, however, very weak and noticeable only in some places. The grain size is 1—2 mm and the texture is granoblastic. The specific gravity is 2.63.

The constituents of the rock are plagioclase (ca. An_{20}), potash feldspar, quartz, biotite, garnet and oxide ore. The modal analysis is presented in Table V on p. 49. With regard to the amount of garnet reported (0.3 vol. %), it should be remarked that the thin section revealed the presence of only one garnet grain, and it is on this basis that the garnet content was determined. According to the megascopic estimation made of the sample, the true garnet content of the rock would seem to be somewhat greater, or around 0.5—1 vol. %.

The plagioclase exhibits distinct polysynthetic twinning. It contains hardly any potash feldspar as antiperthite. The potash feldspar occurs as xenomorphic grains 0.2—0.6 mm in diameter. The majority of the grains are clearly cross-hatched and have a noteworthy content of plagioclase as perthite. In an X-ray examination, potash feldspar separated from the specimen with heavy liquid was found to have two values of triclinicity, namely, 0.0 and 0.9. In this light, the rock contains both orthoclase and microcline.

The biotite is pleochroic from light brown to dark brown. The quartz exhibits very weak undulating extinction. The garnet is xenoblastic and contains quartz and also a bit of oxide ore as inclusions.

The garnet's refraction index $n = 1.796$ and the cell edge $a_0 = 11.528$ Å. (Determined by K. Hytönen).

From a sample weighing several kilograms, some 500 mg of fairly pure garnet was separated by means of heavy liquids, and it was subjected to a silicate analysis (analyst, H. B. Wiik) in spite of the difficulties attendant upon the scant amount of the material.

This analysis gave the following semiquantitative results: $SiO_2 = 38$ %, $TiO_2 = 0.05$ %, $Al_2O_3 = 23$ %, $Fe_2O_3 = 4$ %, $FeO = 27$ %, $MnO = 1.6$ %, $MgO = 6.6$ %, $CaO = 1.2$ %, $H_2O + = 0.2$ % and $H_2O - = 0.1$ % by weight. (Total = 101.75 % by weight).

Calculated according to this analysis, the garnet contains approximately 64—67 mol. % almandine, 26—28 mol. % pyrope, 3.5 mol. % grossular and 3.5—4 mol. % spessartite. The analysis is not closely in agreement with the theoretical composition of the garnet group. (cf. p. 33).

Garnet-bearing granodioritic rock. Specimen No. 14 572. Locality 27 (Plate I), west of the central gneiss between the inner and outer amphibolite rings.

Table V. Modal analyses of quartz dioritic and garnet-bearing granodioritic rocks (vol. %).

	1.	2.	3.	4.
Plagioclase	61.9	70.8	42.3	44.4
Potash feldspar	< 0.1	3.3	19.0	22.4
Quartz	29.1	25.2	36.7	27.6
Biotite	2.3	0.5	1.5	4.2
Garnet	0.0	0.0	0.3	1.0
Sericite	6.3	< 0.1	0.0	< 0.1
Others	0.4	0.2	0.2	0.4
Total	100.0	100.0	100.0	100.0

1. Quartz dioritic rock. Specimen No. 14 554. Description on p. 41.
2. Quartz dioritic gneiss layer in amphibolite. Specimen No. 14 536. Description on p. 46. Chemical analysis in Table IV, p. 42.
3. Garnet-bearing granodioritic gneiss. Specimen No. 14 518. Description on p. 47.
4. Garnet-bearing granodioritic rock. Specimen No. 14 528. Description on p. 49.

Both megascopically and microscopically the rock closely corresponds to the preceding sample. However, the garnet is clearly in a state of alteration into, e.g., biotite. The specific gravity of the rock is 2.63.

According to the chemical analysis, the rock contains 4.60 % Na_2O and 3.12 % K_2O by weight.

Garnet-bearing granodioritic gneiss. Specimen No. 14 528. Locality 28 (Plate I), northeast of the central gneiss between the inner and outer amphibolite rings.

Megascopically, the rock is reddish gray. Its foliation is generally weak and in places can hardly be distinguished. The rock is rather homogeneous, except that to a varying degree it contains roundish knobs about 1—6 mm in diameter composed mainly of garnet and mica. The grain size varies from 0.5 mm to 4 mm, but generally it is 1—2 mm. The texture is granoblastic (Fig. 16 on p. 50). The specific gravity is 2.59.

The principal constituents are plagioclase (ca. An_{20}), quartz, potash feldspar, biotite and garnet. The accessory constituents are sericite, chlorite, ore, sphene and zircon. The modal analysis is presented in Table V. The thin section contains only one garnet-bearing knob and the garnet content reported is based on this fact. It is hard to make accurate estimates of the actual garnet content of the rock.

According to the chemical analysis, the rock contains 4.14 % Na_2O and 5.78 % K_2O by weight.

The potash feldspar contains a considerable abundance of plagioclase as perthite and the plagioclase contains considerable potash feldspar as antiperthite. The mixture of potash feldspar and plagioclase is often so fine-

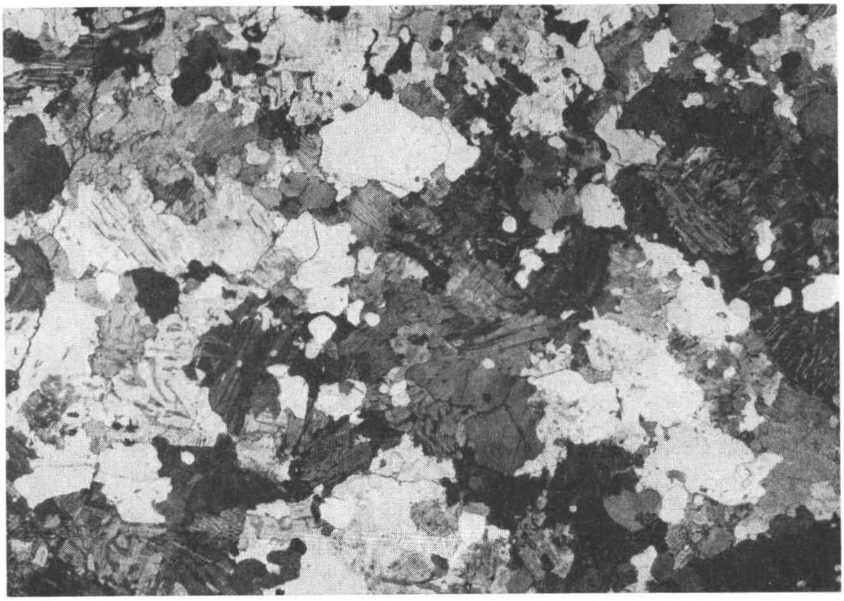

Fig. 16. Texture of garnet-bearing granodioritic rock. Specimen No. 14 528. Nicols +. Magn. 11 x. Photo: E. Halme.

grained that study of the minerals is difficult even under high magnifications. For this reason, also in modal analysis their ratio is apt to be erraneous. The plagioclase has undergone advanced sericitization.

The garnet, ore and main portion of the biotite and and chlorite all occur together in the thin section in an accumulation roughly 5 mm in diameter. Apparently, a garnet grain in a state of alteration is involved.

The garnet's refraction index n = 1.805 and the cell edge $a_0 = 11.567$ Å (determined by K. Hytönen).

Garnet-bearing granodioritic rock. Specimen No. 14 525. Locality 29 (Plate I), south of the central gneiss between the inner and outer amphibolite rings.

Megascopically the rock is light gray. The foliation is very distinct. The grain size is 1—2 mm and the specific gravity is 2.62.

The principal constituents are quartz, plagioclase (ca. An_{20}), and microcline. In addition, the rock contains green biotite, garnet, oxide ore, sericite and a trifle epidote and zircon.

The garnet and the main part of the biotite and the oxide ore occur together in an aggregate about 2 mm in diameter. Apparently, what is involved is a garnet grain that has for the greatest part altered.

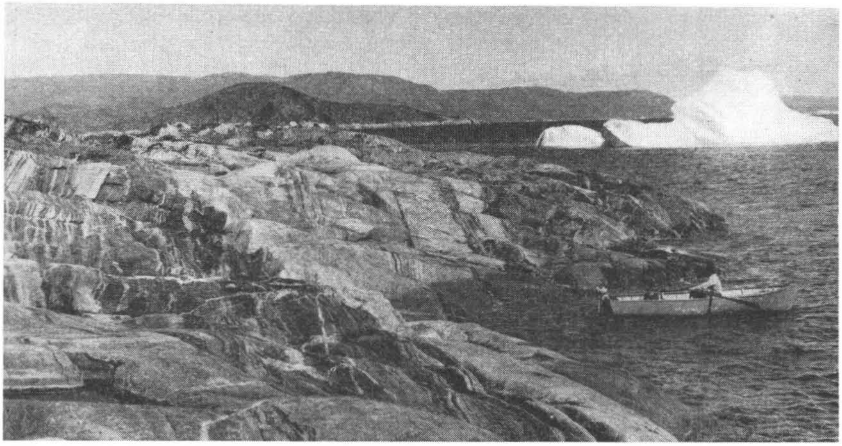

Fig. 17. Migmatic rocks. Niaqornatsiaq, western shore of Godthaabsfjord. Wiew towards north.

According to the chemical analysis, the rock contains 4.40 % Na_2O and 3.20 % K_2O by weight.

The plagioclase contains a fair abundance of sericite, but, on the other hand, scarcely any potash feldspar as antiperthite. The microcline is cross-hatched and in many cases contains abundant plagioclase as perthite. The quartz has a marked undulating extinction.

In the light of the examples described in the foregoing, these garnet-bearing rocks are granodioritic in composition. In three of the four specimens studied, the garnet is obviously in a state of alteration caused by retrogressive metamorphism.

MIGMATITES

The granodioritic and quartz dioritic rocks in the mapped region generally form migmatites with other rocks, especially amphibolites. Often, within a rather limited area, all transitional stages are to be found from amphibolites with only few granodioritic or quartz dioritic veins or schlieren to fairly pure granodioritic or quartz dioritic rocks with only few more or less altered inclusions or schlieren of amphibolite or related rocks. In many places the granodioritic and quartz dioritic rocks also compose more extensive homogeneous areas.

The strike and dip of amphibolitic and other inclusions in the granodioritic and quartz dioritic rocks are generally parallel to the strike and dip of the larger amphibolite layers situated in their proximity.

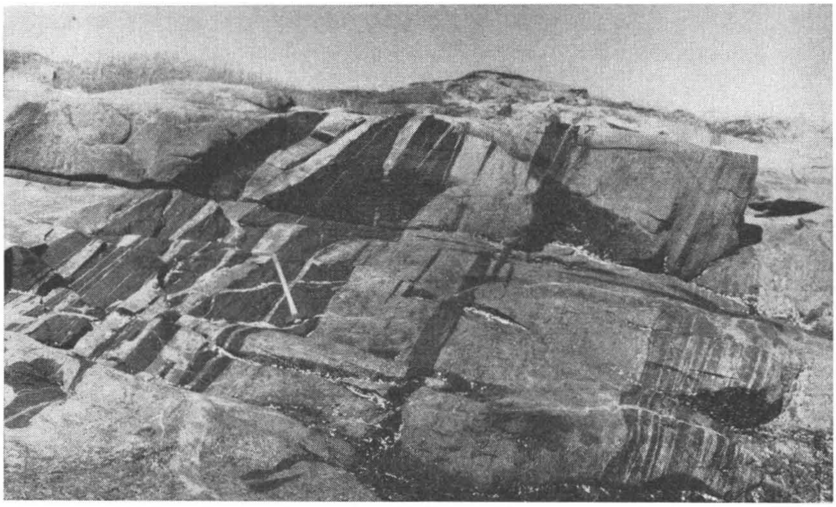

Fig. 18. Migmatic rocks. Western shore of Godthaabsfjord near Niaqornatsiaq. View towards north.

The structure and mode of occurrence of the migmatites in detail are best to be seen and photographed on the sea coast, where the exposures, polished by ice and wholly free of vegetation, are almost ideal for geological study (Figs. 17—19) or then in higher terrain, as, for example, north of the dome. (Fig 20). These figures, 17—20, show migmatite types quite characteristic of the mapped region.

In the following a few typical rock specimens from the migmatite areas will be described.

Migmatic gneiss. Specimen No. 14 574. Locality 30 (Plate I), eastern part of the dome, between the inner and outer amphibolite rings.

The rock is distinctly banded. In it there alternate light and dark bands, the thickness of which varies from a few centimeters to several dozen centimeters. The thin section has been made of the contact between a light and a dark band. The grain size is 1—2 mm and the texture is lepidoblastic, almost granoblastic. The specific gravity is 2.67.

The principal constituents are plagioclase (ca. An_{20}), quartz, biotite and potash feldspar. The accessory contituents are sericite, apatite, ore, sphene and zircon.

Modal analyses have been made of the light and the dark band separately, being presented in Table VI on p. 56. The dark band is seen to contain more biotite but less quartz and potash feldspar than the light band.

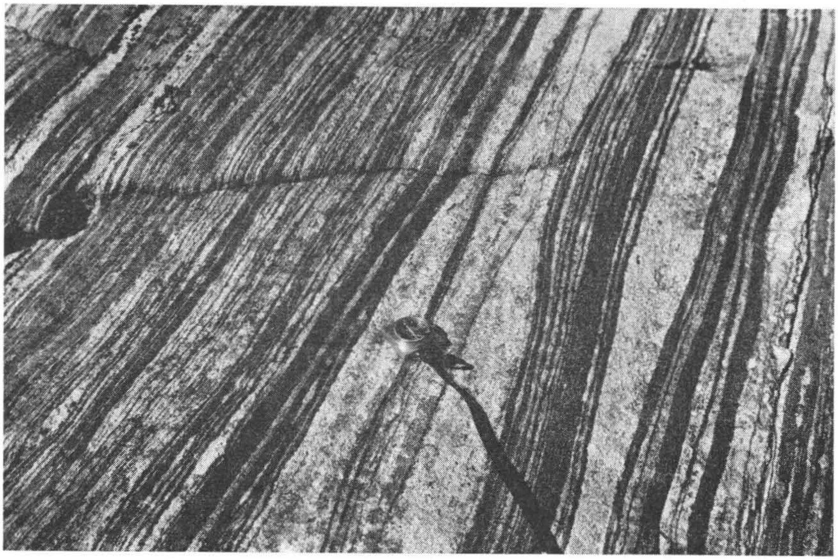

Fig. 19. Migmatic rocks. Western shore of Godthaabsfjord near Utsoqiuse. View towards north.

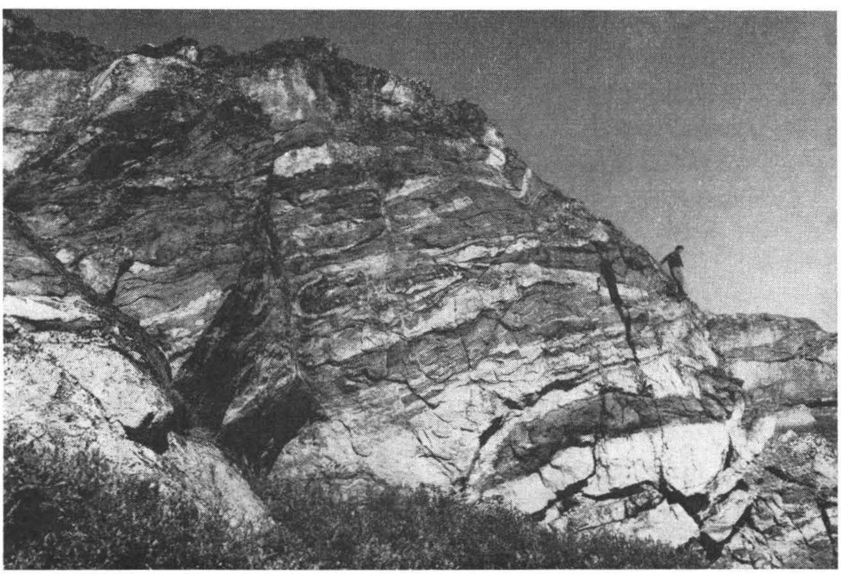

Fig. 20. Migmatic rocks. 4 km north of northwestern corner of dome. View towards northeast.

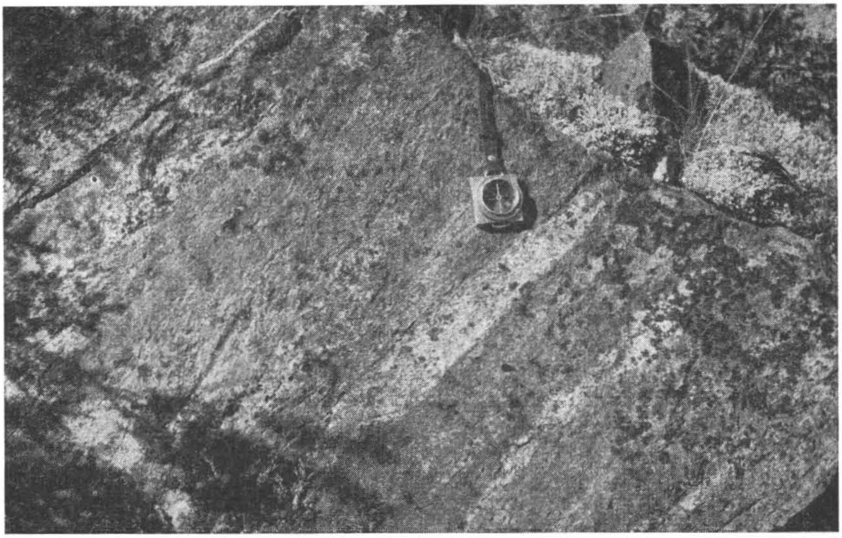

Fig. 21. Light bands in dark gray gneiss. West of inner amphibolite ring. View towards north.

Polysynthetic twinning is visible in the plagioclase only faintly or not at all. The plagioclase contains sericite and a small amount of potash feldspar as antiperthite.

The potash feldspar occurs partly in the plagioclase as antiperthite and partly as interstitial grains 0.1—0.3 mm in diameter. The biotite is pleochroic from light brown to dark brown. The quartz exhibits a marked undulating extinction.

Dark gray portion of migmatic gneiss. Specimen No. 14 516. Locality 31 (Plate I), about 50 m west of the inner amphibolite ring.

In the spot here mentioned there occurs dark gray gneiss, which megascopically resembles the marginal portions of the central gneiss. Deviating from the central gneiss, it has in places light bands a few centimeters broad (Fig. 21). The specimen here described consists of dark gray gneiss while the sample to follow, No. 14 517, represents the light band.

The dark gray gneiss is rather weakly foliated. Its grain size is 1—2 mm and the texture is granoblastic. The specific gravity is 2.68.

The principal constituents of the rock are plagioclase (ca. An_{25}), quartz and biotite. In addition, it contains small amounts of potash feldspar, apatite, sphene, sericite, ore and zircon. The modal analysis is presented in Table VI on p. 56.

Some of the plagioclase grains exhibit a distinct polysynthetic twinning, in others the twinning is only faintly visible or not at all. The entire potash feldspar content of the rock occurs in the plagioclase as antiperthite.

The biotite is pleochroic from light brown to very dark brown. The ore and sphene generally occur as inclusions in the biotite or in the immediate vicinity of biotite. In a few spots the ore grain is in the middle with a rim of sphene around it and biotite situated outermost.

Light band in dark gray gneiss. Specimen No. 14 517. From the same spot as the preceding sample.

The rock is light gray with dark, biotite-rich bands, the thickness of which varies from about 1 mm to 5 cm. The thin section was made from a piece of the sample that contained no biotite-rich bands. The grain size is 1—3 mm and the texture is granoblastic. The specific gravity is 2.65.

The principal constituents are plagioclase (ca. An_{20}), quartz, biotite, feldspar and sericite (muscovite). Accessory constituents are sphene, ore and epidote. The modal analysis is presented in Table VI on p. 56.

Polysynthetic twinning is visible in the plagioclase only faintly or not at all. The potash feldspar of the rock is present partly in the plagioclase as antiperthite, partly as interstitial grains 0.1—0.2 mm in diameter.

The quartz exhibits a marked undulating extinction. The biotite is pleochroic from light brown to dark brown. The sphene occurs mainly in the biotite.

Quartz dioritic gneiss. Specimen No. 14 513. Locality 32 (Plate I), eastern part of the outer amphibolite ring.

This sample has been taken from the same, roughly 200 m broad amphibolite layer as the pyroxene amphibolite specimens No. 14 511 and 14 512 described on pp. 22—24 as well as the hornblende gneiss specimen described on p. 27. According to the field observations, these rocks grade into each other. No sharp boundaries between these varieties of rock have been noticed.

In color this quartz dioritic gneiss is dark gray. It is fairly homogeneous and distinctly foliated. The grain size is 1—2 mm and the texture is granoblastic. The specific gravity is 2.72.

The principal constituents are plagioclase (ca. An_{35}), quartz, biotite and hornblende. In addition, there are small amounts of sericite, clorite, oxide ore, potash feldspar, apatite, epidote, sphene and zircon. The modal analysis is given in Table VI on p. 56.

Polysynthetic twinning is plainly visible in most of the plagioclase grains. Some of the plagioclase grains contain substantial amounts of sericite. The slight potash feldspar content of the rock (0.2 vol. %) occurs as tiny particles in the plagioclase.

Table VI. Modal analyses of migmatic gneisses (vol. %).

	1.	2.	3.	4.	5.
Plagioclase	65.1	64.1	60.4	51.1	50.3
Potash feldspar	0.7	3.5	0.3	3.4	0.2
Quartz	15.1	28.9	21.8	40.9	35.2
Biotite	18.7	3.3	16.4	2.8	11.4
Hornblende	0.0	0.0	0.0	0.0	1.3
Sericite	0.2	0.1	0.3	1.2	0.9
Others	0.2	0.1	0.8	0.6	0.7
Total	100.0	100.0	100.0	100.0	100.0

1. Dark band in migmatic gneiss. Specimen No. 14 574. Description on p. 52.
2. Light band in migmatic gneiss. Same specimen as above.
3. Dark gray part in migmatic gneiss. Specimen No. 14 516. Description on p. 54.
4. Light band in dark gray migmatic gneiss. Specimen No. 14 517. Description on p. 55.
5. Quartz dioritic gneiss. Specimen No. 14 513. Description on p. 55.

The biotite is pleochroic from light brown to dark brown or greenish brown. A portion of the biotite contains ore, sphene and zircon. Around the ore grains there commonly occurs a leucoxene rim and also biotite. The hornblende appears to be in a state of alteration into greenish biotite and chlorite. The quartz has a strong undulating extinction.

The samples of migmatic rocks described in the foregoing are mainly quartz dioritic in composition. The essential petrographic difference between the darker and lighter bands appears to be the fact that the dark bands contain much more biotite but appreciably less potash feldspar and quartz than the light bands do (Table VI).

ULTRABASIC ROCKS

Occurrences of ultrabasic rocks have been met with in considerable abundance in various parts of the mapped region, especially south, east and north of the dome. They are found in other varieties of rocks as conformable lenses. They vary greatly in size; some are only a few square meters or even less in area whereas the largest are over 500 m long and about 200 m wide. Some are situated in amphibolites or hornblende gneisses, some at or near the contact of the amphibolite layers and the granodioritic or quartz dioritic gneisses, while some are situated in the granodioritic or quartz dioritic gneisses. The situation of the largest ultrabasic lenses is shown on the geological map. Fig. 22 shows the mode of occurrence of the ultrabasic rocks in the field north of the dome.

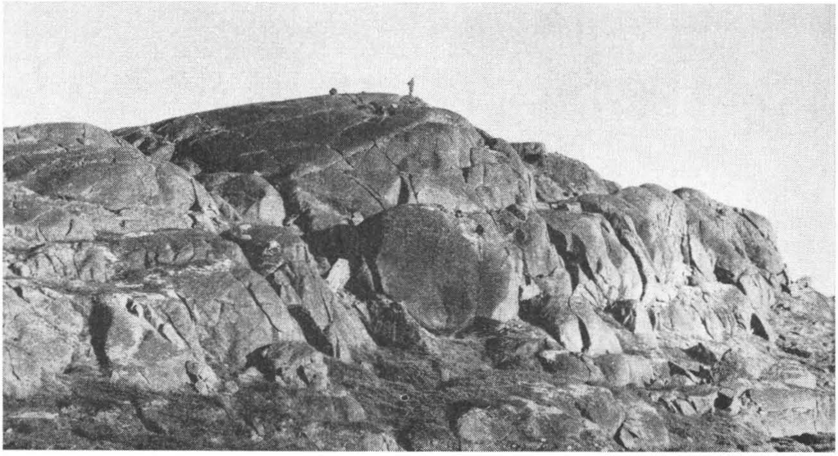

Fig. 22. Part of great lens of ultrabasic rock. 2.5 km north of northern border of dome. View towards southeast.

According to the field observations, some of the ultrabasic rocks contain mainly olivine and serpentine, and the principal constituents of others are hypersthene, hornblende and diopside.

The mode of occurrence, petrography, mineralogy and chemical composition of ultrabasic rocks similar to those of the Ipernat dome region are described in fair detail by Sørensen (1952, 1953, 1954) and Berthelsen (1960).

PEGMATITES AND APLITES

In the region investigated, pegmatites are comparatively rare, and none at all have been found in the area of the dome itself. On the northwestern shore of Godthaabsfjord, especially south of the Niaqornatsiaq peninsula, a few pegmatite dikes cutting across the foliation of the other rocks have been observed (Fig. 23). They vary in breadth from about 10 cm to about 5 m. Their strike varies, too, but mostly it is roughly E—W. The dip is steep or vertical. In addition to feldspars and quartz, the minerals met with in these dikes include muscovite, biotite, garnet and magnetite as well as a few small beryl crystals.

In different parts of the mapped region, some cross-cutting aplite dikes have been found. In general, these dikes are 10 to 50 cm wide but in some cases reach a width of a couple of meters. Their strike varies from N 70°E to E—W and their dip is steep or vertical. Besides feldspars and quartz, some biotite has been found in them. In one aplite dike some 20 cm broad

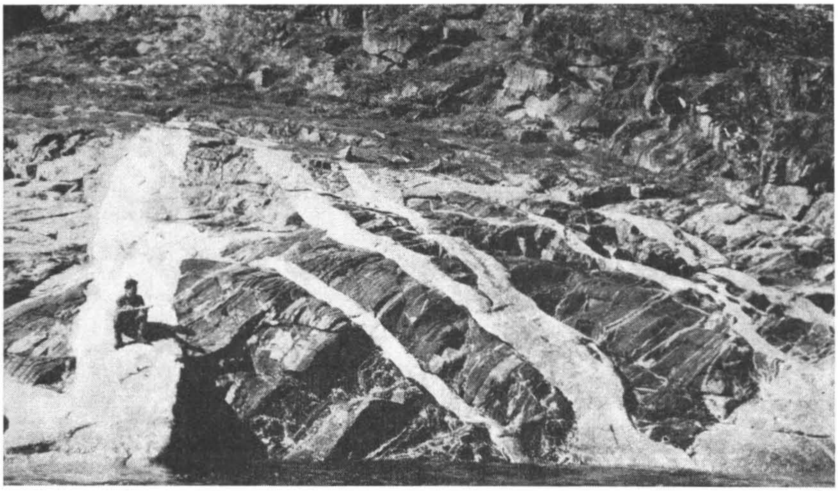

Fig. 23. Cross-cutting pegmatite dikes. Western shore of Godthaabsfjord near Niaqornatsiaq. View towards west.

near the northern edge of the dome there have been found numerous magnetite crystals measuring as much as 1—2 cm in diameter.

Ramberg (1956) has described in some detail the pegmatites on the shores of Godthaabsfjord as well as in other parts of West Greenland.

DIABASE DIKES

Only a few of the diabase dikes commonly occurring in western Greenland have been found in the mapped region. Their situation and strike are best to be seen on the geological map. Their dip is invariably steep or vertical. The dikes vary in breadth from about 20 cm to about 30 m, but their most usual breadth is 10 to 20 m.

In narrow dikes and near the contacts of broader dikes the grain size of the rock is 0.1—0.5 mm or even less, whereas in the middle portions of broader dikes the grain size is about 1—5 mm.

Recently, Berthelsen and Bridgwater (1960) have published a fairly thorough study on the field occurrence and petrography of similar basic dikes in the region immediately north and northwest of the area of the present investigation.

MINERAL FACIES CONDITIONS

As the foregoing petrographic descriptions have made clear, the majority of the rocks in the region investigated crystallized under PT-conditions approaching or corresponding to the granulite facies of Eskola (1939, pp. 360—363). The typical mineral assemblage of the basic rocks, represented here by pyroxene amphibolites, is hypersthene-diopside-hornblende-plagioclase in varying proportions. This mineral assemblage occurs in numerous thin sections in complete equilibrium and devoid of alteration products. In many cases, titaniferous magnetite appears to be associated with this mineral assemblage as an essential member.

The mineral assemblage plagioclase-diopside-hypersthene-hornblende conforms strictly to the hornblende-granulite subfacies of Turner (1958, pp. 232—235). The role of the titaniferous magnetite, which is present in the amount of as much as 10 vol. % (Table I, p. 28), is somewhat uncertain. According to Ramberg (1948 b, p. 553), the stability relationships of some silicates (hornblende, biotite, sphene and titanic augite) require a liberation of Ti and usually Fe from silicate lattices under granulite facies conditions and these elements can form titaniferous iron ore. According to Parras (1958, p. 102), the orthorombic pyroxene seems to remain stable to a certain Fe-content only (65 mol.% ferrosilite), whereupon the excess iron of the bulk composition of the rock becomes distributed between other phases, not least in anhydrous biotite — or it occurs outright as magnetite.

In the region here described, two of the samples dealt with (Nos. 14 512 and 14 524, pp. 23 and 25) reveal the presence of orthopyroxene in apparent equilibrium with titaniferous magnetite, and in both cases the orthopyroxene contains about 52 mol.% ferrosilite. In two other mineral assemblages apparently in equilibrium and including orthopyroxene (Spec. Nos. 14 511 and 14 515, pp. 22 and 24), no magnetite or ilmenite has been observed. The orthopyroxenes in these specimens contain about 45 mol.% and 30 mol.% ferrosilite, respectively.

In the majority of the pyroxene amphibolite samples examined microscopically, however, evidence of the effects of diminishing temperature and/or pressure can be noted. The hypersthene appears to undergo alteration into diopside and the diopside into hornblende. Around or near oxide ore grains, leucoxene often occurs and also in certain cases biotite. In places hornblende also appears to undergo alteration into biotite and/or chlorite.

In one thin section (Spec. No. 14 512, p. 23) sporadic signs of the effects of rising temperature and/or pressure have been noticed.

In rocks of granodioritic composition either hypersthene or garnet occur in many cases, particularly between the inner and outer amphibolite rings. No rocks have been encountered in the region in which these minerals occur together. In the hypersthene gneisses (pp. 28—31) the hypersthene has always been found to be more or less in a state of alteration. In garnet-bearing granodioritic rocks the garnet in some instances at least seems to occur together with biotite in a state of equilibrium without textural evidence of the biotite's having altered into garnet or vice versa (Spec. No. 14 518, p. 47). In other specimens (e.g., Spec. No. 14 528, p. 49) the garnet is clearly in a state of alteration apparently into various kinds of mica and oxide ore.

The occurrence of hypersthene and garnet in granodioritic and quartz-dioritic rocks indicate PT-conditions at or near those of granulite facies. According to Turner (1958, p. 232), who has divided the granulite facies into two subfacies, the pyroxene-granulite and hornblende-granulite subfacies, biotite does not occur in rocks conforming strictly to the specifications of the pyroxene-granulite facies. Accordingly, these hypersthene gneisses and garnet-bearing granodioritic rocks more correctly belong to the hornblende-granulite subfacies. It should be borne in mind, however, that in many instances, though not invariably, at least part of the biotite is distinctly diaphthoretic. Its occurrence in a mineral assemblage does not, therefore, provide evidence against the possibility that these rocks might have crystallized earlier under pyroxene-granulite facies PT-conditions.

According to Eskola (1939, p. 360), a characteristic of granulite facies varieties of rock is the high pyrope content of (Fe, Mg) Al-garnets, a content that in the granulites of Finnish Lapland varies between 47 and 55 mol. %, whereas the pyrope content of rocks belonging to the amphibolite facies rises to no more than 30 mol. %. The garnet of the garnet-bearing granodioritic rock (p. 48) contains, according to chemical analysis, approximately 26—28 mol. % pyrope; the rock would therefore more properly belong, after Eskola's (1939) facies classification, to the amphibolite rather than to the granulite facies. It nevertheless remains questionable to what extent the relatively low pyrope content of the (Fe, Mg) Al-garnet can be used as proof of the lack of PT-conditions of the granulite facies unless the chemical bulk composition of the rock is known. If the rock's MgO:FeO ratio is low, it would be unlikely that the garnet's pyrope content would be high, even where the PT-conditions might allow it.

In the sole garnet-cordierite gneiss occurrence of the region investigated (pp. 31—33) the garnet contains, according to the chemical analysis, approximately 40—45 mol. % pyrope, or nearly as high a content as in the typical granulites of Finnish Lapland, (Eskola, 1939, p. 360). This variety of rock also contains cordierite and biotite. The latter mineral appears to be at least partially diaphthoretic and perhaps does not belong to the assemblage during the crystallization of which the garnet achieved its present chemical composition. On the other hand, it is not known whether at the stage of incipient alteration the chemical composition of the garnet might have changed somewhat from what it was originally. Judging by the rock's mineral composition, its Mg content is fairly high.

In different parts of the region, but especially in the southeastern part of the dome, evidence has been found of rather strong regional retrogressive metamorphism. In such places the amphibolitic rocks no longer contain hypersthene and diopside occurs only as an unstable relict. The hornblende turns unstable, too, and alters partly into biotite and/or chlorite. The oxide ore minerals are in some instances surrounded by leucoxene and/or biotite. The plagioclase is in many cases more or less sericitized. The hypersthene of the hypersthene gneisses is in a state of alteration into serpentine, biotite and oxide ore minerals. The garnet of the granodioritic rocks is in a state of alteration into various micas and oxide ore minerals (pp. 49—50).

Nowhere has this retrogressive metamorphism been observed to have led to the creation of a new mineral assemblage completely in equilibrium. The boundaries of the granulite facies and other facies corresponding to lower PT-conditions are not sharp but represent zones of gradual transition. The most distinct transitional zone is found in the eastern part of the dome, where the hypersthene gneiss area marked on the geological map in yellow gradually passes over into an area of granodioritic and quartz dioritic gneiss. The breadth of this transitional zone is, according to field observations, several hundred meters. In the same area, the outer amphibolite ring reveals plain evidence of a strong retrogressive metamorphism.

The retrogressive metamorphosis has probably been strongly promoted by metasomatic changes, above all, by an increase in water and also at least SiO_2. No quartz-free amphibolites that had crystallized in complete equilibrium under PT-conditions of the amphibolite facies (after the facies classification of Eskola, 1939) whatsoever have been found in the region, but the alteration of pyroxene amphibolites into amphibolite invariably seems to be associated with the occurrence of quartz and also chlorite in the mineral assemblage — at least as minor constituents (Table I, p. 28).

In the great fault zones of the region there are frequent occurrences of epidote and chlorite. In some cases, again, the rocks of the fault zones have been mylonitized and turned into completely aphanitic varieties. The appearance of epidote and chlorite in these zones indicates that, at some stage, at least, during the faulting the regional PT-conditions had permitted the crystallization of such minerals. Faulting of this kind, of course, causes facies changes only of a local character.

CONTENT OF POTASSIUM AND SODIUM IN QUARTZ DIORITIC AND GRANODIORITIC ROCKS

In studies dealing with gneiss domes, the possible roles of potash metasomatism, granitization, granitic magma and other such factors in the formation of the domes have sometimes been the subject of lively discussion. For this reason, efforts have been made in the present investigation to determine the Na_2O and K_2O content of granodioritic and quartz dioritic varieties of rock. For this purpose samples were collected that, according to field observations, represented the average composition of such rocks as closely as possible. By means of megascopic and microscopic comparison, the most representative samples taken from different parts of the region investigated were selected for analyses. In considering the reliability of the results from the standpoint of the representativeness of the samples, it should be noted that all the ones analyzed were taken from places which, in the light of field observations, are free of inclusions, schlieren and veins and are fairly homogeneous. The sample crushed and homogenized for analysis generally weighed about one kilogram. The Na_2O and K_2O contents were determined gravimetrically. The results of the analyses are presented in Table VII as well as in figures 24 and 25, which also show the sampling sites. On the basis of these results, it would appear that K_2O was slightly concentrated in the area between the inner and outer amphibolite rings, but not, considered on the average, in the central gneiss. In the central gneiss the K_2O content is distributed so unevenly that, judging by the six analyses available, no reliable conclusions can be drawn with respect to the distribution of K_2O in the different portions of this gneiss.

The textural mode of occurrence of the potash feldspar varies and appears to depend considerably on the content of this mineral in the rock. The potash feldspar content of certain quartz dioritic rocks is only a fraction of one vol. %. In such cases, the potash feldspar usually occurs as antiperthite in the plagioclase. As the amount of potash feldspar increases, it generally occurs as both antiperthite in the plagioclase and as small interstitial grains. At its maximum, the potash feldspar content of granodioritic rocks rises to roughly 20 vol. %. In these varieties of rock, the

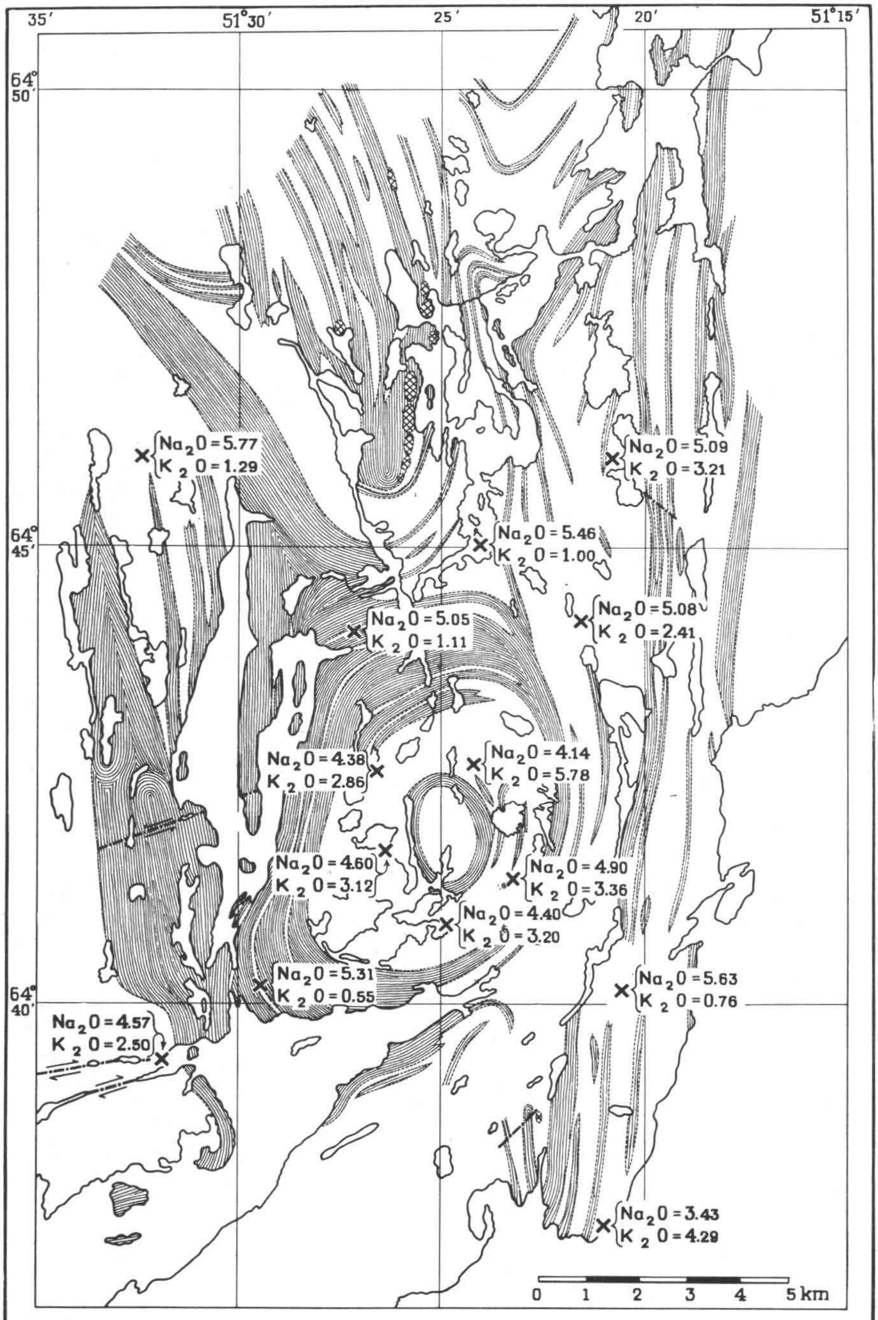

Fig. 24. Content of Na_2O and K_2O in quartz dioritic and granodioritic rocks outside of central gneiss. All values weight %. Compare Fig. 25.

Table VII. Gravimetric determinations of sodium and potassium content of quartz dioritic and granodioritic rocks (weight %). Analyst, H. B. Wiik.

Specimen Nos	Na_2O	K_2O	Na_2O + K_2O	Petrographic description on page
Central gneiss				
14 501	4.83	2.41	7.24	34
14 502	4.86	1.50	6.36	36
14 506	5.50	1.15	6.65	36
14 507	5.52	0.90	6.42	37
14 508	5.83	0.85	6.68	38
14 510	4.50	3.38	7.88	38
Average values	5.17	1.70	6.87	
Area between the inner and outer amphibolite rings				
14 514	4.38	2.86	7.24	29
14 525	4.40	3.20	7.60	50
14 528	4.14	5.78	9.92	49
14 572	4.60	3.12	7.72	48
14 573	4.90	3.36	8.26	44
Average values	4.48	3.66	8.14	
Quartz dioritic gneiss layers in amphibolites				
14 536	5.05	1.11	6.16	46
14 569	5.31	0.55	5.86	47
Average values	5.18	0.83	6.01	
Surroundings on the dome				
14 554	5.46	1.00	6.46	41
14 561	5.77	1.29	7.06	41
14 568	4.57	2.50	7.07	42
14 576	3.43	4.29	7.72	43
14 587	5.09	3.21	8.30	43
14 597	5.63	0.76	6.39	43
14 613	5.08	2.41	7.49	44
Average values	5.00	2.21	7.21	
Average values of all areas	4.96	2.10	7.06	
Average values of all 20 analyses	4.94	2.28	7.22	

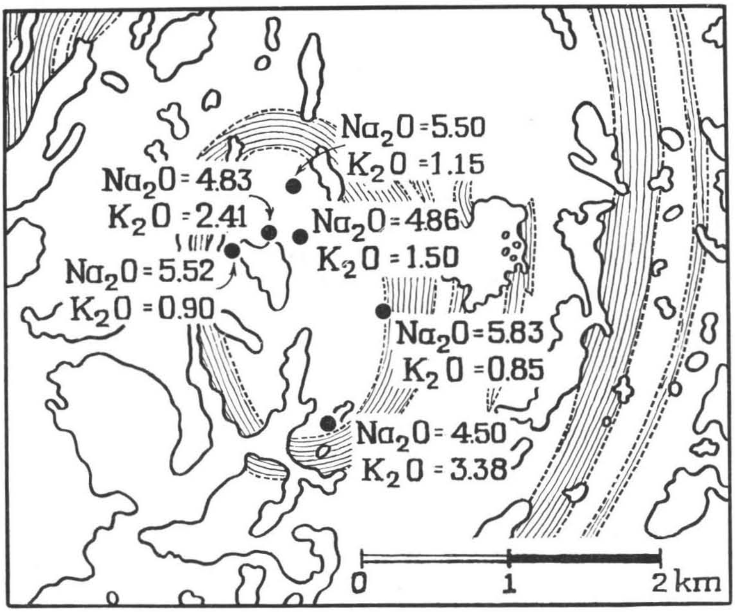

Fig. 25. Content of Na_2O and K_2O in samples from central gneiss. All values weight %. Compare Fig. 24.

textural relations of the potash feldspar and the plagioclase appear to vary greatly. The plagioclase contains potash feldspar and the potash feldspar contains plagioclase in varying amounts. The greatly varying K_2O content of granodioritic and quartz dioritic rocks indicates that the potassium may be, at least in part, of metasomatic origin.

TECTONIC DESCRIPTION

BEDDING AND FOLIATION

In mapping and studying geological structures, it is extremely important to ascertain whether an S-surface appearing in the field runs parallel to the bedding or not. Discriminating reliably between original bedding and foliation of metamorphic origin is perhaps one of the most difficult problems in the analysis of structures of highly metamorphic areas, and erring in this respect is easily apt to lead to a completely wrong conception of the structure of an area (cf. Billings 1954, pp. 344—345, Noe-Nygaard and Berthelsen 1952, pp. 258—262). Also in the region here under consideration, this discrimination has proved to be a highly difficult matter in many places, and in some cases no solution could be reached.

In many places, but mainly in the western and northern parts of the dome, the mode of occurrence of the rocks resembles closely the stratification of younger and far more weakly metamorphosed formations, especially when viewed from some distance (Fig. 13 on p. 44). The mode of occurrence of the amphibolites and hypersthene gneisses — which in places extend as gently dipping or nearly horizontal layers of practically unvarying thickness and petrographic character for several kilometers — gives the distinct impression of original bedding.

The pyroxene amphibolites and amphibolites are seen, upon close examination, to be usually more or less distinctly layered. Dark and light layers alternate and the thickness of the layers varies between a few meters and a few millimeters. The most significant difference between the dark and the light layers appears to be that the darker ones contain a greater abundance of mafic minerals, primarily hornblende (Table I, p. 28). It is, however, hard to determine to what extent this alternation of light and dark layers is due to changes of material occurring during the sedimentation and to what extent to segregation of the rock components in various layers or bands during metamorphism (cf. Turner 1941, pp. 1—16, and Billings 1954, p. 345).

In a large part of the mapped area, particularly in the vicinity of the dome itself, with the exception of its eastern portion, the dip of the layering is 45° or less and in many places only between 5° and 20°. In these areas there was not observed foliation whose plane intersects the bedding plane. The only S-surface to be observed in the field is that of the primary bedding, which is likely to be further accentuated by concordant foliation and/or segregation banding of metamorphic origin, This type of foliation is designated by Mead (1940, p. 1009) »bedding foliation» and by Billings (1954, pp. 340, 343) »bedding cleavage» or »bedding schistosity».

In a few places evidence has been observed of the occurrence of two different S-surfaces in the same exposure. In such cases one S-surface appears to be that of the original bedding and the other one that of foliation of metamorphic origin, which intersects the former.

One such place is situated northeast of the central gneiss approximately half way between the inner and the outer amphibolite ring (Locality 33, Plate I). In addition to a gently dipping, in spots nearly horizontal S-surface, there occurs at this locality steeply dipping or vertical foliation. According to the field observations, two intersecting S-surfaces are involved.

A similar place has been found on the northern shore of the lake Quagssup Taserssua, on a curved peninsula, southwest of the dome (Locality 34, Plate I) The curved form of the peninsula indicates a large fold and the same impression has been obtained through stereoscopic examination of aerial photographs. Field observations made at the western end of the peninsula indicated, however, that a steeply dipping foliation ran straight N 15°E across it, intersecting the plane of the probable bedding. In the same place, another, much weaker and gently dipping S-surface was observed, its attitude conforming quite well to the shape of the probable fold. The rock in which these two intersecting S-surfaces have been noted is a rather fine-grained, pale gneiss. Owing to the scantiness of mafic minerals, these S-surfaces are rather faintly in evidence. In the eastern part of the peninsula, where the rock is amphibolite, two intersecting S-surfaces have not been found.

Similar observations of the existence of two intersecting S-surfaces have been made also elsewhere in the region investigated, particularly to the west and to the north of the dome. In these, as in the foregoing cases, one of the S-surfaces is probably parallel to the original bedding. The other, intersecting S-surface probably represents foliation of metamorphic origin. The relation of the latter to major structures could not be determined with certainty on account of the scantiness of the observations. The probability is, however, that it developed parallel to the axial plane. What we have, therefore, is probably an axial plane cleavage, according to Mead (1940, p. 1010).

During the field work some suitable amphibolitic layers were followed step by step for several kilometers and the attitude of the S-surface to be seen in the amphibolite was measured at regular intervals. In certain places it was observed, however, that the strike of the only S-surface, foliation and/or banding apparent in the amphibolite did not conform to the strike of the amphibolitic layer but deviated from it in many instances by 10° to 20°. In other words, this S-surface does not run parallel to the original bedding.

These observations have been made in areas that are wholly exposed and where the amphibolitic layer is completely visible both in the field and in aerial photographs. Accordingly, the strike of the amphibolitic layer could be examined and mapped quite reliably. The observations have been made chiefly on the northeastern and northern sides of the dome, where the dips of the foliation, layering and probably also the bedding are steep or vertical.

The reasons for this behaviour of the S-surfaces of the amphibolites may be many and differ with each separate case. In some cases the effect of local magnetic anomalies was suspected — this would have meant that the whole phenomenon was illusory, but this could not be demonstrated. It is also possible that the amphibolite is in some places folded, although the layer itself does run rather rectilinearly and the field observations are not sufficiently detailed to verify the folding. Locally the foliation and/or banding is likely to turn more or less parallel with a fault, etc. In some localities the observations were too numerous and consistent, however, to be explained in these ways. A far more plausible explanation is that in the field only foliation and, perhaps, segregation banding of metamorphic origin can be observed and that the original bedding has been nearly or totally obliterated. Detailed study is greatly hampered, however, by the fact that in many places the amphibolites are strongly weathered and the poor quality of the exposures is apt to prevent detection of two intersecting S-surfaces of different ages.

The S-surfaces measured in the field have been marked on the geological map with the same symbol irrespective of wheather they appear, in the light of field observations, to represent original bedding, foliation of metamorphic origin or, perhaps, segregation banding.

FOLDING AND LINEATION

The large folds, which are presented on the geological map, can invariably be recognized in the aerial photographs either directly or under stereoscopic examination. Their existence has then been verified by following

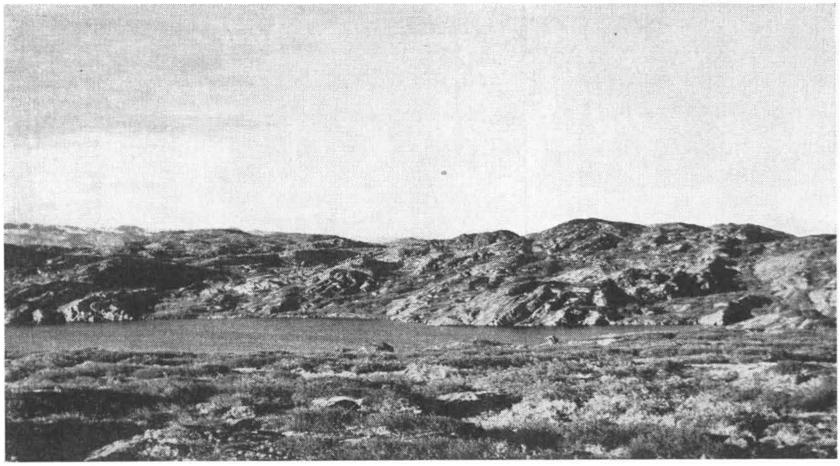

Fig. 26. Gentle anticlines in northern part of dome. Crests of folds are situated at highest tops of ridges, or 20—30 m above level of lake. Site is 1 km north of northern contact of central gneiss. View towards N 60°E.

in the field step by step some suitable layer of rock; almost without exception amphibolites or related rocks have served this purpose. The folds identified in this way have generally been so wide that their observation *in toto* from a single spot on the ground is in most cases impossible. They commonly measure from a few hundred meters to several kilometers across.

In the northern part of the dome, folding has been observed on a slightly smaller scale than that described in the foregoing. This folding is very gentle. Its wave length is a few hundred meters and the crests of the anticlines are at most 10 to 30 meters higher than the troughs of the synclines (Fig. 26). In connection with such folds, there often occurs jointing developed perpedicular to the fold axis (Fig. 27).

The small-scale folding, which varies in width from a few dozen centimeters to several meters, is mostly very gentle, just as are the large folds. In the area of the dome no exceptions to this have been met with, but a few do occur in the surrounding terrain. In some instances the small-scale folding has been so gentle that it has been advisable to apply to it a special tectonic symbol of its own. The axes of these very gentle small folds deviate in some cases distinctly from the attitude of the regional fold axis (Plate II), and the elucidation of the mode of origin of such very gentle small-scale folding is difficult. One possible explanation is that it originated in connection with boudinage.

Some of the small folds are typical drag folds in shape. One of them is situated on the eastern flank of an anticline several hundred meters wide,

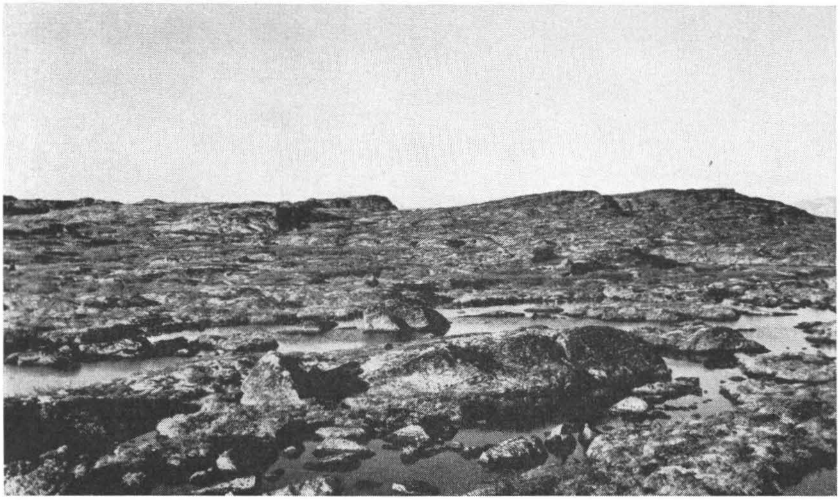

Fig. 27. Ridges running parallel to fold axis and jointing approximately perpendicular to latter. Fold axis plunges 5°—10° north, or to left in picture. View toward N 70° E. Tops of ridges are about 20—30 m higher than foreground.

Fig. 28. Drag folds in migmatic rock. Eastern flank of almost isoclinal anticline several hundred meters wide plunging 15° north. View towards north. Locality 36 in Plate I.

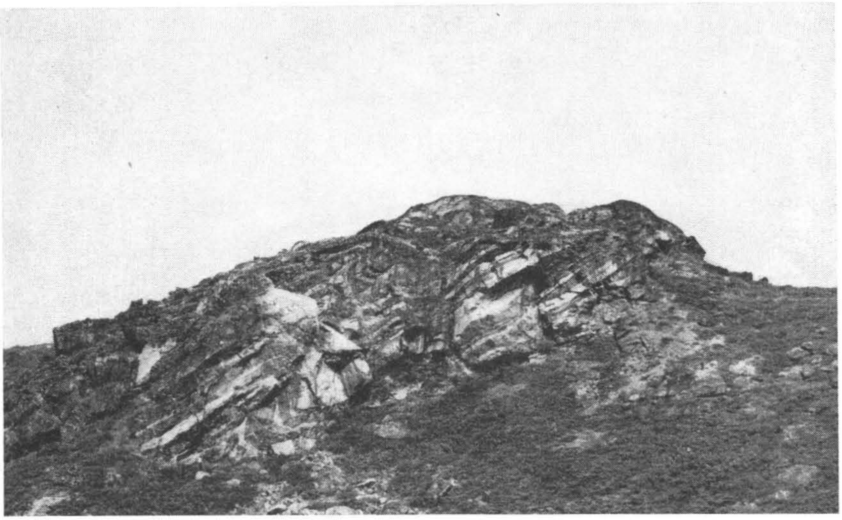

Fig. 29. Great drag fold situated on western flank of anticline about 2 km wide. Height of wall in picture is about 25 m. Northwest of dome. Locality 35 in Plate I.

west of the dome. (Fig. 28, p. 71). Its locality is marked on the Plate I with the number 36.

An exceptionally large and clear-cut drag fold (Fig. 29) is situated northwest of the dome (Locality 35, Plate I). The height of this fold is roughly 20 meters. It rests on the western flank of an anticline about 2 km wide.

Especially in the study of the structures of high-metamorphic complexes, the term »lineation» has been very much used and misused. The term itself has fallen into considerable disrepute because it has been applied to a great variety of linear structures without any clear definition of its signification in each separate case. In the present study the lineation has been measured on the bedding or foliation planes, where it is due to parallel alignment of mineral grains, mainly hornblende prisms or biotite flakes. It has been endeavored to make observations of this type of lineation, but the results achieved have remained slight. It has proved difficult to measure reliable lineations from the principal rock types of the region, namely, granodioritic and quartz dioritic gneisses, hypersthene gneisses and amphibolites.

The few lineation observations that seem reliable are presented in Plate II. In the light of these observations, this kind of lineation agrees most often but by no means invariably with the conception of the attitude of the regional fold axes arrived at by other means.

On the other hand, observations of the lineations have not been appreciably needed in the elucidation and mapping of larger geological structures, for other procedures have proved far more reliable. The trend and plunge of the fold axis of large folds has generally been determined from those points where the crest of the anticline or the trough of the syncline »rises into the air», i.e., intersects the surface of the ground, and where beds of the opposite limbs converge and the beds show a maximum curvature. By means of a stereographic projection, the attitude of the fold axis can be constructed from a few measurements of the strike and dip of a bed in such a place much more reliably than is possible on the basis of any observations of small folds or lineations (cf. Wegmann, 1929 a, b; Phillips, 1954).

JOINTING AND FAULTING

During the field work, the jointing of the region investigated proved so complicated and difficult to study that no systematic field observations or measurements were made of it. A far clearer and more unified picture was observed to be obtainable by photogeological means (Plate III).

The relation of a few joint sets to other geological structures appears rather plain. One joint set is always parallel to the bedding or foliation. Another joint set has a strike parallel to the bedding or foliation, but the dip of the joints is approximately perpendicular to the dip of the bedding or the foliation. These two joint sets appear very clearly in, e.g., the marginal parts of the central gneiss, especially close to its northern contact (Fig. 8 on p. 34) and in association with gently dipping amphibolite and other layers. In certain cases the jointing perpendicular to the fold axis of major folds is also highly developed (Fig. 27 on p. 71).

Numerous large fault zones have been observed in the mapped region, being clearly visible in aerial photographs and topographic maps as well as in the field. Cutting across high ridges and mountains, they frequently form deep gorges. According to photogeological investigations and field observations, there are at least several dozen smaller fault zones. In most of the cases studied, the fault plane appears to have a steep or vertical dip.

Insofar as possible, efforts were made both in the field and by photogeological means to obtain data on the nature of the movement along the faults and on their age relations.

In the southwestern corner of the region covered by the geological map, two fault zones intersect two nearly parallel diabase dikes with a N—S trend. The trend of the more southern fault zone is N 70°E and that of the one farther north N 80° E. Each of them appears to be approximately vertical. These dikes have enabled determining the direction and magnitude of the strike-slip component in each of the fault zones. In both cases the more northern block has moved eastward compared to the more southern block. In other words, according to Anderson's terminology (1951, p. 59), they both are dextral wrench faults. The strike-slip in the more southern fault zone is about 70 to 80 meters and in the more northern one about 10 to 20 meters. These strike-slip components are fairly definitely established and not merely apparent, for both diabase dikes are just about vertical. No reliable data exist on the magnitude of the possible dip-slip component, but it seems to be small or negligible. The magnitude and direction of the given displacements naturally apply only to the time after the intrusion of the diabase dikes. No other reliable marker layers besides these dikes have been found. In the light of photogeological studies, however, other morphological features yield the same result as the diabase dikes. Accordingly, at least the greatest part of the movement along the fault zone took place after the intrusion of the diabase dikes. In the morphology both fault zones occur as gorges the depth of which is mostly from 5 to 20 meters. The breadth of the gorges is mostly 10 to 30 meters.

Some 5 km north of these fault zones there is located a third, which runs fairly parallel to the other two across the end of the Usuk peninsula. It trends N 70°E and seems to be nearly vertical. In the topography it occurs as a slightly shallower and narrower ravine. No point has been found in the mapped area where this fault intersects a diabase dike or some other dependable marker layer. Consequently, it has not been possible to determine the direction and magnitude of the relative movement as readily as in the cases described earlier. Some field observations indicate, however, that the more northern block has moved westward, in comparison with the more southern one, roughly 10 to 20 meters. According to photogeological investigations, the direction of the displacement is the same and the magnitude of the strike-slip about 20 to 40 meters. In contrast to the previously described faults, this one thus appears to be sinistral. The magnitude of the possible dip-slip component is unknown.

Between Godthaabsfjord and the southeastern part of the dome, there are two large fault zones, which have been followed for a distance of about 10 km by means of aerial photographs and in the field. The trend of the more northern fault zone is N 60°E and that of the more southern one N 45°E. The dip appears to be vertical. No reliable observations have been obtained of the direction and magnitude of the displacement along these faults. One somewhat unreliable observation has been made at a point where the more northern fault zone runs across Niaqornaq peninsula. In the southern part of the peninsula there is a diabase dike which appears to be cut off at the point of intersection with the fault. No extension of the dike, properly speaking, has been found. In an area of a few square meters on the northwestern wall of the gorge formed by the fault zone, over 200 m from the northern end of the dike toward the northeast, the remains of a strongly crushed and mylonitized diabase dike have been found. If these remains derive from the same dike, the northern block has moved northeast in comparison with the more southern one and the magnitude of the strike-slip is over 200 m. No other data supporting this view are available.

In the region investigated a few other faults as well have been found, the direction of displacement and the magnitude of the strike-slip component of which could be determined principally by photogeological means. They are also shown on the geological map. In all these cases the magnitude of the strike-slip appears to be 10 to 30 meters.

As the appended aerial photographs and the Plate III show, the entire region under investigation is morphologically exceedingly broken up. In most cases, however, it is difficult to determine whether a zone or line observed in aerial photographs or in the field intersecting other geological structures, is a fault with a remarkable displacement, a fault in which the total displacement is so small — a few meters or less — that it is hard to recognize either in the field or by photogeological means, or just a joint possibly enlarged by erosion. One is apt to walk across the smallest zones in the field without noticing them, although they can be clearly distinguished in aerial photographs under stereoscopic examination. The larger fault zones usually occur in the topography as quite clear-cut gorges.

The central parts of the larger fault zones are very seldom exposed. Ordinarily they are covered by boulders, water or vegetation. For this reason, the reliable elucidation of petrographic and mineralogical changes that possibly occurred in connection with faulting is difficult. Usually observations relating to it have to be made from the walls of the fault zone, where the rocks have probably not undergone such far-reaching alterations as in the middle of the zone.

In the three fault zones described in the foregoing — the ones situated to the west and southwest of the dome — the rocks appear to have broken only mechanically in connection with the faulting. Furthermore, weak epidotization has been observed in places in the walls of the fault zones. In none of these three fault zones, however, is the middle of the zone exposed.

In many places in the great fault zones situated to the southeast of the dome, marked mylonitization (Fig. 30) and, in addition, some epidotization have been observed. In certain places the rocks have altered completely aphanitic. The middle portions of the fault zones are seldom exposed here, either.

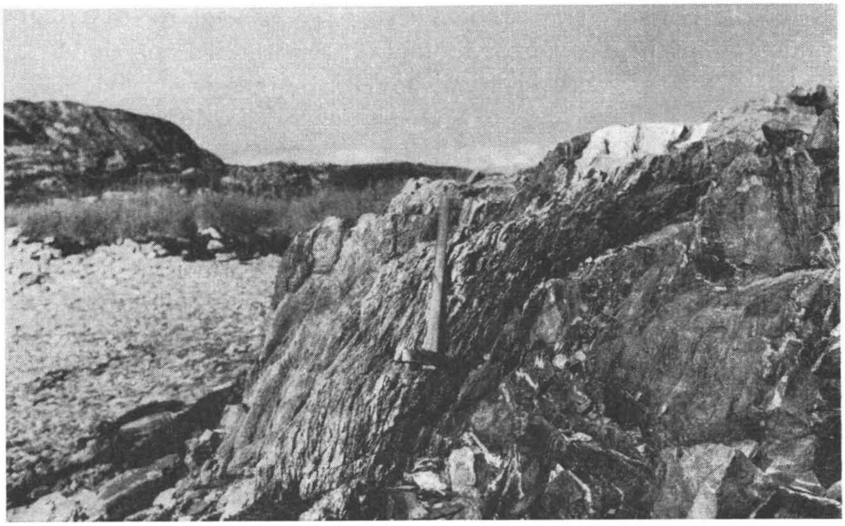

Fig. 30. Southeastern border of great fault zone. Niaqornaq. View towards northeast.

The trend of the great fault zones discussed in the foregoing is approximately the same as that of the Fiskefjord faults (Berthelsen and Bridgwater, 1960, pp. 14—15; Berthelsen, 1962). With but a single exception, they show dextral displacement, they cut up the diabase dikes and the rocks in them have undergone by and large the same kind of petrographic alterations as in the Fiskefjord faults. Accordingly, it seems probable that the great fault zones of the region investigated are approximately of the same age as the Fiskefjord faults.

ON THE STRUCTURE OF THE DOME

The main features of the structure of the dome and its surroundings are best to be seen from the geological map and Plate II, which includes various data on the fold axes together with related observations. Matters concerning the structure of the dome have also been discussed in the previous chapters.

As previously the middle of the dome can be described as a homogeneous quartz dioritic and granodioritic gneiss area. This central gneiss is conformable overlain by an amphibolitic layer, which dips radially outwards from the center of the gneiss at 20°—45°. The thickness of the amphibolitic layer varies, according to estimates made on the basis of its outcrop breadth

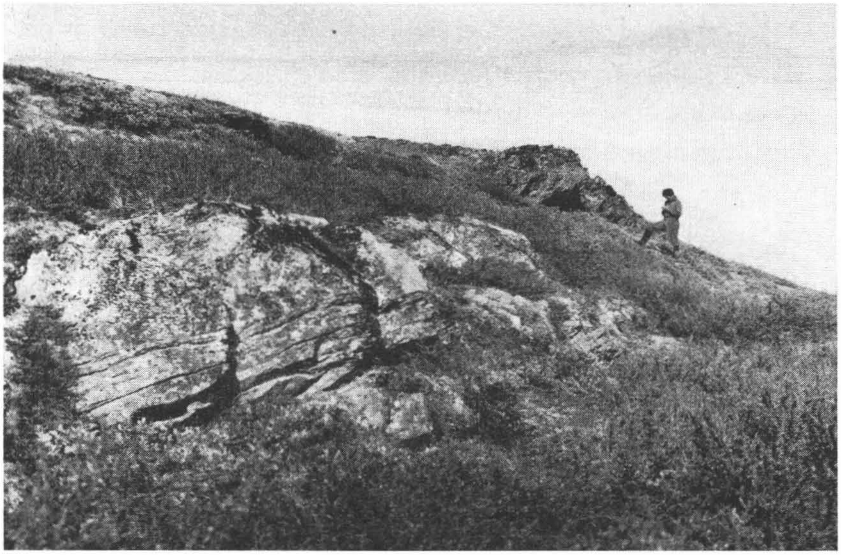

Fig. 31. Northern contact of central gneiss. Central gneiss in foreground and amphibolite on upper slope. Person at right is standing at contact. View towards northeast.

and dip, from approximately 100 m in the western part of the layer to slightly over 200 m in the eastern part of the layer. The form of the central gneiss and the amphibolitic layer surrounding it is a dome or doubly plunging anticline, which is slightly tilted eastward. Figs. 6—8 on pp. 32—34 offer a general view of the vicinity of the central gneiss.

The foliation to be seen in the marginal portions of the central gneiss, the contact between the gneiss and the amphibolite and the S-surfaces noticeable in the amphibolite — probably bedding — are all conformable. The foliation of the central gneiss is weak in its marginal portions and in the center of the gneiss it is scarcely detectable. The reason for this may be only or mainly the fact that the gneiss does not contain more than very small amounts of mafic minerals.

Apophyses possibly penetrating the amphibolite from the central gneiss have been sought but not found. At all the points investigated where the contact of the amphibolite and the central gneiss is exposed, it appears to be a shear surface or a shear zone. The thickness of this shear zone reaches 10 to 50 cm. In many spots it contains abundant biotite, whereas in other places it tends to be cataclastic.

In certain places near the northern and northeastern contact of the central gneiss, cataclastic zones and biotite-rich shear surfaces have been found in the gneiss parallel to the contact. Their distance from the contact

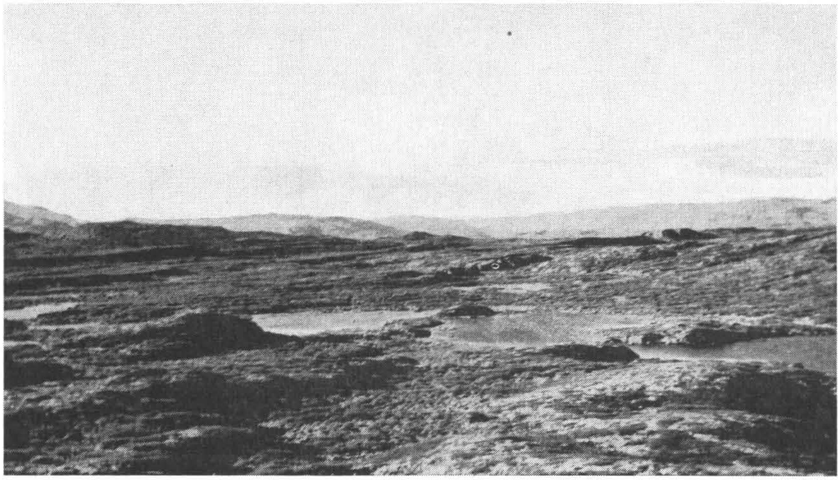

Fig. 32. View towards northeast over northermost part of dome while sun is setting in northwest. Bedding dips very gently north and northwest or is nearly horizontal.

is as much as 10 to 20 meters, measured over the ground. The amphibolite layer thus appears to have slipped along the contact between the amphibolite and the central gneiss, and shearing and cataclastic movements have taken place to a smaller extent in the gneiss itself as well. Fig. 31 shows the mode of occurrence of the northern contact of the central gneiss and the amphibolite in the field.

Evidence of possible diapiric movements has been sought in the central gneiss and its surroundings but with scanty results. The dip of the contacts of the central gneiss is not appreciably steeper than the dip of the bedding and/or the foliation in its surroundings.

Boudinage has been met with at a number of places in the lower part of the inner amphibolite layer near the western contact. It usually occurs in the immediate vicinity of the contact, i.e., 20 to 50 cm away. The length of the boudins along the strike of the amphibolitic layer is generally in the range of 10 to 20 cm, but the direction of their longest axis seems to be radially outward from the central gneiss. This boudinage is of too local development and insufficiently clear, however, to be regarded as reliable evidence of the diapiric rise of the central gneiss.

The structure of the outer portions of the dome is also fairly simple, and it can best be seen on the appended maps and on the Figures 32—33. The tilting of the dome toward the east is far more conspicuous, however, in its outer parts than in the central gneiss and its surroundings. The outer amphibolite layer is slightly overturned in the eastern part of the dome.

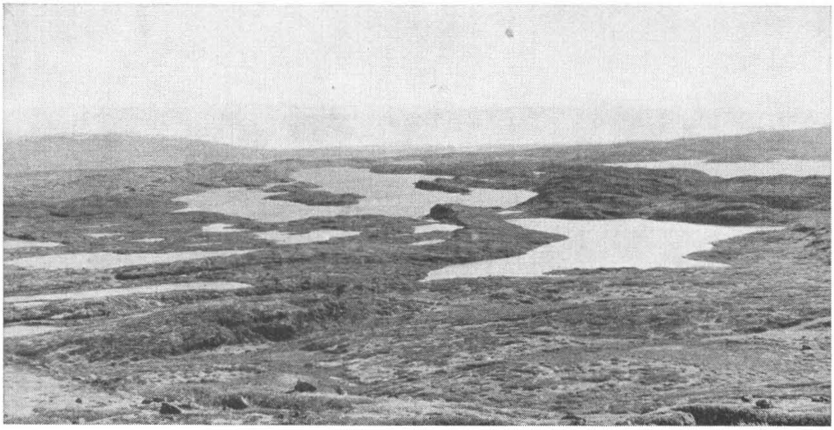

Fig. 33. View towards S 15° W over western part of dome (at left) and great lake west of it. Northwesternmost corner of dome is situated on eastern shore of nearest lake in right foreground.

ON THE STRUCTURE OF THE SURROUNDINGS OF THE DOME

The area north of the dome is structurally perhaps the most complicated part of the entire mapped region. The main structural features are clearly evident on the geological map and on the Plate II, i.e., a syncline, or, perhaps, synclinorium several kilometers wide opening toward the north. West of this syncline there is an anticline, likewise several kilometers wide, which plunges gently southeast. On the eastern side of the syncline the folding is more complicated. Inside the nose of the syncline, a couple of kilometers north of the northern border of the dome, the axis of the syncline appears to plunge roughly 40° toward the NW; but 1 to 2 km northwest the axes of small folds plunge gently SE. Some 5 to 10 km northward from the northern border of the dome, ridges formed mainly by amphibolite layers rise at an angle of about 5°—10° toward the north to altitudes of between 450 and 550 meters (Fig. 35). The morphology of the area and tectonic observations indicate here a fold axis plunging gently toward the south or southeast.

In the northwestern corner of the geological map (Locality 37, Plate I), three parallel arching amphibolite layers may be observed. They are situated on a fairly steep slope and a two-dimensional map is apt to give a misleading conception of their true form. The true mode of occurrence of these layers is best revealed by Fig. 34 where the northernmost layer is to be seen. The layers form part of the eastern flank of a great anticline plunging gently southeastward.

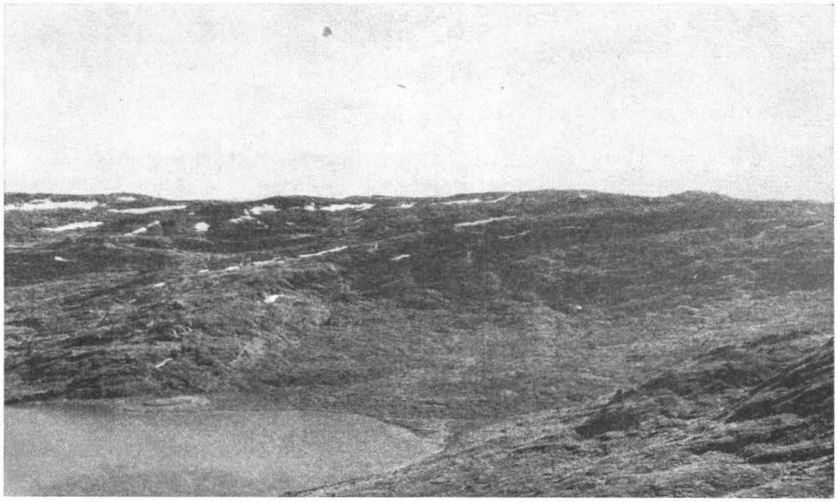

Fig. 34. Portion of eastern flank of anticline about 2 km wide. Northwest of dome. Locality 37 in Plate I. Fold axis plunges 20° to left or S 20° E. View towards west. Top of ridge is about 200 m higher than level of lake.

West of the southwestern part of the dome there is a gentle syncline, the width of which probably reaches a few hundred meters. It appears to be largely situated underneath the lake, and the reliability of the tectonic observations is not entirely free of doubts. In certain places it seems possible that the S-plane appearing in the field represents a foliation that does not run parallel to the bedding.

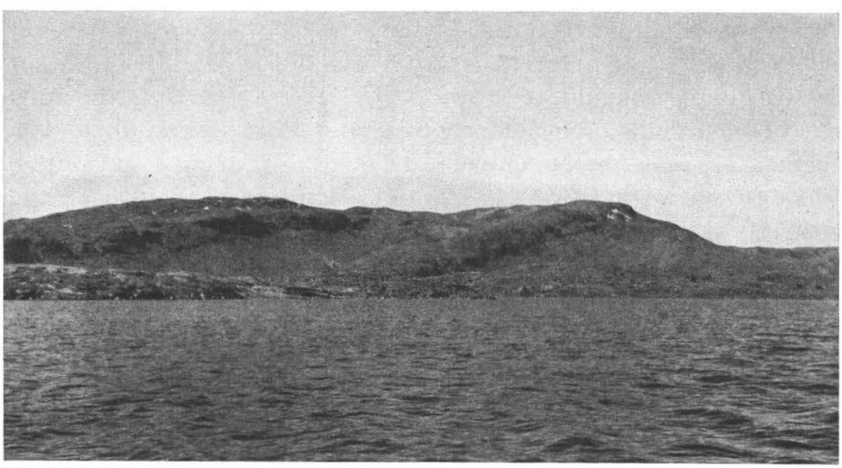

Fig. 35. Ridges formed by amphibolite zones rising gently towards north. Highest tops are situated about 8—10 km north of northern border of dome and rise 400—500 m above level of lake. View towards northwest.

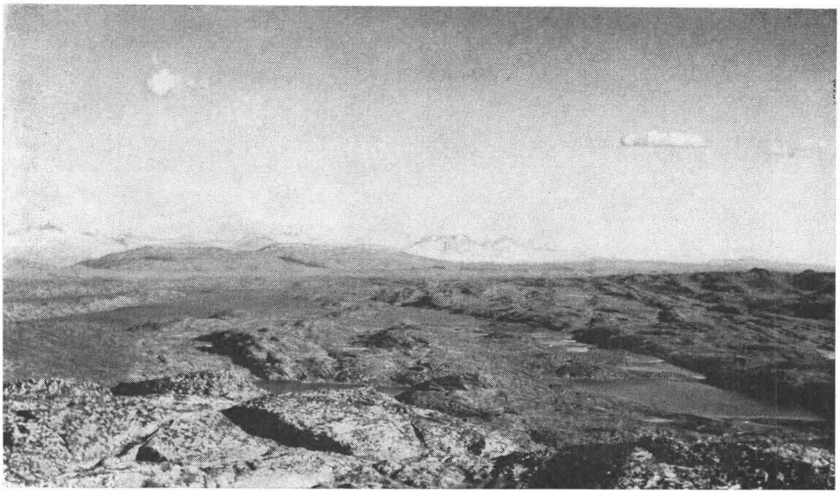

Fig. 36. View towards south over area west of dome.

Two kilometers west of the southern end of Usuk peninsula, an anticline several hundred meters wide has been recognized. Its locations is marked on the Plate I with the number 36. The fold axis plunges about 15°—20° north. The fold is almost isoclinal and its axial plane dips about 70°W. In addition, a few other probable folds have been found west of the dome. The field observations do not, however, suffice to draw reliable conclusions concerning their size, attitude of the fold axes and the axial planes, etc. The rocks are fairly homogeneous and the bedding and/or foliation generally hard to detect, in addition to which suitable marker beds are lacking. Furthermore, in certain places, fracture cleavage tends to hamper the investigation of these folds. The total structure of the area is best grasped from appended maps and Fig. 36.

The area southeast, east, and northeast of the dome appears to be rather simple in structure. It is characterized by numerous fairly parallel amphibolite layers with N—S strike and, for the most part, a 70°—85° W or vertical dip. These layers generally vary in thickness from 20 to 200 meters. Several of these layers run very nearly or quite parallel for a distance of as much as 20 km. In spite of the fact that they have been usually followed step by step along the strike of the layers, they have not been observed to converge anywhere; thus, at no place does the crest of the anticline or the trough of the syncline appear to intersect the surface of the ground. Accordingly, the regional fold axis ought to be nearly horizontal and to trend N—S. No evidence is available to show whether the amphibolite layers are stratigraphic beds situated on top of each other or whether isóclinal folding causes certain layers to recur at least in some cases.

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SUMMARY AND DISCUSSION

The area of the Ipernat dome with its surroundings, which is described in this paper, belongs to the extensive Precambrian bedrock region of western Greenland. Geologically it belongs to a pre-Nagssugtoqidian orogeny, which is probably the same as the Ketilidian orogeny in southern Greenland described by Wegmann (1938). The pre-Nagssugtoqidian rocks have been divided into several complexes. The area described here belongs to the Nordland complex.

By far the greatest part of the area not covered by lakes and ponds is exposed. This circumstance has made it possible to map the structure of the dome and its surroundings with unusual reliability, mainly by photogeological methods and by tracing suitable rock layers in the field step by step.

The main rock types of the mapped area are pyroxene amphibolites, amphibolites, hornblende gneisses and quartz dioritic to granodioritic rocks. In addition, there are occurrences of garnet-cordierite gneisses and ultrabasic rocks as well as pegmatite, aplite and diabase dikes.

At least for the greatest part, the rocks, excluding the dikes, crystallized under PT-conditions corresponding to the hornblende-granulite subfacies of Turner (1958). Subsequently, a large proportion of the rocks retrogressively metamorphosed under PT-conditions corresponding to amphibolite and/or epidote-amphibolite facies according to the facies classification of Eskola (1939). Nowhere has this retrogressive metamorphism been observed to have led to the creation of a new mineral assemblage completely in equilibrium.

The main rock types of the area all show a conformable mode of occurrence. Saksela (1953), who has described the tectonic occurrence of plutonic rocks in Finland, divides them into two classes: synorogenic and late-orogenic. The tectonic mode of occurrende of quartz dioritic and granodioritic rocks of the mapped area corresponds closely to the synorogenic group described by Saksela. Only the pegmatite, aplite and diabase dikes are discordant in their mode of occurrence.

The folding in the whole mapped area appears to be rather gentle. The breadth of a few of the anticlines and synclines is several kilometers. Noe-Nygaard and Berthelsen (1952) regard a gentle mode of folding to be a typical feature of rocks of the granulite facies. Similar observations have been made with respect to the area of highly metamorphic pyroxene-gneisses in southwestern Finland (Härme, 1954, p. 48; 1960, pp. 42—45 and 62). Wegmann (e.g., 1953, 1956) has greatly emphasized the different structural styles in the different structural levels (Stockwerke).

The Ipernat dome is some nine kilometers long and six wide. It consists of a quartz dioritic to granodioritic central gneiss and several concentric rings. The two most conspicuous rings consist in the main of amphibolite. The central gneiss is two kilometers long and 1.2 kilometers wide. The foliation to be seen in its marginal portions, the contact between the gneiss and inner amphibolite ring and the S-surface noticeable in the amphibolite — probably bedding — are all conformable. No apophyses intruding from the central gneiss into the surrounding amphibolite have been detected. The contact between the central gneiss and the amphibolite is, at least in most places, a shear surface. The bedding and/or foliation dip outward from the center of the dome, in most cases at angles of 10° to 45°. In the eastern part of the dome the dips vary from 45° to 90°. The dome is slightly tilted to the east. Evidence of the diapiric rise of the central gneiss has been sought but with very scanty results.

In the light of field observations, microscopical investigations and chemical analyses of twenty representative samples of quartz dioritic and granodioritic rocks taken from different parts of the area, the potassium seems to be slightly concentrated into the area between the inner and outer amphibolite ring, but not into the central gneiss.

Gneiss domes and their mode of origin have been described and discussed by numerous geologists from different countries. Some well-known examples of domes are the Vredefort dome in South Africa (e.g., Hall and Molengraaff, 1925) and the domes in Maryland and New Hampshire, U.S.A. (e.g., Cloos and Broedel 1940; Chapman 1939; Billings 1945). Wegmann (1930) has elucidated the problems of diapirism, which are closely related to the mode of origin of gneiss domes. Attention in the following will be mainly given the studies on the Precambrian domes in Finland and Greenland with which the present author is more familiar.

Very well known is Eskola's (1949) conception of so-called mantled gneiss domes, the origin of which he attributes to two successive orogenies. In his view, these domes apparently represent earlier synkinematic intrusions related to an orogenic period. The plutonic mass is later eroded and levelled, and thereafter follows a period of sedimentation. The domes are heaved up in connection with the granitization of the old pluton during a second orogenesis. Eskola has also made later contributions (1951, 1952) to the problem of mantled gneiss domes.

Saksela (1951, 1952, 1953) emphasizes that the Karelidic gneiss domes are situated at axial culminations and it is to this circumstance that he principally attributes their origin.

Preston (1954) has studied in detail the Karelidic gneiss domes of the Kuopio district in eastern Finland. His description fits well with the examples of the type at Pitkäranta that had been described and discussed by Trüstedt (1907), Eskola (op.cit.) and Saksela (op.cit.). Preston's conception of the origin of mantled gneiss domes comes quite close to that of Eskola's, except that the gneiss of the Kuopio domes is not wholly orthogneiss in origin, and he concludes that also other rocks than the old pluton have been mobilized and swollen by this same geological process. The chief reason for the swelling of the domes is, in his view, an expansion of volume brought on by a potash metasomatism.

A comparison between the Karelidic domes and the Ipernat dome brings out many differences. According to Eskola (1949, 1951) the lowest horizon of the mantle of Karelidic domes consists in some cases of basal conglomerates with boulders of the same gneiss that forms the dome; in others the basement stratum is a layer of quartzite, overlain by dolomite and mica schists; and in still others the dolomite or amphibolite forms the basement stratum. In the Ipernat dome the layer conformably lying on the central gneiss consists of amphibolite overlain by rocks of quartz dioritic to granodioritic composition and amphibolite layers. Furthermore, investigations so far made have brought to light no marked petrographic difference between the central gneiss and other homogeneous granodioritic and quartz dioritic rocks in the area of the Ipernat dome and its surroundings.

Härme (1954) has studied the Mustio cupola in southwestern Finland. No gneiss core similar to that which in Karelidic domes is supposed to be the basement of sedimentation is to be seen here. The lowest member detected in the stratigraphic sequence is a leptite interbedded with limestone and penetrated by microcline granite. Above this follow subsilicic volcanic rocks and mica schists. According to Härme the updoming of this cupola was caused by the upheaval of the microcline granite in a liquid state, the overlying strata having been pushed up.

The closest object of comparison to the Ipernat dome is the Tovqussaq dome described by Berthelsen (1950, 1960). It is situated roughly forty kilometers northwest of the area mapped in connection with the present study. The size, structure and petrography of these two domes are in broad outline rather similar, but the structure of the surroundings of the Tovqussaq dome seems to be more complicated. According to Berthelsen's descriptions

(1960, pp. 58—60) and my own observations, made one day in the summer of 1954, there are also rather marked petrographical differences between the core rocks (or central gneisses) of the two domes. The core rocks of the Tovqussaq dome contain nebulitic bands, schlieren and veins in noteworthy amounts. None at all have been found in the central gneiss of the Ipernat dome, which as a whole is highly homogeneous.

According to Berthelsen (1960, pp. 214—215), the formation of dome structures in gneissic terrains depends on three factors designated by him physico-chemical disharmony, tectonic disharmony, and true double folding.

The Ipernat dome forms a natural and well-fitting part of a rather gently folded complex of highly metamorphic rocks. In this view it could simply be described as a gentle doubly plunging anticline. Even in cases like the Jura mountains, however, the folding is a very complicated process, where many factors and mechanisms act simultaneously and in succession (e.g., Wegmann, 1962). A still more intricate process is the folding in connection with high-metamorphic complexes, where also intrusions as well as metamorphic and metasomatic reactions take place. In this light it is also difficult to estimate to what extent incipient diapiric movements, metasomatic processes and other such possible factors have contributed to the origin of this dome.

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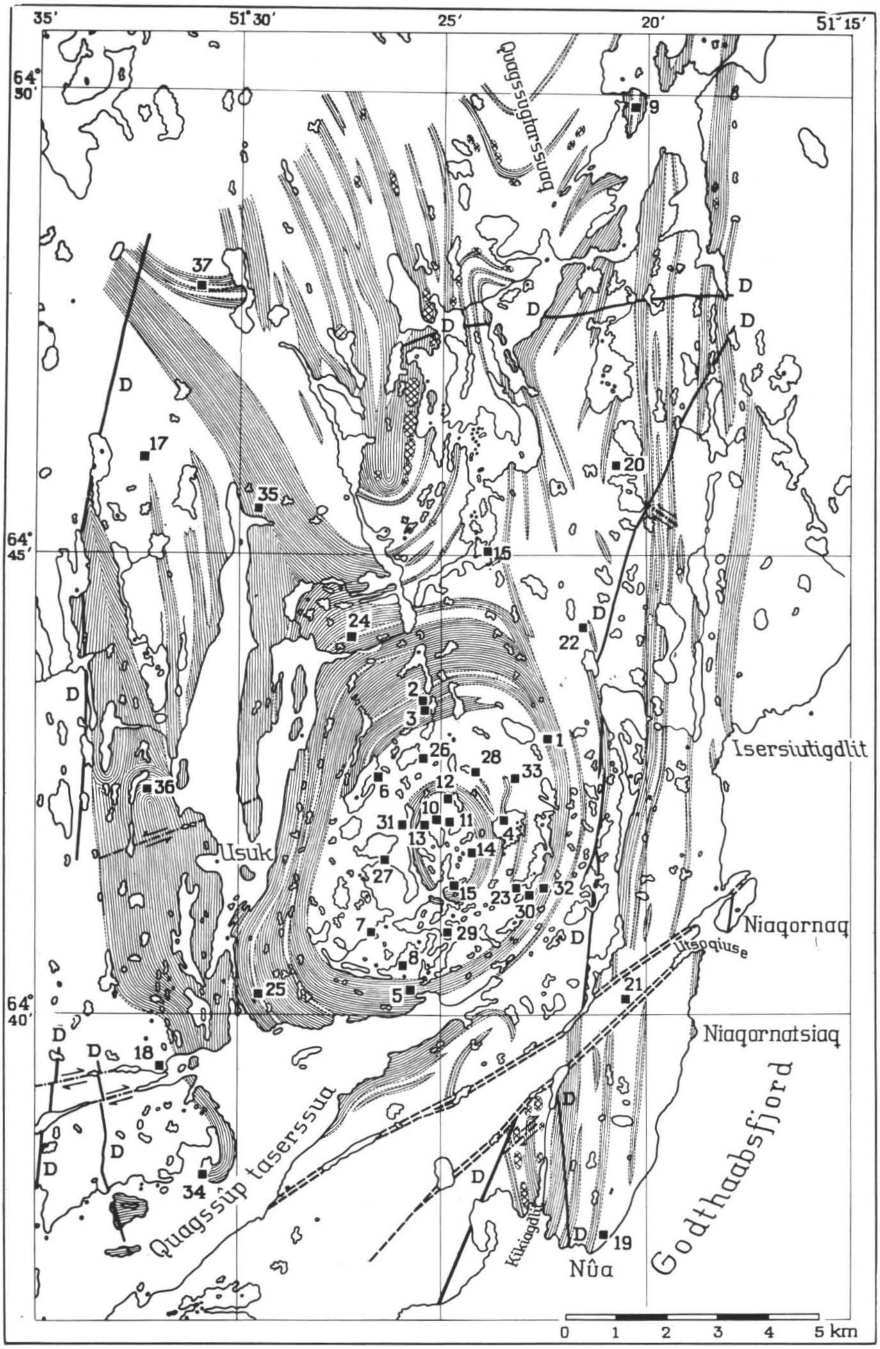
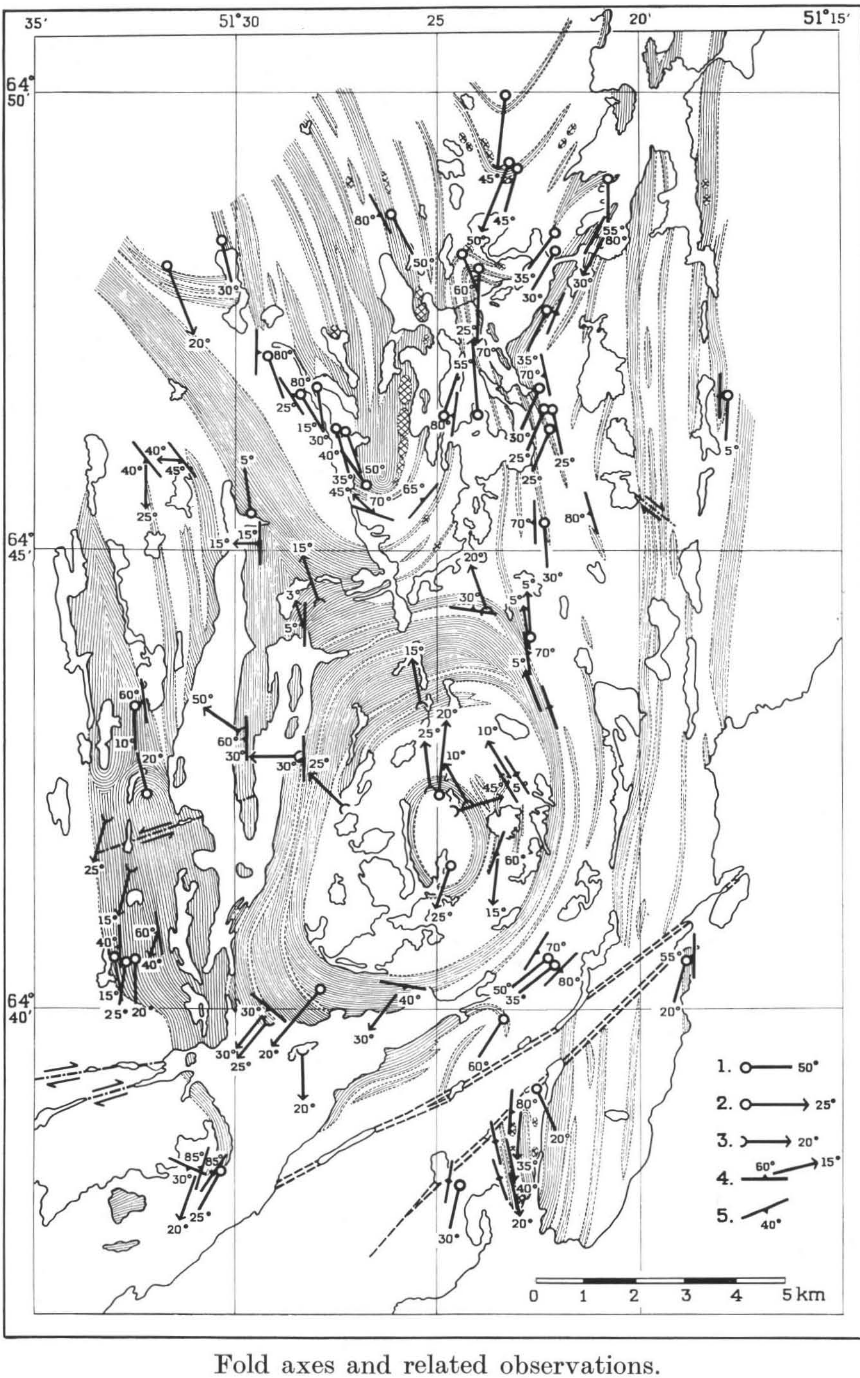
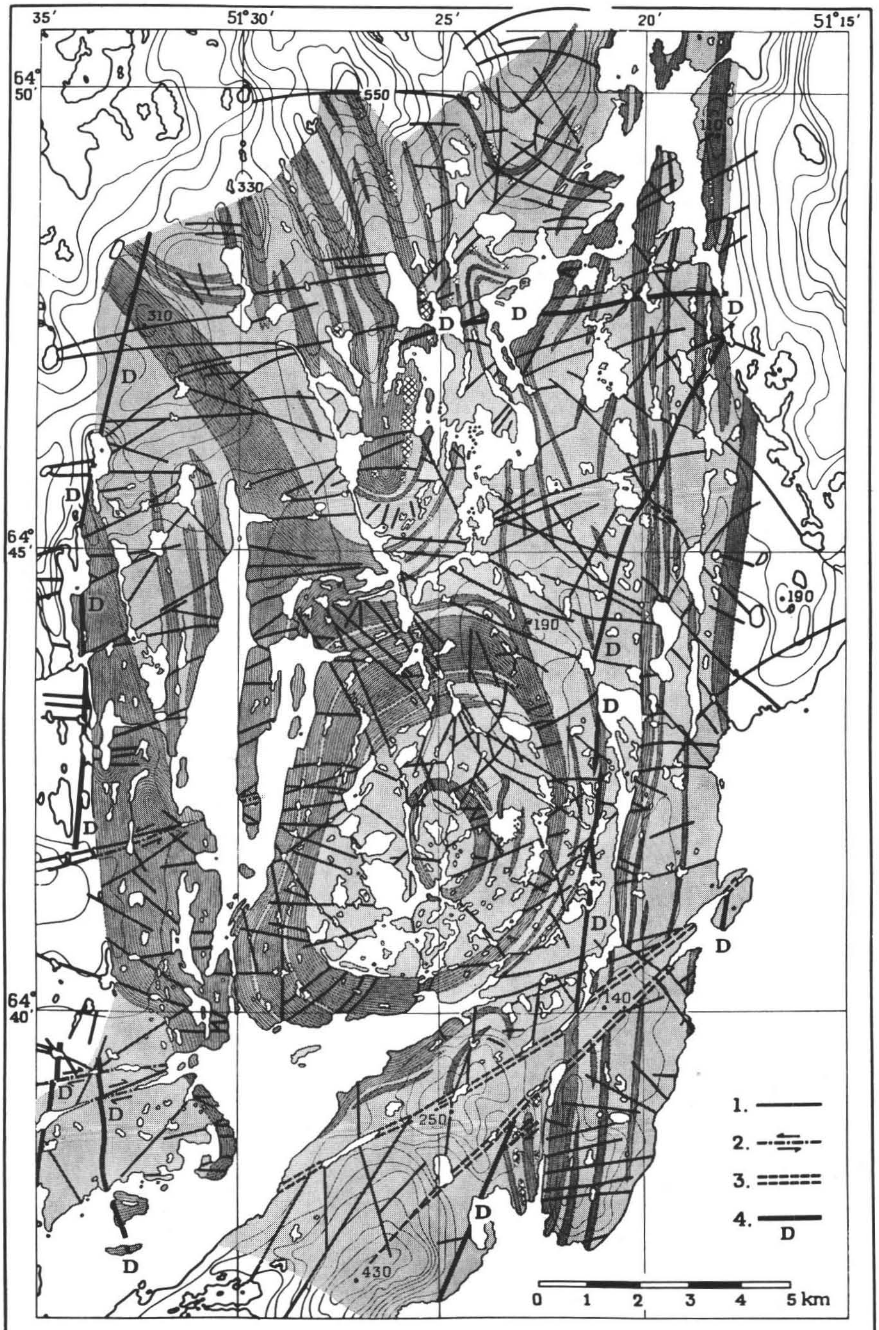


Index map of localities referred to in text.



Fold axes and related observations.

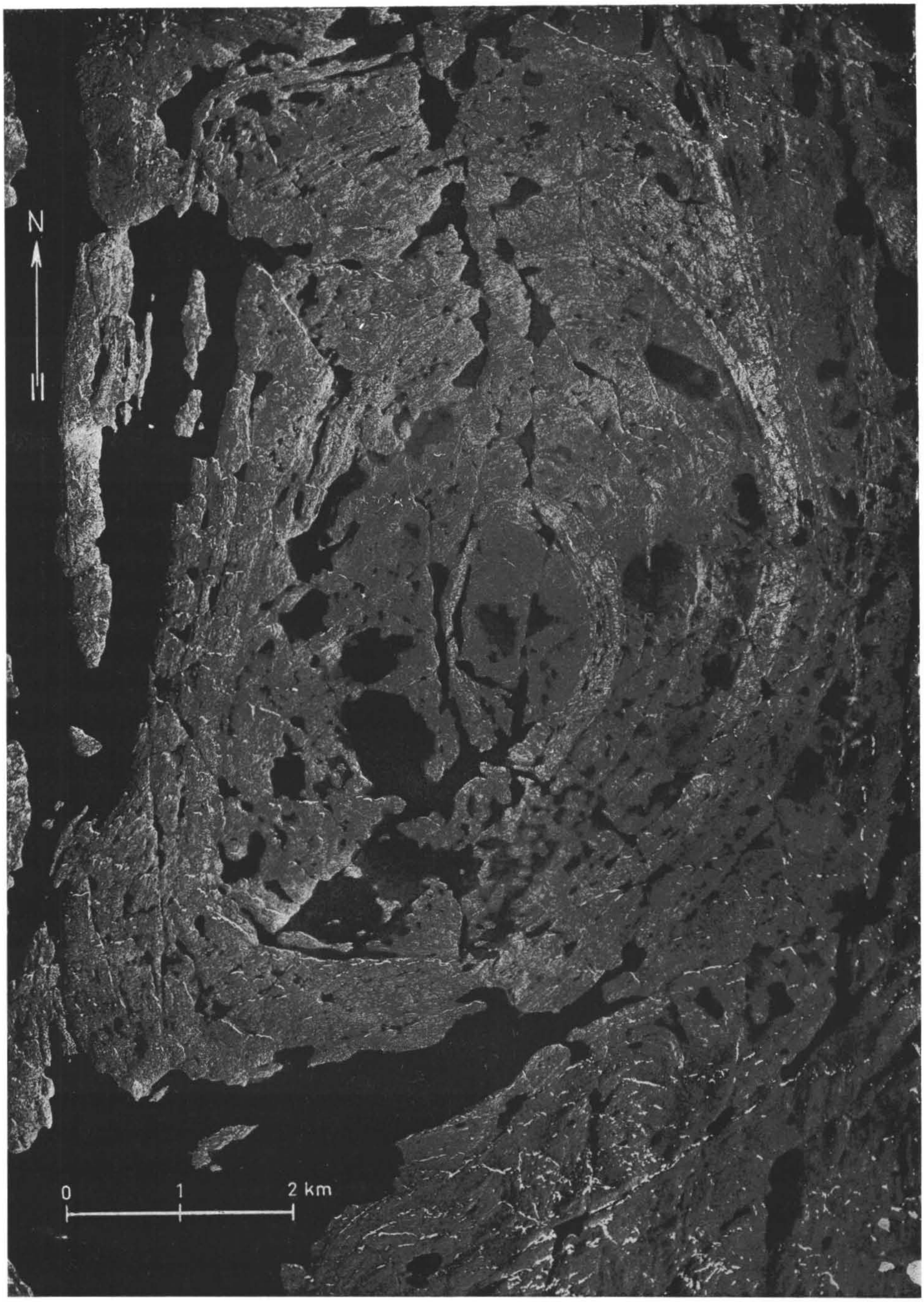
1. Fold axis measured on small fold.
2. Fold axis of large fold constructed by means of stereographic projection.
3. Fold axis measured on very gentle small fold.
4. Lineation and foliation. See text p. 72.
5. Bedding and/or foliation.



Joints, fractures and faults according to photogeological investigations. Joints parallel to strike of bedding and/or foliation are omitted.

1. Joint, fracture, or fault with very small displacement.
2. Fault with remarkable displacement.
3. Large fault zone.
4. Diabase dike.

Heights in meters. Contour interval 50 m. Base map reproduced by permission of Geodetic Institute, Copenhagen.

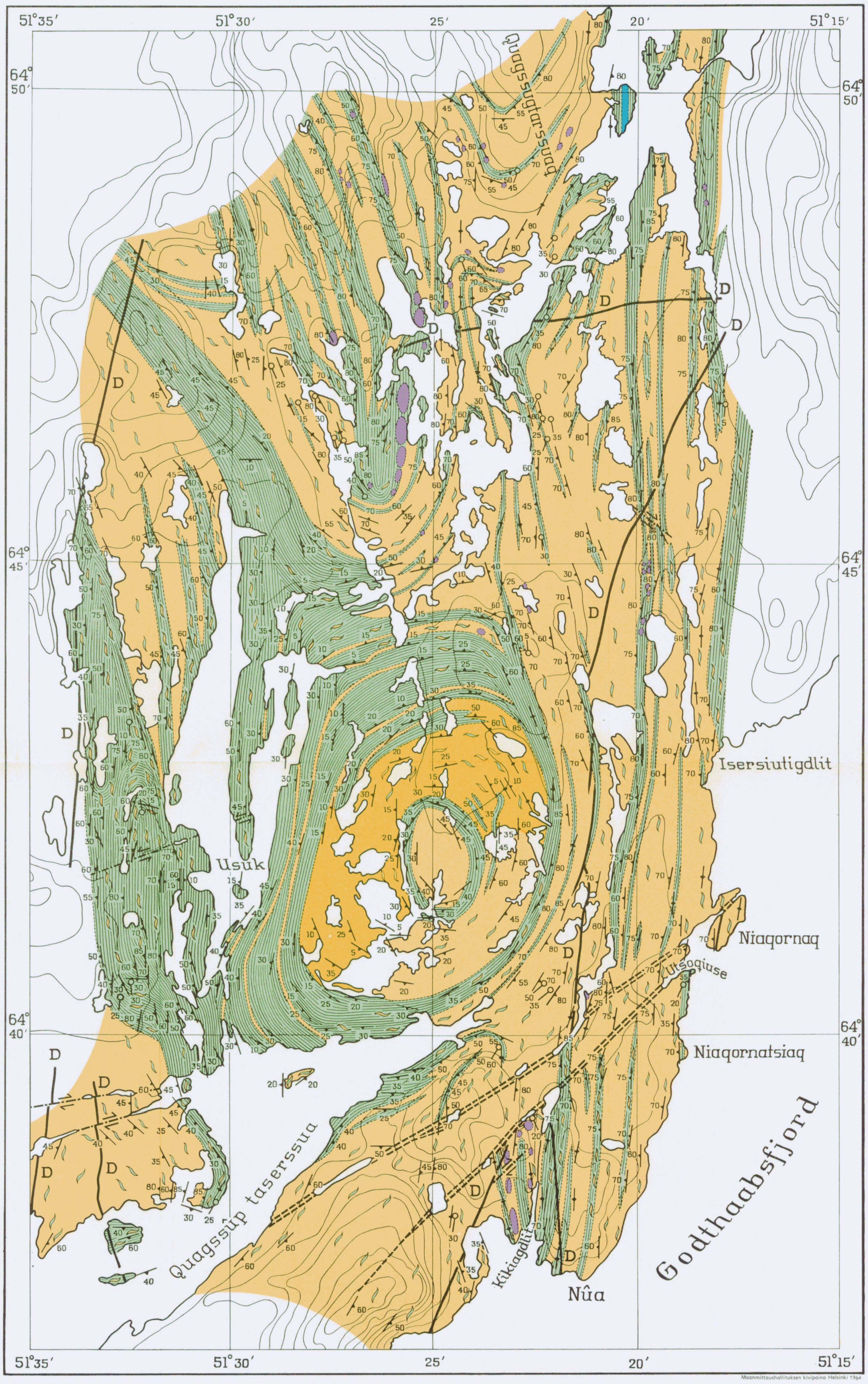


Aerial photographs A 31 B/69 and A 31 B/71 showing Ipernat dome in summer.

Uncontrolled mosaic of aerial photographs taken in spring. Same area as in geological map (Plate VI). Tones of some ice-covered lakes are slightly retouched.

GEOLOGICAL MAP OF IPERNAT DOME

by
RAIMO LAUERMA
1964



LEGEND

- Pyroxene amphibolites, amphibolites, and hornblende gneisses
- Hypersthene gneiss
- Quartz dioritic and granodioritic rocks
- Garnet-cordierite gneiss
- Ultrabasic rocks
- Quartz dioritic and granodioritic veins and schlieren
- Amphibolitic inclusions and schlieren
- D Diabase dike
- 50 Strike and dip of foliation, banding, and/or bedding,
- 20 Fold axis
- Fault
- Large fault zone

Scale 1:75 000

0 1 2 3 4 5 km

Contour interval 50 meters

Based on topographic map, scale 1:50 000
Reproduced by permission of Geodetic Institute, Copenhagen

2

OBSERVATIONS ON PREHNITE AND ITS OCCURRENCE IN KANGASNIEMI PARISH, FINLAND [1]

BY

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ABSTRACT

The author describes the occurrence of prehnite in two different places in Kangasniemi parish: chlorite bearing schist, rich in plagioclase and pyroxene gabbro. On the prehnite occurring in the first mentioned rocks, the author gives the chemical composition, physical properties and x-ray data.

CONTENTS

INTRODUCTION

Kangasniemi parish circles the northern shore of Lake Puulavesi in the province of Mikkeli, about 220 km to NNE from Helsinki as the crow flies. In the summer of 1954 the writer conducted geological researches there.

[1]) Received January 2, 1964.

About 2,5 km northwards from the main village of Kangasniemi Parish, a road to Mikkeli town was being built and in the roadside ditch there was a shallow, recently blasted rock section of a few meters width. In this section the schist was brecciated with prehnite veins. In another rock, about 1 km to the north from the spot mentioned above, prehnite also occurred in narrow veins in the metamorphic pyroxene gabbro.

Previously Laitakari (1914) has described the prehnite he found in the town of Helsinki. He maintains that this was the first prehnite finding in Finland. This prehnite (Laitakari, 1914, p. 432) was found brecciated on amphibolite together with a quartz pegmatite vein. Prehnite has also been found in Finland, among other places, near the town of Hämeenlinna, in the parish of Kalvola and in connection with limestone in Parainen. Another study on prehnite has been published in Finland (Eskola, 1934) describing boulders of prehnite amygdaloid, in which the origin and conditions of genesis of prehnite are discussed.

CHLORITE-BEARING SCHIST RICH IN PLAGIOCLASE AND THE OCCURRENCE OF PREHNITE IN THIS SCHIST

The wall rock of the prehnite vein in the finding place near the main village of Kangasniemi is a schist of fine texture (size of grains abt. 0,3—0.02 mm) and of a greenish, black-grey colour. It is chlorite-bearing and rich in plagioclase. The rock is strongly metamorphic and its structure is granoblastic. Its mineral grains are distinctly of two different sizes. The small grains are inclusions in the big ones, and in their interstices. All the minerals are usually xenomorphic, and the chloritized biotite and chlorite usually appear as grains of a rather irregular shape. The main constituents of the rock are plagioclase, quartz, potash feldspar, biotite and chlorite. There are also small quantities of epidote, sericite, apatite, titanite, zirkon and opaque minerals.

Plagioclase ($\gamma > 1{,}547$, $\alpha < 1{,}531$; An_{5-25}) occurs abundantly as dull, xenomorphic grains. There is not nearly so much microcline and this, too, appears in dull grains. In the microcline grains the »tartan twinning» is not visible, but plagioclase is there visible as perthite. There is not much quartz, either, and the grains show distinct undulating extinction, especially the big ones.

The biotite ($\gamma = \beta = 1{,}634$ = green, α = light green) is chloritized, and its interference colour is decreased. The rock also contains green chlorite ($\gamma = 1{,}634$, $\alpha' < 1{,}634$, interferences colour violet blue), which cannot be recognized as a product of the alteration of biotite. In connection with the chloritizing biotite there are here and there small epidote and titanite grains.

This schist is intersected by light grey pegmatite veins, rich in quartz, of varying thickness. Here and there in these veins there are small segregations of red-brown microcline ($\gamma = 1.525$, $\alpha = 1.519$) which is beautifully tartan twinned and not dull. The vein also contains some clear, albitic plagioclase and greenish-brown biotite ($\gamma = \beta = 1.637$). The quartz grains of these veins mostly show a very pronounced undulating extinction, they are stretched. At least in some cases the quartz grains which they contain are stretched almost perpendicularly to the longitudinal direction of the vein and their ε' is parallel to the elongation of the grains.

These pegmatite veins and the schist near the veins are brecciated by prehnite veins, about 6—0.2 mm thickness, of a very lightly greenish white colour. The veins branch off into thin apophyses, both in the pegmatite and in the schist (Fig. 1, Table 1). The prehnite veins contain both schist and pegmatite in small fragments. The fragments are partly very angular, partly eye-shaped. The outcrop also shows slickensides which at least partly intersect the prehnite veins, and often have a thin layer of pyrite on the surface. Here and there the slickensides part the prehnite veins into wedge-shaped pieces, and in a microscope it is possible to see that they are slightly displaced with regard to each other. Pyrite sometimes also occurs at the contacts between the prehnite veins and the schist forming the wall rock of these veins. The last-mentioned occurence of the pyrite is evidently often slightly older than one appearing on the slickensides.

The structure of the prehnite veins varies as follows:

1) Small (0.06—0.2 mm in size) prehnite grains form the ground mass of the veins, which here and there contain prehnite, the diameter of which is 1—0.5 mm.
2) Prehnite grains of 0,3—1,5 mm size alone form veins
3) The slickensides show prehnite veins of 0.05—0.3 mm width, where the size of the prehnite grains is 0.01—0.2 mm.
4) Tortuous very thin veins of varying thickness consisting of considerably ragged prehnite grains.

The prehnite grains of the types mentioned under item 1 and 2 often distinctly show undulating extinction in the manner of quartz. In the same vein sometimes occur alternately both type 1 and type 2 in longitudinal direction. These types of veins (1 and 2) are also zoned in certain spots. In such cases there is usually on the edge of the vein a very fine-grained zone of 0.06—0.2 mm width. Usually this zone is not continued in an unbroken succession through the whole vein but is interrupted in some places to start again elsewhere. The finely grained edge zone is generally strictly confined to the centre. It is either sharply limited against the wallrock or else it transits this while the quantity of the prehnite grains decreases, to disappear completely on a very short distance. As there is no border zone of fine texture, the big prehnite grains adjoining by the wallrock have a torn edge against the wallrock minerals. They have penetrated between the mineral grains and even into the grains.

Sometimes the veins (type 1) contain relics of the wall rock minerals, among other plagioclase and quartz which are visible in the middle of the vein. In his study Eskola (1934, p. 8) describes pseudomorphic plagioclase twinning in a phenocryst now consisting of prehnite with grains of quartz.

The veins of group 3 cut both quartz pegmatite veins and prehnite veins (groups 1 and 2), very often almost vertically across the veins. The most common direction of the veins (group 3) is almost identical with the direction of the schistosity. The veins of group 3 pierce the prehnite veins in the direction of the vein as well. The same vein (types 1 and 2) can contain up to three veins (type 3) alongside of each other in the longitudinal direction of the vein.

The veins of group 4 usually are between two prehnite veins which are at a distance of 0,5—3 mm from each other, joining these together. They join veins both of groups 1 and 2 and of group 3.

In pegmatite and prehnite veins apatite, magnetite, chalcopyrite, ilmenite and hematite have been found as accessory minerals. Apatite occurs in big xenomorphic grains in the quartz pegmatite veins and in the fissures of its grains there is often prehnite.

CHEMISTRY, OPTICAL AND PHYSICAL PROPERTIES AND X-RAY POWDER PATTERN OF THE PREHNITE IN THE SCHIST

The prehnite in the schist was separated using the Clerici solution in a zentrifuge and a Franz type isodynamic separator.

Table 1 shows the chemical composition of the separated prehnite, its optical properties and density. In accordance with Winchell (1951, p. 359) and Deer, Howie and Zussmann (1963, p. 262) the variation in the composition of the prehnite is insignificant. It is mainly only iron which replaces aluminium to some extent, whereas the contents of alkalies, manganese and magnesium are constantly low. The composition and properties of the prehnite in Table 1 equal in the first place those of the prehnite of New South Wales (Coombs et al., 1959, p. 102): SiO_2 43,7; Al_2O_3 24,05; Fe_2O_3 0,93; FeO 0,03; MgO 0,11; CaO 26,85; Na_2O 0.04; K_2O n. f.; H_2O+ 4,54; H_2O- 0,03; Ag 0.01; Pb 0,04; Sn 0,01; BaO $<$ 0,003; Total 100,36; $\alpha =$ 1,615; $\beta =$ 1,624; $\gamma =$ 1,643; $2V\gamma = 69°$.

Table 2 shows the indexed x-ray powder pattern of the analyzed prehnite. The lattice dimensions will be as follows: $a_0 = 4{,}62$ Å, $b_0 = 5{,}47$ Å, $c_0 = 18{,}47$ Å. The structure of prehnite has been elucidated by Gossner and Mussgnug (1931), Berman (1937), Nuffield (1943) and Malčić and Preisinger

Table 1. Chemical composition and physical properties of prehnite. Kangasniemi, Finland. Analyst, A. Heikkinen.

	Weight per cent				
SiO_2	43.75	Si	6.01	Z = 8.00	Si = 6.01; Al = 1.99
TiO_2	0.14	Ti	0.02		
Al_2O_3	24.10	Al	3.90	Y = 2.03	
Fe_2O_3	0.96	Fe^{3+}	0.10		
FeO	0.14	Fe^{2+}	0.02		
MnO	0.04	Mn	0.00		
MgO	0.10	Mg	0.02	W = 3.98	
CaO	26.45	Ca	3.90		
Na_2O	0.06	Na	0.02		
K_2O	0.14	K	0.02		
P_2O_5	0.01				
CO_2	0.00				
H_2O+	4.35	(OH)	3.98	OH = 3.98	
H_2O-	0.03				
	Σ 100.27				

Physical properties:

α = 1.616
β = 1.626
γ = 1.646
2 V γ = 66°
γ—α = 0.029 (with Berek compensator)
D = 2.93

Table 2. X-ray powder diffraction data of prehnite. Kangasniemi, Finland. (Copper K α radiation, Nickel filter.)

No.	I/I_0	2θ	d Å	hkl
1	22	16.88	5.25	011
2	29	19.21	4.62	100, 004
3	43	25.21	3.55	110
4	90	25.72	3.46	111
5	50	27.04	3.29	112
6	59	27.31	3.26	104
7	100	29.12	3.06	015
8	38	31.92	2.801	114
9	19	34.12	2.626	022
10	88	35.12	2.553	106, 115
11	43	38.19	2.355	120
12	26	38.51	2.336	121
13	34	38.91	2.313	200, 116
14	25	43.82	2.064	204
15	27	47.03	1.931	214
16	17	49.36	1.845	0010, 215
17	31	51.60	1.770	220
18	21	55.42	1.657	217
19	16	56.20	1.635	133
20	23	60.13	1.538	135, 300
21	12	64.52	1.443	313

(1960). It is orthorhombic pyramidal according to Deer et al. (1959), its space group is P2cm and the dimensions of its unit cell are $a_0 = 4{,}61$ Å, $b_0 = 5{,}47$ Å, $c_0 = 18{,}48$ Å. Malčić and Preisinger (1960) have obtained the result $a_0 = 4{,}63$ Å, $b_0 = 5{,}49$ Å and $c_0 = 18{,}48$ Å. They carried out the Patterson and Fourier synthesis of prehnite and give the following structure schema of prehnite: $^2_\infty Ca_2^{[7]}\,Al\,(OH)_2\,[Al^{[6]}\,Si_3^{[4]}\,O_{10}]$ r

PREHNITE AND ITS OCCURRENCE IN PYROXENE GABBRO

The pyroxene gabbro body where prehnite occurs is about 4 km^2 large. Prehnite is found in an almost NS directed, nearly vertical shear zone which is situated on the SE edge of the body. The gabbro is medium-grained, and its structure is ophitic. Along the edges of the body it is much more finely grained than in its central parts. The main minerals in the gabbro are plagioclase (An_{60}), augite and hypersthene. Accessory minerals contained in the gabbro are magnetite, ilmenite, apatite and small quantities of biotite, potash feldspar and quartz. There are comparatively small quantities of secondary minerals.

In the above mentioned shear zone the gabbro is highly metamorphic. Its structure has become blastophitic. There is no orthorhombic pyroxene to speak of and augite occurs as a relic only. The composition of plagioclase is more albitic ($An_{30}-_{40}$) than in the non-metamorphic gabbro. It contains considerable quantities of secondary minerals: epidote, chlorite, amphibole, sericite and titanite. Apatite occurs in big xenomorphic grains. This rock also contains remarkable quantities of quartz and potash feldspar.

In the ground gabbro described above, slightly greenish white prehnite occurs in 0,01—1 mm wide veins. There are also quartz veins in this rock. Both veins follow the direction of the fault approximately.

For the prehnite occurring in these veins, the following properties have been determined: $2V\gamma = 65°$, $\gamma - \alpha = 0{,}021$ and $\gamma - \beta = 0{,}015$ (both with Berek compensator), $\alpha = 1{,}611$, $\beta = 1{,}617$, $\gamma = 1{,}632$. According to Tröger (1956) the properties determined are adaptable to an Al-prehnite. This prehnite is evidently poorer in Fe^{+++} than the prehnite described above, which was found in the schist.

DISCUSSION

Prehnite is often found in the capacity of a hydrothermal mineral in cavities and in amygdules of effusive rocks and in contact deposits of igneous rocks. All of these often contain zeolite minerals in addition to the prehnite. It has also been found in veins in many rocks of different composition and as pseudomorphs after various minerals. It has also been encountered in comparatively acid rocks containing but little calcium (*e. g.*, Hjelmquist, 1937, and Coombs et al., 1959). In these papers the host rock of the prehnite described at the beginning is comparatively poor in calcium. According to Coombs et al. (1959, p. 66) prehnite occurs in veins, either alone or together with quartz, and the prehnite veins may in addition contain some of the minerals calcite, sericite, albite and chlorite.

According to Brothers (1956, p. 478) and Turner and Verhoogen (1960, p. 532) prehnite seems to be a characteristic mineral in rocks of the zeolitic facies elsewhere. Coombs et al. (1959, p. 92) consider that the prehnite subfacies, transitional between the zeolitic facies and greenschist facies may come in the temperature around 300° C ($\pm$ 50° C) in regions where water and load pressure are approximately equal and at lower temperatures if osmotic conditions occur. Eskola (1934, p. 11) has found that in the case which he has studied the epidote seems to be rather earlier than the prehnite, or both may be contemporaneous, and that the zeolite minerals, on the other hand, generally succeed the prehnite. Fyfe et al. (1958, p. 170) says, that the prehnite may also form by alteration of feldspar, and probably represents metamorphism of even lower grade than do the epidote minerals. According to Eskola (1939, p. 383) prehnite generally occurs in quartz veins in postcrystalline-deformed tectonites. In accordance with Coombs et al. (1959, p. 64) prehnite occurs as coatings and veins in crush zones and it is considered to be a product of shearing. Prehnite found in gabbro has the same manner of occurrence, which is described in this paper. Prehnite described in the schist also occurs rather abundantly in deformed, tectonized rocks.

Crush-zone laumontite associated with prehnite has been described (Coombs, 1952, p. 815; Coombs et al., 1959, p. 63). Coombs (1952, p. 815) describes among others a crush-zone in albitic metagabbro and writes that a sheet of pulverized and slickensided rock several inches thick has been almost completely replaced by laumontite associated with a little prehnite and chloritic matter and containing broken fragments of augite, albitized feldspar and iron ore from the country. Coombs et al. (1959, p. 66) describes how greywacke is traversed by numerous quartz-prehnite veins which are cut by later fractures cemented by laumontite or lined with dusty analcime, and by still later joints filled with stilbite. In Finland laumontite has been found in many places and often from crush-zones (Eskola, 1935, 1960; Pipping, 1961). Pipping (1961) maintains that prehnite occurs in connection with laumontite in Viitasaari, even if it only occurs as an accessory mineral. In New Zealand these minerals are found in large areas (Coombs et al, 1959). Perhaps in Finland, as well, they will yet be found together in many places and in crush-zones.

Acknowledgement — The author wishes to express his indebtedness to the University of Helsinki on whose staff he was allowed to work and whose research equipment he has been at liberty to use. At the same time the author wishes to thank very sincerely the State Board of Natural Sciences (Valtion luonnontieteellinen toimikunta) for the financial support he has received.

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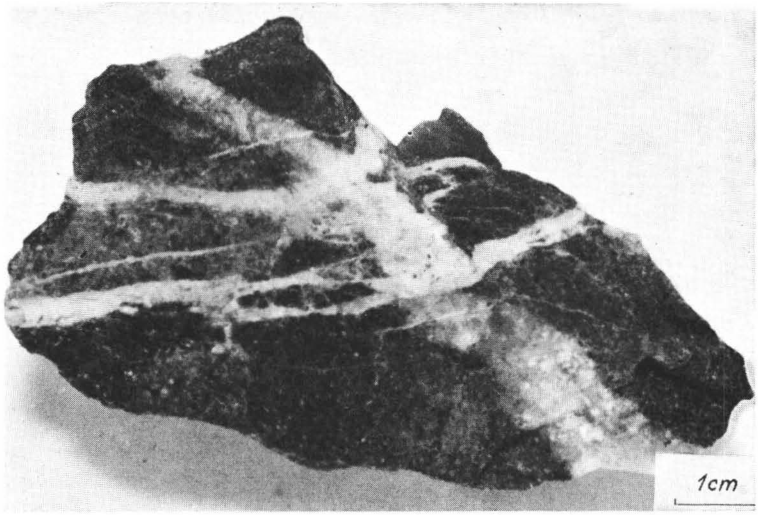

Fig. 1. Chlorite-bearing, plagioclase-rich schist, containing quartz veins (light grey). Both schist and veins are cut by prehnite veins (white) Kangasniemi, Finland.

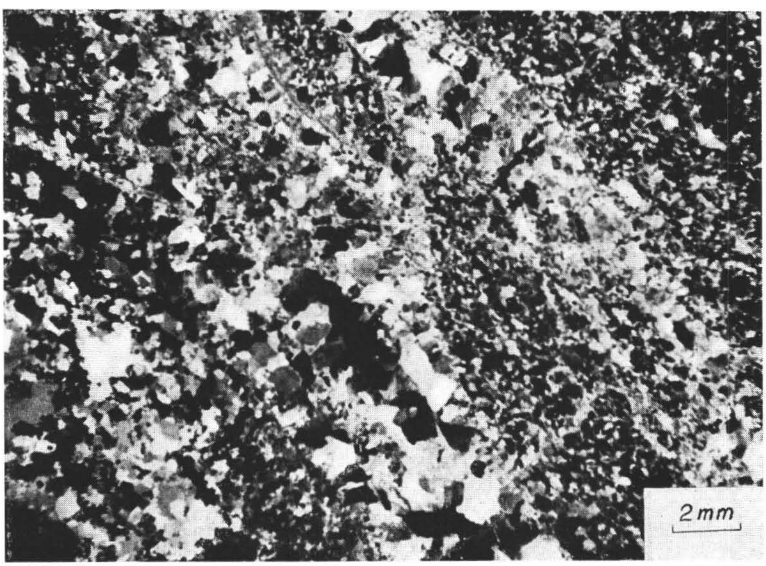

Fig. 2. Prehnite veins (light, coarsely grained spots) in the schist. Both are intersected by narrow finely grained prehnite veins (also light). + nic. Kangasniemi, Finland.

3

OLIVINE DIABASE DIKE OF ANSIO IN PADASJOKI, FINLAND [1]

BY

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ABSTRACT

The author describes an olivine diabase dike and gives a petrographic exposition of it. The rock has been chemically analyzed and its modal composition has been stated. The main minerals of this rock are plagioclase, olivine and augite. A chemical analysis of the augite is given as well as x-ray diffraction powder data. In addition the age of the olivine diabases is discussed and a description of the shearing zones occurring in this diabase dike is given.

CONTENTS

INTRODUCTION

Padasjoki parish is situated in the province of Häme on the western shore of the lake of Päijänne, about 150 km to the north of Helsinki. The diabase

[1]) Received, January 23, 1964

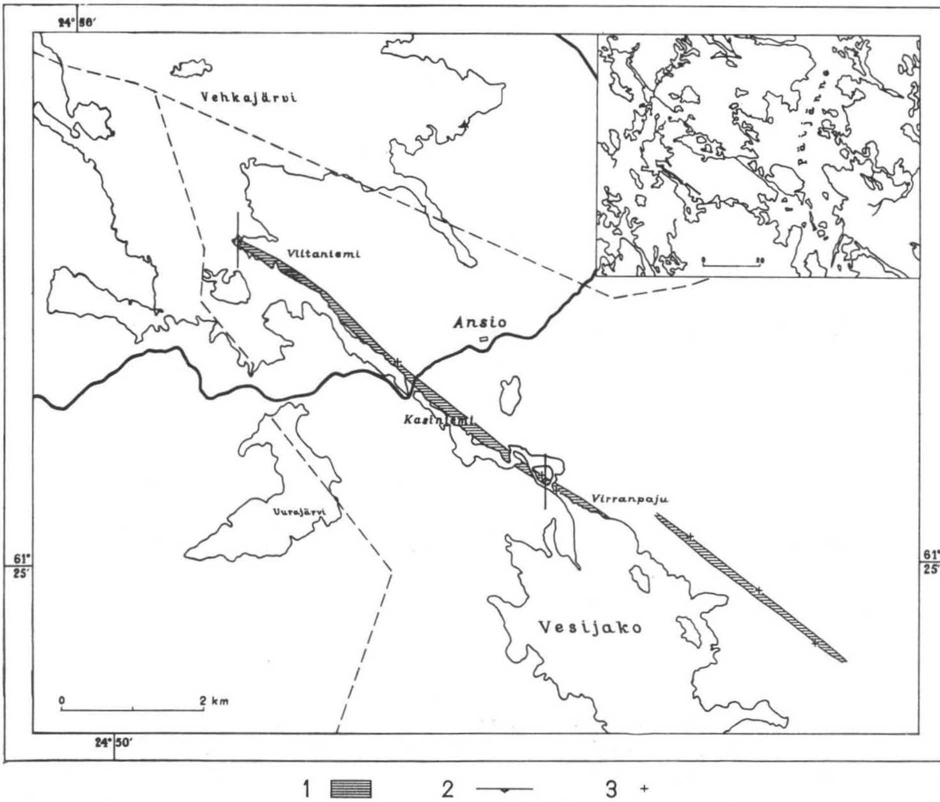

Fig. 1. The map shows the olivine diabase dike of Ansio (1), some shear zones (2) and the contact exposures (3). In the right hand corner is a map of the southern part of Lake Päijänne and the tract west of it. The fracture lines are visible as the course of waterways. The diabase dike is also shown on the small map as a line.

dike of Ansio is situated in the NW-corner of the parish between lakes Vehkajärvi and Vesijako. The dike begins in Viitaniemi village, on the shore of Vehkajärvi and runs along the NE-shore of the river which connects the said lakes; later it is either the same or perhaps even a completely different dike which runs on the EN side of Vesijako at a small distance from the shore of this lake. This diabase dike does not appear any more on the NW-shore of Vehkajärvi, and a esker formation begins at its SE-end. (Fig. 1).

The general strike of foliation of the bedrock is approximately from the west to the east. The direction of the diabase dike is N 60°W. The last mentioned trend is at the same time the direction of the fracture valleys and fault lines, which is quite common in the south of Finland and in central

Finland (Sederholm, 1911, 1912, 1913, 1932; Tuominen, 1957, Eskola, 1960, Härme, 1961). This diabase dike is comparatively well exposed. Small, shallow exposures run successively, at small distances, as already reported by Frosterus (1902). Contact between the surrounding country rocks and the dike has been found in several places, and there is a number of exposures of the surrounding rock near the dike. It is possible to form a clear conception of the thickness of this dike, even where the wall rock is not near, nor the contact of the dike visible, for in diabase exposures the rock often shows a gradual transition into fine-grained texture, exactly in the same manner as in the diabase contacts. Thus it has been possible to determine rather accurately the direction of the dike, its width and the place of its contacts.

To our knowledge, the olivine diabase dike of Ansio appears for the first time on a geologic map in a publication of Sederholm (1893). *Cf.* also maps (Sederholm, 1897 and 1930). On this map it is shown in the same colour as the olivine diabase of Satakunta. In this publication (Sederholm, 1893, p. 105) there is also a short petrographic description which shows the rock to be olivine diabase. In the same connection the author says that similar dikes are also found in Orivesi. Later on the diabase dikes of Orivesi have been examined once more and it was found that their rocks are entirely similar to the Jotnian olivine diabase in Satakunta. They are consequently considered to be of the same age as the olivine diabase dikes of Satakunta (Seitsaari, 1951, p. 88).

According to Frosterus (1902, p. 74) the diabase dike of Ansio is augite -bearing amphibole gabbro, and it is older than the microcline granites which occur in this district. Frosterus (1902, p. 32, 74) and Väyrynen (1954, p. 97) count it among the gabbros of Keuruu intersected by microcline granites.

The olivine diabase of Satakunta is the youngest member of the Jotnian complex, to which, in addition, the rapakivi granites and sandstone also belong (Sederholm, 1911 a, p. 107; Laitakari, 1925, p. 17; Eskola, 1941, p. 461; Kahma, 1951, p. 18; Simonen, 1960, p. 44). The K^{40}—A^{40} age of the unmetamorphosed Jotnian sandstone (Simonen, 1960, p. 44 and 1962, p. 259) is 1300 million years. The maximum K^{40}—A^{40} age of the Satakunta olivine diabase has been estimated at 970 million years[1].) These calculations have been made on the basis of two specimens. The first was taken from the olivine diabase of Sorkka and the second from that of Säppi. Corresponding ages are known even in other places in Fennoscandia (Neumann, 1960, p. 179). According to Magnusson (1960, p. 140) the absolute age determinations of Polkanov have given 964 million years for the Jotnian Almesåkra series and 985 million years for the Visingö series. Magnusson (1960, p. 140) himself believes that

[1]) The determinations have been made at the Institute of Physics in Freiburg, whose principal is Prof. W. Gegnert, by Kley, Kiessling, Schmiedling and Zähringer, Ph. Drs.

Jotnian and Eocambrian must be much younger, and that the age of the Dalslandian is 964 million years.

According to the observations of the writer, the olivine diabase intersects its wallrock of tourmaline-bearing microcline granite among other places at a point to the north of Kasiniemi. As a matter of fact, narrow veins of granite composition are found on the edge of the diabase dike, but they can be explained as the product of partial remelting of the contry rock, induced by diabase intrusion, as Kahma (1951) has proved with regard to the Satakunta olivine diabase. If it is permissible to guess at the age of the Ansio olivine diabase on the basis of petrographic facts, it must be coordinated with the olivine diabase of Satakunta. Indeed the age interval between the microcline granite and the olivine diabase of Satakunta is long, about 700 million years, and during this period of time diabase dikes may have erupted at other times as well.

DESCRIPTION OF OLIVINE DIABASE FROM ANSIO

The Ansio olivine diabase is a hypabyssal dike, which has distinct chilled contacts. At the borders of the dike the rock is much more fine grained than in the middle of it. The marginal variety (Fig. 1, Plate 1) which contains some stray grains of plagioclase phenocryst shows a gradual transition into the main rock type of this dike (Fig. 2, Plate 1). Winkler and Weitz (1956) have made a detailed study of the crystallisation of the Hardeberga basalt dike and of the variations of the grain size there.

The main type of diabase is a rock of bluish dark-grey colour on its fresh surface. On slightly weathering surfaces of the rock the plagioclase laths are light grey and the femic minerals are either brownish gray or black. The ophitic texture of the rock is beautifully exposed on these surfaces. In some places the plagioclase laths show ophitic texture with subparallel arrangement. The plagioclase laths are of the same type as in the main rock they are about 10—20 mm long and 2—8 mm thick.

The main constituents of the olivine diabase are plagioclase (An_{65}), olivine (Fa_{44}) and augite (about 35 per cent Fe- component). Accessory minerals are titanomagnetite, biotite, apatite, potash feldspar, quartz, some pyrite and zircon and green spinell. Secondary minerals are serpentine, chlorite, opaque minerals, pale amphibole, hornblende, epidote. Except at dynamometamorphic places there is not a great amount of secondary minerals. For instance in the Satakunta olivine diabase they often occur in much greater quantities. The average modal mineral composition of the rock in volume per cent is:

Plagioclase	60.0 ± 3
Olivine	22.0 ± 7
Augite	10.0 ± 7
Opaque minerals	3.5 ± 1
Apatite	1.0 ± 0.3
Serpentine, Chlorite, Biotite, Other minerals	3.5 ± 0.5
	100.0

The plagioclase laths are zonal. Three zones can be distinguished: central part, intermediate zone and narrow border zone. (Fig. 3, Plate 2.) The central part is not found in all grains, but only in the large ones. This central part of the plagioclase grains is the one most rich in An. It contains about 10 % more anorthite than the intermediate zone. The central part contains a considerable number of small inclusions, which are augite, olivine, plagioclase and opaque minerals. Around the inclusions is a plagioclase sphere which is poorer in anorthite than the plagioclase of the central part at a greater distance from the inclusion. Augustithis (1963) describes a somewhat divergin but most interesting oscillatory zoning of plagioclase phenocryst of olivine basalt from Debra-Sina. The olivine diabase of Ansio has evidently crystallized in more immobile circumstances than the Debra-Sina basalt. This zone is generally sharply distinguished from the intermediary zone, the composition of which is approximately An_{65} ($\alpha = 1.560$, $\gamma = 1.570$). The greater part of the plagioclase grains has this composition. The intermediary zone does not show any important number of inclusions and its extinction between the crossed nicols is regular, even if it is zonal. Oscillatory zoning also occurs in this region. The intermediate zone changes over — as a matter of fact on a very short distance — by and by into a narrow border zone which sometimes is altogether missing and the composition of which is $An_{(40-30)}$. The composition of the plagioclase in the olivine diabase of Satakunta is the same, and there, too it varies comparatively much (Kahma 1951, p. 13).

It is difficult to give a perfectly clear explanation of the difference between the crystallisation of the central and the intermediate zone of the plagioclase. Considerable changes must have occurred in the circumstances in which this crystallization took place. One or several of the following properties have undergone a change: the form energy of the plagioclase grains, its capacity to acquire inclusions and its rate of crystallization as well as the stability of the plagioclase lattice and the supersaturation of the magma. See also Vance (1962). Evidently this change has occurred while the composition of the plagioclase was between 65—70 per cent An.

Olivine and augite are not evenly distributed throughout the rock, in the same way as plagioclase, and their relative amounts vary. Here and there are large groups of either the one or the other of these minerals between the plagioclase laths. The olivine occurs in grains the diameter of which is 0.3—1.5 mm. It is often euhedral, but anhedral grains are not unusual; sometimes it is even found as poikilitic grains among the plagioclase laths, in the same ways as augite. Its composition varies between 35—52 % Fe-component ($\alpha = 1.711$—$1,721$, $\gamma = 1.751$—1764, $2V\alpha = 81°$—$73°$). The value of the axial angle shows that the euhedral grains are richer in Mg,

Table 1. Chemical composition for olivine diabase from Ansio. (Analyst, A. Heikkinen). Padasjoki, Finland

Weight per cent		Weight norm		Molecular norm		Niggli values	
SiO_2	47.36	or	3.9	Or	1.0	si	103.
TiO_2	1.00	ab	22.5	Ab	23.5	al	25
Al_2O_3	19.70	an	39.8	An	39.5	fm	48
Fe_2O_3	0.75	Σ	66.2	Σ	67.0	c	20
FeO	10.46	wo	1.2	Wo	1.0	alk	7
MnO	0.14	en	3.0	En	3.8	k	0.14
MgO	8.34	fs	2.4	Fs	2.2	mg	0.57
CaO	8.77	fo	12.4	Fo	14.4	ti	2
Na_2O	2.68	fa	11.3	Fa	9.0	o	0.03
K_2O	0.63	mt	1.2	Mt	0.9	qz	—25
P_2O_5	0.15	il	2.0	Il	1.4		
CO_2	0.0	ap	0.3	Ap	0.3		
H_2O+	0.43						
H_2O-	0.03						
Σ	100.44		100.0		100.0		

Modal composition

Plagioclase	62.0
Olivine	27.5
Augite	3.7
Opaque minerals	3.0
Serpentine	1.3
Chlorite	0.5
Apatite	0.8
Other minerals	1.2
Σ	100.0

Plagioclase (An_{64})

$\alpha = 1.560$
$\beta = 1.564$
$\gamma = 1.569$

Olivine (Fa_{40})

$\alpha = 1.751$
$\beta = 1.735$
$\gamma = 1.711$
$2V\gamma = 77°$
$d_{130} = 2.806$ Å

but the statistics are insufficient to permit to make conclusions with any certainty. In the fractures of olivine grains there are oxide minerals, especially in those cases when the olivine is altered. In such cases other secondary minerals are present as well. On the fracture faces of the olivine there is also dark green chlorite, light green serpentine and brown iddingsite (Brown and Stephen, 1959). Uniform broad reaction rims are not found in the surroundings of the olivine grains.

The colour of the augite is a violet brownish grey. It occurs as allotriomorphic, often large poikilitic grains, whose the parts do not always touch each other in the thin section, but are extinguished simultaneously between the crossed nicols. The composition of the augite also varies ($\alpha = 1.698$—1.718, $\gamma = 1.729$—1.742; $2V\gamma = 41°$—$49°$; $c \wedge \gamma = 32°$—$42°$). The chemical composition and physical properties of the augite are shown in Table 2.

The opaque minerals are generally allotriomorphic. They occur in connection with olivine, and also in augite and its surroundings and, in addition, in independent grains in which, for instance, plagioclase inclusions are found. Sometimes there are small quantities of reddish brown biotite around the opaque minerals.

The potash feldspar is allotriomorphic in the interstices. Secondary minerals are often found beside it. Where quartz is found, it often forms graphic textures together with the potash feldspar.

Table 2. Chemical composition and physical properties for augite from Ansio. Padasjoki, Finland. (Analyst, A. Heikkinen.)

	Weight per cent				
SiO_2	49.38	Si	7.52	Z = 8.00	Si = 7.52; Al = 0.48
TiO_2	1.25				
Al_2O_3	2.28	Ti	0.15	Y = 0.34	
Fe_2O_3	2.24	Al	0.40		
FeO	11.62	Fe^{3+}	0.26		
MnO	0.29	Fe^{2+}	1.48	X = 4.67	
MgO	13.86	Mn	0.04		
CaO	17.97	Mg	3.15		
Na_2O	0.50	Ca	2.93	W = 3.08	
K_2O	0.05	Na	0.15		
P_2O_5	0.19				
CO_2	0.00				
H_2O+	0.23				
H_2O-	0.02				
	100.00				

Physical properties

$\alpha = 1.699$
$\beta = 1.704$
$\gamma = 1.729$
$2 V \gamma = 43°$
$c \wedge \gamma = 39°$
$D = 3.38$

There are but minute amounts of apatite in the Ansio diabase and inclusions of apatite have been found for instance in the potash feldspar in interstices.

Table 1 shows the chemical composition of the olivine diabase of Ansio, its weight and molecular norms as well as Niggli values determined from a specimen which was taken from an island situated north of Virranpaju in a river connecting both lakes. The chemical composition of the rock corresponds rather well with that of the olivine diabase of Satakunta (Kahma 1951, p. 15). The only difference is that the TiO_2 and H_2O+ contents are somewhat lower and the Fe_2O_3, Al_2O_3 and MgO contents somewhat higher than in the olivine diabase of Satakunta.

Table 2 gives the chemical analysis and physical properties of the augite from the rock in Table 1. Eskola (1954, p. 11) has published a phase analysis of the olivine diabase of Satakunta which also contains an analysis of augite. At the same time the optic properties of augite are given. Both these and the analysis correspond pretty nearly with the values established for the Ansio olivine diabase. Table 3 shows the x-ray diffraction powder data of the augite of the Ansio olivine diabase. In the handbook of Deer *et al.* (1963) there is a

Table 3. X-ray diffraction powder data for augite from Ansio. (Copper Kα radiation, Nickel filter.) Padasjoki, Finland.

No.	d	Q	I/I_0	No.	d	Q	I/I_0
1	4.76	.0441	11	21	2.042	.2398	21
2	4.60	.0473	11	22	2.015	.2463	11
3	4.22	.0562	10	23	1.969	.2579	10
4	4.19	.0567	16	24	1.932	.2707	11
5	4.08	.0600	13	25	1.837	.2967	11
6	3.275	.0935	18	26	1.797	.3100	11
7	3.218	.1023	53	27	1.754	.3250	21
8	2.994	.1116	100	28	1.626	.3787	26
9	2.971	.1133	18	29	1.529	.4277	11
10	2.945	.1153	78	30	1.522	.4317	11
11	2.901	.1188	37	31	1.488	.4516	10
12	2.721	.1351	11	32	1.472	.4615	11
13	2.567	.1518	37	33	1.444	.4796	11
14	2.516	.1580	47	34	1.411	.5023	11
15	2.296	.1897	21	35	1.394	.5146	10
16	2.232	.2007	13	36	1.350	.5487	11
17	2.211	.2046	13	37	1.324	.5705	11
18	2.162	.2139	21	38	1.296	.5954	11
19	2.134	.2198	26	39	1.265	.6249	10
20	2.113	.2240	11	40	1.249	.6410	11

summary of the research concerning the structure of the augite, its chemical composition, the variations in the lattice constants as a function of its composition, and a description of the history of its research.

The border variety of the Olivine diabase dike is a dark grey, fine grained rock of porphyritic and ophitic texture. The essential minerals are the same as in the main types: plagioclase, augite and olivine. The remaining minerals are: opaque minerals, potash feldspar, quartz, biotite, hornblende, chlorite, serpentine and apatite.

The composition of the main part of the plagioclase is An_{60} ($\gamma = 1.568$, $\alpha = 1.559$) and its grains are zonal $An(_{69}—_{40})$. The following properties have been found for the augite: $2V\gamma = 48°$; $c \wedge \gamma = 44°$; $\alpha = 1.718$, $\gamma = 1.742$. The composition of the olivine is Fo_{32} ($\alpha = 1.768$, $\gamma = 1.813$; $2V\alpha = 68°$). The augite and especially the olivine are greatly altered at the borders and cleavages. There are great amounts of opaque minerals, more than in the specimens taken from the central part of the dike, as elongated, almost ophitically ranged crystals. Biotite is found in connection with the opaque minerals. There is considerably more potash feldspar in the contact variety, and it shows graphic textures. Quartz is also found in separate grains. There is more graphic feldspar in this border variety than in the main type of the rock. Kahma (1951) has discussed its genesis at great length. Apatite (Kerr, 1959, p. 233) occurs abundantly in long, narrow, prismatic crystals, especially in the graphic feldspar (Fig. 4, plate 2).

The modal mineral composition of the marginal variety of olivine diabase from the specimen taken on the island to the NW of Virranpaju is:

Plagioclase (phenocrysts)	9.1
Plagioclase (groundmass)	29.8
Augite	27.9
Olivine	2.9
Hornblende	5.1
Biotite	6.4
Quartz	5.7
Potash feldspar	4.5
Opaque minerals	6.2
Apatite	1.2
Apatite (inclusions in graphic feldspar)	1.2
	100.0

The border variety contains plagioclase and olivine in considerably lesser quantities, but opaque minerals, secondary minerals and apatite more than the main rock. About one half of the apatite occurs in inclusions in graphic potash feldspar. The dark minerals are richer in Fe, the plagioclase slightly richer in Ab than the main type of rock.

SHEAR ZONES IN THE OLIVINE DIABASE DIKE

The olivine diabase of Ansio contains here and there shear zones of different stages of development. Their thickness varies from 1 mm to several metres. The trend of the zones is generally NS, the dip rather vertical (Fig. 1, p. 100). The microscopically thin zones are almost exclusively formed of light green fibrous serpentine regardless of the minerals through which it goes.

The serpentine fibres run vertically to the trend of the shear zone. The different minerals of the rock behave differently to the serpentinization caused by the dynamometamorphism. Both remaining halves of the augite crystals pierced by the shearing zone are generally fresh and unaltered. On the other hand, the halves of olivine grains, which are cleft by the shear zone and which remain outside the zone, have mostly also altered into light green serpentines, though there are fresh olivine crystals in the rock outside of the shear zone. The fragments of plagioclase grains are quite well preserved, even if the serpentines have pushed their way into the cleavages.

The broadest shearing zone of olivine diabase is in Viitaniemi near the shore of Lake Vehkajärvi. The weathering surface of the rock of this zone is of a greenish light grey. There is an abundant quantity of shear surfaces intersecting each other, so that it is difficult to find a sufficiently big uniform sample in this exposure. The ophitic texture of this rock has disappeared, (Figs. 5 and 6, Plate 3) and its minerals have altered. The lamellae of the plagioclase are bent and broken, the mineral has become more albitic than in the fresh diabase. The composition of the plagioclase in this shear zone is, according to the refractive indices, usually between An_{40} and An_{50}, but more albitic fragments have also been found, among the rock meal, even down to An_{15} (α=1.535). A certain quantity of augite also occurs, forming rather large relics, but no olivine. Secondary and accessory minerals in the rock are epidote, chlorite, serpentine, actinolite, carbonate, biotite, opaque minerals apatite and titanite.

The plagioclase grains are strongly epidotized, they are completely covered by small epidote grains. There is also carbonate, which very easily dissolves in cold diluted HCl. The carbonate appears in rather large allotriomorphic grains in connection with the plagioclase. Epidote is also found in quite large grains elsewhere in the rock. The biotite is partly chloritized and its flakes are bent. The chlorite flakes (β = 1.625) are twinned resembling the albite twinning in plagioclase and the lamellae of the grains show an oblique extinction (11°) with regard to the twinning plane. Greenish serpentine (β = 1.564) occurs in large groups of grains. Light amphibole crystals ($c \wedge \gamma = 18°$, α = 1.631, β = 1.643, γ = 1.652) in radial groups also occur. The grains of the opaque minerals are very allotriomorphic, and there seem to be more opaque minerals than in the fresh diabase. The apatite occurs in large euhedral grains, abundantly.

The diabase of Ansio has erupted into a fracture valley, which evidently is older than the diabase dike. After the settling of the diabase, shearing movements the trend of which is very nearly from the north to the south have occurred. Here the fracture valley and fault line direction from the north to the south is rather common. Even the lake Päijänne (Fig. 1) follows this trend (Sederholm, 1932).

Since the Ansio diabase dike is not found any more on the islands of Lake Vehkajärvi, nor on its NW- shore, we must ask ourselves if the reason for this fact are the shear zones, or if the dike terminates before reaching the other shore of the lake. Near Virranpaju, in the NE- corner of Lake Vesijako the map (Fig. 1) shows the dike to be severed, and its SE end is displaced more than one hundred m to the east. It is not probable and on the basis of field observations it seems even impossible that the dike should make such a bend near Virranpaju, or that it should be so much thicker here that it would now form one continuous dike. A fault has occurred or there are two different dikes. It is possible to assume a fault because of the shear zones occurring in the dike, even if the existence of two different dikes is not an impossibility.

No differences in texture or in mineral composition, which would not find an explanation in the normal changes occurring in the dike, were, however, found in the different halves of the dike. If there have been two different dikes from the very beginning, later movements have in any case broken up the diabase dikes into pieces, which are separated from each other by a rock of a wholly different texture and composition — the »amphibole gabbro», as Frosterus (1902) has said. The last mentioned rock, which is a small minority, was generated by crushing and mylonitization from olivine diabase. Seitsaari (1962, p. 20) has written about the fracturization and brecciation of basic dikes considerably older than the Ansio diabase. The formation of faults, the period of time of their formation and their importance have been discussed by, among others, Sederholm (1913), Asklund (1923), Sundius (1928), Hills (1955) and Kukkamäki (1963).

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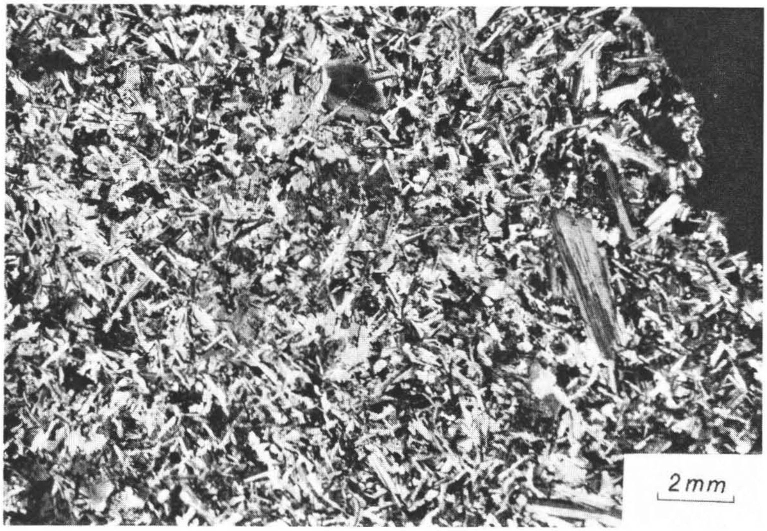

Fig. 1. Texture of marginal variety of olivine diabase from Ansio. (Microphoto. Thin section, + nic..) Padasjoki, Finland.

Fig. 2. Ophitic texture of olivine diabase from Ansio. (Microphoto. Thin section, + nic..) Padasjoki, Finland.

Fig. 3. Texture of diabase and zonal structure in plagioclase phenocryst. Olivine diabase from Ansio. (Microphoto. Thin section, + nic..) Padasjoki, Finland.

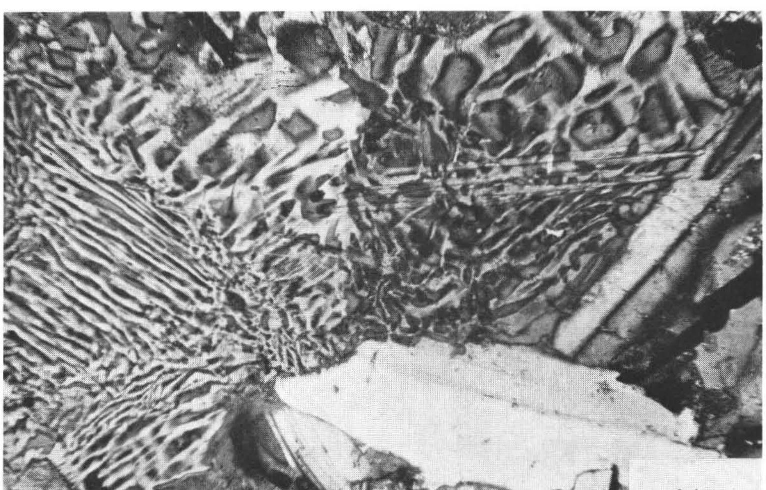

Fig. 4. Graphic feldspar in the interstices of Ansio olivine diabase and apatite in the graphic feldspar. (Microphoto. Thin section, + nic..) Padasjoki, Finland.

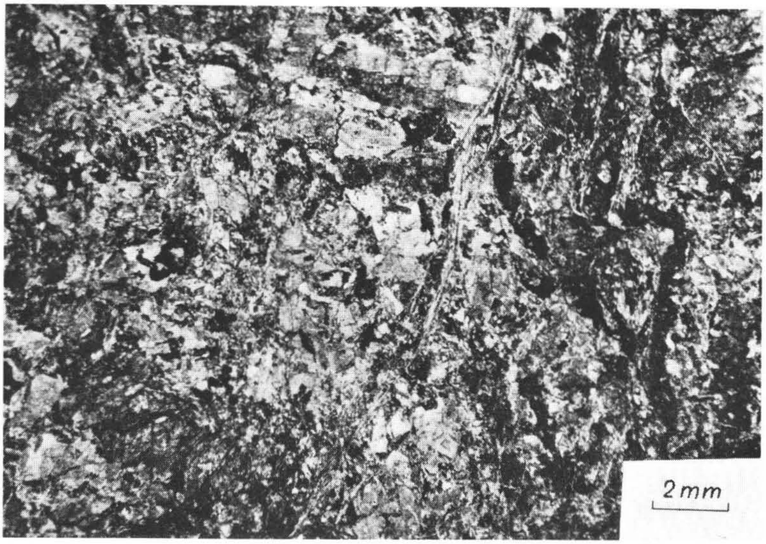

Fig. 5. Mylonitized olivine diabase from Ansio. (Microphoto. Thin section, + nic..) Padasjoki, Finland.

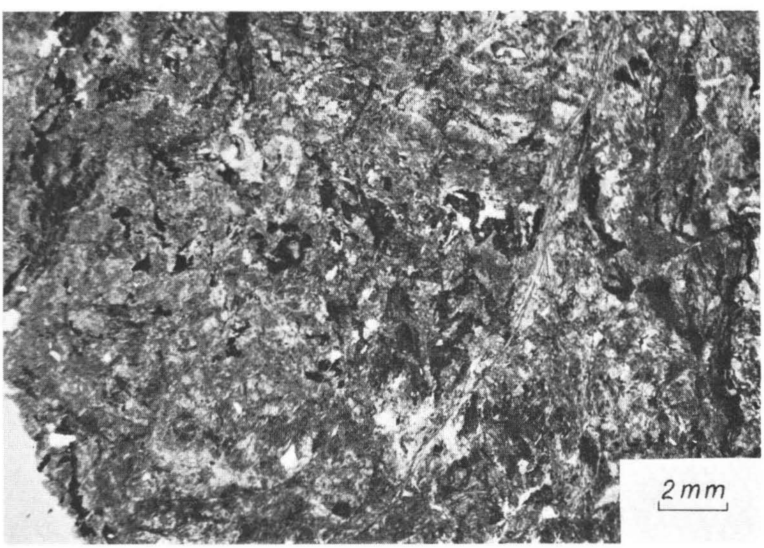

Fig. 6. Mylonitized olivine diabase from Ansio. (Microphoto. Thin section, one nic..) Padasjoki, Finland.

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SEDERHOLMITE, WILKMANITE, KULLERUDITE, MÄKINENITE and TRÜSTEDTITE, FIVE NEW NICKEL SELENIDE MINERALS [1)]

BY

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ABSTRACT

Five new nickel selenide minerals from Kuusamo, NE-Finland are described. The minerals occur in veinlets in albitites associated with uranium mineralisation.

Sederholmite, which is identical with the synthetic hexagonal β-NiSe phase, has $a_0 = 3.62 — 3.65$ Å and $c_0 = 5.29 — 5.34$ Å depending upon composition.

Wilkmanite corresponds to the artificial monoclinic Ni_3Se_4 phase. Its $a_0 = 6.22$ Å, $b_0 = 3.63$ Å, $c_0 = 10.52$ Å and $\beta = 90.55°$.

Kullerudite $NiSe_2$ is an ortho-rhombic nickelian analogue of ferroselite with $a_0 = 4.89$ Å, $b_0 = 5.96$ Å and $c_0 = 3.76$ Å.

Mäkinenite, which is the same as the artificial γ-NiSe phase, has a trigonal symmetry with $a_0 = 10.01$ Å and $c_0 = 3.28$ Å.

Trüstedtite is a cubic spinel type Ni_3Se_4 having $a_0 = 9.94$ Å. It forms a solid solution series with polydymite. Members of this series with 40—70 mole per cent polydymite component have also been found.

CONTENTS

1) Received May 6, 1964.

INTRODUCTION

The system nickel-selenium has been studied by various authors. According to the literature the following selenides of nickel are known.

α-NiSe was obtained by Levi and Baroni (1935) by conducting ammonium selenide into a neutral nickel salt solution. The product was amorphous.

β-NiSe, which has a hexagonal structure of the NiAs type (Grønvold and Jacobsen, 1956, Hiller and Wegerner, 1960) has been prepared either by heating appropriate amounts of the elements in evacuated and sealed silica tubes (Grønvold and Jacobsen, 1956) or by precipitating it with hydrogen selenide in a nickel acetate solutiòn containing acetic acid (Levi and Baroni, 1935).

γ-NiSe has been obtained by Levi and Baroni (1935) by conducting hydrogen selenide into a nickel sulphide solution containing sulphuric acid. Hiller and Wegener (1960) have also prepared γ-NiSe by heating powdered nickel and selenium at 225—250 C°. The compound is trigonal with a millerite type structure.

$NiSe_2$ has been prepared by de Jong and Willems (1928) by heating a mixture of NiSe and selenium at 230 C° for 48 hours. It has also been synthetised by e.g. Grønvold and Jacobsen (1956), Hiller and Wegerner (1960) and Klemm (1962). The $NiSe_2$ phase has a pyrite structure.

Ni_3Se_2 has been reported by Hiller and Wegener (1960). The phase is hexagonal with $a_0 = 6.029$ Å and $c_0 = 7.249$ Å.

The Ni_3Se_4 phase has also been prepared by Hiller and Wegenar (1960). According to them it is monoclinic with $a_0 = 12.15$ Å, $b_0 = 3.633$ Å, $c_0 = 10.45$ Å and $\beta = 149°22'$. Grønvold and Jacobsen (1956) have derived the structure of Ni_3Se_4 from that of the deformed hexagonal β-NiSe cell and obtained different unit cell dimensions: $a_0 = 6.196$ Å, $b_0 = 3.634$ Å, $c_0 = 10.464$ Å and $\beta = 90.78°$.

Only two nickel selenide minerals have been reported so far. Blockite $(Ni,Cu)Se_2$ was described by Herzenberg and Ahlfelt (1935) from Bolivia. In fact the mineral is the same as penroseite which was discovered in the same locality and studied by Gordon (1925). Blockite is structurally similar to the artificial $NiSe_2$.

Robinson and Brooker (1952) reported a new nickel-bearing selenide, tyrrelite $(Cu, Co, Ni)_3Se_4$ from Beaverlodge Lake, Saskatchewan. They suggested that the mineral is structurally similar to pentlandite but Machatschki (1952) has pointed out that the powder pattern of tyrrelite resembles those of the spinel and linnaeite groups of minerals.

In connections with the prospecting work carried out by the Outokumpu Co in Kuusamo, in NE Finland during 1961—1963 quite a few new nickel selenide minerals were encountered. In this paper five of them i.e. β-NiSe, monoclinic Ni_3Se_4, γ-NiSe, ortho-rhombic $NiSe_2$ and cubic Ni_3Se_4 are described.

GEOLOGY OF THE SELENIUM MINERALISATION AREA.

In Kuusamo in NE Finland the Se-mineralisation occurs in albite diabases which form sills in a schist formation. The schist formation is composed of quartzites, dolomites, marls, mica schists and black schists. Also green-schists are encountered. Quartzites are either clastic ortho-quartzites or albite-bearing arkose quarzites grading into dolomites. The marls are fine-grained, predominantly grey, dolomite-bearing sediments which contain, apart from dolomite, quartz, micas and albite as major minerals.

Ripple marks, fossil mud cracks and clastic dikes are found in sediments. Graded bedding which permits the determination of the sedimentation bottom is observed in mica schists and phyllites. In some localities phyllites contain carbon-bearing »bags» with calcite-rich cores. The size of the »bags» is variable but the biggest attain a diametre of about 30 cm. They are believed to be of fossil origin.

The green-schists, which may be, at least partly, of volcanic origin are chlorite-albite-epidote rocks often showing a distinct bedding. The bedding of the schist formation dips subvertically. Folds as well as faults are common. The formation has locally tectonised intensively giving rise to transverse schistosity.

Albite diabase is a hornblende-albite rock of variable grain size. Especially in strongly tectonised places it changes into more fine-grained biotite-bearing rock which contains only a small amount of hornblende or is completely free from it. Pink albitite is associated with these biotite-albite diabases forming dikes or irregular bodies in the latter. Albitites occur preferably at the ends or at the flexure points of the albite diabase sills, or near by the country rock contacts i.e. generally in places where tectonisation has been strongest.

Small amounts of sulphide, selenide and telluride minerals have been found in albitites, among them some hitherto unknown in geological literature. These minerals are intimately associated with low-grade uranium mineralisation and especially nickel selenides seem closely to follow uranium. Outside the uranium mineralisation the nickel content diminishes rapidly, whereas the amount of cobalt relative to that of nickel increases considerably. A similar correlation has been reported from Shinkolobwe by Derriks and Vaes (1955). In the wall rock cobalt is incorporated in the lattice of pyrite which can contain up to 3 per cent cobalt.

The uranium mineralisation, which is confined to albitites, is not homogeneous but the uranium minerals occur as isolated spots and veinlets. The most important mineral is uranite, but brannerite, davidite, kasolite and uranoanatase have also been identified as well as some alteration products of primary uranium minerals.

OCCURRENCE

Among the new nickel selenides β-NiSe and monoclinic Ni_3Se_4 form a natural and interesting pair. The artificial counterparts of these minerals have been prepared and reported by e.g. Levi and Baroni (1935), Grønvold and Jacobsen (1956) and Hiller and Wegener (1960). Since β-NiSe changes into monoclinic Ni_3Se_4 by decreasing nickel content, both minerals occur associated most intimately with each other. They are found almost exclusively in blockite clusters of a few millimetres in diametre, which seem to favour calcite veinlets in the uranium-bearing parts of albitite dikes. β-NiSe and monoclinic Ni_3Se_4 brecciate blockite forming micro-veinlets and flame-like segregations in it as is seen in Fig. 1, Plate I. β-NiSe has also been detected as independent grains in clausthalite. Monoclinic Ni_3Se_4 seems to occur either as a primary or secondary mineral. The latter is an alteration product of β-NiSe. The secondary monoclinic Ni_3Se_4 often gives a diffuse powder pattern with the consequence that only the strongest reflections are visible. Monoclinic Ni_3Se_4 is frequently followed by native secondary selenium and sometimes also by ferroselite. Selenian vaesite and selenian cattierite also belong to the mineral paragenesis but they are younger than blockite, β-NiSe and monoclinic Ni_3Se_4, for they form veinlets cutting the latter. (See Fig. 1, Plate I).

The ortho-rhombic $NiSe_2$ seems to occur almost exclusively as an alteration product of monoclinic Ni_3Se_4. The mineral grains often contain relics of the latter as indicated by Fig. 2, PlateI, but in some cases the primary mineral has completely disappeared. The formation of the ortho- rhombic $NiSe_2$ has taken place soon after the crystallisation of β-NiSe and monoclinic Ni_3Se_4 but before the appearance of selenian sulphides, for they are seen to cut ortho-rhombic $NiSe_2$. In spite of the fact that ortho-rhombic $NiSe_2$ is a nickelian analogue of ferroselite it has never been observed in association with the latter. Ferroselite seems to be primary. It is distinctly yellow in colour and contains about 5 per cent nickel.

γ-NiSe was first found combined with clausthalite (see Fig. 3, Plate I) in stringers at the ends of the veins filled with carbonate, clausthalite, NiTeSe (a new mineral to be described in a forthcoming paper), selenian melonite, hematite and spinel type selenian sulphides. It greatly resembles millerite but is considerably softer. Later γ-NiSe was also found as submicroscopic grains in the cracks of selenian melonite associated with clausthalite.

The spinel type Ni_3Se_4 also belongs to the mineral paragenesis of the aforementioned calcite- and uranium-bearing veinlets. It occurs as euhedral crystals in clausthalite in association with blockite and β-NiSe (Fig. 4, Plate II), Cubic Ni_3Se_4 seems to form a solid solution with polydymite with the consequence that $Ni_3(Se,S)_4$ minerals have also been detected. They occupy, together with clausthalite, the very ends of the stringers filled by carbonate, NiTeSe, hematite and Ni-, Co- and Se-bearing pyrite (Fig. 5, Plate II). They have also been found in NiTeSe where they are commonly enveloped by idiomorphic apatite crystals.

PHYSICAL AND OPTICAL PROPERTIES

The optical properties of β-NiSe vary somewhat depending upon composition. In oil the stoichiometric variant is orange yellow in colour. Pleochroism is distinct with different hues of yellow. Anisotropism is strong: pinkish — greenish. The nickel-deficient type is yellow in colour. Its pleochroism is weak: yellow — greyish yellow and anisotropism distinct: pink — greenish. Monoclinic Ni_3Se_4 is pale greyish yellow in colour. Pleochroism is distinct: pale yellow — greyish yellow and anisotropism strong: pink — yellowish green. Both minerals take an easy polish and have a high reflectivity.

The colour of the ortho-rhombic $NiSe_2$ is very similar to that of blockite except that it is slightly paler. Pleochroism in oil is distinct: grey — pale grey and anisotropism very strong: yellowish grey — grey to almost black. The mineral seldom polishes well and is rather soft.

In oil γ-NiSe is pure yellow in colour but in air orange yellow. Pleochroism is strong: pure yellow — greenish yellow and anisotropism extremely strong: pale green — pale orange yellow. In air anisotropism is different: glowing cinder red — blue green or green. The mineral is almost as soft as clausthalite.

Cubic Ni_3Se_4 is yellow in colour and completely isotropic. Its reflectivity is higher than that of β-NiSe and blockite and it is only slightly softer than blockite. The $Ni_3(Se,S)_4$ variant is different in colour. It is olive grey and takes an easy polish. Reflectivity is slightly less than that of sulphur-free Ni_3Se_4. The mineral is considerably softer than selenian linnaeite.

CHEMICAL COMPOSITION

Due to their rarity it was impossible to separate the selenide minerals in quantities sufficient for ordinary wet chemical analyses and their composition had to be determined by X-ray fluorescence methods using chemically analysed nickel selenides as standards. The analyses were carried out by a Philips's universal vacuum spectrograph provided with a tungsten tube. The composition of β-NiSe was found to vary from $Ni_{1\cdot05}Se$ to $Ni_{0\cdot85}Se$ i.e. from the stoichiometric NiSe to the point at which the β-NiSe phase changes into monoclinic Ni_3Se_4. A typical analysis for a nickel deficient β-NiSe is as follows: 36.8 % Ni, 1.9 % Co and 61.3 % Se. As a rule the mineral contains about 1—2 per cent cobalt, whereas iron and copper are completely lacking or exist only as impurities.

According to an X-ray fluorescence analysis, the most nickel-rich monoclinic Ni_3Se_4 contained 37.6 % Ni and 61.1 % Se, which gives $Ni_{0\cdot83}Se$ as the composition for that particular mineral sample. Cobalt is, of course, added to nickel in the above formula. The transition point between hexagonal β-NiSe and monoclinic Ni_3Se_4 phases represented by the above specimen agrees excellently with the composition of the transition point reported by Grønvold and Jacobsen (1956) i.e. $Ni_{0\cdot82}Se$.

The most nickel deficient monoclinic Ni_3Se_4 contained 33.7 % Ni, 1.0 % Co and 65.3 % Se which corresponds closely to the formula Ni_3Se_4. Again, only traces of copper and iron were detected in this mineral.

An X-ray fluorescence analysis showed ortho-rhombic $NiSe_2$ to contain 23.1 % Ni, 1.4 % Co, 1.91 % Fe, 0.5 % Cu and 73.1 % Se. The chemical formula for the mineral is thus $NiSe_2$ within limits of analytical errors.

The composition of γ-NiSe was also determined by the same method and the mineral was found to contain 41.1 % Ni, 1.0 % Co and 57.9 % Se. Only traces of copper were detected.

An analysed sample of cubic Ni_3Se_4 mineral contained 29.5 % Ni, 6.4 % Co and 64.1 % Se. Traces of copper were also found. Due to the extremely small amount of the material available it was impossible to determine sulphur accurately. Its concentration was in any case very low, if not altogether nil.

According to the length of the a-axes the sulphur-bearing $Ni_3(Se,S)_4$ minerals contain about 40—70 mole percent of polydymite component. As a rule cobalt percentage seems to be lower than that of the sulphur-free Ni_3Se_4.

CRYSTALLOGRAPHY

The unit cell dimensions and powder patterns for the minerals were obtained with a 114.6 mm diametre camera using silicon as an internal standard. The lattice parametres of β-NiSe and monoclinic Ni_3Se_4 vary somewhat depending on the composition. The values for a nickel-rich β-NiSe are: $a_0 =$ 3.65 Å and $c_0 = 5.34$ Å and for a nickel-deficient one: $a_0 = 3.624$ Å and $c_0 = 5.288$ Å, respectively. These values are plotted in Fig. 1. as a function of the composition together with the values for the artificial compounds

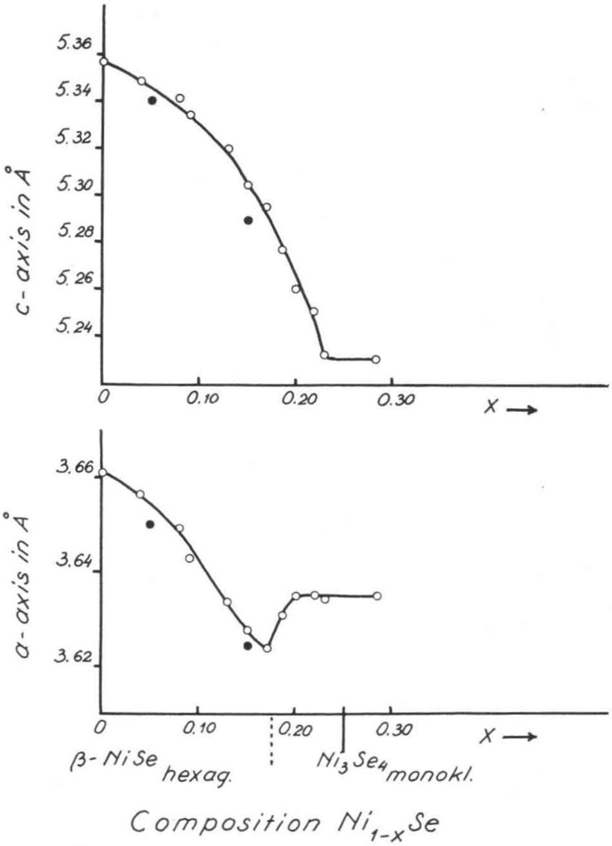

Fig. 1. The unit cell dimension of hexagonal β-NiSe plotted against the composition. The white circles: values reported by Grønvold and Jacobsen (1956). The black circles: values obtained from the β-NiSe minerals from Kuusamo.

reported by Grønvold and Jacobsen (1956). The lattice parametres of the β-NiSe minerals are slightly smaller than those for the synthetic compounds but the difference is obviously insignificant and may be entirely due to the analytical errors. On the other hand it may also be caused by a small amount of cobalt incorporated in the lattice of the mineral. The space group for β-NiSe is $P6_3/mmc$. In Table 1 the X-ray powder data for the β-NiSe mineral of the composition $Ni_{0.85}Se$ is given.

The indexing of monoclinic Ni_3Se_4 has been based on the information reported by Grønvold and Jacobsen (1956). The powder data for the mineral is shown in Table 2. The data represent the composition of $Ni_{0.78}Se$ which approaches Ni_3Se_4 very closely. The unit cell dimensions for that particular

Table 1. X-ray powder data and lattice dimensions for β-NiSe mineral from Kuusamo, NE-Finland. The composition of the mineral is $Ni_{0.85}Se$. Camera diametre 114.59 mm, Cu/Ni radiation.

hkl	I	$d_{obs.}$	$d_{calc.}$	
001	w	5.27	5.29	a_o = 3.624 Å
101	vvs	2.70	2.699	c_o = 5.288 Å
102	vs	2.015	2.022	Z = 2
110	s	1.806	1.812	V = 60.14 Å^3
103	ms	1.535	1.537	$D_{calc.}$ = 7.06 g/cm^3
112	ms	1.50	1.495	
202	m	1.348	1.350	
004	w	1.32	1.322	
203	mw	1.172	1.173	
211	m	1.155	1.158	
212	m	1.082	1.082	
114	m	1.067	1.068	

Table 2. X-ray powder data and lattice dimensions for monoclinic Ni_3Se_4 mineral from Kuusamo, NE-Finland. Camera diametre 114.59 mm, Cu/Ni radiation.

hkl	int.	$d_{obs.}$	$d_{calc.}$	
002	m	5.25	5.26	a_o = 6.22 Å
011	w	3.42	3.422	b_o = 3.63 Å
$\bar{1}03$	w	3.06	3.078	c_o = 10.52 Å
$\bar{1}12$	vsI	2.70	2.700	β = 90.53°
$\bar{2}02$			2.700	Z = 2
112	m	2.66	2.677	V = 237.5 Å^3
013	w	2.52	2.518	$D_{calc.}$ = 6.96 g/cm^3
211	vw	2.29	2.295	
$\bar{1}14$	vsII	2.02	2.024	
$\bar{2}04$			2.025	
114	s	2.00	2.004	
020	s	1.815	1.815	
310	vsIII	1.800	1.799	
$\bar{1}16$	ms	1.532	1.535	
222	ms	1.497	1.495	
$\bar{4}04$	m	1.343	1.348	
008	w	1.31	1.315	
$\bar{2}26$	wB	1.170	1.172	
$\bar{4}06$			1.173	
$\bar{1}32$	wB	1.158	1.156	
132			1.154	
406			1.154	
512	wB	1.147	1.144	
$\bar{1}34$	wB	1.081	1.081	
$\bar{3}18$	wB	1.066	1.069	
028			1.064	
600	vwB	1.037	1.037	

Table 3. X-ray powder data and lattice dimensions for ortho-rhombic $NiSe_2$ mineral and ferroselite from Kuusamo, NE-Finland. Camera diametre 114.59 mm, Cu/Ni radiation.

Ferroselite $FeSe_2$				Ortho-rhombic $NiSe_2$ mineral				
hkl	I	d		hkl	I	$d_{obs.}$	$d_{calc.}$	
110	vw	3.69	$a_0 = 4.79$ Å	110	f	3.79	3.799	$a_0 = 4.89$ Å
				011	w	3.12	3.125	
101, 020	s	2.86	$b_0 = 5.72$ Å	020	w	2.98	2.980	$b_0 = 5.96$ Å
				101	s	2.935	2.937	
111	vs	2.565	$c_0 = 3.58$ Å	111	vs	2.64	2.634	$c_0 = 3.67$ Å
120	vs	2.47		120	vs	2.545	2.546	
200	w	2.395	Z = 2	200	w	2.44	2.449	Z = 2
210	vw	2.21		210	w	2.26	2.265	
121	w	2.02	V = 98.2 $Å^3$	121	mw	2.095	2.092	V = 107.0 $Å^3$
211	vs	1.88		211	s	1.925	1.928	
130	s	1.79	$D_{calc.}$ =	130	s	1.84	1.841	$D_{calc.}$ =
			7.22 g/cm^3	002			1.835	6.72 g/cm^3
031	ms	1.69		031	w	1.75	1.747	
221	w	1.63		221	mw	1.685	1.682	
112	w	1.595		112	m	1.648	1.651	
				131			1.646	
310, 022	w	1.54		310, 022	wB	1.57	1.575	
				320	w	1.435	1.434	
240	w	1.238		240	w	1.275	1.273	
312	w	1.162		312	w	1.195	1.195	

sample are: $a_0 = 6.22$ Å, $b_0 = 3.63$ Å, $c_0 = 10.52$ Å and $\beta = 90.53°$. These values are again in good accordance with the values given by Grønvold and Jacobsen.

As pointed out by Hiller and Wegener (1960) the β-NiSe and monoclinic Ni_3Se_4 phases form a continuous solid solution series with variable concentration of the empty Ni-spaces. These empty spaces are statistically distributed in the hexagonal solid solution range, for there are no reflections in the powder pattern indicating a change of the lattice symmetry. With the increasing concentration of the empty spaces the hexagonal β-NiSe phase changes into the monoclinic Ni_3Se_4 phase, in which the defiency of nickel obviously reaches its maximum. The densities for the β-NiSe and monoclinic Ni_3Se_4 phases have been calculated under the assumption that the changes in composition are caused by subtraction of nickel atoms from the β-NiSe structure. The validity of this assumption has been verified e.g. by Grønvold and Jacobsen (1956) and Hiller and Wegener (1960). Due to the increase of the empty Ni-spaces in the β-NiSe — monoclinic Ni_3Se_4 solid solution the density tends to decrease from β-NiSe towards Ni_3Se_4.

The indexing of the ortho-rhombic $NiSe_2$ has been based on the structure of marcasite. The powder data for the mineral is given in Table 3 together with those for ferroselite. The mineral is ortho-rhombic with $a_0 = 4.89$ Å,

Table 4. X-ray powder data and lattice dimensions for γ-NiSe mineral from Kuusamo, NE-Finland compared with those for millerite. Camera diametre 114.59 mm, Cu/Ni radiation.

Millerite NiS ASTM 12—41				γ-NiSe mineral				
hkl	I	$d_{calc.}$		hkl	I	$d_{obs.}$	$d_{calc.}$	
110	60	4.81	$a_0 = 9.620$ Å	110	m	4.99	5.00	$a_0 = 10.01$ Å
101	40	2.946		101	vw	3.07	3.070	
300	100	2.777	$c_0 = 3.149$ Å	300	vs	2.88	2.890	$c_0 = 3.28$ Å
021	65	2.513		021	vs	2.63	2.617	
220	12	2.406		220	vw	2.49	2.502	$Z = 9$
211	55	2.228		211	vs	2.325	2.319	
131	95	1.8631		131	vs	1.95	1.940	$V = 284.6$ Å
410	45	1.8178		410	w	1.892	1.892	
401	40	1.7372		401	mw	1.81	1.809	$D_{calc.} = 7.22$ g/cm^3
321	18	1.6340		321	mw	1.71	1.701	
330	35	1.6037		330	mw	1.665	1.669	
				002	mw	1.64	1.642	
012	25	1.5470		012			1.612	
				331	vw	1.48	1.487	
600	8	1.3884		600	vw	1.445	1.445	
520	4	1.3343		520			1.388	
312	10	1.3008		312			1.355	
042	8	1.2560		042			1.308	
440	6	1.2023		440			1.251	
161	4	1.1783		161			1.226	
502	6	1.1447		502			1.191	
422	16	1.1133		422			1.159	
701				701			1.158	
710	8	1.1033		710			1.148	
152	8	1.0846		152			1.129	
621				621			1.129	
342	12	1.0333		342			1.076	
541	6	1.0104		541			1.051	
612	6	0.9888		612	mw	1.035	1.030	

$b_0 = 5.96$ Å and $c_0 = 3.67$ Å. The space group is probably Pnnm. The X-ray powder pattern of ortho-rhombic $NiSe_2$ has great similarity with that of ferroselite (Kullerud and Donnay, 1958) and marcasite, even though the unit cell parametres for the former are somewhat bigger. The density of the ortho-rhombic $NiSe_2$ has been calculated based on the assumption that its structure is similar to that of ferroselite.

The unit cell dimensions for γ-NiSe have been reported by Levi and Baroni (1935) and Hiller and Wegener (1960). According to the latter $a_0 = 10.007$ Å and $c_0 = 3.333$ Å for the compound. The values obtained from the γ-NiSe mineral are: $a_0 = 10.01$ Å and $c_0 = 3.28$ Å which agree well with the values of Hiller and Wegener (1960). The powder pattern for γ-NiSe is given in Table 4 together with that for millerite. The mineral is trigonal and its powder pattern very similar to that of millerite except that the d-values for

Table 5. X-ray powder data for Ni_3Se_4 and $Ni_3(Se,S)_4$ from Kuusamo, NE-Finland compared with those for polydymite. Camera diametre 114.59 mm, Cu/Ni radiation.

Polydymite ASTM 8—106				Ni_3 (Se, S)$_4$ Kuusamo			Ni_3Se_4 Kuusamo		
hkl	I	d		I	d		I	d	
111	20	5.50		mw	5.60		mw	5.75	
220	40	3.34	a_o =	mw	3.40	a_o =	mw	3.52	a_o = 9.94 Å
311	90	2.85	9.48 Å	vs	2.90	9.65 Å	ms	3.00	
222	—	—		w	2.79		ms	2.87	Z = 8
400	90	2.36		vs	2.41		vs	2.48	
422	30	1.941		v	2.00		—	—	V = 982.1 $Å^3$
511, 333	90	1.820		s	1.86		m	1.905	
440	100	1.674		vs	1.706		vs	1.755	$D_{calc.}$ =
531	10	1.600		—	—		—	—	6.62 g/cm^3
026	10	1.499		—	—		—	—	
533	50	1.444		—	—		—	—	
540	—	—		vw	1.50		—	—	
622	—	—		—	—		w	1.498	
444	60	1.369		vw	1.395		w	1.434	
711, 551	20	1.330		vw	1.355		—	—	
642	30	1.269		f	1.29		—	—	
731, 553	80	1.232		vw	1.255		vw	1.293	
800	70	1.185		vw	1.21		—	—	
660, 822	10	1.117		—	—		—	—	
751, 555	70	1.095		—	—		—	—	
662	—	—		—	—		w	1.149	
840	60	1.055		vw	1.078		w	1.11	
842	—	—		f	1.055		—	—	
911, 753	5	1.041		—	—		—	—	
664	5	1.010		—	—		—	—	
931	50	0.994		—	—		—	—	

γ-NiSe are invariably somewhat bigger. The powder pattern shows the mineral to be isotype with millerite which has a space group R3m.

The powder data for cubic Ni_3Se_4 resemble those for tyrrelite and also for polydymite very closely. For the sake of comparison the d-values and intensities of polydymite, of a $Ni_3(Se,S)_4$ variant with about 63 mole per-cent of polydymite component and of Ni_3Se_4 are given in Table 5. The Table shows that the length of the a-axis increases from polydymite towards Ni_3Se_4 with increasing Se-content reaching a = 9.94 Å at Ni_3Se_4.

The space group for the cubic Ni_3Se_4 is obviously the same as that for tyrrelite, i.e. Fd3m. The mineral can be considered as the nickelian analogue of bornhardite (Ramdor, 1955) which has only a slightly longer a-axis than Ni_3Se_4.

The density of the cubic Ni_3Se_4 i.e. 6.62 g/cm^3 has been obtained by calculation based on the assumption that it is structurally similar to tyrrelite. The density for tyrrelite is 6.6 g/cm^3, as reported by Robinson and Brooker (1952).

NOMENCLATURE

The facts presented above suggest the existence of five new nickel selenide minerals. The β-NiSe mineral is named in honour of the late Dr. J. J. Sederholm, the former Director of the Geological Survey of Finland, in recognition of his unique and fundamental contribution to Pre-Cambrian geology. The name sederholmite should be applied to hexagonal β-NiSe phase from the stoichiometric NiSe composition to the point at which β-NiSe changes into monoclinic Ni_3Se_4 phase.

The monoclinic Ni_3Se_4 mineral is named after the late Dr. W. W. Wilkman as an appreciation of his great contribution to geological knowledge about Finland, and the name wilkmanite is proposed to be used for a mineral the composition of which falls within the range of $Ni_{0.82}Se$ — $Ni_{0.7}Se$.

For the ortho-rhombic $NiSe_2$ mineral the name kullerudite is suggested in honour of Dr. Gunnar Kullerud, Carnegie Institution of Washington, Geophysical Laboratory, in recognition of Dr. Kullerud's valuable contributions to mineralogy.

The γ-NiSe mineral is named in honour of the late Dr. Eero Mäkinen, the former President of the Outokumpu Co. as an acknowledgement of his great merit in geology and mineralogy, as well as of his achivements in the Finnish mining industry.

The name trüstedtite is proposed for the spinel type Ni_3Se_4 mineral in honour of the late Dr. O. Trüstedt in recognition of his pioneering work in the development of prospecting methods which in 1910 led to the discovery of the Outokumpu deposit.

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To all the aforementioned persons the authors wish to express their sincere gratitude.

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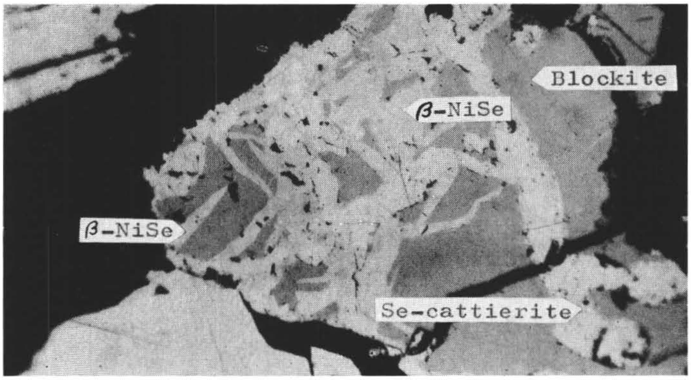

Fig. 1. β-NiSe mineral brecciating blockite. Younger selenian cattierite forms veinlets cutting the β-NiSe — blockite association. Magn. 100 x.

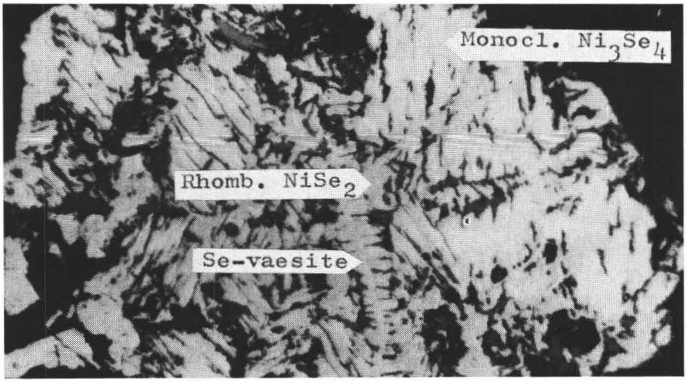

Fig. 2. Monoclinic Ni_3Se_4 mineral partly changed into rhombic $NiSe_2$. A narrow selenian vaesite veinlet cuts rhombic $NiSe_2$. Magn. 300 x.

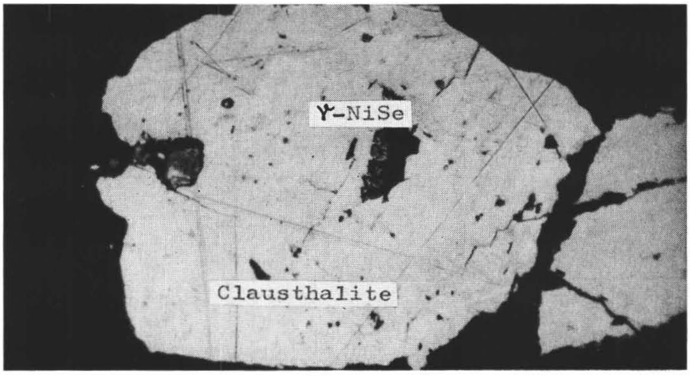

Fig. 3. γ-NiSe mineral in association with clausthalite. Magn. 300 x.

Y. Vuorelainen, A. Huhma and *A. Häkli:* Sederholmite, Wilkmanite . . .

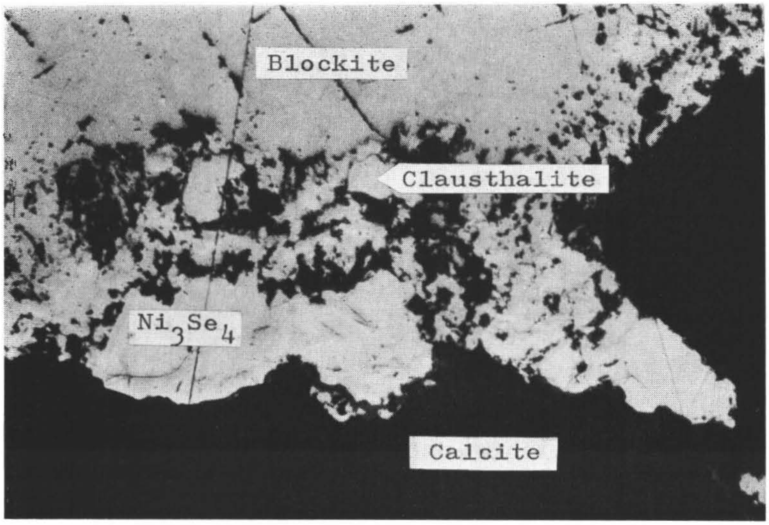

Fig. 4. Cubic Ni_3Se_4 crystals in clausthalite which surrounds a blockite ovoid. Magn. 100 x.

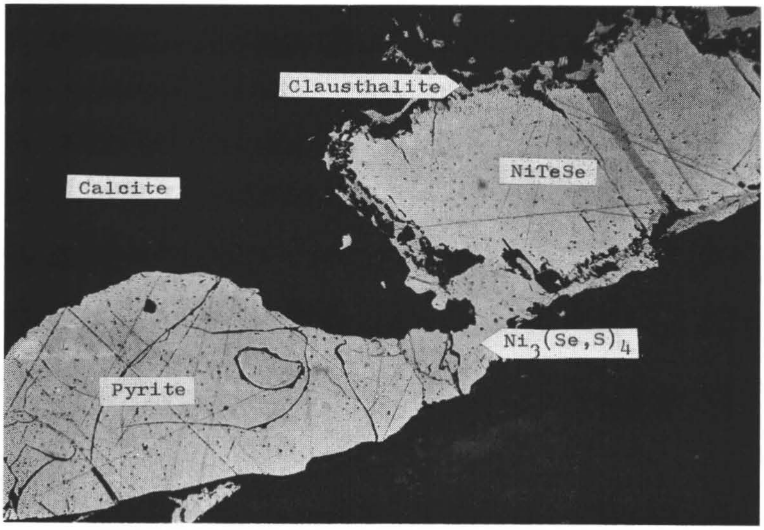

Fig. 5. $Ni_3(Se,S)_4$ mineral in association with Ni-, Co- and Se-bearing pyrite, NiTeSe and clausthalite at the end of a stringer. Magn. 200 x.

5

THE HORNBLENDE RAPAKIVI DIKE OF JAALA-IITTI [1])

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ABSTRACT

The authors petrographically describe a hornblende rapakivi dike that penetrates rapakivi and migmatic granite, and they present the chemical composition of the various rocks contained in the dike. The origin of the dike rocks is also dealt with briefly.

CONTENTS

[1]) Received May 16, 1964.

INTRODUCTION

In the explanatory text accompanying Map Sheet No. 8 (Lahti), K. Ad. Moberg (1885) mentions that on the northern side of Märkjärvi in the commune of Iitti as well as in the area between Märkjärvi and Kirkkojärvi, rapakivi granites occur that deviate from the common rapakivi types. He refers to them as quartz- and feldspar-porphyries. In connection with the mapping operations carried out by one of the present writers (ML) in this area in 1955, it became evident that the occurrences mentioned by Moberg belong to a dike formation over 20 km long, which begins near Kamposenjärvi in the commune of Jaala and extends to Kirkkojärvi in the commune of Iitti (Plate I; Ahti Simonen-Mauno Lehijärvi 1963). The eastern part of the dike trends approximately eastwest and it penetrates porphyritic rapakivi for a distance of approximately 7 km. The end of the dike curves toward the southwest, penetrating the migmatic granite of the surroundings, and in the northwestern corner of Kirkkojärvi, Iitti, again joins the great Viipuri rapakivi region. Between the dike and the rapakivi region there occurs an area of migmatic granite about 20 km^2 in breadth. The dike varies in breadth from a couple of hundred meters to three kilometers. Because, in the light of preliminary investigations, the dike formation proved exceptionally interesting, the other of the two writers (AL) undertook to map the area in detail in the years 1956—1958. The results of his work were presented by him in 1960 at the University as his thesis for a licenciate degree.

DIABASE

The dark rock containing potash feldspar phenocrysts which occurs in fragmentary form in the area between Kirkkojärvi and Pyhäjärvi has been designated in this paper as diabase. Its chemical composition (Table 1, No. 1) does not really correspond to that of any ordinary plutonic rock, but it constitutes some sort of potash- and silica-rich diabase-like rock. In a classification of rocks, it might come closest to being identified as a quartz-monzonite porphyry.

The dense, dark groundmass is weakly ophitic in structure and the grains are on the average 0.06 mm in diameter. The roundish phenocrysts occurring in it are chiefly potash feldspar, and they vary in diameter between 0.5 and 7.0 cm. In certain cases the phenocrysts have a plagioclase mantle. Some of the ovoids contain potash feldspar, quartz and plagioclase. Plagioclase is also present as laminae, which may be 1.5 cm in length. In places the groundmass also contains quartz as small phenocrysts. With the exception of the fine grain and very dark color of the groundmass, the rock outwardly bears a close resemblance to the hornblende rapakivi that brecciates it.

Table 1. Chemical analyses of rocks of the Jaala—Iitti hornblende rapakivi dike.

	Weight per cent			
	1	2	3	4
SiO_2	57.00	68.05	67.96	73.30
TiO_2	0.66	0.68	0.67	0.20
Al_2O_3	15.68	12.77	12.96	13.36
Fe_2O_3	2.31	1.75	1.92	0.95
FeO	8.70	4.36	4.20	1.45
MnO	0.12	0.10	0.09	0.00
MgO	2.80	0.58	0.54	0.26
CaO	5.50	2.05	2.49	0.86
Na_2O	2.90	2.91	3.04	4.00
K_2O	2.63	5.12	4.92	4.92
P_2O_5	0.51	0.29	0.28	0.22
F	0.03	—	—	0.19
H_2O+	1.01	0.92	0.56	0.59
H_2O-	0.10	0.14	0.07	0.04
	99.95	99.72	99.70	100.34
$-O = F$	0.01			0.08
	99.94			100.26
	Weight norms			
Q	9.9	24.9	24.7	28.9
or	15.6	30.3	29.1	29.1
ab	24.5	24.6	25.7	33.8
an	22.5	6.6	7.4	2.8
C	—	—	—	0.4
wo	0.6	0.7	1.3	—
en	7.1	1.6	1.5	0.7
fs	12.9	5.4	5.0	1.5
mt	3.3	2.5	2.8	1.4
il	1.3	1.3	1.3	0.4
ap	1.2	0.7	0.7	0.5
	98.9	98.6	99.5	99.5
	Niggli values			
si	173.0	314.0	307.5	405.0
al	28.4	34.5	34.5	44.0
fm	40.1	27.0	26.0	13.0
c	18.2	10.0	12.0	5.0
alk	13.6	28.0	27.5	39.0
k	0.37	0.55	0.51	0.44
mg	0.31	0.14	0.14	0.16
c/fm	0.45	0.37	0.46	0.38
qz	+17.0	+102.0	+97.5	+149.0

1. Potash feldspar phenocrysts containing diabase. About 200 m N of Kirkkojärvi, Iitti. Analyst, E. Lindqvist.
2. Hornblende rapakivi. Sahaniemi, Jaala. Analyst, P. Ojanperä.
3. Tirilite. Kokkomäki, Jaala. Analyst, P. Ojanperä.
4. Even-grained rapakivi granite. Hausanniemi, Jaala. Analyst, E. Lindqvist (CaO and MgO determined by P. Ojanperä).

The degree of triclinicity of the potash feldspar varies between 0.0 and 0.6. A micrographic structure is to be observed in many spots around individual potash feldspar crystals as well as in the large ovoids containing potash feldspar and quartz.

The plagioclase laminae are well preserved around the middle, but toward their ends they are generally sericitized. A zoned structure is a common feature. With respect to composition the plagioclase is andesine An_{45}.

Hornblende ($\alpha = 1.672$, $\gamma = 1.697$) occurs both in the form of flakes and as large phenocrysts. In the latter form this mineral is exceedingly tattered. In many places it has partly altered to biotite. Around the potash feldspar phenocrysts and near the contacts the biotite content is in excess of the usual amount.

In connection with the biotite ($\gamma \approx \beta = 1.679$) one will often observe zircon, sphene and small magnetite grains. As accessory minerals the groundmass also contains quartz, potash feldspar, apatite and, in spots, carbonate.

HORNBLENDE RAPAKIVI

For the most part the Jaala-Iitti rapakivi dike consists of hornblende rapakivi. Outwardly it varies in appearance. The mineral composition and structure of the different varieties do not, however, appreciably differ from one another. Between them there are no contacts or short-distance transitions. In view of these differences in appearance, the most distinct varities are separated in the following according to the areas of their occurrence.

In the eastern parts of Pyhäjärvi the hornblende rapakivi outwardly resembles granite porphyry. The potash feldspar phenocrysts have no plagioclase mantle and they are situated a few centimeters apart. In form they are either roundish or angular and approximately between 1 and 2 cm in diameter. Also plagioclase and dark quartz occur as phenocrysts. The grains in the matrix measure about 0.4 mm in diameter.

In the Pyhäjärvi area and to the southwest of it, near the eastern end of Märkjärvi an even-grained variety prevails. The grains average 2 mm in diameter. Potash feldspar phenocrysts occur at intervals of about half a meter from each other, and the grains generally have no plagioclase mantle.

In the area between Märkjärvi and Kirkkojärvi the hornblende rapakivi is unhomogeneous. Here too the groundmass is fine-grained, the diameter of the grains being roughly 2 mm. The diameter of the potash feldspar phenocrysts varies between 1 and 10 cm and they are frequently encircled by a plagioclase mantle. Around the quartz grains one can occasionally notice a dark rim. Fig. 1 presents this unhomogeneous variety.

Fig. 1. Unhomogeneous variety of hornblende rapakivi. Northern shore of Märkjärvi, Iitti. Photo: M. Lehijärvi.

In a few places there has crystallized around the potash feldspar and plagioclase phenocrysts a dark, fine-grained substance with an ophitic structure (Fig. 2). In addition to the small plagioclase laminae, the hornblende

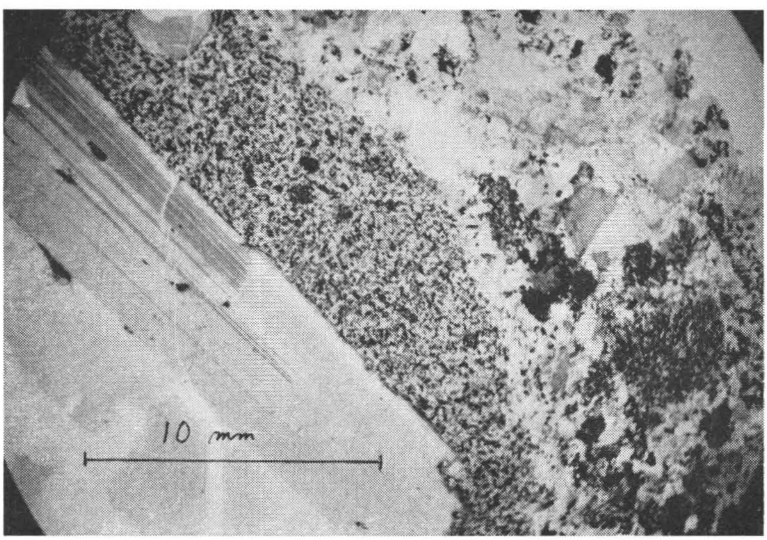

Fig. 2. Dark matter around plagioclase. Approx. 200 m to the north from Kirkkojärvi, Iitti. Photo: P. Virtanen.

likewise occurs in a flake-like form. Another mafic mineral is biotite. Accessory minerals are potash feldspar, quartz, apatite and magnetite. Similar dark, basic accumulations are present here and there in the hornblende rapakivi rock as isolated schlieren.

The hornblende rapakivi is throughout a reddish brown in color. The variation in type is caused principally by the grain size of the matrix and by the differences in porphyritic character. Microscopically, however, the different varieties look very much alike. The chief minerals are quartz, potash feldspar, plagioclase, hornblende and biotite. The accessory minerals are fluorite, pyroxene, apatite, sphene and magnetite.

The quartz in spots exhibits idiomorphic features, but for the most part it is allotriomorphic. In Fig. 3 there is a quartz grain with a dark rim. The grain's weakly angular form suggests an originally idiomorphic quartz grain. The dark rim contains principally hornblende and small amounts of biotite and pyroxene. Together with the potash feldspar, the quartz tends to exhibit a beautiful micrographic structure.

The degree of triclinicity of the potash feldspar varies from 0.2 to 0.7. At Hausanniemi, near the contact with the country rock, there occurs nearly pure orthoclase, with a triclinicity degree of 0.0—0.2.

The plagioclase is weakly zoned. In the central portions of the grains the composition represents andesine, while in the marginal portions it borders on oligoclase and andesine. To some degree the plagioclase has sericitized.

The relative amounts of hornblende and biotite vary locally, but in any case the hornblende content invariably exceeds that of biotite. The biotite is dark and typical of the biotite contained in rapakivi granite, the index of refraction varying in different parts of the area relatively little ($\gamma \approx \beta = 1.679$ —1.682). The hornblende ($\alpha = 1.684$, $\gamma = 1.709$, $c \wedge \gamma = 28°$) appears to be considerably more ragged in the western parts of the area than in the eastern parts.

The even-grained hornblende rapakivi closely resembles the granite of Lappee (Hackman 1934) as well as the hornblende rapakivi of Ahvenisto (Savolahti 1956). The chemical composition of this type is given in Table 1, No. 2.

In the eastern part of the dike, where it intersects the porphyritic rapakivi, a dark green variety of rapakivi occurs in a couple of places along the margin as a strip a few dozen meters wide. Wahl (1925) called it tirilite. The groundmass is fine-grained, the diameter of the grains averaging 0.03 mm. Present are phenocrysts of quartz, plagioclase, potash feldspar and hornblende. The borders of the phenocrysts are eroded and filled with fine-grained minerals from the groundmass. In connection with the quartz grains one may observe a beautiful micrographic structure.

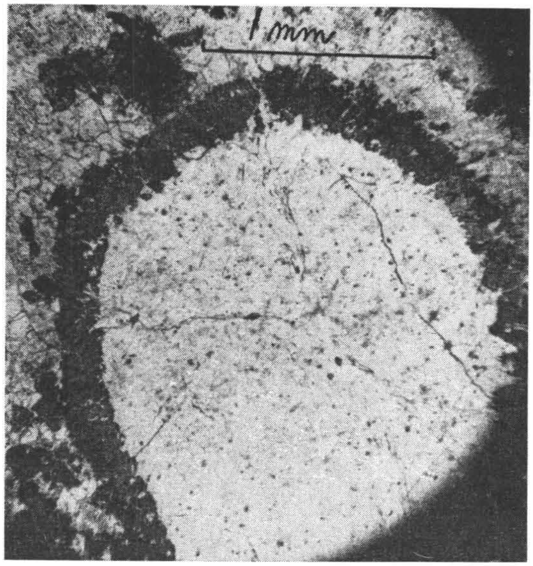

Fig. 3. Quartz grain in hornblende rapakivi. Around the grain occur hornblende, biotite and pyroxene. On the northern shore of Märkjärvi, Iitti. Photo: O. Näykki.

The plagioclase is andesine. With the exception of small sericite particles situated parallel to the cleavage the plagioclase phenocrysts are pure.

The potash feldspar phenocrysts contain inclusions of plagioclase, quartz and hornblende. No cross-hatching is to be observed and the degree of triclinity is very low (0.0—0.2).

The hornblende ($\alpha = 1.686$, $\gamma = 1.705$, $c \wedge \gamma = 29°$) is quite tattered, and only the total picture suggested by the ragged outlines indicates the crystal surfaces that had possibly previously existed. Quite commonly distributed through the central portions of the grains are remnants of fayalite. In addition the hornblende contains inclusions of quartz, biotite, magnetite and apatite.

The groundmass is composed of the same minerals as the phenocrysts. Compared to the content of light minerals there is, however, a remarkably small amount of biotite and hornblende particles. Also present in the groundmass is fluorite.

The chemical composition of this tirilite (Table 1, No. 3) proved to correspond to that of green rapakivi varieties met with elsewhere in the Viipuri rapakivi area. It is further interesting to observe that with respect to its chemical composition this tirilite is pretty nearly the equivalent of the hornblende rapakivi just discussed.

EVEN-GRAINED RAPAKIVI GRANITE

At Hausanniemi, Hiidensaari, Kettusaari and the western part of the peninsula of Lyöttilä, in the area of Marlebäck, there occurs fine- and even-grained rapakivi granite. In color the rock is for the most part gray, but in spots it gradually turns reddish. The grain size does not vary. As a slight deviation from homogeneity, a weak tendency toward porphyritic texture is to be observed in the Marlebäck area. The gray rapakivi granite there contains plagioclase phenocrysts measuring about 0.5 cm in diameter at an average distance of between 5 and 10 cm from each other.

The structure of the even-grained rapakivi granite is hypidiomorphic. There are approximately equivalent amounts of plagioclase An_{15}, potash feldspar and quartz. The potash feldspar commonly is cross-hatched. The degree of triclinicity varies between 0.2 and 0.7. The sole mafic mineral is biotite, $\gamma \approx \beta = 1.669$. The indices of refraction are lower than in the case of typical rapakivi but nevertheless higher than those met with in older granites. The accessory minerals are zircon and fluorite.

The chemical composition of the even-grained rapakivi granite is presented in Table 1, No. 4. The amounts of both silica and calsium suggest potash granite rather than rapakivi. The natrium content is likewise substantially higher than in rapakivi. On the other hand, the low mg figure and the fluorine content apply admirably to rapakivi.

PEGMATITE

It is generally known that there is a scant occurrence of pegmatite in rapakivi granites. The same observation has been made with respect to the Jaala-Iitti rapakivi dike. The only slightly broader pegmatite dike (0.5 m) has been encountered on the western shore of Sahaniemi. The large potash feldspar crystals are in many cases full of plagioclase laminae. The composition of the plagioclase is An_{30}. The quartz content is substantial and it frequently forms beautiful crystals. In spots there is a considerable amount of calcite. Accessory minerals are muscovite and fluorite.

THE RAPAKIVI DIKE AS RELATED TO COUNTRY ROCKS AND THE MUTUAL RELATION OF DIKE ROCKS

Nowhere has any contact between the diabase containing potash feldspar phenocrysts and the migmatic granite been found. The diabase occurs only as fragmentary breccia or as isolated inclusions in hornblende rapakivi. It must therefore be older than the hornblende rapakivi.

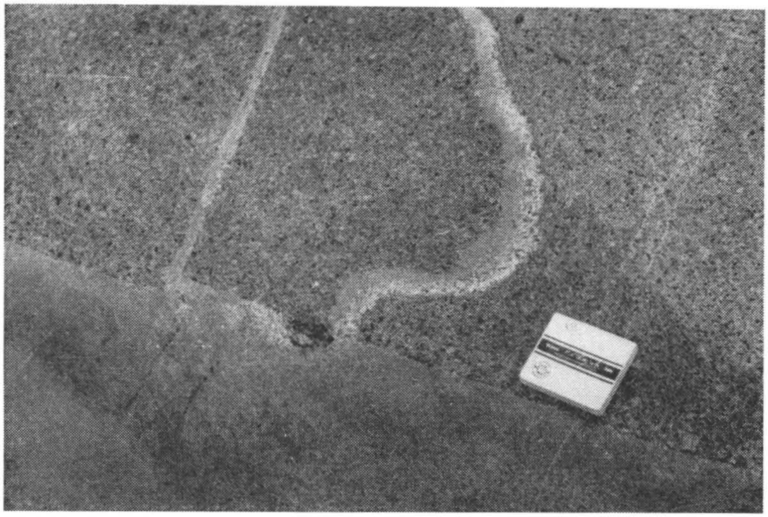

Fig. 4. Contact between even-grained rapakivi granite and hornblende rapakivi. Hausanniemi, Jaala. Photo: M. Lehijärvi.

The hornblende rapakivi intersects both the porphyritic rapakivi granite and the migmatic granite. Wherever the contacts are visible, they always appear sharp against these country rocks.

The hornblende rapakivi contains considerable amounts of both wiborgitic and porphyritic rapakivi fragments as well as migmatic granite fragments. It is in the eastern part of the dike that rapakivi fragments occur in greatest abundance, whereas in the western part of the dike there is a prevalence of migmatic granite fragments.

No contact appears between the tirilite and the hornblende rapakivi. With respect to mineral composition as well as chemical composition, the tirilite nearly corresponds to hornblende rapakivi. The fine-grained groundmass indicates a rapid cooling down of the tirilite, and it must accordingly be regarded simply as a contact variety of hornblende rapakivi. The contact between the tirilite and the porphyritic rapakivi is invariably a sharp one. Types of rock corresponding to the hornblende rapakivi and tirilite of Jaala-Iitti are to be met with in the Ahvenisto area (Savolahti 1956 and 1962).

Even-grained rapakivi granite brecciates migmatic granite in the Marlebäck area. At Hausanniemi it sharply cuts across hornblende rapakivi (Fig. 4). From it apophyses also run over to the hornblende rapakivi side. The even-grained rapakivi granite, moreover, contains hornblende rapakivi fragments. In this area it represents a very late product of crystallization.

Fig. 5. Detailed map some 200 m north of Kirkkojärvi, Iitti. 1. Diabase. 2. Hornblende rapakivi. 3. Aplite. 4. Pegmatite. 5. Potash feldspar ovoid with a plagioclase mantle as well as without the mantle. 6. Pegmatitic dike cuts through potash feldspar ovoid.

The sparse pegmatite dikes that occur in the area intersect all the aforementioned types of dikes.

THE ORIGIN OF THE DIKE ROCKS

As it became evident in connection with the petrographic description, the diabase-like rock containing phenocrysts is a somewhat unusual rock. The phenocrysts present in large amounts in the basic ophitic groundmass are chiefly potash feldspar. The question as to whether the genesis of these phenocrysts is the result of metasomatosis or whether the rock as a whole has evolved out of magma requires closer examination.

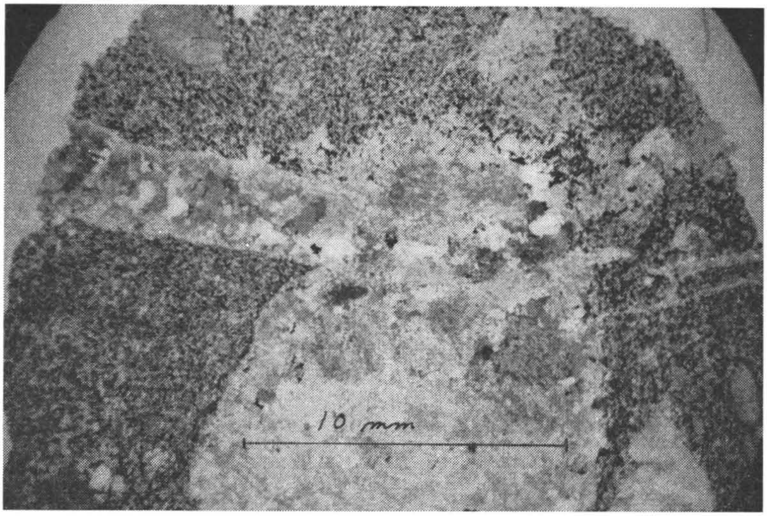

Fig. 6. Pegmatitic dike penetrates potash feldspar ovoid in diabase. Approx. 200 m north of Kirkkojärvi, Iitti. Photo: P. Virtanen.

The detailed map (Fig. 5), which has been drawn of the breccia area on the northern side of Kirkkojärvi, reveals the mutual relation between the diabase and the hornblende rapakivi, pegmatite and aplitic granite that brecciate it. The hornblende rapakivi thrusts dikes and apophyses into the diabase. Pegmatite and aplitic granite cut across both the hornblende rapakivi and the diabase and spread a dense dike network through both varieties of rock. These highly potash-rich dikes are not, however, anywise involved with the potash feldspar phenocrysts of the diabase. Fig. 5 shows in sections 1 and 2 how a pegmatite dike intersects a potash feldspar ovoid in the diabase. The same thing is illustrated in photomicrograph Fig. 6, which means that the phenocrysts must have evolved prior to the pegmatitic stage.

Insofar as hornblende rapakivi magma caused the genesis of the potash feldspar phenocrysts, it must have happened under the influence of acid magma in a more basic rock existing in a crystalline state. It is understandable that an acid magma is incapable of melting a rock formed from minerals that melt at a higher temperature. Partial melting may, to be sure, occur. According to Bowen's assimilation theory, a certain ion exchange also occurs. These effects of assimilation reactions are generally far more strongly developed on the acid rock side of the contact (Turner-Verhoogen 1960).

In clarifying in the ligth of the foregoing the relation between the hornblende rapakivi and the diabase of Jaala-Iitti, it should first be noted that the contact between these rocks is quite sharp (Fig. 7). On the diabase side,

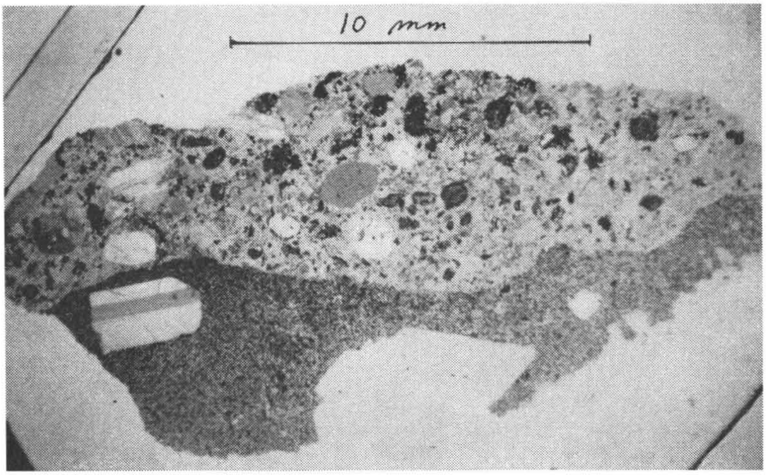

Fig. 7. Contact between hornblende rapakivi and diabase. Approx. 200 m to the north from Kirkkojärvi, Iitti. Photo: P. Virtanen.

one might observe quite next to the contact the slightly more abundant occurrence of biotite at the expense of hornblende. Similarly, in the immediate proximity of a number of potash feldspar phenocrysts, the amount of biotite increases while there is a total absence of hornblende. The zoned character of the plagioclase may represent the diminishing of the anorthite content toward a state of equilibrium, starting at the edges, but it is more likely that the causes of the zoned structure must be sought in magmatic crystallization. In only a few spots can one observe on the hornblende rapakivi side a slight concentration of mafic minerals and then at a distance of at most 1—2 cm from the contact. The changes thus appear to be too slight on the sides of both hornblende rapakivi and diabase to justify regarding such a considerable amount of potash feldspar in diabase as having been produced by assimilation or, in general, through metasomatic action.

In addition to the foregoing, it should be pointed out that potash feldspar phenocrysts are not met with in the Jaala-Iitti region in association with any other basic rock. Insofar as rapakivi magma, generally considered, had caused potash metasomatosis in the Jaala-Iitti region, one would expect the rocks occurring in different parts of the region to exhibit its effects at least to some degree.

A megascopic examination of the hornblende rapakivi occurring in the Jaala-Iitti region and the diabase found there with its fragmentary potash feldspar phenocrysts will reveal their close structural resemblance. This suggests that if the hornblende rapakivi is of magmatic origin, the diabase, too, together with its potash feldspar phenocrysts must have crystallized wholly from magma. The micrographic structures appearing in both, the high indices of refraction of the biotite — typical of rapakivi —, the ragged hornblende, the predominance of the orthoclase in the potash feldspar and, above all, the occurrence of diabase exlusively in association with hornblende rapakivi indicate their possible crystallization from the same magma.

In connection with the description it was mentioned that in spots around the feldspar phenocrysts or also quite in isolation, the hornblende rapakivi contains small inclusions corresponding in composition to diabase. It is not possible to draw a line between these small instances of unhomogeneity occurring in the hornblende rapakivi and the larger inclusions or fragments occurring in the form of brecciae.

The hornblende rapakivi in the Jaala-Iitti region appears to be very unhomogeneous. The same kind of observation has been made by Högbom (1893) concerning the dike types of Ragunda. This indicates that the rapakivi magmas were not of very even quality, either. Accepting the premise that hornblende rapakivi and diabase derived from the same magma, the crystallization in the case of this magma would have begun with potash feldspar, just as Savolahti (1956) observed it to begin in rapakivi. After the basic components of the magma had started to crystallize, differentiation occurred in the crystallizing magma, and the partially crystallized basic fraction separated under pressure into a fissure dike, carrying along with it already crystallized potash feldspar grains. Thereafter the crystallization process advanced rapidly, as is typical expressly of dike rocks. The rock took on its characteristic porphyritic structure. Differentiation was not, however, complete. Evidence of this is provided by the small basic accumulations around the feldspar grains or in isolation in the hornblende rapakivi.

After the crystallization of the basic rock new fissures developed. The partially crystallized hornblende rapakivi magma erupted and brecciated the previously separated basic fraction. Crystallization was once again rapid, although the structure is not so reminiscent of the hypabyssal one as was the preceding type, for a considerably larger intrusion, if still a dike, is in question. At the edges quite hypabyssal structural features resulted, as exhibited by the tirilite.

Considering magmatic origins, one further possibility should be taken into account. It has previously been noted that hornblende rapakivi magma is apt to cause partial melting in basic fragments. This means that as the hornblende rapakivi magma crystallized, the constituents undergoing partial melting could have accumulated in the basic rock to form quartz and potash feldspar porphyroblasts. The question nevertheless remains whether quartz and potash feldspar can crystallize in a solid basic rock. According to Szadeczky-Kardoss (1955), the bonding potential of quartz and orthoclase is one of the lowest among common minerals. Partly on the basis of the bonding potential research of Szadeczky-Kardoss, Eskola (1956) states that it is unreasonable to regard potash feldspar crystals achieving their proper form as porphyroblasts, for their idiomorphism can be explained only by an early crystallization from magma.

With respect to chemical composition and outward appearance as well, the Jaala-Iitti diabase may be compared to some extent to the diabase hybrids of Satakunta studied by Kahma (1951). In Kahma's view the potash feldspars of these diabase hydrids are porphyroblasts, which evolved as a result partly of the mutual reactions of molten rapakivi and diabase magma and partly of the migration at a later stage of solid matter from the granite to the diabase. In the Satakunta region, however, the succession of the rapakivi and the diabase hydrids is opposite to what it is in the Jaala-Iitti region.

In the light of the foregoing discussion, the diabase-like rock of the Jaala-Iitti region represents in its entirety a magmatic rock differentiated from hornblende rapakivi magma. One must not, however, forget what Kaitaro (1956) wrote in elucidating the similarity between granites and lamprophyres occurring in association with them: »To discuss the petrogenetic development as a whole, it is necessary to know the nature of parent magma, or generally speaking parent material, and the character of the differentiation process. The differentiation of these rocks, especially the basic rocks with a high content of alkalies, cannot be very simple. By normal crystallization and gravitational differentiation it is not possible to bring about such products.»

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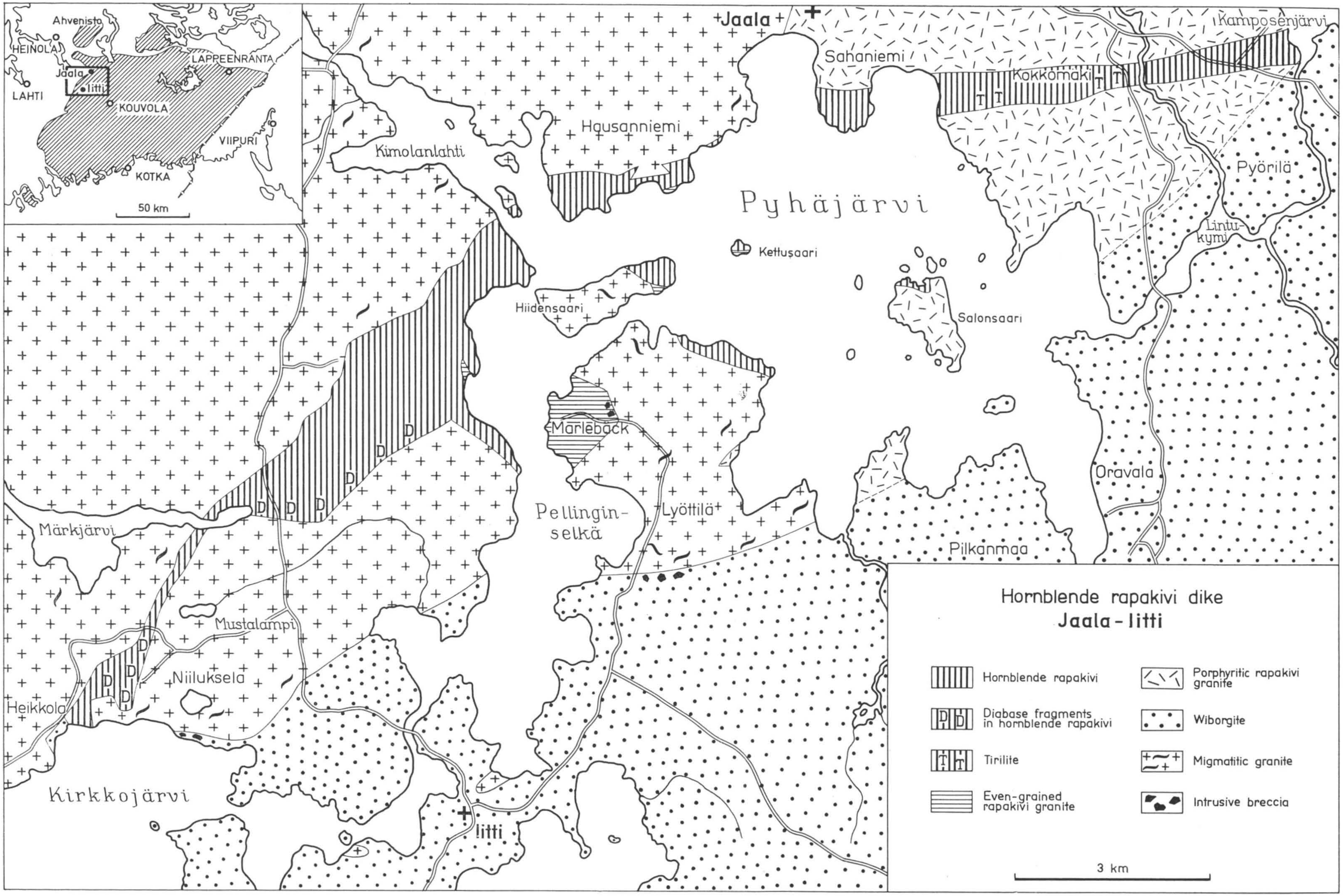

Mauno Lehijärvi and Anssi Lonka: The hornblende rapakivi dike . . .

6

A NOTE ON EUCLASE FROM MUIANE MINE, ALTO LIGONHA, MOZAMBIQUE [1]

BY

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On a visit in 1963 to the Muiane beryl producing pegmatite mine in Alto Ligonha, Zambezia, Mozambique, one of us (Th. G. S.) received from the manager of the mine, Mr. Eurico Lopes da Silva, two crystals of a colorless transparent mineral for identification. Later in the laboratory, the mineral proved to be euclase that hitherto was not known from that pegmatite occurrence nor has been mentioned by Behier (1957) from any other pegmatite locality in Mozambique. For that reason it seems appropriate to give a brief note on the occurrence of euclase in the Muiane pegmatite.

Only the two crystals mentioned have been found in Muiane so far. The smaller crystal having been used up for study, the larger one weighing ca. 20 g is all that is available for reference. The monoclinic crystal with the faces (010), (210), (120), (111) and (211) exhibits a perfect cleavage parallel to (010). A cross-section of the crystal with the optical orientation is reproduced in Fig. 1. The optic axial plane is parallel to (010). The refractive indices measured with the immersion method are: $\alpha = 1.652$, $\beta = 1.657$, $\gamma = 1.674$. For determining β that is perpendicular to the cleavage, a cleavage flake was mounted on a glass rod rotatable

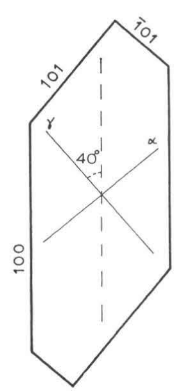

Fig. 1.

[1]) Received May 25, 1964.

in the immersion liquid. The optic axial angle was measured from the interference figure with the result $2V\gamma = 50°$. The specific gravity, determined by suspending the crystal in water, was found to be 3.08.

The powder pattern that will not be reproduced here is virtually indentical with that reported by Mrose and v. Knorring (1959). The following unit cell dimensions were determined from precession and Weissenberg photographs: $a_0 = 4.62 \pm 0.02$ Å, $b_0 = 14.31 \pm 0.04$ Å, $c_0 = 4.77 \pm 0.02$ Å, $\beta = 100°$ 15' $\pm$ 5'. Space group P $2_1/c$. These data agree with those given for euclase by Mrose and Appleman (1962) except for an interchange of the a- and c-axes.

A chemical analysis of the mineral (analyst: O. v. K.) gave the following result:

SiO_2	41.20 %	MnO	none
TiO_2	none	MgO	none
Al_2O_3	35.00	CaO	none
BeO	17.25	H_2O+	6.40
FeO, Fe_2O_3	none	H_2O-	0.09
		Total	99.94

The composition of the mineral corresponds very closely to the ideal formula of euclase, *viz.* $AlBeSiO_4(OH)$. In addition, Mr. A. Löfgren, of the Geological Survey Finland, very kindly determined the gallium and germanium contents of the mineral by means of optical spectroscopy. The following results were reported by him: Ga_2O_3 0.013 % and GeO_2 0.026 %.

The authors are indebted to Mr. Lopes da Silva for the euclase material and to Mr. Löfgren for the spectroscopic determinations.

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7

OBSERVATIONS ON THE JYVÄSKYLÄ ICE-MARGINAL FORMATION IN CENTRAL FINLAND [1]

BY

REINO REPO

Geological Survey of Finland, Otaniemi

ABSTRACT

The observations on the Jyväskylä ice-marginal formation deal with the relationship between the location of the formation, the regional topography, and the directions of flow of the ice sheets. Pollen and diatom analyses have also been used to elucidate the initial conditions of the formation.

CONTENTS

INTRODUCTION

The retreat of the continental ice sheet within Quaternary Finiglacial time across South and Central Finland toward the northwest was interrupted, causing both annual and more prolonged regional standstills of the ice margin.

[1]) Received, May 30, 1964.

During these periods the ice flow generally compensated the melting at the margin. The material transported by the ice was able to accumulate in front of the margin, thus showing that the location of the ice margin then prevailing was comparatively stabile.

The best known of the marginal formations of Finland are the extensive Salpausselkä systems I and II, whose continuations extend across Central Sweden to the Oslo region in Norway. The Salpausselkä formations in Finland have been studied not only geologically and geographically, but also from the angles of other natural sciences. Well known although much shorter marginal formations are Jaamankangas and Selkäkangas in East Finland and Hämeenkangas in the western part of the country. The subject of the present study, the Jyväskylä marginal formation, is best classified with the latter. The formation begins in the commune of Muurame (partly Korpilahti) in the southwestern part of Central Finland. With occasional breaks it runs NNE through the town of Jyväskylä to the commune of Laukaa where it peters out. The length of this Muurame—Jyväskylä—Laukaa marginal formation is some 60 km.

Brief references to this marginal formation of Central Finland have been made in various connections, *e.g.* in explanations to map sheets (Frosterus 1911, Brander 1934) and in chronological and other general investigations (Sauramo 1923, 1929 among others). More comprehensive studies of the morphology and structure of the formation were made by Saksela (1930) and Leiviskä (1951). These present observations of mine were mainly made on numerous new sections exposed in the town of Jyväskylä and its surroundings as far as Vihtavuori to the north-northeast.

ON THE MOVEMENTS OF THE CONTINENTAL ICE SHEET

The research area is topographically a peneplain, as is most of Finland. It is noteworthy, however, that the elevations around Korpilahti, Muurame and Jyväskylä are for the most part distinctly higher than those in the surroundings (*cf.* Saksela 1929, the topographical map). Although the elevations generally slope gently, some bedrock hills are still rather steep. Particularly around Jyväskylä a number of elevations rise to 200 — ca. 250 m above sea level. Especially the great lake basins east of the marginal formation display a considerable range of altitudes. These include Leppävesi 80.7 m (altitude of niveau), Lievestuoreenjärvi 84.5 m, and Päijänne (N end near Jyväskylä) 78.2 m. Among these, Päijänne is one of the deepest lakes in the lake district of Finland (the deepest location 96 m at the border between Korpilahti and Muurame) and at the same time the second greatest in area (1090 km²) of our lakes. No doubt the rather high bedrock hills and the

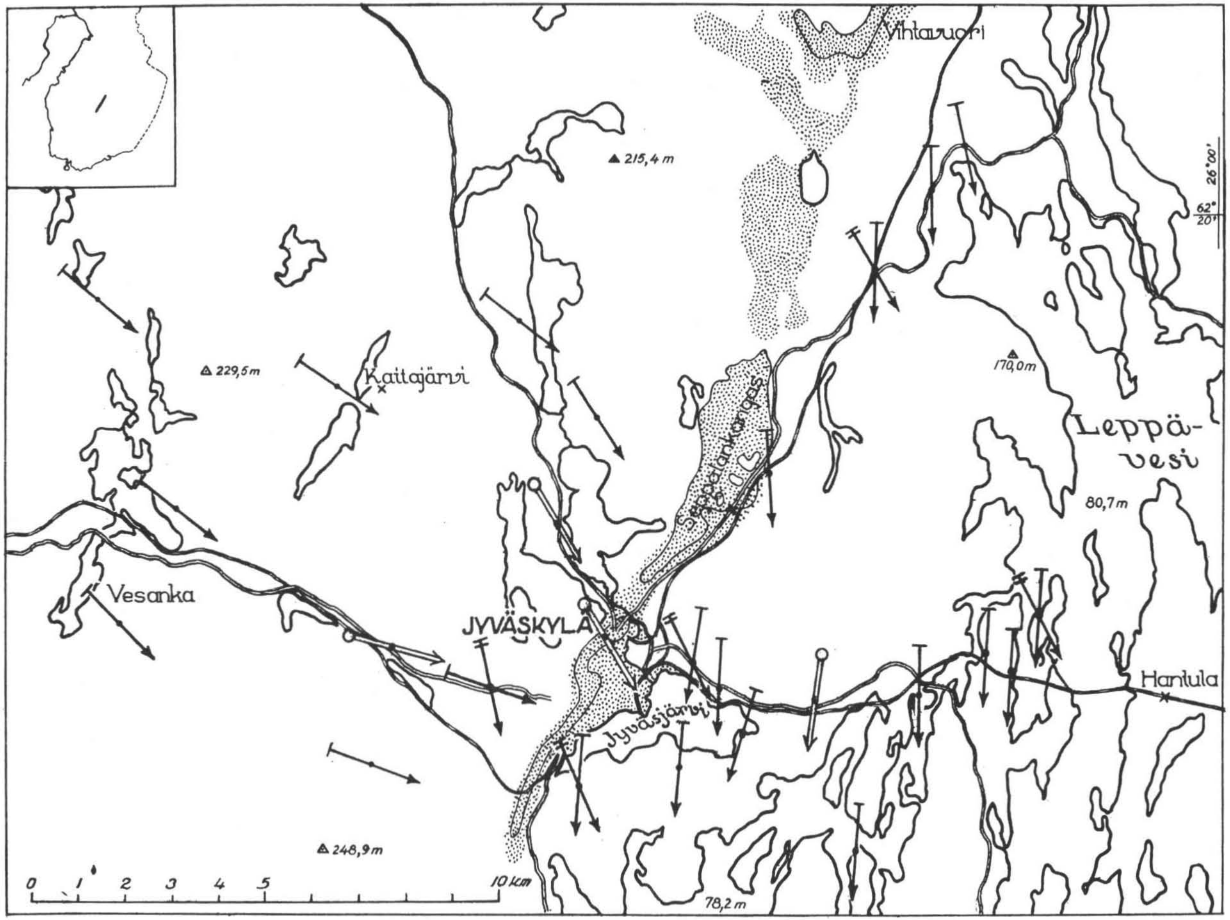

Fig. 1. The location (shaded areas) of the Jyväskylä marginal formation. Striae indicated with black arrows. At the sampling points the prevailing younger striations are marked with one and the older ones with two crosslines. Other arrows correspond to the main axis of the stones in the till.

depths of the nearby lake basins greatly influenced the directions of movement of the continental ice sheet and the location of the margin, specifically during the retreat phase when the ice sheet was already considerably thinned.

Glasial erosion features are presented in the map (Fig. 1). These are mainly based on measurements made by the author. Striation maps showing the directions of movement of the continental ice sheet over larger areas have been presented by Frosterus (1911), Brander (1934) and Saksela (1949).

The prevailing N-S-directed striation system east of the marginal formation is probably in part older than the system west of the formation, which runs roughly from NW to SE. This is attested among other things by the fact that the northerly direction dominates only east of the formation, *i.e.* in the outer area which was earlier released from the ice. The ice movement indicated by the northerly system of striae was evidently determined by the Päijänne basin. According to the glacigenous marks west of the marginal formation, the ice oscillated strongly during the final phase of retreat in the area.

The oldest direction shown in Fig. 1 is from the north-northwest. The ice flow with this trend seems not to have been influenced by the Päijänne basin or the topography of the surroundings. In all probability the ice sheet was at that time comparatively thick.

The Jyväskylä marginal formation is located at the border between the two dominating flow directions during the retreat phase of the ice, and runs west of the Päijänne and Leppävesi basins, and parallel to them. The marginal formation originated in a crevasse formed when the thinning continental ice sheet, owing chiefly to topographical causes, broke at the site. Initially the lobe of the ice sheet east of the crevasse evidently remained connected with the continental ice sheet at its northern end and was still active, at least in its deeper portions.

At first glaciofluvial material was transported to the crevasse, but, as is seen from Fig.2, the bulk of the deposit is structurally not like that of a crevasse formation but resembles a marginal one. The eastern part of the marginal formation was released from the ice during the final phase when the ice sheet still extended to it in the west. At that time the youngest local striae originated perpendicularly to the marginal formation. As was mentioned earlier and as will be seen later from the data to be presented, before its final retreat from the site the ice sheet still oscillated above the marginal formation as far as the distal part. According to the till fabric analysis made from the till sheet at the top of the marginal formation this movement came roughly from the northwest.

In Ostrobothnia, west of the Jyväskylä region, the old westerly or west-northwesterly striations occur (*cf. e.g.* Hyyppä 1948, Mölder & Salmi 1954). This older direction has not been observed in the Jyväskylä region.

ON THE STRATIGRAPHY OF THE MARGINAL FORMATION

GENERAL SURVEY OF THE MUTUAL PROPORTIONS BETWEEN THE FLUVIOGLACIAL MATERIAL, VARVED SEDIMENTS AND TILL

The Jyväskylä marginal formation rises some 40 m above its surroundings. Within the town limits the highest ridge reaches 144 m above sea level. Seppälänkangas, northeast of the town, forms a plateau 1—2 km broad at an altitude of ca. 150—160 m.

Sections exposed by the excavation of gravel and sand occur in various places along the Jyväskylä marginal formation, in both the proximal and distal parts. They show that the bulk of the deposit is of fluvioglacial gravel and sand. Till has frequently been met with in the summit areas of the formation; in places it is distinct and wide, whilst in places there are lesser occurrences. The thickness of the till sheet is usually less than 2 m, but may exceed this figure on the slopes.

Saksela (1930) drew attention to the fairly abundant occurrence of till in certain parts of the marginal formation. Further excavations carried out later have made it possible to follow the mutual location of fluvioglacial and till material. With the aid of data from most of the observation points their distribution can be presented schematically (Fig. 2). Further details of these phenomena can be seen in photographs (Fig. 3).

The structure of the till varies considerably. In places it is of the normal basal type, in places more or less washed. The original structure is in many places distorted and the till was mixed with coarser fluvioglacial sediments during the movements. According to particle size analyses, the clay content of the till (Fig. 5) is ca. 1—7 %. The clay content of washed till is usually lower than that of typical till. Particularly in the sections of Seppälänkangas the biggest boulders and cobbles seem to be concentrated in the upper and surface parts of the marginal formation. Quite often this material represents the remains of strongly washed till.

The schematic section at the right in Fig. 2 shows the distinct SE slope of the bedding of the fluvioglacial material. This shows that during deposition the flow was away from the continental ice sheet, radially, toward the basin

Fig. 2. Schematic profiles showing the location of the till in the Jyväskylä marginal formation. Movement of the ice sheet occurred in the NW-SE direction.

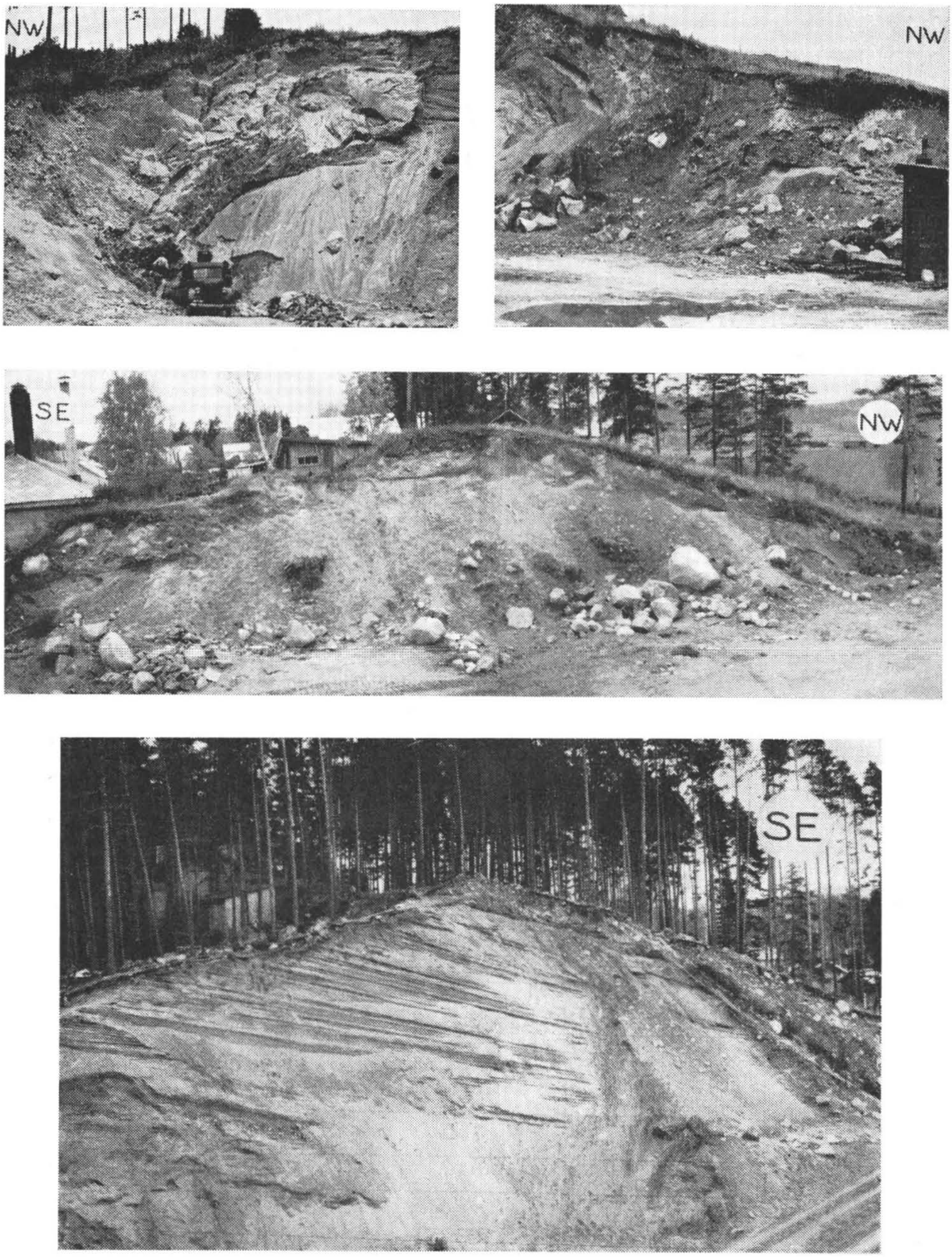

Fig. 3. Sections of the marginal formation in the town of Jyväskylä. Schematic profiles showing the location of the till sheets above the fluvioglacial beds are presented in Fig. 2. The left top picture shows a sorted gravel sand bed (diagonally above the truck) with maximum thickness 7 m above the till sheet inclined parallel to the slope.

of the ancient Päijänne. Since the distal part is covered by till, the ice margin, during the final phase of sedimentation, must have oscillated over the marginal formation in places.

The varved clay-silt sediments of the Jyväskylä marginal formation occur as incoherent, areally restricted deposits when compared with the till sheet. In my opinion, they were deposited on the whole simultaneously with the onthrusting till. In two places in the proximal part (the Nisula section) varved sediment has been found as lenses deeper down. These lenses are ca. 0.5 m and 1 m thick at the center and occur at depths of some 20 m and 10 m inside the fluvioglacial gravel sand. It is thus also possible to estimate the originating phase to a certain extent on the basis of the microfossils of the clay-silt sediments.

Brenner's description (1927) of the till section at Hantula offers a significant comparison for the clarification of the stratigraphy of the Jyväskylä marginal formation. The till accumulation, 28 m thick, is the thickest found so far in Finland, and covers a varved clay sediment some 2 m thick. No observations were made on the microfossiles of the deposit. The varved clay comprises at least 100 annual varves (the uppermost ones were probably worn away during the later advance of the ice sheet, and thus the time interval for sedimentation before the advance must have exceeded 100 years). According to the map of the retreat of the ice margin by Sauramo (1923), the margin had taken some 100 years to retreat from Hantula to the west of the Jyväskylä marginal formation. West of Jyväskylä sand layers under till as well as till lenses inside sand have also been found (Frosterus, Brenner). According to Brenner, the abundance of fine material in the till of the Jyväskylä region would point to interglacial fine-grained sediments as the source of the till. According to Brenner, the varved clay of Hantula originated during the retreat phase of the ice sheet of the last interglacial period. He based his opinion on the thickness (28 m) of the overlying till, the accumulation of which must, so he thought, have taken a considerable time. In contrast to Brenner, I believe that the thickness of the till is due more to local conditions. The Hantula till may well have accumulated at the retreat phase of the latest glaciation. Located at it is, the interglacial or interstadial deposit could not have withstood the subsequent advance of the ice sheet but would have survived the oscillation which occurred locally in deep water during the retreat phase. From the fact that the till accumulation is thicker than usual it is also therefore possible that Hantula belongs to the same zone as some marginal moraines, according to Leiviskä (1951). The striae however do not lend support to the view that a north-south marginal formation was located at Hantula. On the contrary, there seems to be a radially directed moraine formation that originated simultaneously with the northern striation. This striation is younger than the older direction of the ice flow in the region from

Fig. 4. A close-up of a till bed in the distal part of the marginal formation. Fluvioglacial sand imbedded in the lower part of the till indicates thrusting of the ice sheet.

the north-northwest. Had information on the microfossils of the clay beds been available, it would have been possible to compare them with the microfossils of the varved clay of Jyväskylä and thus estimate the age relations of these till sheets.

According to the striae and the eastermost till fabric analysis (Fig. 1), it seems plausible that the latest ice flow occurred from north to south in the direction of the Päijänne-Leppävesi basin. It is probable that the till above the Hantula clay accumulated during the oscillation of the ice sheet while the climate was turning colder. On the other hand, the till sheet above the Jyväskylä marginal formation in all probability accumulated in connection with the simultaneous ice flow from the northwest.

ON THE POLLEN COMPOSITION OF THE CLAY AND SILT LAYERS

To clarify the initial conditions of the varved layers of the Jyväskylä marginal formation the standard profile was taken west of the marginal formation in the village of Palokka, east of Kaitajärvi. The layer series taken from a slope section is located some 7 km northwest of the Jyväskylä mar-

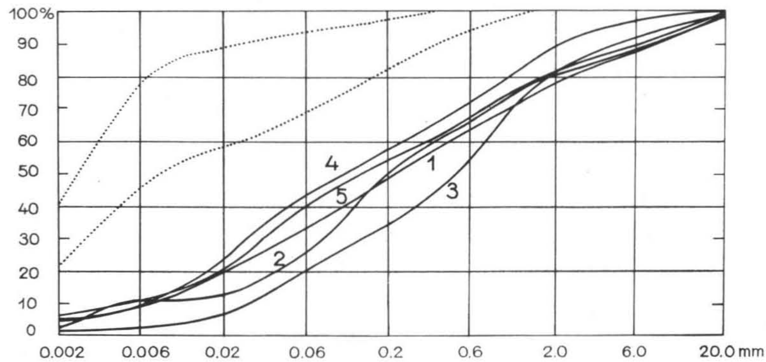

Fig. 5. Size analyses of the till (1—5) in the area investigated and of the varved lenses of the marginal formation. Grain size curve 1 is from the west of the Jyväskylä marginal formation, curve 2 is taken east of the formation and curves 3—5 are from the till sheets of the marginal formation. The upper dotted line is from the clay lens at 10 m depth, the lower dotted line at 20 m depth from the silt lens.

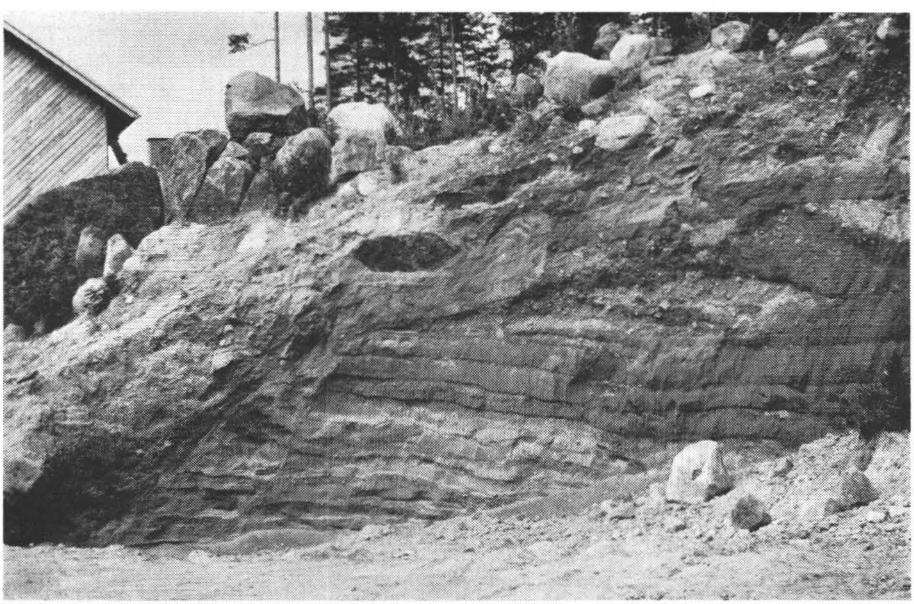

Fig. 6. Layer sequence from the proximal part of the marginal formation (Seppälänkangas, S part). The repeated layers of gravel and sand below were in all probability deposited close to the ice margin. Regular alternations of the grain size of the sediment result from an annual rhythm of melting. The uppermost till cutting the strata marks the last local oscillation of the ice margin.

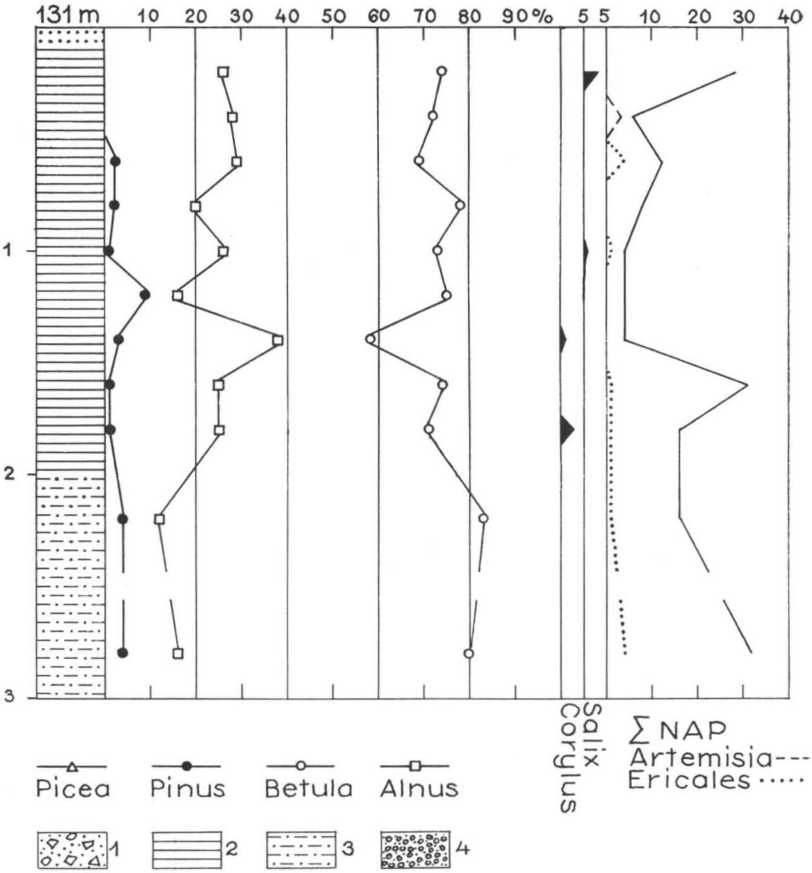

Fig. 7. Pollen diagram of Kaitajärvi. Explanations: 1 = till, 2 = varved clay, 3 = varved silt, 4 = fluvioglacial gravel and sand.

ginal formation. The layer series comprises varved silt at a depth of 2—3 m, above the stratified fine sand, the varved silt-clay at the depth of 2—0 m, and uppermost 10 cm of sand (pollen diagram 1, Fig. 7).

The pollen frequency proved to be very low in the layer series. Only a few samples yielded 100 arboreal pollen grains. The pollen flora type of the series is throughout dominated by deciduous trees, with a maximum of *Betula* pollen and a high percentage of *Alnus*. Some *Corylus* pollen found in clayey strata further augments the evidence that the flora type was dominated by deciduous pollen. Of the coniferous pollen *Pinus* alone is represented by less than 10 %. At its maximum at 2.8 m depth the NAP is composed of *Juniperus, Polemonium, Ericales,* and *Cyperaceae* pollen. A peculiar feature of the NAP is the fairly general abundance of *Ericales (Ericaceae* and *Empetrum)*

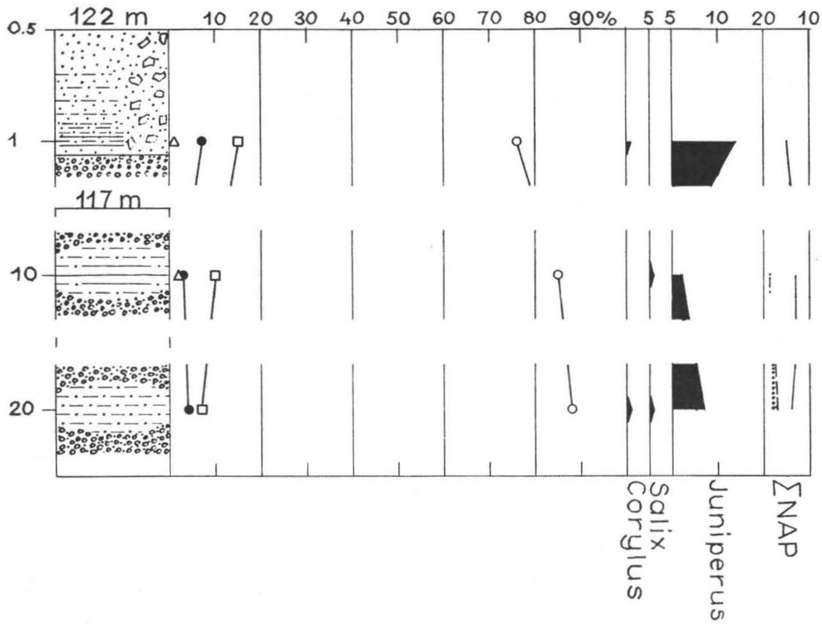

Fig. 8. Pollen diagram from the varved clay of the surface part of the Jyväskylä marginal formation, and from the varved clay and silt lenses imbedded in the fluvioglacial layers.

and the scarcity of *Artemisia* except at a single site in the upper part of the clayey strata.

The flora type presented above comprises pollen transported long distances by air and running water as well as from the nearer land areas. The sediment was deposited, according to the pollen flora, during a comparatively humid Preboreal phase when the vegetation was still rather open. This is shown, in my opinion, by the high maximun of *Betula* and proportionally small amount of NAP. Furthermore, the composition of the NAP, *e.g.* the occurrence of *Ericales*, *Polemonium*, *Cyperaceae* and *Juniperus* and the scarcity of *Artemisia*, is in good agreement with the Preboreal flora type.

The pollen proportions of the varved sediments of the marginal formation are shown in Fig. 8, pollen diagram 2. The pollen frequency in these samples was comparatively low, but 100 arboreal pollen grains could be counted from each sample. The pollen composition from the two thin lenticular clay-silt strata located comparatively deep in the marginal formation proved greatly to resemble the one presented in the Kaitajärvi profile above. Evidently the composition of the pollen flora caught up in sediments did not change while the ice margin retreated from the marginal formation to the east of Kaitajärvi.

As the sediments in question are inorganic, their exact dating by the pollen method is not possible. Thus the pollen from the inorganic sediments described in the diagrams might belong either to the Preboreal or to the final phase of the Younger Dryas. For the reasons presented below, however, these sediments can be assumed to be Preboreal.

When the pollen spectrum presented above is compared with the pollen diagrams from the Onttola region, Northern Karelia (Repo 1963), there are considerable differences east of the Jyväskylä marginal formation except in the upper part of the Onttola profile 3 (2 uppermost analyses). The majority of the varved sediments from Onttola display quite NAP-dominated pollen, peculiar to the Younger Dryas. In the upper part of the Onttola profile 3, however, the pollen flora is of the same type as in the profiles described from the Jyväskylä region. The distinct differences between the pollen floras of the lower parts of the varved sediments in the two regions, *i.e.* in Northern Karelia and Central Finland, are significant from the dating point of view. They suggest that in the Jyväskylä region the retreat of the ice sheet was not followed by a tundra vegatation, at least not on a scale to be evidenced by the pollen flora. According to Donner (1958), within the initial phase of the Finiglacial retreat of the ice sheet margin there followed a narrow tundra zone. It is probable, in the case of the Jyväskylä region, that the pollen flora imbedded in the sediments in the vicinity of the ice margin reflects the climatic and vegetational conditions of a wider area rather than any local floral successions. The pollen datings of the varved sediments of the Jyväskylä and Jaamankangas marginal formations agree well with the development of the vegetation based on pollen research (Firbas 1949, Sauramo 1958 and others). The results demonstrate the significance of the pollen method in the dating of inorganic sediments also.

According to the study of varve chronology by Sauramo (1923), the margin of the ice sheet reached the Jyväskylä region about 500 years after it retreated from the Salpausselkä II. This coincided with the early Preboreal. The varve chronology by Sauramo (1929) can be connected with the C^{14} dating of the Preboreal deposits from Lapaneva bog in Kihniö, presented by Salmi (1962). Lapaneva is located at about the latitude of Jyväskylä and some 130 km to the west. The Lapaneva district was released from the ice sheet approximately 750—800 years later than Salpausselkä II, according to the map of the retreat of the ice margin by Sauramo (1929), corresponding to ca. year 7630 B. C. (Sauramo 1958, p. 445). According to Salmi, the oldest preboreal organic layers are dated to ca. 7900 $\pm$ 320 B. C. On the basis of the locations of the ice margin, as presented by Sauramo, the retreat occurred some 200—250 years earlier in Jyväskylä than in Lapaneva. The C^{14} dating above is somewhat older than the dating derived from the varve chronology by Sauramo. I would believe the C^{14} dating to be more correct within the error limits than the varve chronology by Sauramo in the area in question. According to the correction of the latter chronology by Hyyppä (1963), Sauramo's zero year would correspond to 8413 B. C. On this basis there is a much better agreement between the varve chronology and C^{14} dating. In any case, according to the stratigraphical observations the varved clay-silt lenses imbedded in the Jyväskylä marginal formation were deposited in the vicinity of the ice sheet margin, whilst the pollen flora shows that this bedding occurred in Preboreal time.

OBSERVATIONS ON THE DIATOMS OF THE SEDIMENTS

The diatom analyses are from the same sampling points as the pollen analyses above. The first sample is from the Kaitajärvi profile at the depth of 1 m (diatom diagram 1, analysis point 130 m above sea level). The following three samples are from the Jyväskylä marginal formation. The uppermost one was taken at a depth of 1 m from the varved clay-silt sediment at the top of the section (pollen analysis 2, altitude ca. 122 m). The other two samples are from the clay-silt lenses in the nearby gravel pit (Nisula) at depths of 10 m and 20 m (diatom analyses 3 and 4, the altitude of the pit top 117 m). The diatoms are practically the same in all samples. *Melosira islandica* ssp. *helvetica* dominates in three samples, and in one is the next in abundance. In addition to this plankton diatom peculiar to big waters, the plankton form *Melosira granulata* and the littoral forms *M. distans* and *Pinnularia* sp. (fragments) are also quite common. Compared with the fresh-water forms, the salt-water forms are of secondary importance. The salt-water group is composed both of forms at present common in the Baltic and of some rarer individuals such as *Endictya oceanica* and *Pyxidicula mediterranea* (Mölder 1962 among others). The diatom composition in Table I is in all probability primary belonging to the marine phase.

The main diatom composition refers to the Yoldia Sea at the margin of the ice sheet, influenced by the meltwaters. The varved structure of the sediments also bears witness to the proximity of the ice margin. From the nearby continental ice sheet coarser material accumulated in the marginal formation in summer and silt in winter. In these wintry conditions the finer material was able to accumulate even in the upper parts of the marginal ridge. When silt-bearing sediments accumulated close to the ice margin, the flow of meltwaters must have been comparatively slow. In other words, one must assume a temporary cooling of the climate. This assumption is sustained, in the case of the Jyväskylä marginal formation, by the till sheets associated with the higher located varved sediments.

The proportional abundance of the saline and planktonic diatom forms shows that the Yoldia Sea phase in Central Finland was directly connected

Table I

s = salt-water forms b = brackish-water forms f-b = fresh- and brackish-water forms = fresh-water forms	Ecological group	K. 131 m 1 m	Jyväskylä 122 m 1 m	Jyväskylä 117 m 10 m	Jyväskylä 117 m 20 m
Amphora ovalis	f	1	—	—	—
— var. *pediculus*	»	1	—	—	—
Coscinodiscus radiatus	s	1	—	—	—
C. sp. fragm.	»	—	1	3	16
Cymbella aspera	f	1	—	—	—
C. austriaca	»	—	—	—	2
C. sp. fragm.	»	—	1	—	—
Didymosphenia geminata	»	—	—	1	—
Endictya oceanica	s	3	—	—	—
Epithemia sp. fragm.	f-b	—	—	—	2
E. turgida	»	1	1	3	—
E. zebra	»	—	1	—	—
Eunotia papilio	f	1	—	—	—
E. pectinalis	»	—	—	2	—
E. sp. fragm.	»	1	2	—	—
Fragilaria pinnata	f-b	1	—	—	—
Grammatophora marina	s	—	—	—	2
G. oceanica	»	—	9	21	2
Gyrosigma kützingii	f	—	—	1	—
Melosira ambigua	»	1	—	—	—
M. arenaria	»	3	—	1	—
M. distans	»	16	6	17	16
— var. *lirata*	»	1	—	—	2
M. granulata	»	16	11	20	10
— var. *angustissima*	»	13	—	—	—
M. islandica ssp. *helvetica*	»	20	32	19	36
M. italica	»	—	—	1	—
M. westii	s	—	—	1	—
Navicula sp. fragm.	f	—	1	—	—
Pinnularia sp. fragm.	»	9	21	8	8
Pyxidicula mediterranea	s	—	2	—	—
Rhabdonema arcuatum	»	2	2	1	2
Stauroneis phoenicenteron	f	—	—	1	—
Stephanodiscus dubius	f-b	3	—	—	—
Stephanopyxis turris var. *arctica*	s	1	—	—	—
S. turris var. *polaris*	»	3	—	—	—
Synedra tabulata	b	—	3	—	—
Thalassionema nitzschioides	s	—	1	—	—
Thalassiosira baltica	b	1	6	—	2
salt-water forms		10	15	26	22
brackish-water forms		1	9	—	2
fresh- and brackish-water forms		5	2	3	2
fresh-water forms		84	74	71	74
		100	100	100	100%

with the more saline central parts of the Baltic Sea. It implies a comparatively high level of the Yoldia Sea phase in Central Finland. Investigations on the shores of the Yoldia Sea phase in question have been carried out by Ramsay (1917, 1927), Tolvanen (1922), Sauramo (1958) and Hyyppä (1963), among others. According to Sauramo, the highest Yoldia shore (Yoldia I) is around 155 m in Jyväskylä. According to the isobase map and shore diagram by Hyyppä, the shore of Yoldia I in Kuhmo, which lies on approximately the same isobase as Jyväskylä, is at the elevation of 160 m. The dating of this shore line is approximately 8200 B. C.

SUMMARY

The broken marginal formation sequence through Jyväskylä runs roughly in a SSW-NNE direction. In the west, esker sequences are associated with it almost perpendicularly. The marginal formation is located, according to striation observations, at the border between two ice flow movements which occurred during the retreat phase of the ice sheet. On the eastern side the north-south flow parallel to the Päijänne basin and influenced by it was dominant. On the western side of the marginal formation the ice flow ran perpendicularly to it.

The marginal formation originated in a crevasse in the thinning ice. The stratification of the fluvioglacial sediments parallel to the distal direction proves that at the time of deposition the area immediately to the east of the formation was ice-free. The location of the ice margin did not shift during the melting phase, because the ice flow at right angles to the marginal formation compensated melting.

The varved »deep water» sediments imbedded in the marginal formation and on its surface afford evidence that the sedimentation conditions were such as are found in fairly deep water. Pollen dating points to the preboreal sedimentation of these layers which, according to the diatom flora, took place within the Yoldia phase of the Baltic Sea. At the same the ice margin remained fairly close to the marginal formation and later oscillated above it from the northwest. The oscillation reached at least as far as the distal part. Simultaneously east of the marginal formation the ice sheet correspondingly oscillated in the north-south direction to the Hantula area.

The oscillations of the ice sheet within the Yoldia period are associated with the temporary cooling of the climate during the Preboreal phase after the main sedimentation of the marginal formation had already taken place.

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A DIOPSIDE FROM THE NORDRE STRÖMFJORD AREA, WEST GREENLAND [1]

BY

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ABSTRACT

This article describes the petrography and paragenetic properties of two samples collected from Precambrian dyke-like occurrences in West Greenland. One of the samples is characterized by the presence of association diopside-spinel-anorthite. The composition of the obviously pure $MgCa(SiO_3)_2$ has been estimated with the aid of X-ray optical data.

CONTENTS

INTRODUCTION

The samples which are the subject of this study were collected by the author from a pegmatite-like occurrence situated on the north shore of Nordre Strömfjord, at Akuliaruserssuaq at approximately lat. 67°17′N, long. 52°50′W. The occurrence was briefly visited by boat from an anchorage close to the south shore of the fjord which was held for one night during the

[1]) Received June 6, 1964

»Helicopter Expedition» in summer 1962. The expedition was organized by the Kryolitselskabet Øresund A/S, Copenhagen.

The first view of the occurrence in the field gives an impression of the usual light-colored feldspar pegmatites, which form a system of several parallel dyke-like bodies visible for a length of more than a kilometer. The width of the individual dykes varies with the limits of 0.5—5 meters in the locality visited. The prevaling rocks in the area are Precambrian granulite gneisses (*cf.* Noe-Nygaard and Ramberg, 1961) showing a schistosity in the direction WSW-ENE. The system of dykes conforms to this direction.

Though the formation did not arouse any special interest as such, attention was excited by the abundance of a bluish spinel in one of the dykes and by the presence of a pinkish component in another dyke. Hence, representative sample of each dyke was taken for later mineralogical determinations.

A preliminary microscopic study of the sample containing the bluish spinel revealed that the rock itself is composed chiefly of a colorless diopsidic material. This mineral, which forms the main subject of the present paper, has been subjected to closer study with microscopic and X-ray powder methods. For the last-mentioned determinations the procedure developed by Zwaan (1954) has been applied to compositional estimations of the diopside studied. The naturally occurring pure diopside analyzed by Juurinen and Hytönen (1952) from Juva, Finland, has been used for reference. The röntgenographic properties of this diopside, needed for the present purpose, were determined by the author. It should be mentioned that in the course of later study the pinkish mineral of the other dyke turned out to be composed of micaceous material.

Because no systematic and detailed geological survey was carried out in the locality of sampling, the origin of the dyke-like bodies will be discussed only tentatively.

ON THE PETROGRAPHY OF THE SAMPLES

Sample 1.

Macroscopically, the specimen of diopside rock is medium-grained, grayish-white in color and contains relatively evenly distributed grains of bluish spinel. The latter are 1—10 mm in diameter.

In thin section only diopside, spinel and calcic plagioclase (An_{80}) were identified. Small grains of diopside occur as irregularly shaped inclusions in the spinel. The plagioclase occurs only as small, rounded inclusions in the poikilitic diopside host. The spinel is probably pleonast, with an $a_0 = 8.108$ Å, as was revealed by an X-ray powder run. For this determination the reflections (311), (400) and (511) were used.

Sample 2.

This sample, characterized by the presence of pinkish, micaceous material, shows a mineral composition clearly deviating from that of the first sample. The main components are mica, carbonaceous material and diopside, in this order of abundance. The bluish spinel occurs sparsely as small grains. An optically positive uniaxial mineral, so far unidentified, has been observed occasionally as minute grains in the carbonates. The mineral shows birefringence clearly greater than that of quartz. The pinkish material, which according to X-ray powder determination belongs to the group of di-octahedral micas, is the dominant component. In places it is followed by some flakes of flogopite.

THE OPTICAL AND X-RAY POWDER DATA OF THE DIOPSIDE IN SAMPLE 1.

Sample 1.

The refractive index determinations have been made from crushed and sieved material. The diopside for X-ray study was drilled directly from the sample under a binocular microscope. The runs were performed with an X-ray spectrometer using CuK_{α_1}-radiation and a Ni filter. In all runs a silicon internal standard was applied.

Besides the diopside from Nordre Strömfjord, the diopside from Juva was examined with X-rays as well. From the last-mentioned locality the pure material used for analysis by Juurinen and Hytönen (1952) was kindly offered for the present purpose by Th. G. Sahama, Ph. D., Department of Geology and Mineralogy, the University of Helsinki. The d-spacings of both diopsides studied are shown in Table 1, including also the corresponding data determined from diopside at Zillertal, Austria (Zwaan, 1954).

The optical data of the diopside from Nordre Strömfjord are presented in Table 2, which also contains the optical data reported from diopsides at Juva (Juurinen and Hytönen, *op. cit.*) and at Zillertal (Zwaan, *op. cit*).

The d-values are very close in all three cases, as is seen from Table 1, and the similarity of values belonging to specimens from Nordre Strömfjord and Juva is conspicuous.

The $Z \wedge c$ reported for the diopsides at Juva and Zillertal approximates to the value 38.5° generally accepted for a pure diopsidic composition (*cf.* Winchell and Winchell, 1951). $Z \wedge c$ for the diopside of Nordre Strömfjord was determined as 33° and is thus clearly smaller than the value usually found. By contrast, $2V_z = 61°$ for the Greenlandic diopside corresponds to the value assigned to a pyroxene of rather pure diopsidic composition. This is also shown by the values determined in the two other instances considered here.

Table 1. d-spacings of Diopsides from Nordre Strömfjord, Zillertal (Zwaan, 1954) and Juva (Juurinen and Hytönen, 1952)

hkl	Nordre Strömfjord, Greenland	I	Zillertal Austria	I	Juva Finland	I
220	3.2200	s	3.23	m	3.2245	s
221	2.9838	s	2.994	vs	2.9838	s
310	2.9415	s	2.957	mw	2.9434	s
311	2.8895	s	2.896	sm	2.8877	s
131	2.5600	w	2.562	sm	2.5600	w
202, 002, $22\bar{1}$	2.5115	s	2.514	mw	2.5115	s
400	2.2973	w	2.298	mw	2.3018	w
$11\bar{2}$	2.1483	w	2.157	w	2.2035	w
331	2.1290	w	2.134	mw	2.1252	w
421	2.1026	w	2.111	w	2.1044	w
420	2.0377	w	2.035	mw	2.0412	w
332	1.8322	vw	1.835	w	1.8349	vw
150	1.7483	vw	1.754	mw	1.7527	vw
511	1.6243	s	1.624	s	1.6247	ms
442	1.5038	s	1.503	mw	1.5038	s

Table 2. Distances between the Reflections in Hundreths of Degrees [1]) and in Å [2]).

X-ray reflections used	Nordre Strömfjord, Greenland	Zillertal Austria	Juva Finland
Q (221)—Q (220)	1.13 [1])	—	1.14
d (220)—d (221)	0.241 [2])	0.238	0.241
Q ($22\bar{1}$)—Q (131)	0.36	—	0.35
d (131)—d ($22\bar{1}$)	0.050	0.048	0.049
Q (310)—Q (221)	0.23	—	0.21
d (221)—d (310)	0.044	0.037	0.040
Optical data			
$2V_z$	61	61	60
$Z \wedge c$	33	38.5	38
n_x	1.665	—	1.664
n_y	1.677	—	1.673
n_z	1.697	—	1.695

ON THE COMPOSITION OF THE DIOPSIDE STUDIED

Because no chemical analysis of the diopside under study has been available, an attempt has been made to evaluate the composition by means of optical and X-ray powder data. For this purpose special use has been made of the relative distance in degrees (θ) or in Å (d-values) of definite reflections

Table 3. Chemical Analyses of the Diopsides from Zillertal (Zwaan, 1954) and Juva (Juurinen and Hytönen, 1952).

	Zillertal, Austria	Juva, Finland
	Weight %	Weight %
SiO_2	53.42	54.66
TiO_2	tr.	0.00
Al_2O_3	0.24	0.00
P_2O_5	tr.	n.d.
Fe_2O_3	1.72	0.68
FeO	2.56	0.07
MgO	16.82	18.78
MnO	0.18	0.02
CaO	24.89	25.85
Na_2O	0.15	n.d.
K_2O	tr.	n.d.
H_2O+	0.06	0.22
H_2O-	0.00	0.00
	100.04	100.35

vs. the variation in composition, an application especially demonstrated for the pyroxene group by Zwaan (1954). Accordingly, the reflections (220), (221), (131), (310) and (22$\bar{1}$) have been determined most accurately and the differences calculated from the diopsides at Nordre Strömfjord and at Juva have been compared with those presented by Zwaan (*op. cit.*) for the diopside at Zillertal. All these results, with the optical data, are shown in Table 2. With regard to further compositional correlations the analyses of the diopsides at Juva and at Zillertal are shown in Table 3.

According to Table 2, the properties of the diopsides at Nordre Strömfjord and at Juva, as determined by the X-ray method, are very similar. Of the diopside at Zillertal only the differences of the d-spacings have been available (Zwaan, *op. cit.*). These values deviate slightly from the values determined for the other two specimens. The deviations, however, are within the limits of errors in determination.

In view of the similarity of the results arrived at it seems plausible that the composition of the diopside at Nordre Strömfjord is close to the compositions shown in Table 3 and especially to that of the Juva diopside. As pointed out by Zwaan (*op. cit.*), the method, as applied, requires a consideration of some important factors, which may essentially affect the results as follows:
— The substitution of Fe for Mg has relatively little influence on the unit cell dimensions or on the shape of the unit cell, whereas the substitution of Ca for Mg essentially affects the unit cell dimensions.
— The 2V depends on the shape of the unit cell, which changes greatly when Ca is substituted for Mg.

In agreement with the analytical results obtained from the diopsides at Juva and at Zillertal, it might be presumed that the iron content in the diopside at Nordre Strömfjord is relatively small or negligible. Besides the results of X-ray study this conclusion gains support from the facts that the mineral is colorless and is not pleochroic at all. On the basis of the values determined in the course of the present study there are no direct means to evaluate the content of aluminum. A comparison of the data shown in Table 2, however, makes it relatively safe to assume that even in this case the content of aluminum must be negligible. Consequently, the composition of the Greenlandic diopside would be fairly pure $CaMg(SiO_3)_2$ lacking aluminum and iron.

CONCLUDING REMARKS

As stated at the beginning of this article, both types of samples studied were collected from dyke-like formations occurring within an area of gneissose granulitic rocks. In spite of the lack of wieder field geological observations on the occurrences considered, some tentative speculations may be ventured with regard to the interpretation of the interesting petrographic features observed in the samples investigated.

In view of the paragenetic properties noted in the two samples collected, it is presumed that in the system of dyke-like formations met with reactions of the following character may have taken place:

$$(1)\quad \underset{\text{Anorthite}}{CaAl_2Si_2O_8} + MgO \rightarrow \underset{\text{Diopside}}{CaMg(SiO_3)_2} + \underset{\text{Spinel}}{MgAl_2O_4}$$

$$(2)\quad \underset{\text{Diopside}}{4CaMg(SiO_3)_2} + 2H_2O + 2CO_2 \rightarrow \underset{\text{Mica}}{Mg_6\,[Si_8O_{20}/(OH)_4]} + 2(Mg,\,Ca)CO_3$$

A process such as is represented by the first equation supposes reactions in relatively high P, T conditions and absence of hydrous and vaporous phases. The author is inclined to believe that the process, principally representing one of the amphibolite facies, occurred close to the P, T conditions of a granulite facies, which has determined the lithological properties of the surrounding basement formations as well.

A process such as is represented by the second equation is characterized by the presence of vaporous and hydrous phases and suggests reactions in considerably lower P, T conditions *i.e.* possibly those presumed to give rise to the greenschist facies.

As is revealed by the geological mapping carried out during later years, the anorthositic bodies and layers are widely distributed within the zones of gneissose rocks of Precambrian terrain in West Greenland. With regard to the genesis of these rocks, the investigations by Ramberg (1948), Sörensen (1955) and Berthelsen (1957) are especially pertinent. On the basis of the textural relations observed under the microscope it might be suggested that the pegmatite formations here considered were originally anorthositic in composition and underwent paragenetic changes by metamorphic-metasomatic processes as illustrated by equations 1 and 2. On the other hand, one may infer that the conformable system of dyke-like formations, as observed in the Nordre Strömfjord area, originally represented carbonate rich beds in a series of sediments reworked by metasomatic changes and pegmatite activity.

Acknowledgment — The author is indebted to the Kryolitselskabet Øresund A/S, Copenhagen, for permitting the publication of the data presented in this study. Likewise sincere thanks are due to Mrs. Toini Mikkola, M.Sc., of the Geological Survey of Finland, for the universal stage determination of the diopside studied.

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OBSERVATIONS ON THE NICKELIFEROUS ORE BLOCK OF NIEMILAHTI IN KANGASLAMPI AND ON SMALL ULTRABASIC AND BASIC BODIES IN THE RAUHAMÄKI — KURONLAHTI DISTRICT, FINLAND [1]

BY

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ABSTRACT

The samples of two nickeliferous ore blocks are described. Thereafter a quantity of small ultrabasic and basic bodies which occur in migmatite in the district named in the heading to the NW of the ore block, are presented, the optical properties of the minerals of the described rocks are shown, as well as their modal mineral compositions.

CONTENTS

[1]) Received August 1, 1964.

INTRODUCTION

In the autumn of 1959, the sisters Riitta and Reeta Räisänen found in Niemilahti an ore block containing considerable quantities of pentlandite. Niemilahti is situated in Pisamaniemi, about 5 km to the south of Kangaslampi village. As a result of the find, the Exploration Department of the

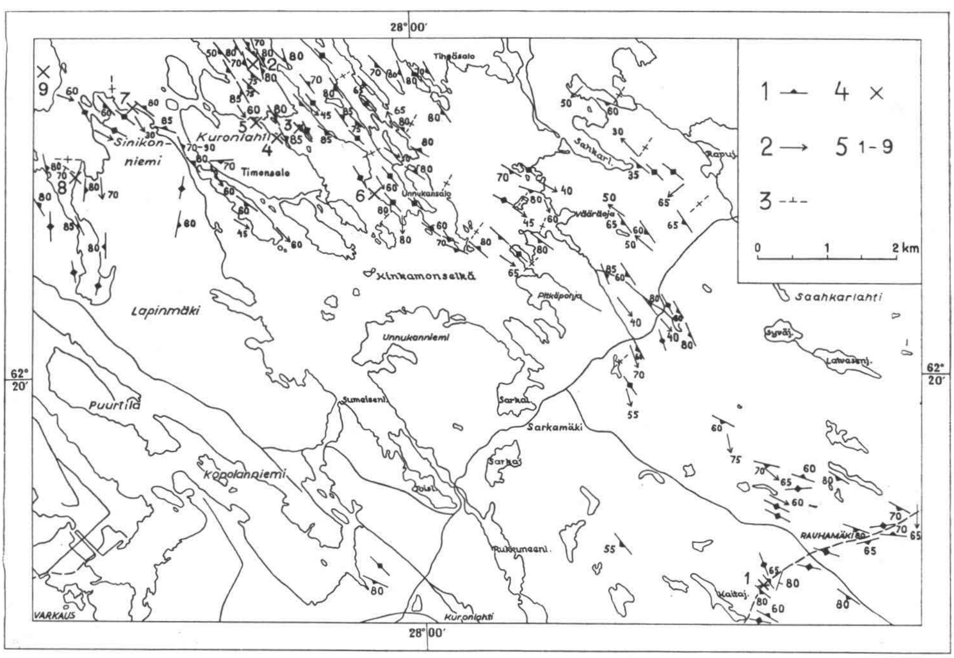

Fig. 1. Sketch map of the Rauhamäki -Kuronlahti area. 1) Strike and dip. 2). Lineation. 3) Cross joint. 4) Location of the ultrabasic and basic bodies. 5) Number of the bodies [1) Railway junction, Rauhamäki, Kangaslampi, 2) North end of Timonsalo, Kuronlahti, Leppävirta, 3) Island between Timonsalo and Unnukansalo, Kuronlahti, Leppävirta, 4) Timonsalo, Kuronlahti, Leppävirta, 5) Timonsalo, Kuronlahti, Leppävirta, 6) North end of Unnukansalo, Kuronlahti, Leppävirta, 7) Sinikonniemi, Kuronlahti, Leppävirta, 8) Sinikonniemi, Kuronlahti, Leppävirta and 9) Mäntyniemi, Kuronlahti, Leppävirta].

Outokumpu Company began to explore the district in the summer of 1960. The authors of the present paper participated in this work. Several more sulphide bearing blocks were found in the course of these investigations.

In the course of the investigations, small ultrabasic and basic bodies were found, sometimes in rather significant quantities, in the district situated to the north-west of Niemilahti. This area (Fig. 1) is formed mainly by the Rauhamäki village in the northern part of the Kangaslampi commune and by Kuronlahti village in the southern part of the Leppävirta commune.

The present paper describes two samples of the Niemilahti ore block and in addition a quantity of ultrabasic and basic bodies from the area of Fig. 1, which megascopically deviate very much from each other. The authors wanted to ascertain whether the ore block of Niemilahti or part of the ore blocks found later can have originated from these bodies.

The denominations used in this paper and the texture terms are based almost without exception on the microscopic and petrographic properties of the rocks and on observations, which indicates that the denominations have no genetic meaning, being merely descriptive. The compositions of the minerals are mainly those of Tröger (1959), though sometimes they are given according to Deer *et al.* (1963).

PETROGRAPHIC DESCRIPTION OF THE NICKELIFEROUS ORE BLOCK IN NIEMILAHTI

The Niemilahti ore block is a dark grey, massive, even- and medium-grained rock. In texture it resembles hypidiomorphic texture. The ore forms twisted bands in the interstices of the silicate mineral grains. The authors present two samples, deviating to a certain degree from each other, one of which bears a rather close resemblance in composition and texture to diorite, and the other one to quartz diorite. The modal mineral composition of both samples is shown in Table 1. Both contain the same ore minerals, but the latter is poorer in chalcopyrite and richer in pentlandite than the former; however, the pentlandite has been altered into bravoite to a greater degree than the former.

P l a g i o c l a s e ($\alpha = 1.546$, $\gamma = 1.555$, An_{32-36}) occurs in the diorite sample as grains of 0.5—1.5 mm in length, and in the sample of quartz diorite type as anhedral grains about 3 mm long. In the plagioclase of diorite small inclusions are found, mainly of sericite; in the plagioclase grains of the quartz diorite type mainly carbonate inclusions were found.

H o r n b l e n d e ($\alpha = 1.651 =$ is almost colourless, $\beta =$ greenish or brownish, $\gamma = 1.674$ brownish green, $c \wedge \gamma = 18°$, 37 mol. — % Fe^{2+} —component) occurs as anhedral, ragged grains containing abundant inclusions.

Table 1. Modes in volume per cent of the ore block samples. 1) Diorite, 2) quartz diorite. Niemilahti, Pisamaniemi, Kangaslampi.

	1	2
Plagioclase	17.0	59.0
Hornblende	42.6	8.8
Cummingtonite	6.0	—
Biotite	17.5	—
Quartz	—	3.2
Ore minerals	16.6 [1]	6.5 [1]
Accessory minerals	0.5 [2]	0.5 [3]

[1]) Pyrrhotite, chalcopyrite, pentlandite, bravoite, (sphalerite, ilmenite, magnetite and graphite.)
[2]) Quartz and apatite.
[3]) Diopside, apatite, carbonate and cummingtonite.

C u m m i n g t o n i t e ($\alpha = 1.644$, $\gamma = 1.673$, 50 mol. — % Fe^{2+} — component) is slightly zonal and polysynthetically twinned according to (100); it occurs in the marginal parts of the hornblende; the boundary between them is often very irregular. Some of the cummingtonite perhaps originates from hornblende.

B i o t i t e ($\gamma = \beta = 1.631$, 37 mol. — % Fe_6 Al_2-component) occurs as small flakes which are partly found in the cleavages of hornblende. Apatite is often found as an inclusion in biotite. Some of the biotite and hornblende shows an abundance of opaque pigments.

P y r r h o t i t e occurs as xenomorphic grains (1—3 mm) together with chalcopyrite and pentlandite. It is not lamellaric. It is clearly anisotropic and is reddish brown in colour.

C h a l c o p y r i t e occurs, together with pyrrhotite- and pentlandite grains as grains about 1 mm in diameter. Grains of sphalerite occasionally form inclusions in these grains. It also occurs as veins in the fissures of silicate minerals. Its colour is yellow and it is slightly anisotropic.

P e n t l a n d i t e is found in two generations: grains of 1 mm in diameter in association with pyrrhotite and chalcopyrite, and sometimes as small exsolution lamellae in the pyrrhotite. The lamellae, which are not frequent, are parallel, and have the same direction as the surfaces of the pyrrhotite (0001), as shown *e.g.* by Saksela (1960). The big pentlandite grains are mainly xenomorphic and quite often cataclastic. Euhedral pentlandite grains have also been found in the quartz diorite sample (Fig. 1, Plate I). According to Ramdohr (1955), pentlandite crystallizes somewhat later than pyrrhotite, seldom at the same time or somewhat earlier. It is quite evident that the chalcopyrite has crystallized later than these. The colour of the pentlandite is a slightly yellowish white. It is completely isotropic and its hardness is distinctly less pronounced than that of the pyrrhotite. It is common Ni-pentlandite since a test with the Gaudin solution stained it blue.

The pentlandite alters into b r a v o i t e, starting at the fractures. Many of the grains have altered to a considerable extent, the quartz diorite type more than the diorite type. The colour of the bravoite is a slightly brownish grey. It is darker and about as hard as pyrrhotite. It is isotropic with a comparatively slight polish. According to Ramdohr (1955) the properties of bravoite, for instance its polish, vary, depending on the chemical composition.

I l m e n i t e occurs in the form of very small grains in a silicate mineral. It has a very slight reflectivity and it is clearly anisotropic with its bluish and brownish colour tints.

There is very little m a g n e t i t e and it occurs together with pyrrhotite.

G r a p h i t e occurs as small flakes in connection with the silicate minerals. Its colour is brownish, the pleochroism is distinct and the anisotropy is strong in yellow to bluish colour tints.

According to Ramdohr (1955) almost all the pyrrhotite deposits have a high formation temperature in common. The lack of pyrite, as in this case, also indicates a high temperature. The lack of cubanite and vallerite exsolution lamellae makes it difficult to determine the exact minimum temperature limit. The exsolution temperature of flame-shaped pentlandite varies with the various research workers (among others Schwarz, 1938; Hewitt, 1938; Edwards, 1954; Marmo, 1955; Ingerson, 1955; Pauly, 1958 and Kullerud 1963) from 310°C to 800°C.

An ore mineral assemblage, such as is found in the described Niemilahti samples, is inevitably early magmatic in nature, according to Schneiderhöhn (1941) and Ramdohr (1955).

DESCRIPTION OF THE ULTRABASIC AND BASIC BODIES OF THE RAUHAMÄKI — KURONLAHTI DISTRICT

INTRODUCTION

The Kuronlahti-Rauhamäki district is classified as belonging to the so-called Savo schist zone. The southern district of the mapped region has been investigated by Hackman (1933), the northern by Wilkman (1938).

The principal rock of this region is migmatite. Its texture mostly looks like veined gneiss, and it is comparatively poor in mica. The rocks of this district are mostly strongly folded. Sometimes well preserved schist layers are met with.

Acid veins and bodies of migmatite have revealed a great paucity of potash feldspar. For instance in Rauhamäki there is a small body of rock,

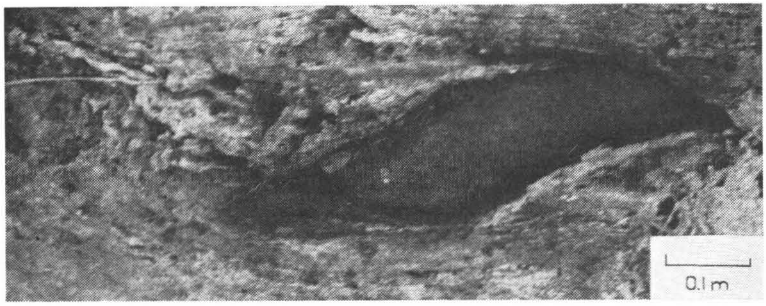

Fig. 2. Amphibolite lens in veined gneiss. Timonsalo, Kuronlahti, Leppävirta.

which Hietanen (1945) has called dioritetrondhjemite and which has the following modes in volume per cent: hornblende 53, plagioclase (An_{25}) 29, biotite 10, quartz 3 and accessories (apatite, titanite, sericite, chlorite, rutile and opaque minerals) 5, and in the northern end of Timonsalo there is a small boby of trondhjemite: plagioclase (An_{30}) 50, quartz 30, biotite 15, muscovite 5 and accessories (potash feldspar, apatite, chlorite, zircone, carbonate, sericite and opaques) 5.

Ultrabasic and basic bodies occur in the migmatites. These have not been described earlier. However, Wilkman (1938) has described the peridotites and gabbros of the northern and central parts of the Leppävirta commune, and Hackman the gabbro granites about 10 km south of Rauhamäki. The bodies are unevenly distributed over the area. In places they occur in abundance, while elsewhere only a few are found. Especially the gabbro bodies usually occur in clusters.

In size the bodies are small, 0.3 m to 6 m long, and 0.1 m to 3 m broad. Only one exception was found, a body 250 m long and 150 m wide. The form of the bodies is usually that of a lenticular lens or an ellipsoid (Fig. 2 in text and Fig. 8, Plate IV), but even some roundish and some subangular bodies are known. The longitudinal direction of the bodies is often the same as the direction of the lineation. All bodies have sharp contacts against their wall rocks and are also conformable.

The bodies are divided into two groups: the ultrabasic and the basic. The rocks of the ultrabasic bodies are composed of various peridotites. Bodies, whose rocks contain gabbros, quartz gabbros, quartz diorites and amphibolites have been regarded as basic bodies. Various bodies belonging to different groups may occur, and often do, in the same outcrop. It has not been possible to establish with full certainty that bodies belonging to any definite species of rock would have become enriched in any definite area.

Table 2. Modes in volume per cent of the peridotite (Central part of the body (1), border variations (2, 3 and 4), 4 occurs nearest to the wall rock). Railway junction, Rauhamäki, Kangaslampi

	1	2	3	4
Olivine	20	—	—	—
Hypersthene	10	—	—	—
Augite	3	—	—	—
Serpentine	5	—	—	—
Actinolite	60	—	—	—
Hornblende	—	74.1	95.0	12.0
Biotite	—	—	—	41.4
Plagioclase	—	—	—	37.7
Quartz	—	—	—	3.8
Acessory minerals	2 [1]	6.8 [2]	1.9 [2]	3.3 [3]
Opaque minerals	[4]	6.1 [4] [5]	3.1	+
	100	100.0	100.0	100.2

[1]) Cummingtonite, biotite, talc, carbonate, plagioclase and rutile.
[2]) Biotite, apatite, carbonate and chlorite.
[3]) Apatite, epidote and opaques.
[4]) Pyrrhotite, pentlandite, chalcopyrite, tetragonal iron sulfide, magnetite, pyrite, chromite and graphite.
[5]) Cubanite.

ULTRABASIC BODIES

Four ultrabasic bodies are presented. Rauhamäki peridotite (1),[1] Mäntyniemi augite-bearing hornblende peridotite (9), Timonsalo olivine-bearing augite-hornblende peridotite (4), and the plagioclase-bearing augite-hornblende peridotite of Timonsalo (2). The former is situated in the Rauhamäki village of the Kangaslampi parish, the others in the Kuronlahti village of the Leppävirta parish.

THE PERIDOTITE OF RAUHAMÄKI

The peridotite of Rauhamäki (5 × 3 m), situated at a railway junction (Fig. 1, No. 1), is ellipsoid in form. It is a comparatively homogeneous rock, only the mineral composition of the border zone deviates from the main type. The body has some narrow pegmatitic veins, and near the western margin there is an 0.5 m long pyrrhotite vein. There are comparatively few opaque minerals, and they occur in small grains scattered on the rock.

In the central part of the peridotite body there is a greyish-black medium- or coarse-grained, fairly even-grained, strongly metamorphosed rock with a blastohypidiomorphic texture. Its essential minerals are actinolite, olivine, hypersthene, serpentine and augite (No. 1, Table 2). The marginal zone of the rock deviates from the central part in its mineral composition (Nos 2, 3 and

[1]) The numbers in brackets show the position of the bodies in Fig. 1.

4, Table 2). There are zones rich in hornblende, the hypersthene is lacking and there is very little olivine. According to Hess (1933), olivine, pyroxene and serpentine turn during their steatitization into actinolite to begin with and if Al is present, they change into hornblende. In the contact the rock contains quartz and large quantities of biotite and plagioclase.

Olivine ($\alpha = 1.689$, $\gamma = 1.729$, $2V\alpha = 84°$, Fa_{27}) occurs as grains 2—3 mm in length, which are partly serpentinized and which during the process of serpentinization, have become rounded and separated into pieces. The serpentine is often like a network, in which the meshes are formed by an olivine or a pleochroic, almost isotropic reddish brown or yellowish fine crystalline mass. Part of the olivine is completely serpentinized. The serpentine is mainly colourless, flake-formed antigorite ($\alpha = 1.569$, $\gamma = 1.574$). In the fissures of some olivine grains there occurs a fibrous, slightly pleochroic chrysotile of a greenish colour. Instead of the original fissures we often find bands formed by small magnetite grains in the antigorite (Fig. 2, Plate I). On page 182 another form of occurrence of magnetite in connection with the olivine is shown. Here and there the actinolite is also slightly serpentinized. DuRietz (1935) has described similar serpentinization phenomena.

The hypersthene ($\alpha = 1.663$, $\beta = 1.667$, $\gamma = 1.673$) is euhedral. It contains a considerable number of small, brown or black, plate-formed opaque inclusions, which occur in the way described by Rosenbusch and Mügge (1925, p. 344).

Augite ($\alpha = 1.695$, $\gamma = 1.722$, $c \wedge \gamma = 42°$, $2V\gamma = 59°$) occurs as a relict in actinolite. Here and there the relictic augite grains and the actinolite together form myrmekite-like textures. Fig. 3 shows the manner of intergrowth of augite and actinolite. The c- and b- axes of both intergrown minerals are parallel.

Actinolite ($\alpha = 1.623$, $\beta = 1.632$, $\gamma = 1.652$, $c \wedge \gamma = 18°$, $2V\alpha =$ 83°, 28 mol. — % $CaFe_5$ component) contains some small carbonate grains and talc flakes, black opaque pigment is present in its fissures. Biotite and chlorite flakes form abundant inclusions in actinolite. The flakes have the same direction as the cleavages. In the investigated cases, the γ of chlorite and the α of biotite have been almost parallel to the normal of the (100) and the γ of biotite and β of chlorite are parallel to each other while both may often have the directions of the c-axes of the actinoiite (Fig. 4, Plate II).

Hornblende ($\alpha = 1.642$, $\gamma = 1.662$, $2V\alpha = 84°$, $c \wedge \gamma = 22°$, 22 mol. % Fe component) occurs in abundance in the marginal varieties of the bodies (Nos 2 and 3, Table 2) instead of actinolite, in grains which are 2—5 mm, sometimes even 1—2 cm long. In the hornblende grains there are augite, sometimes even olivine relicts. In the cleavages of hornblende, opaque or almost opaque, brown or black, needlelike or rounded inclusions are found in abundance. According to Rosenbusch and Mügge (1925) they are products

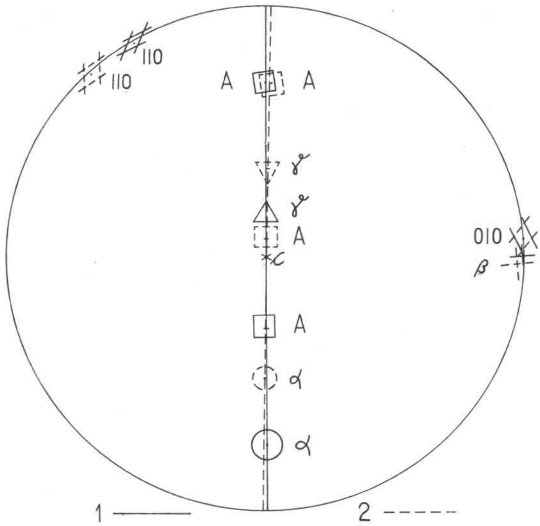

Fig. 3. Stereographic projection of intergrowth of augite (2) and actinolite (1). The projection shows the indicatrixes, cleavages and c-axes of both minerals according to Reinhard (1931). Peridotite. Rauhamäki, Kangaslampi.

of exsolution. According to Johannsen (1952) they lie in the direction of the c-axis the prismatic cleavage. According to Williams *et. al.* (1954) they are either rutile, ilmenite or magnetite. The hornblende of the marginal variety (No. 3, Table 2) is slightly carbonitized and chloritized so that magnetite lamellae have exsolved from it, and run in different directions. Rogers and Longshore (1960) called this structure also a schiller structure.

The chlorite is chlinochlore ($a \wedge \alpha = 3°$, $2V\gamma = 50°$, the elongation is negative). It is also found outside of the actinolite. In the marginal varieties of the peridotite body (Nos 2 and 3, Table 2), radial chlorite is also found in the hornblende.

According to Hess (1933), olivine pyroxene and serpentine in the process of steatitization are first changed into actinolite, and if Al is present, into hornblende. If Al is present, talc is also formed. According to Haapala (1936), amphiboles are formed partly as an alteration of the pyroxenes. Not much talc and carbonate has been generated in the Rauhamäki peridotite. According to Turner and Verhoogen (1951) steatitization has taken place after serpentinization.

Of cummingtonite ($\alpha = 1.656$, $c \wedge \gamma = 17°$, $2V = +$) there are a few small, colourless, multiple, polysynthetic twinning grains.

There is very little plagioclase (An_{25}). In the contact variations (No. 4, Table 2) there is more plagioclase and it is richer in An (An_{74}, $\alpha = 1.568$, $\gamma = 1.576$, $2V\alpha = 87°$). As a result of the alteration, the plagioclase contains epidote as small grains and groups of grains.

Biotite of a light grey colour, which occurs as inclusions in actinolite and in hornblende as well as outside of them, is present in abundance in the border varieties (No. 2, Table 2).

Carbonate occurs in border variety (No. 2, Table 2) as an aggregate with serpentine and opaque minerals. They are perhaps alteration products of olivine.

Pyrrhotite occurs as veins in the body, and in addition as roundish, anhedral grains of about 2—1 mm in diameter, forming mixed grains with chalcopyrite and pentlandite (Fig. 3, Plate II). In places it has altered into a slightly polished mass, in which small, better polished, white grains occur. These resemble marcasite rather closely. Here and there these products of alteration have formed concentric spheres. They are »hydropyrrhotite».

Pentlandite occurs in two generations, since it is also found as exsolution lamellae in pyrrhotite. Its colour is slightly yellowish and it is isotropic, its hardness slightly less than that of the pyrrhotite. In its fissures it has to a certain degree altered into bravoite. Etching with Gaudin's solution gave a blue stain.

Tetragonal iron sulfide occurs in the pentlandite as small, longish, sometimes flame-shaped inclusions (Fig. 3, Plate II). These are distinctly pleochroic, from the light grey of the pyrrhotite to the dark grey of magnetite. Their anisotropy is very pronounced with bluish and white tints. The properties correspond to the properties of tetragonal iron sulfide (Kouvo *et. al.* 1963). According to Fleischer (1962) kansite has the same physical properties.

The chalcopyrite evidently replaces the pyrrhotite. Around a big pyrrhotite grain small relict pyrrhotite grains are found, separated by silicate minerals, which are strongly cataclastic, and the interstices of which are filled with chalcopyrite and magnetite.

Magnetite occurs in pyrrhotite, pentlandite and chalcopyrite as veins, which replace the above mentioned minerals (Fig. 3, Plate II). It is isotropic and grey.

Chromite occurs in the olivine as small, roundish, anhedral grains, which sometimes have a magnetite border. The hardness of the chromite is greater than that of magnetite, its reflectivity is smaller, however. Both are isotropic.

Ilmenite occurs in the hypersthene and augite as narrow, long, plate shaped, evidently exsoluted grains, and in actinolite, which has evidently generated from augite by alteration.

Table 3. Modes in volume per cent of the augite bearing hornblende peridotite (1) from Mäntyniemi and the olivine bearing augite-hornblende peridotite (2) from Timonsalo, Kuronlahti, Leppävirta.

	1	2
Olivine	—	10.8
Augite	18.2	33.1
Hornblende	75.0	46.1
Opaque minerals	2.3	8.8 [2]
Acessory minerals	4.5 [1]	1.2 [3]
	100.0	100.0

[1] Epidote, plagioclase, apatite and chlorite.
[2] Pyrrhotite, chalcopyrite, pentlandite, tetragonal iron sulfide, ilmenite and magnetite.
[3] Serpentine.

G r a p h i t e occurs as twisted flakes of less than 1 mm in length. It is pleochroic, its colour fluctuating from a light brownish grey to dark brown, and anisotropic from yellow to blue. Magnetite occurs in its cleavages and on its borders. Between the magnetite and the graphite there are small amounts of pyrrhotite and yellowish pyrite.

C u b a n i t e is found in the chalcopyrite in the border zone of the body (No. 2, Table 2). Its colour is darker than that of the chalcopyrite and more yellowish than pyrrhotite. It is isotropic.

THE AUGITE-BEARING HORNBLENDE PERIDOTITE OF MÄNTYNIEMI

In Mäntyniemi (Fig. 1, No. 9) there occur some grey-black small bodies, which are surrounded by 1—10 cm wide, coarse-grained, light grey acid veins. In one of the bodies there is a winding vein about 2 mm wide containing quartz grains about 1 mm in size, and slightly smaller augite, hornblende and carbonate grains. Their interstices are filled with quartz, carbonate and an opaque and dark brown dense mass, in which small augite and hornblende grains occur as relicts.

The bodies are coarse-grained, massive, augite-bearing hornblende peridotite of hypidiomorphic texture. The mineral composition of one of the bodies is shown in Table 3, No. 1.

H o r n b l e n d e ($\alpha = 1.643$ = colourless, $\beta = 1.656$, $\gamma = 1.660$ = light grey, $2V\alpha = 82°$, $c \wedge \gamma = 21°$, 20 mol. — % Fe-component) occurs here as anhedral grains 3—5 mm long. Sporadically it is altered into chlorite.

A u g i t e ($\alpha = 1.680$, $\beta = 1.685$, $\gamma = 1.706$, $2V = 55°$, $c \wedge \gamma = 42°$) occurs as small (<1 mm) prisms, which are situated in the interstices of the hornblende grains. The augite is to some extent altered into hornblende.

The rock contains very little p l a g i o c l a s e ($\alpha = 1.555$, $\gamma = 1.562$, An_{55}).

THE OLIVINE-BEARING AUGITE-HORNBLENDE PERIDOTITE OF TIMONSALO

In the southern part of the Timonsalo island (Fig. 1, No. 4), there is a coarse-grained, grey, roundish body (20×30 cm) of olivine-bearing augite-hornblende-peridotite. It is encircled by a plagioclase-rich pegmatite vein, which also contains quartz and biotite. Its mineral composition is shown in Table 3, No. 2.

The h o r n b l e n d e ($\alpha = 1.648$ = colourless, $\gamma = 1.660$ = greenish $2V\alpha = 78°$, $c \wedge \gamma = 19°$, 28 mol. — % Fe-component) is xenomorphic, 1—2, cm in size. It contains augite and olivine inclusions. Hornblende also occurs within the augite as a result of alteration.

The o l i v i n e ($2V\alpha = 83°$) is anhedral and occurs as grains of about 1 mm. In its fissures it is altered to a greenish or yellowish fibrous serpentine. Olivine exhibits lamellaric twinning according to surfaces (120) and (110). These are possibly twinning planes as described by Rosenbusch (1925). Such a twinning may be a sign of tectonic movement (Turner and Verhoogen, 1951).

The a u g i t e ($2V\gamma = 58°$, $c \wedge \gamma = 42°$) occurs as short prismatic grains and is slightly zonal.

The mode of occurrence of the o p a q u e m i n e r a l s is almost the same as in the Rauhamäki peridotite. Pyrrhotite predominates, and chalcopyrite and pentlandite occur as mixed grains together with pyrrhotite. Pentlandite of two generations is also found. It has not changed into bravoite. Ilmenite occurs both as small grains and as lamellae in the silicate minerals. Magnetite encircles the olivine as small vermicular grains.

THE PLAGIOCLASE-BEARING AUGITE-HORNBLENDE PERIDOTITE OF TIMONSALO

The largest peridotite body (approx. 250—150 m) is situated in the Kuronlahti village, in the northern part of the Timonsalo island. The wall rock foliation follows the outlines of the body and is intersected by narrow acid pegmatite veins. The peridotite is a black, medium- or coarse-grained, massive, comparatively homogeneous rock. The mineral composition of the rock of the body is shown in Table 4. In its central parts the rock (No. 1) is olivine-bearing. The olivine is not present in the border zones of the body. There is less hypersthene and more hornblende or plagioclase and biotite than in the central parts. (Table 4, Nos 2 and 3).

The p l a g i o c l a s e (An_{70-65}) cannot be called zonal. Along the borders of the body it is abundantly sericitized. Moreover, the border zones exhibit completely unaltered, polysynthetically twinned, non-zonal plagioclase grains (An_{50}).

Table 4. Modes in volume per cent of the plagioclase-bearing augite-hornblende peridotite. Main type (1), border variations (2 and 3). Timonsalo, Kuronlahti, Leppävirta.

	1	2	3
Plagioclase	5.0	9.6	3.0
Olivine	5.3	—	—
Hypersthene	14.0	5.7	3.0
Augite	22.6	34.4	20.7
Hornblende	47.0	41.5	61.0
Actinolite	+	—	+
Biotite	+	6.3	5.7
Quartz	+	+	+
Chlorite	—	+	+
Serpentine	+	—	—
Carbonate	—	+	+
Apatite	+	+	+
Sericite	—	—	+
Opaques	+ [1]	+	+
Accessories	6.1	2.9	6.8
	100.0	100.4	100.2

[1]) Pyrrhotite, tetragonal iron sulfide, pentlandite, bravoite, ilmenite, limonite and hydropyrrhotite.

The olivine is idiomorphic and has altered to fibrous serpentine beginning with the fissures. In the fissures there are also inclusions of opaque minerals.

The augite ($2V\gamma = 55°—60°$, $c \wedge \gamma = 42°$) occurs as short prisms which are often twinned according to the augite law. The augite has changed into hornblende and also to serpentine far more than the hypersthene (Fs_{25}). Often it occurs in the hornblende as irregular relict grains.

The hornblende ($\alpha = 1.655$, $\gamma = 1.675$, $c \wedge \gamma = 26°—18°$, $2V\alpha = 82°—88°$) is xenomorphic and poikilitic. There are inclusions of olivine, augite and hypersthene. It is partly an alteration product of augite and hypersthene. As an alteration product of augite, its colour is brownish. On the other hand, as an alteration product of hypersthene it has a greenish colour. Here and there the hornblende turns first into bluish green amphibole and then into colourless lamellaric actinolite ($\alpha = 1.631$, $\gamma = 1.652$, $2V\alpha = 80°$, $c \wedge \gamma = 15°$). Opaque mineral is found in the cleavages of hornblende, and in the border zones of this body biotite and chlorite also occur in the cleavages.

The pyrrhotite is xenomorphic and has in many cases changed into hydropyrrhotite, and even to limonite. Pentlandite often occurs as flame-shaped lamellae in pyrrhotite (Fig. 5, Plate III). It has partly changed into bravoite, especially where the pyrrhotite has changed into hydropyrrhotite. Ilmenite occurs exsolved in pyroxene (Fig. 6, Plate III). There is very little tetragonal iron sulfide.

BASIC BODIES

Bodies which belong to this groups are found in abundance on the area of Fig. 1. Their length varies from 0.3 to 10 m. They are roundish, elliptic or lens-shaped. In the small bodies a slight foliation is more common than in the big ones. Sometimes the border zone of a body is foliated while the centre is massive. The rock in the lens-shaped bodies often looks like amphibolite (Fig. 2). The roundish bodies are frequently coarse-grained independently of their size. Very often the bodies are surrounded as well as pierced by narrow, acid pegmatite veins. The main rock species of the bodies are biotite-hornblende gabbro, quartz gabbro, quartz diorite and amphibolite.

THE BIOTITE-HORNBLENDE GABBRO OF RAUHAMÄKI

The Rauhamäki biotite-hornblende gabbro is situated about 20 m to the west of the respective peridotite body (Fig. 8, Plate IV). This gabbro is medium- and even-grained, dark grey and non-foliated. In Table 5, No. 1 the modal composition of the rock and the optical properties of the minerals are given. The gabbro is comparatively rich in biotite.

The hornblende (γ = of a light bluish green, 15 mol. % Fe-component) occurs as 1—2 mm long grains, which contain biotite and epidote inclusions.

The biotite ($\gamma = \beta$ = reddish brown) generally occurs as flakes of less than 1 mm in length. It occurs in the hornblende, both in the directions of the cleavages, in which case it is anhedral, and also situated at random when it is euhedral. On the boundary of the biotite and hornblende small titanomagnetite grains are often found, in some of these the ilmenite has changed into leucoxene.

The plagioclase (An_{65-80}) is to some extent distinctly zonal. It contains an abundance of inclusions: quartz, epidote, and especially hornblende.

THE QUARTZ GABBRO FROM AN ISLAND BETWEEN TIMONSALO AND UNNUKANSALO

On the small island between Timonsalo and Unnukansalo (Fig. 1, No. 3) there occurs a body of quartz gabbro (2.5×2.5 m). It is surrounded by narrow pegmatite veins. The average grain size of the gabbro minerals is 2 mm. The rock is dark grey and massive. Its texture is hypidiomorphic. The modal mineral composition of the gabbro and the optic properties of its minerals are shown in Table 5, No. 2. The hornblende of the gabbro is richer in Fe than the hornblende of the Rauhamäki gabbro.

The hornblende grains (γ = light green, 33 mol. % Fe-component) are 1—2 mm long, sometimes as much as 5 mm. They contain biotite inclusions

Table 5. Modes in volume per cent of the basic rocks. Rauhamäki-Kuronlahti district.

	1	2	3	4	5
Plagioclase	10.0	21.2	14.6	27.3	44
Quartz	+	8.6	4.4	13.4	10
Hornblende	52.8	52.2	65.2	39.4	30
Cummingtonite	—	—	—	—	5
Biotite	33.4	13.0	11.6	17.6	10
Apatite	+	+	+	+	+
Titanite (leucoxene)	+	+	+	+	+
Carbonate	+	+	+	+	+
Sericite	—	+	+	—	+
Opaque minerals	+	+	+	+	+
Epidote	+	—	—	—	—
Zircon	+	—	—	—	—
Chlorite	—	—	—	+	+
Rutile	—	—	+	—	—
Acessories	3.8	5.0	4.2	2.3	1
Σ	100.0	100.0	100.0	100.0	100
Plagioclase α	1.569	1.567	1.567	—	1.547
» γ	1.578	1.576	1.576	—	1.554
Hornblende α	1.640	1.648	1.647	1.655	—
» γ	1.657	1.666	1.665	1.673	—
» $2V\alpha$	80°	76°	76°	76°	—
» $c \wedge \gamma$	21°	17°	18°	19°	—
Biotite $\gamma \approx \beta$	1.610	—	—	1.645	—

1. Biotite-hornblende gabbro. Railway junction, Rauhamäki, Kangaslampi.
2. Quartz gabbro. Island between Timonsalo and Unnukansalo, Kuronlahti, Leppävirta.
3. Biotite-hornblende gabbro. Island between Timonsalo and Unnukansalo, Kuronlahti, Leppävirta.
4. Quartz gabbro. North-western end of Sinikonniemi peninsula, Kuronlahti, Leppävirta.
5. Quartz diorite. Unnukansalo, Kuronlahti, Leppävirta.

in the same way as the hornblende of the Rauhamäki gabbro. In addition there is a small content of carbonate in the hornblende.

The p l a g i o c l a s e (An $_{70-75}$) is strongly altered to sericite and contains inclusions of carbonate and quartz.

The gabbro contains an abundance of t i t a n i t e as small xenomorphic grains, which occur both in biotite and in hornblende. Part of the titanite is leucoxene. In these cases it occurs around the opaque minerals.

The a p a t i t e occurs in abundance in big, approx. 1 mm long, xenomorphic prisms, containing inclusions of hornblende. The pyrrhotite is fine-grained and xenomorphic. It contains very small, occasional grains irregular in form, which are whiter and slightly harder than the pyrrhotite, which has sporadically changed into hydropyrrhotite.

The p y r i t e occurs in connection with the pyrrhotite as small xenomorphic grains. It is harder than pyrrhotite and isotropic. The chalcopyrite which occurs together with the pyrrhotite is anhedral. Ilmenite occurs as small anhedral grains in silicate minerals.

Other kinds of bodies are also found on the same island. Table 5, No. 3 shows, *e.g.*, the modal mineral composition of a biotite- hornblende gabbro, and the optic properties of its minerals. The biotite of the gabbro contains shagenite needles. (Fig. 7, Plate IV), which form a networklike texture. According to Rosenbusch (1925) (301) and (101) appear as twinning planes.

THE QUARTZ GABBRO IN THE NORTH-WESTERN END OF SINIKONNIEMI

In the north-western end of the Sinikonniemi peninsula there is a quartz gabbro body (Fig. 1, No. 7). The border zone of this body is slightly foliated and narrow pegmatite veins run crosswise through it. The quartz-gabbro in this body is homogeneous, medium- and even-grained, its texture is hypidiomorphic. The modal mineral composition of the gabbro is shown in Table 5, No. 4.

The plagioclase (An_{44}) occurs as grains 0.5—1.5 mm in length. The grains are only slightly sericitizised.

There are two kinds of hornblende: light green (38 mol. % Fe-component) and green hornblende ($\alpha = 1.661$, $\gamma = 1.682$, $c \wedge \gamma = 17°$, $2V\alpha = 76°$, 47 mol. % Fe- component). Its grain size is 1—3 mm, but some big (5 mm) grains also occur. The hornblende contains biotite and some chlorite.

The biotite (γ = brown) occurs in the hornblende both in the direction of the cleavages and as irregular flakes the diameter of which is about 1 mm.

The quartz occurs as small grains the other minerals.

The pyrrhotite forms small, anhedral grains, which have not altered.

The pentlandite occurs together with pyrrhotite as small, xenomorphic grains. Here and there small magnetite grains are found and exsoluted ilmenite is found in one of the silicates.

The graphite occurs as flake-like aggregates.

THE QUARTZ DIORITE OF UNNUKANSALO

Small quartz diorite bodies occur in Unnukansalo (Fig. 1, No. 6). In one of the bodies the quartz diorite is medium- and even-grained, massive, and hypidiomorphic in texture. Table 5, No. 5 shows the modal mineral composition of the rock.

The composition of the plagioclase is An_{32}. It contains only small quantities of alteration products.

The hornblende (25 mol. % Fe component) occurs as xenomorphic grains, which sometimes are slightly bigger than average.

The cummingtonite (46 mol. % Fe- component) occurs as small grains which are polysynthetically twinned.

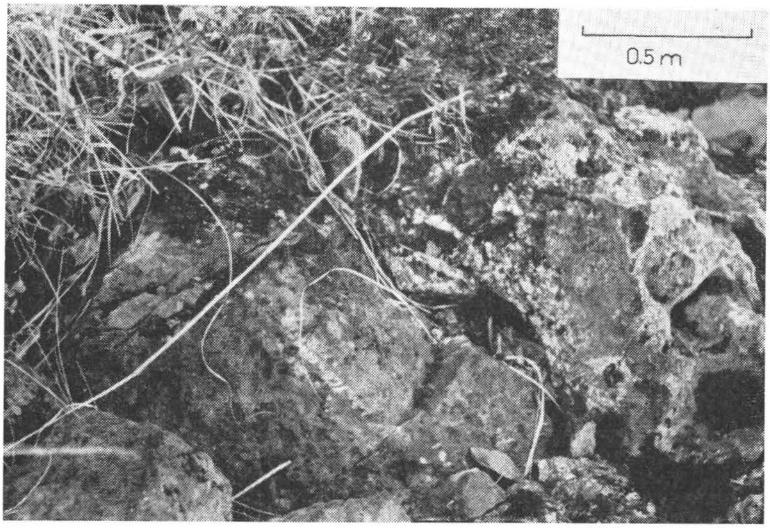

Fig. 4. Biotite-hornblende gabbro body, showing different varieties of basic rocks and pegmatite veins. Timonsalo, Kuronlahti, Leppävirta.

THE BIOTITE-HORNBLENDE GABBRO FROM TIMONSALO

In Timonsalo (Fig. 1, No. 4) almost all the various bodies occurring in the Rauhamäki-Kuronlahti district are met with in close vicinity to each other, including the narrow amphibolite layers.

The largest body (6 × 5 m) is rounded and very non-homogeneous. There are two different varieties of biotite-hornblende gabbro, a medium-grained, and a coarse-grained variety, as well as quartz diorite. The body is encircled and pierced by acid pegmatite veins. The different types of rock in the bodies are sharply limited against each other, and between them a narrow pegmatite vein is often found. The rocks which are the most coarse-grained appear to be more cutting than the others. Figure 4 shows part of this body. Table 6 shows the modal mineral composition of the basic types of rock and the optic properties of the minerals. All three types of rock: the medium-grained biotite-hornblende gabbro (No. 1), the medium-grained quartz diorite (No. 2) and the coarse-grained biotite-hornblende gabbro (No. 3), contain the same minerals though in different quantities and in composition slightly deviating from each other.

Depending on the different types of rock the composition of the plagioclase changes from An_{42} to An_{52}. The grain size in the medium-grained types is 0.5—1.5 mm, and in the coarse grained types 3—6 mm. Especially the latter type contains sericite and carbonate.

Table 6. Modes in volume per cent of the basic rocks. Rauhamäki-Kuronlahti district.

	1	2	3	4	5
Plagioclase	28.9	38.7	30.6	33.4	24.7
Quartz	2.9	10.7	—	3.9	6.7
Hornblende	53.0	31.4	50.5	51.4	48.1
Biotite	10.1	16.4	16.8	11.0	19.1
Apatite	+	+	+	+	+
Titanite (leucoxene)	+	—	+	—	+
Carbonate	—	+	+	—	+
Sericite	+	+	+	—	—
Zircon	—	+	—	+	—
Chlorite	+	+	+	+	+
Rutile	+	+	—	—	—
Opaque minerals	+	+	+	+	+
Accessory minerals	5.1	2.8	2.1	0.5	1.6
	100.0	100.0	100.0	100.2	100.2
Hornblende α	1.647	—	1.651	1.657	1.647
» γ	1.666	—	1.666	1.678	1.678
» $2V\alpha$	—	—	75°	—	76°
» $c \wedge \gamma$	19°	—	22°	17°	19°
Biotite $\gamma \approx \beta$	—	—	1.637	—	—

1. Medium-grained biotite-hornblende gabbro. Timonsalo, Kuronlahti, Leppävirta.
2. Medium-grained quartz diorite. Timonsalo, Kuronlahti, Leppävirta.
3. Coarse-grained biotite-hornblende gabbro. Timonsalo, Kuronlahti, Leppävirta.
4. Amphibolite. Timonsalo, Kuronlahti, Leppävirta.
5. Amphibolite. Timonsalo, Kuronlahti, Leppävirta.

The hornblende contains 27—37 mol % Fe- component. In coarse-grained types of rock it occurs as grains of 3-8 mm in length and in its cleavages opaque minerals, leucoxene, biotite and some chlorite are met with. In medium grained types the hornblende occurs as needles about 1—2 mm long.

The biotite (γ = brown) occurs mainly as xenomorphic flakes in the hornblende, *i.e.* as flakes orientated in the direction of the hornblende cleavages and as flakes which are arbitrarily situated. The biotite sporadically contains shagenite needles. Titanite tends to occur on the boundary between hornblende and biotite, an opaque mineral is usually met with in connection with it. The pyrrhotite occurs as small xenomorphic grains which are unaltered. Magnetite occurs scattered in the silicate mineral, and graphite is found as band-like flake aggregates.

THE AMPHIBOLITE FROM TIMONSALO

In Timonsalo (Fig. 1, No. 4) we have lens-shaped amphibolite bodies. They are often somewhat twisted and rounded (Fig. 2). Table 6, Nos 3 and 4 shows the modal mineral composition of the amphibolite and the optic properties of the minerals occurring in the two bodies. The rock in the bodies is foliated, evenly small-grained, sometimes medium-grained.

The h o r n b l e n d e (γ = brownish green) is xenomorphic. It contains 28—42 mol. % Fe- component. Biotite occurs in the hornblende and in the fissures there are small quantities of chlorite.

The composition of the p l a g i o c l a s e varies from An_{45} to An_{58}. There are quartz inclusions in the plagioclase.

Carbonate, apatite and titanite are present in comparatively large quantities. The c a r b o n a t e is anhedral and it occurs, as does the a p a t i t e, in plagioclase. The t i t a n i t e occurs in the biotite and especially in connection with opaque minerals.

The rocks contain comparatively small quantities of o p a q u e m i n e r a l s. There is some pyrrhotite, which has partly altered into hydro-pyrrhotite. There is also magnetite and some graphite.

ON THE ORE BLOCK OF NIEMILAHTI CONTRA ULTRABASIC AND BASIC BODIES

Ore minerals are more abundant in the ultrabasic than in the basic bodies. Pentlandite in particular occurs to a greater extent, in ultrabasic than in basic rocks so that in this respect the ultrabasic rocks bear a closer resemblance to the Niemilahti ore block. There is very little pentlandite in basic rocks and very often it is almost completely absent. This geochemical mode of occurrence of nickel is in accordance with the results which Häkli (1963) has obtained with Finnish rokcs.

On the other hand, the silicate mineral assemblage of the Niemilahti nickeliferous ore block corresponds to the basic rocks, which have been described in greater detail than the ultrabasic ones. No rocks showing the same silicate mineral composition as the Niemilahti ore block have been found in the ultrabasic bodies. Has the mineral paragenesis of the ore block changed in the process of ore mineralization? Erratic boulders may have detached themselves from the ultrabasic and even from the basic bodies, which possibly makes it more difficult to find the site of origin of the Niemilahti ore block. Probably all the small bodies of which less than one half were under the surface of the non-disintegrating rock before the continental glaciers passed over this surface, have been completely detached and now occur in blocks only. Statistically the blocks must have existed in equall quantities, since of the blocks visible today, one half or even more of the body remains under the surface of the rock.

ON THE GENESIS OF THE ULTRABASIC AND BASIC BODIES

It is difficult to form any idea of the genesis of these ultrabasic and basic bodies. They are found as small conformable bodies with sharp boundaries against the wall rock, usually in veined gneiss. As a rule they occur in groups: different bodies are often found quite close to each other. There are no magmatic rocks in their vicinity and facts clearly indicating a magmatic origin are lacking making it difficult to presume a magmatic genesis. Neither it is easy to assume a so-called cold intrusion in this case (Turner and Verhoogen, 1951).

In the area of Orijärvi, Mikkola has described ultrabasics with internal layering, which sometimes occur as boudins in diopside-amphibolites and in diopside gneiss. According to Mikkola (1955) the ultrabasics of Orijärvi originated through metamorphoses. Sørensen (1953) describes ultrabasic rocks in western Greenland which occur in parent as conformable bands in gneisses pointing to a possible »tectonic origin of the ultrabasic».

According to Erdmannsdörffer (1947) gabbros and diorites can originate even in situ, in metamorphic amphibolites and hornfelses.

According to Seitsaari (1951) the gabbro found in the area south east of Savonjärvi, in the Tampere district, has originated through ultrametamorphosis. The gabbro has gradually transformed into amphibolite (Seitsaari, 1951).

According to Nickel (1948, 1952), especially the diorites and gabbros that occur in schists in parallel alternating layers, are rocks originated from amphibolites and biotite plagioclase schists.

According to Nickel (1952) changing conditions generate mineral facies whose metamorphic and magmatic parageneses are fairly similar.

The formation of the ultrabasic and basic bodies of Rauhamäki-Kuronlahti can best be seen as the result of metamorphic and tectonic processes. They may have originated during a process of migmatitization, for instance from amphibolites, diabase dikes and similar rocks.

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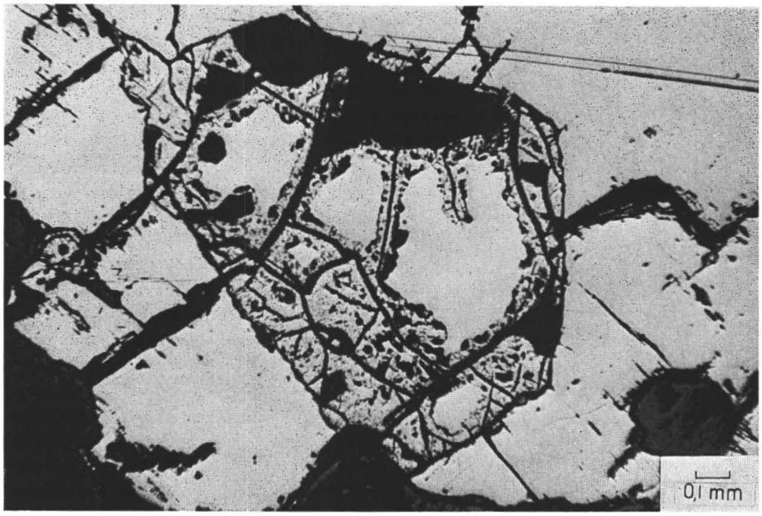

Fig. 1. Euhedral pentlandite crystal in pyrrhotite. The pentlandite has partly altered into bravoite. Quartz diorite. Niemilahti ore block. Polished section, 1 nic., Pisamaniemi, Kangaslampi.

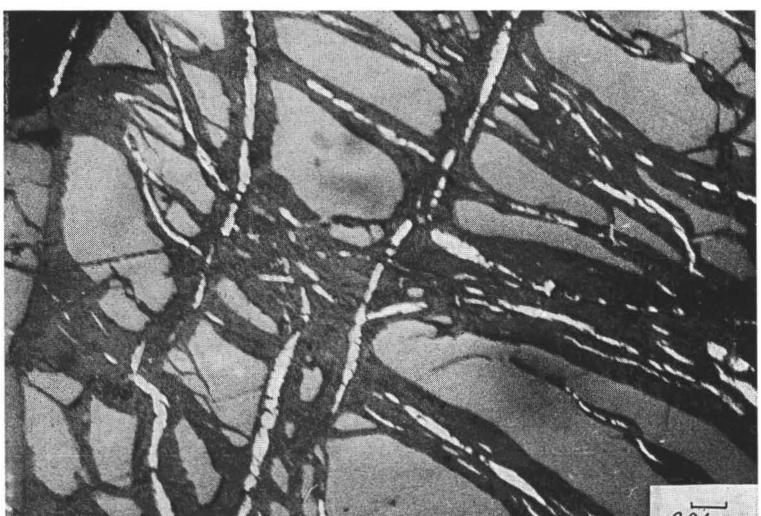

Fig. 2. Magnetite bands (white) in serpentine (dark grey). The serpentine is in the fissures of olivine. Polished section, 1 nic., Rauhamäki, Kangaslampi.

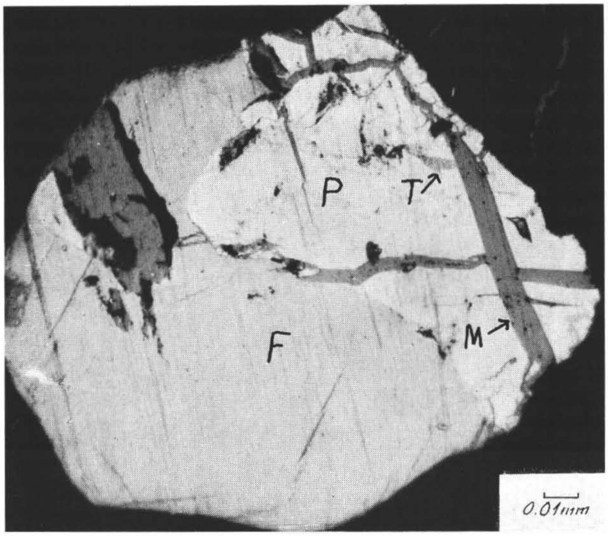

Fig. 3. Pyrrhotite (F) and pentlandite (P) form together mixed crystals, intersected by magnetite veins (M). The pentlandite contains tetragonic iron sulfide (T). Peridotite. Polished section, 1 nic., Rauhamäki, Kangaslampi.

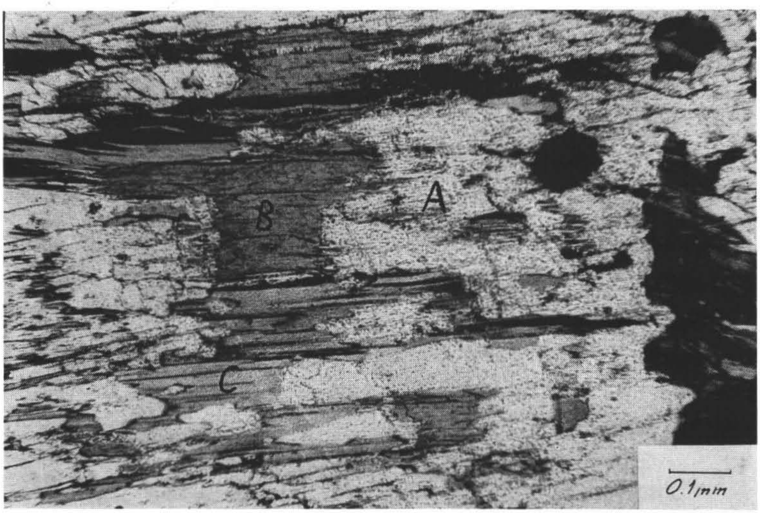

Fig. 4. Actinolite (A). In the direction of its cleavages biotite (B) and chlorite flakes (C). Peridotite. Thin section, crossed nicols. Rauhamäki, Kangaslampi.

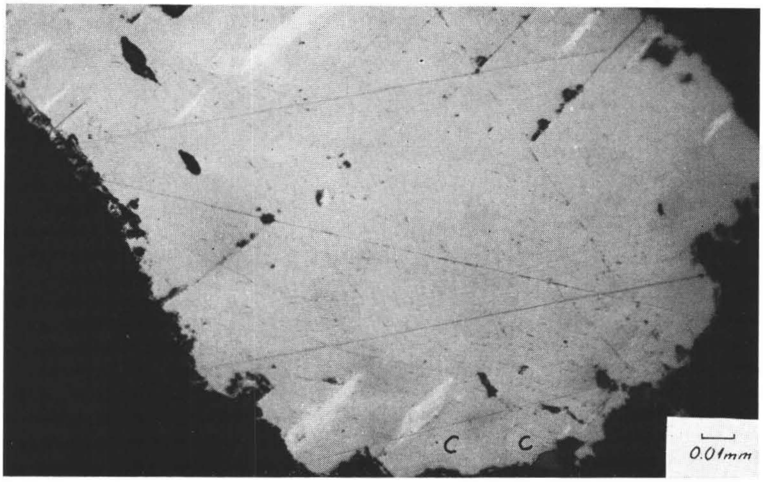

Fig. 5. Exsolution lamellae of pentlandite in pyrrhotite. The lamellae have the direction of the pyrrhotite (1000) surfaces. (C) is chalcopyrite. Plagioclase-bearing augite-hornblende peridotite. Polished section. Timonsalo, Kuronlahti, Leppävirta.

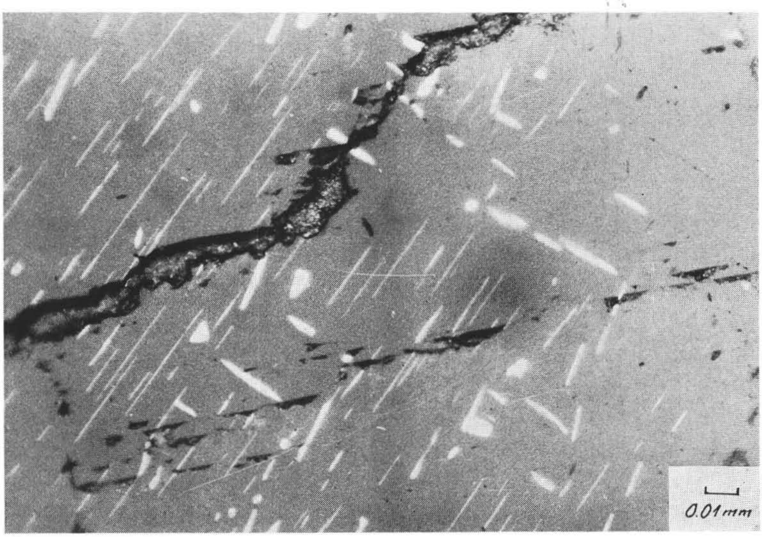

Fig. 6. Augite with ilmenite in its cleavages. Plagioclase bearing augite-hornblende peridotite. Polished section, 1 nic., Timonsalo, Kuronlahti, Leppävirta.

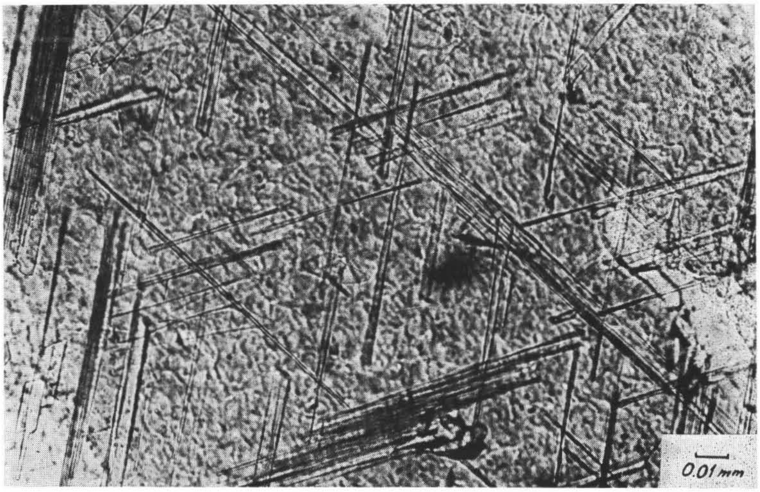

Fig. 7. Shagenite needles in biotite running in three different directions at angles of almost 60°. Thin section. 1 nic., Island between Timonsalo and Unnukansalo, Kuronlahti, Leppävirta.

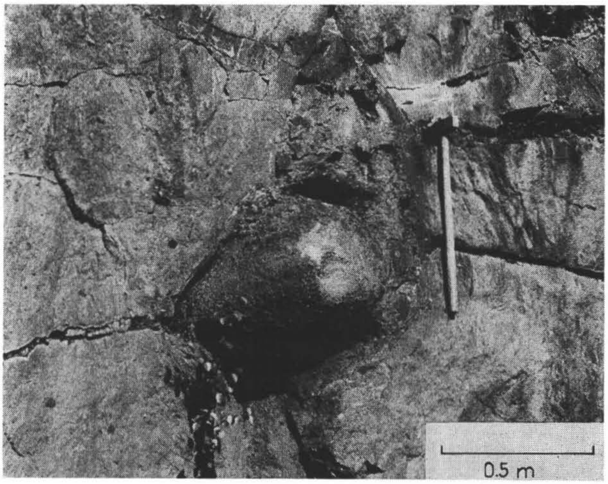

Fig. 8. Small biotite-hornblende gabbro body. Rauhamäki, Kangaslampi.