GRØNLANDS GEOLOGISKE UNDERSØGELSE
BULLETIN No. 4.

THE ULTRABASIC ROCKS AT TOVQUSSAQ, WEST GREENLAND

A CONTRIBUTION TO THE PERIDOTITE PROBLEM

BY

HENNING SØRENSEN

WITH 22 FIGURES IN THE TEXT
AND 9 PLATES

REPRINTED FROM "MEDDELELSER OM GRØNLAND" BD. 136. No. 4

KØBENHAVN
BIANCO LUNOS BOGTRYKKERI
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CONTENTS

Abstract.

Ultrabasic rocks of bahiaitic type occur as bands in amphibolite at Tovqussaq. It is concluded that the ultrabasics are formed *in situ* in the amphibolite in zones of stress concentration.

Stress of high order in connection with chemically, and mechanically generated heat is regarded to be the cause of the metamorphic differentiation that was responsible for the formation of the ultrabasic rocks.

From the results of the examination of the hypersthenes of the Tovqussaq region it seems reasonable to assume that the hypersthene may be used as a geological thermometer.

Ultrabasic rocks are described from other pre-cambrian areas and the peridotites of orogenic zones are discussed. The interpretation of the ultrabasic rocks as being the products of metamorphic, rather than magmatic processes seems to have general validity.

Finally, the peridotites are compared with eclogites which were probably formed in a similar way.

1. INTRODUCTION

During the geological mapping of South-West Greenland, carried out under the leadership of Professor, Dr. phil. A. Noe-Nygaard for the Geological Survey of Greenland, I had, in the summers of 1949 and 1951, the opportunity of investigating a considerable number of occurrences of ultrabasic rocks in the Sukkertoppen, and in the Godthaab district.

On the small island of Langø, on which the fishing station Tovqussaq is situated, the mode of occurrence of the ultrabasic rocks was so interesting and instructive that a general discussion of the formation of this rock type seems to be justified.

The structure of the area has been described by A. Berthelsen (1950). The formation of the ultrabasic rocks of Langø appears, as it is discussed in the present paper, to be determined by the structural evolution of the Tovqussaq region. The occurrences in question seem to be of particular significance for our understanding of the origin of ultrabasic rocks in orogenic zones.

I have previously described the rocks of the Sukkertoppen district in an unpublished prize-dissertation, now kept at the University of Copenhagen. An outline of this paper was presented to the Scandinavian geologists assembled in Copenhagen in May, 1951 (Sørensen, 1951a).

The greatest part of the laboratory work was carried out at the Mineralogical Museum of the University of Copenhagen. Besides, I have had the opportunity of working at the Geological Department of the University of Durham, England in the spring of 1950 and for the time being at the Mineralogical Museum in Oslo, Norway. I wish to thank the directors, the staffs, and collaborators of these institutes for much kind help and many valuable discussions.

Professor, Dr. phil. A. Noe-Nygaard, director of the Mineralogical Museum in Copenhagen, made my working in Greenland possible and placed the facilities of his institute at my disposal. I am indebted to Professor Noe-Nygaard for the interest he has taken in my work, for his constructive criticism, and not least for his inspiring colloquies, in which the ideas presented in this paper have been discussed at several occasions. My best thank is due to all participants in these colloquies.

Dr. philos. HANS RAMBERG introduced me to the geology of precambrian areas. I have profitted much from his insight in petrology and our many discussions in the field have been of invaluable help to me. I wish to thank Dr. RAMBERG, Mrs. M. L. RAMBERG, and all my colleagues in the field work for the wonderful summers I have spent in Greenland.

Mr. A. BERTELSEN, who undertook the structural analysis of the Tovqussaq area, has offered me much invaluable help during the work. He placed in the most unselfish way his observations and collections at my disposal and kindly allowed me to reproduce his maps of the Tovqussaq peninsula.

Professor, L. R. WAGER, Sc. D. now Oxford, provided me with samples of ultrabasic rocks from East Greenland and offered me excellent working conditions during my stay in Durham in 1950.

I have learnt much from discussions with Professor, Dr. philos. T. F. W. BARTH, Mr. G. KULLARUD, civil engineer, Mr. A. MUAN, civil engineer, and Mr. T. GJELSVIK, cand. real., all of the Mineralogical Museum in Oslo. Mr. KULLARUD and Mr. GJELSVIK have critically read parts of the manuscript.

My visit to the Outer Hebrides was spent in the agreeable company of Mr. R. PHILLIPS, B. Sc., who also critically read the chapter on South Harris.

My best thanks are also due to Mr. HANS PAULY, mag. scient, who kindly examined the ore-minerals of some of the rocks, to Dr. SVEN PALMQVIST, Fagersta, Sweden, who undertook three chemical analyses, to Mr. J. BONDAM, Copenhagen for valuable discussions, and to Miss R. HANSEN, Copenhagen, and Miss R. GULLIKSEN, OSLO, who helped me with the preparation of the drawings.

The microphotographs were taken by Miss B. MAURITZ, Oslo, and by Mr. CHR. HALKIER, Copenhagen. Mr. HALKIER kindly prepared the field photographs for reproduction.

Parts of the manuscript were translated into English by Mrs. AASE HOLST, Copenhagen, and Mr. J. CRADDOCK, Northampton, kindly corrected the English language of a great part of the manuscript.

The thin sections were prepared by Mr. J. LOMAX, Bolton, Mr. G. O'NEILL, Durham, Mr. C. A. JENSEN, Copenhagen, and Mr. A. GRANLI, Oslo.

The preparation of the maps was made possible through the kindness of the Geodetic Institute, Copenhagen.

May, 1952.

GEOLOGICAL DESCRIPTION OF THE ULTRABASIC ROCKS ON LANGØ

A. BERTHELSEN (1950 pp. 559 and 568) has briefly described the geological structure of the country north of Tovqussaq. A few supplementary notes should be added, in relation to the following.

To the north, north of the Sdr. Isortoq fiord (see map, fig. 1) the rocks were formed under granulite facies conditions. Further south between the fiords of Sdr. Isortoq and Alángua a para-gneiss series occurs which is built up of alternating layers of amphibolite (composed of green hornblende, diopside, plagioclase (andesine to bytownite) and quartz), garnet-sillimanite-mica-schists and various types of gneissic rocks. One of these, a hornblende-bearing, dark gneiss (quartz-dioritic), conformably encloses lenticular masses of ultrabasic rocks. The parageneses of the para-gneiss series show that its rocks were formed under physico-chemical conditions corresponding to the upper part of the amphibolite facies of regional metamorphism. The area is highly folded, the folding being intensified to the south where the para-gneiss series passes smoothly into the homogenous, isoclinally folded Finnefjeld-complex, which is most typically developed between Alángua fiord and the Uivfaq peninsula. This area consists almost entirely of the quartz-dioritic "Finnefjeld gneiss". Further south again there is a comparatively smooth transition into an area the conditions of which are on the whole similar to those north of the Finnefjeld complex. Continuous layers of amphibolite again appear south of Uivfaq.

The Tovqussaq peninsula is characterized by a peculiar closed structure (figs. 2 & 3), the amphibolite layers being arranged almost concentrically around the Tovqussaq mountain. A. BERTHELSEN (1950, pag. 559) describes the structure as a tilted, elevated dome. The dome structure has caused granulite-facies conditions locally, indicated e. g. by the presence of hypersthene in gneisses and amphibolites.

The district south of the Tovqussaq peninsula was also formed under granulite facies conditions.

The ultrabasic rocks on Langø. Unfortunately, it was only possible to visit the ultrabasic rocks on the island itself, but according to A. BERTHELSEN ultrabasic rocks have also been observed in the area north of Langø.

At high tide Langø is divided into a northern and a southern part. The former has been more closely studied although no detailed investigation has yet been made.

On the east coast of the northern part of Langø continuous bands of amphibolite occur which can be followed northwards on the peninsula. The amphibolite is a medium-grained, foliated rock bearing hypersthene. It carries occasionally garnet, but this mineral is present only in small quantities as small grains. The amphibolite has thin, light-coloured schlieren rich in plagioclase and also more prominent light-coloured veins with concentrations of the dark minerals, viz. hornblende and hypersthene. These pegmatitic secretions are usually elongated parallel with the foliation of the amphibolite, and the material rich in plagioclase occurs especially at the ends of the basic concentrations.

The most remarkable feature of the amphibolites is the conformable bands of ultrabasic rocks (fig. 4). Two such bands were observed in the amphibolite layer at the beach on the east coast of the northern island. At their broadest part they are one meter and two meters respectively. They can be traced for about 50 meters and then they pass into normal-looking amphibolite. This happens as follows: At the ends of the ultrabasic bands pegmatites occur, both as pegmatitic masses and as schlieren in the ultrabasic rock which consequently appears striated with alternating light and dark bands. In other words, the ultrabasic bands wedge out at the ends.

As was to be expected, the marginal border between amphibolite and ultrabasite is somewhat different. The transition from foliated amphibolite to more coarse-grained and fairly massive ultrabasite covers about one centimeter. The ultrabasite contains a little plagioclase in its outer part, and towards the ultrabasite, the amphibolite contains a few thin pegmatitic veins which, normally, may be traced for only a few centimeters, but locally, they may occur as small pegmatitic masses in the border itself. The pegmatite consists chiefly of dark plagioclase and hypersthene grains up to a size of one centimeter. On the border between pegmatite and amphibolite concentrations of hypersthene and hornblende may be found.

The ultrabasic rock varies somewhat in appearance. It may have linear schistosity as a result of the orientation of the hornblende grains. The rock is more coarse-grained than the amphibolite but in certain parts it is similar to the latter differing from this only in its lack of plagioclase. Locally the rock is very massive with fairly large individuals of hypersthene and a groundmass of a dark hornblende and a clear green diopside.

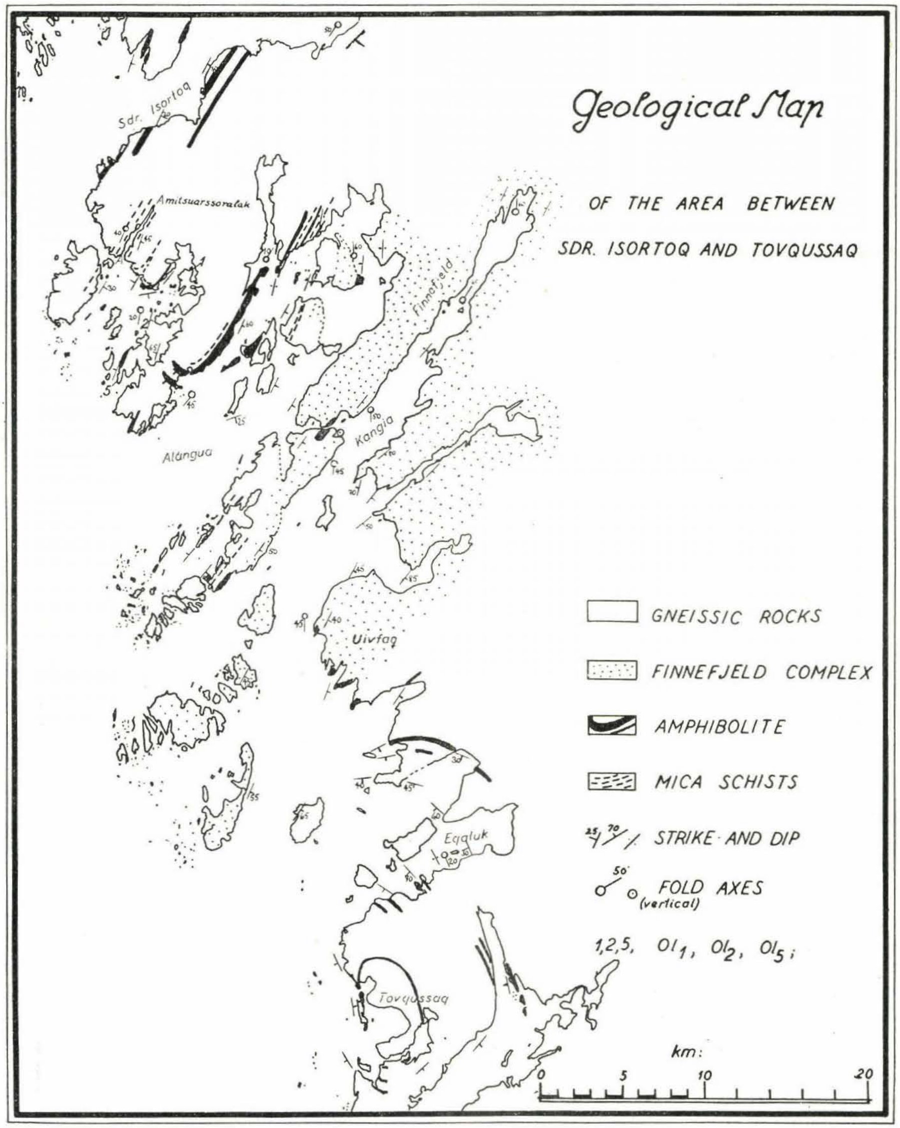

Fig. 1. Geologic map of the Sdr. Isortoq-Tovqussaq region.

Quartz-feldspar pegmatites, intersecting both amphibolite and ultrabasite, are present and, as a rule, quite thin. Their presence shows that the great stress prevailing during the deformation has locally exceeded the tensile strength of the two rocks.

Further west on the northern island the amphibolite no longer occurs as continous layers but is intimately mixed with the hypersthene-carrying gneiss. Consequently the amphibolite is much deformed and broken up into smaller fragments which occur as thin bands in the

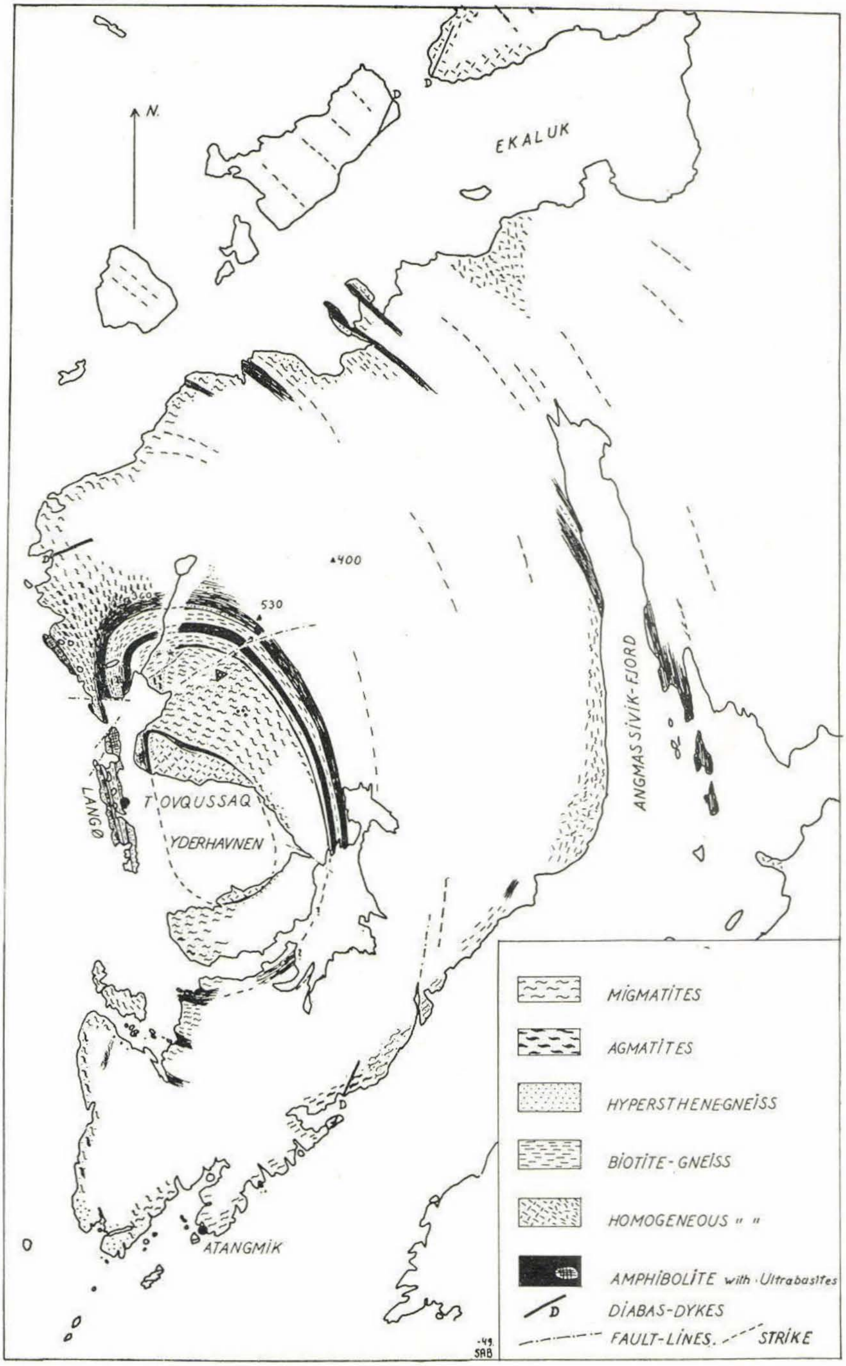

Fig. 2. Geologic map of the Tovqussaq peninsula. Scale 1 : 125.000.

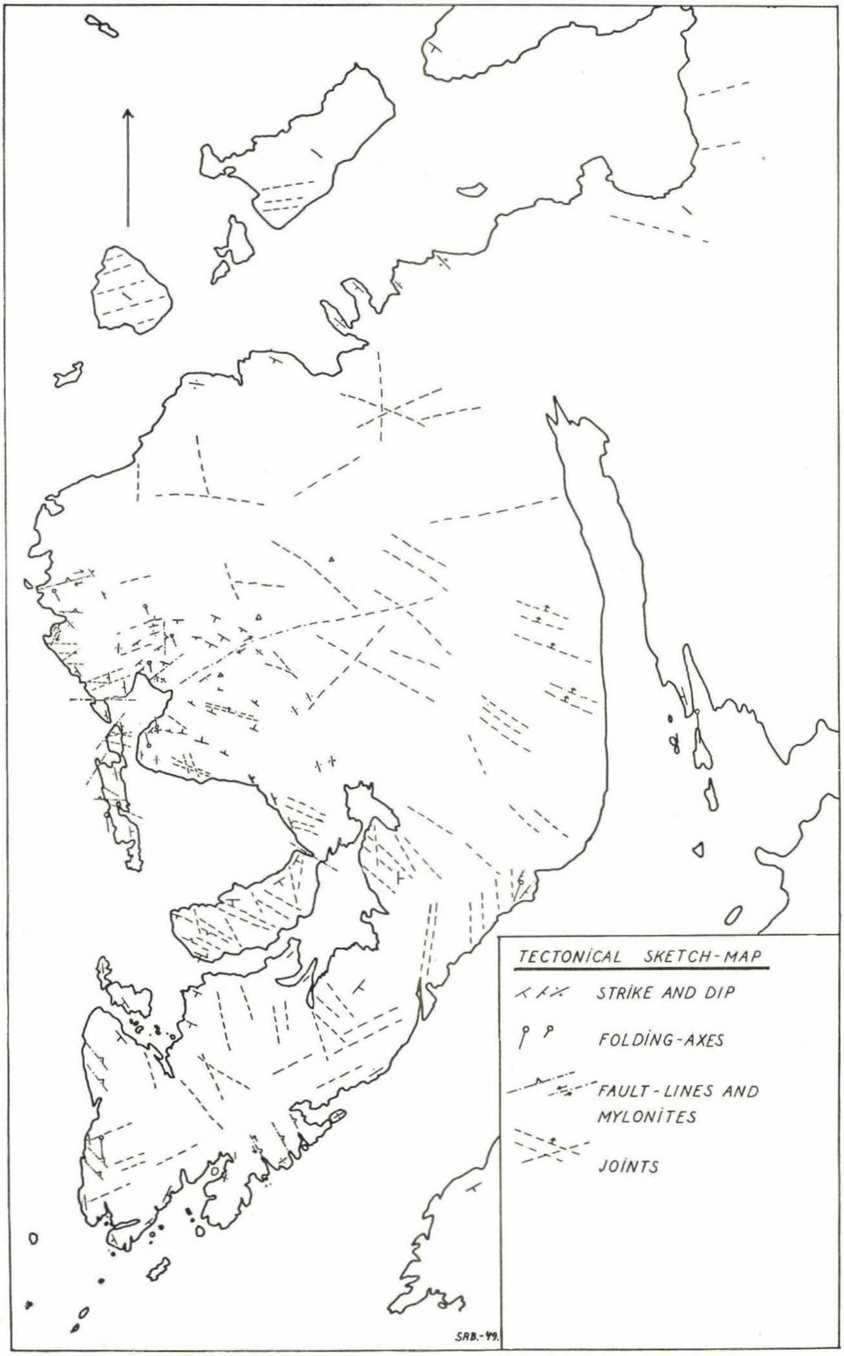

Fig. 3. Tectonic sketch map of the Tovqussaq peninsula.
(Figs. 2 and 3 from A. BERTHELSEN (1950) with kind permission).

gneiss. These bands may be folded. An ultrabasic rock is present and occurs in "strings of beads", the individual "beads" being fairly small. It is curious that the size of the "beads" shows a smooth decrease northwards. Each string has its largest "bead" to the south, the smallest to the north. The smallest "beads" are of amphibolitic appearance, presumably on account of material added from the surrounding gneiss, from which the ultrabasites are separated by a thin, black rim of horn-

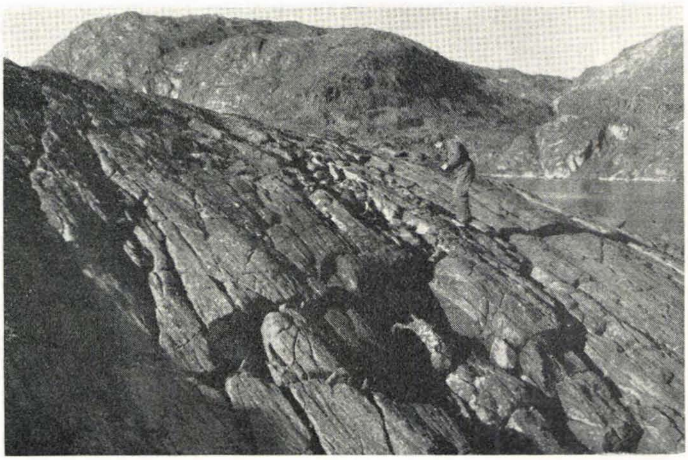

Fig. 4. Ultrabasic band in amphibolite. The east coast of Langø. (HANS RAMBERG phot.).

blende. These occurrences can be studied particularly well just north of the small "sound" which, at high tide, divides Langø into two parts.

On the west coast of the island still another mode of occurrence may be observed. Here the amphibolite has been reduced to thin, dark, conformable bands in the hypersthene-bearing, almost pegmatitic gneiss (fig. 5). Ultrabasic rocks occur as lenticular bodies conformably enclosed in the gneiss (fig. 6). They are usually larger than those mentioned above though rarely more than two meters in diameter. The lenses are sharply separated from the gneiss against which they have thin reaction rims. Here, too, there is a decrease in size towards the north and the northernmost small lenses are usually of amphibolitic appearance. In a row of ultrabasic bodies it may be seen that the southernmost consist of fairly large hypersthene grains with numerous inclusions of black hornblende. Diopside occurs as small, transparent, green grains. Then come masses where the rock is more fine-grained and equigranular consisting of hypersthene, hornblende, diopside and a little plagioclase. The northernmost bodies are pure amphibolites, which consist chiefly of hornblende and plagioclase. A small quantity of hypersthene also occurs, especially in fairly large "secretions" together with hornblende. Thus, there are two types of amphibolite in this locality, the one, a component of the gneiss is the remains of the original amphibolitic bands, the other is formed at the expense of the ultrabasic rocks.

Fig. 5. Ultrabasic body in gneiss which has amphibolitic bands. The west coast of Langø.

A probable explanation of the mode of occurrence of the ultrabasics is that there was, originally, a series of amphibolite bands which like the east coast amphibolite have

Fig. 6. Ultrabasic lenses enclosed conformably in banded gneiss. The west coast of Langø.

had ultrabasic schlieren of varying magnitude. In consequence of the strong shearing stress prevailing during the structural development of the area, the amphibolite bands were subjected to a very pronounced deformation.

As mentioned above, traces of the deformation may be seen in the amphibolite band on the east coast in the pegmatite-filled cross joints. The tensile strength of the amphibolite was, however, only exceeded quite locally.

Towards the west, in the central part of the island, the deformation was more pronounced and the amphibolite was divided into fragments which were incorporated in the gneiss. The more rigid ultrabasic bands were broken into boudins and arranged in more or less parallel rows of small lenses.

On the west coast the deformation was even more pronounced. While amphibolitic material still plays an important rôle in the central part of the island, it has been almost completely digested by the gneiss occuring as a subordinate component of the latter on the west coast. The ultrabasic boudins, on the other hand, were more resistant to the "homogenisation" accompanying the deformation. Only the smaller lenses and the margins of the larger ones were affected. The rigid ultrabasic bands may have had a controlling influence on the deformation. The amphibolite bands have probably acted in a similar way during the deformation of the Tovqussaq area (see page 34).

Conditions on the west coast of the southern part of Langø correspond to those on the west coast of the northern part, but the ultrabasic bodies are still larger than those in the north, attaining a size of 10 meters across. As in the northern part of Langø, the lenses decrease in size northwards, indeed even more so here. The deformation may have brought this outstanding feature about in the following way. First, the ultrabasic bands are boudined, next, the "tectonically most exposed" fragments are metasomatically transformed and partly digested by the gneiss. In accordance with this the northernmost fragments in each row are the most transformed. On the other hand, the progressive decrease in size northwards may be original, thus indicating that the continuous ultrabasic bands were wedge-shaped. This may be compared with the wedge-shaped peridotites of South Harris (see page 8 and page 57).

The ultrabasic rocks of this southern part of Langø vary somewhat in appearance. Massive rocks with hypersthene porphyroblasts of as much as two centimeters are common. The hypersthene grains are embedded in a fine-grained groundmass consisting of hypersthene, hornblende and, in certain occurrences, diopside. Plagioclase and, occasionally, biotite occur in the marginal parts of the lenses.

The ultrabasics are often schistose. This structure may be formed when the minerals are arranged in more or less monomineralic layers. Biotite is prominent in these rocks. A few lenses are foliated in consequence of a linear arrangement of the hypersthene grains. The schistose rocks are occasionally folded. The hypersthene layers in particular appear as beautiful small "ptygmatic" folds.

The ultrabasic bodies are conformably embedded in hypersthene-bearing gneiss which here, as in the western part of the northern island, contains thin bands of amphibolite.

The gneiss has cross-cutting as well as conformable pegmatites along which displacement may have taken place. A more powerful movement in later stages of the deformation has resulted in the formation of mylonites (v. A. BERTHELSEN, op. cit.).

The ultrabasics are often cut by pegmatites. The boundaries against the latter vary greatly in development, this will be demonstrated by a few examples:

1. A fairly large ultrabasic mass is enclosed in and penetrated by pegmatites composed of blue quartz, purple feldspar and hypersthene. The intersecting pegmatites are rather fine-grained. The ultrabasic rock is clearly separated from the pegmatite without development of a border facies.

2. Another ultrabasic body, which in its central part contains the usual hypersthene grains in a fine-grained groundmass, has, towards the pegmatite, a stratified appearance with alternating broad layers of hypersthene and narrow ones of hornblende. Right against the pegmatite is a fine-grained rock consisting of plagioclase and hypersthene. In certain parts of the border there is no layering, and the plagioclase is seen to have "forced" its way in between the dark minerals, apparently by replacement processes.

3. The transition from ultrabasite to pegmatite may be more or less gradual or almost diffuse. This means that the ultrabasite, as it approaches the pegmatite, contains more and more fine-grained pegmatitic material. At first this is present in small quantities only, but it gradually increases at the expense of the ultrabasite which furthest out is preserved merely as thin stripes almost at right angles to the border. The fine-grained, light-coloured rock passes into a normally coarse-grained pegmatite. Hypersthene is very common in the fine-grained rock, less frequent in the pegmatite itself, in which a small quantity of biotite occurs.

4. An ordinary border type is a hornblende-rich zone between ultrabasite and pegmatite. An instructive example of this can be seen in the southern part of the island in a rather small ultrabasic inclusion in pegmatitic material. The greater part of the border has a narrow black rim of hornblende (cf. p. 31, no. 13434). A thin intersecting pegmatite has a similar rim. The northern border of the ultrabasite is, however,

of quite a different appearance. The ultrabasite there has schlieren of pegmatitic material (cf. example 3). The end-product is an amphibolitic rock with a gradual (diffuse) transition into the pegmatite. Traces of the transformed ultrabasic rock can be observed in the pegmatite in continuation of the dark layers of the "amphibolite". These traces appear as shadowy stripes in the pegmatite, providing a beautiful example of replacement.

5. Finally, in certain places there is a narrow biotite zone between ultrabasite and pegmatite. Small flakes of biotite may occasionally be

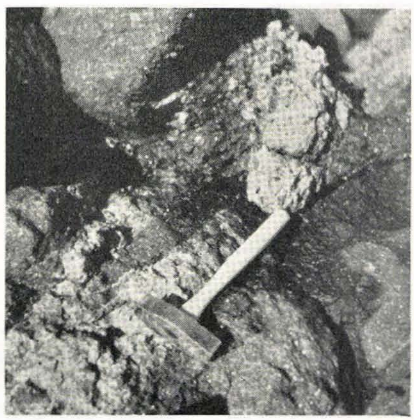

Fig. 7. Biotite border between ultrabasite and cross-cutting pegmatite. The ultrabasic inclusion in the pegmatite is described as no. 13436 on page 31. The west coast of Langø.

found a few centimeters inside the ultrabasic rock. The border is not monomineralic but contains, in addition to biotite, hornblende and hypersthene. The pegmatite is very rich in hypersthene towards the ultrabasic rock and also contains large flakes of mica. Hornblende is present in a very narrow zone inside the biotite but does not otherwise occur in the pegmatite. The pegmatite may have small inclusions of ultrabasite which, apart from containing biotite, are of the usual ultrabasic appearance. The inclusions also have biotite adjacent to the pegmatite (see fig. 7 and p. 31, no. 13436).

Although the petrogenesis is discussed in greater detail below it may be useful at this point to examine the processes responsible for a) the very different development of the borderzones between ultrabasite and pegmatite and b) the fact that the ultrabasic rocks in some cases are almost unaltered, while they, in other cases, are transformed to a great extent into amphibolitic rocks or are even replaced by salic minerals.

The explanation seems to be found in differences in tectonical environment during the deformation, just as the pegmatites themselves are probably tectonically determined. Thus RAMBERG (e. g. 1944, pag. 109) states that the pegmatites are formed by consolidation of the migrating ions of the dispersed phase at places of low mechanical pressure, for instance in the present example in fissures in the rigid ultrabasites (see fig. 8).

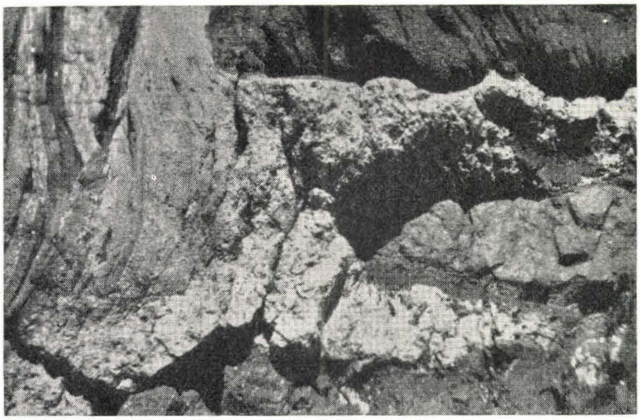

Fig. 8. Pegmatite in fracture in ultrabasic body. The west coast of Langø. (A. BERTHELSEN phot.).

As the different border-types between ultrabasites and pegmatites were formed in the same level of the crust during the same period of deformation, the varying appearances may be explained by assuming a close connection between tectonical processes and border-type. At the tectonically "most exposed" places where "frag-

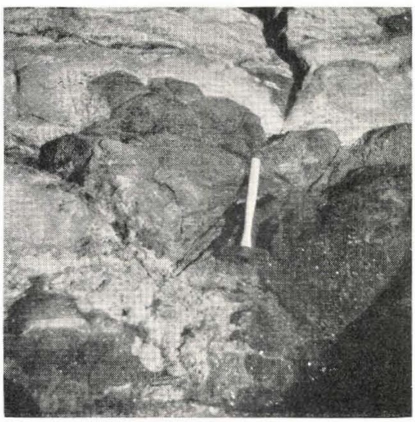

Fig. 9. Transition from ultrabasic rock to amphibolite in an inclusion in gneiss. The west coast of Langø.

mentation" is most pronounced, numerous fissures are formed into which selected ions of the dispersed phase migrate (RAMBERG, op. cit. p. 104). Thus, the replacement processes are accelerated at these places.

In addition migration of material along pressure gradients from places of high, to places of low, chemical activity is determined by the kind of deformation.

The connection between tectonics and e. g. the border types described seems to be beyond doubt. Unfortunately, the time available for the field-work did not allow a closer study of this very interesting relationship.

In some ultrabasic bodies the gradual change from ultrabasite to amphibolite from south northwards can be observed in one continuous body. The ultrabasite of fig. 9 contains in its southern end hypersthene grains of 0,5 centimeter in a hypersthene-hornblende-groundmass. Towards the north the grain size decreases, the rock becomes equigranular, and plagioclase appears. Finally, in the northern end, we have a typical, schistose amphibolitic rock with a great deal of plagioclase and with grains of hornblende and hypersthene in a linear arrangement. This transition will be dealt with in detail on p. 37, no. 13437. Thus, the transition, which on page 13 was described as occuring successively in several subsequent small lenses, is here found in one lense.

PETROGRAPHICAL DESCRIPTION OF THE MOST IMPORTANT ROCK TYPES

Some of the amphibolites of the area will be dealt with first. The ultrabasic rocks will be described next and finally examples of the retrograde transformation will be given.

Hypersthene Amphibolite from Eqaluk. (No. 13415).

This rock is found at one of the western points of the south coast of the bay Eqaluk due north of Tovqussaq. It is one of the most northerly hypersthene-bearing amphibolites and is thus situated in the outer part of the region disturbed by the dome.

The rock has hypersthene porphyroblasts of half a centimeter in a finer and more equigranular groundmass with grains of about 0.1 centimeter. The groundmass consists of app. 50 $^0/_0$ hornblende, 45 $^0/_0$ diopside, and less than 5 $^0/_0$ plagioclase and hypersthene.

Hornblende is present as prismatic, rather irregular and, on the whole parallel grains. $2\,V = (\div)\ 80°$, $c : \gamma$ ab. $15°$ and $n\gamma = 1.668$. The pleochroism is: γ dark green, β olive, and α yellowish. There is a great deal of ore pigmentation particularly in the outer parts of the grains.

Diopside occurs in equidimensional grains often in small aggregates in which the mutual borders of the grains are almost rectilinear. $2V = (+)\ 57°$, $c : \gamma = 41°$, and $n\beta = 1.694$. The aggregates are, from the rim, intergrown with hornblende against which the diopside has rather lobed boundaries. Thus, the amount of hornblende increases at the expense of the diopside. This is also proved by the numerous small rounded, pigmented inclusions of diopside in the hornblende and by "growth lines" of ore-pigmentation in the latter.

A very small amount of plagioclase occurs in tiny, rounded grains. The anorthite content varies from 45 to 50 $^0/_0$. Small needles in the plagioclase are probably apatite. It should be noted that other parts of the rock have a greater amount of plagioclase.

Apart from the small amount of plagioclase the groundmass of this rock resembles the para-amphibolites from the Isortoq-Alangua district (see page 59). This means that the rock was probably formed at the expense of an original sedimentary rock rich in Ca-Mg. As diopside in most cases has sharp contact against the plagioclase, the small amount of the latter is not due to this mineral having been used up by reaction with diopside, but to a poor supply of its components during the transformation of the sediment. Consequently, the hornblende may perhaps be regarded as an Al-poor, Fe-rich, actinolitic hornblende, an assumption which is in accordance with the optical data (compare BARTH 1930, page 226 and REYNOLDS 1947, page 415).

Hypersthene occurs in irregular grains up to half a centimeter in size which enclose all the other minerals. It is highly pleochroic with $2V = (\div)\ 49°$, $n\gamma = 1.719$, which according to POLDERVAART (1950) corresponds to 42 $^0/_0$ ferrosilite (fs). The inclusions in the hypersthene are principally hornblende of almost the same appearance and orientation as the hornblende of the groundmass, but occurring as smaller, more corroded grains (see plate 1, fig. 1). Diopside is present as a few rounded, pigmented grains. It should be noted that the greatest number and the largest grains of plagioclase occur either enclosed in the hypersthene or in the amphibolite just outside it. The hypersthene is clearly the last formed mineral.

Hypersthene also occurs as a few independent small grains.

Amphibolites from the Tovqussaq Peninsula.

The amphibolite (4040) south of the one described above (see map, fig. 10) resembles the latter apart from its larger amount of plagioclase. The rock is composed of plagioclase (53 $^0/_0$ an), hornblende, diopside, hypersthene, a little quartz and ore. As in no. 13415 hypersthene ($2V = (\div)\ 59°$, $n\gamma = 1.714$) is the last formed mineral and encloses the other minerals of the rock. Similar rounded hornblende inclusions are seen and in addition irregular diopside inclusions, the cleavages of which are parallel with those of the hypersthene.

The amphibolite (4048) south of 4040 has hypersthene ($n\gamma = 1.704$) in some parts while other parts lack this mineral. The rock consists then of hornblende, diopside, plagioclase and ore. Remnants of diopside aggregates are seen, the diopside is clearly under transformation into hornblende and has e. g. hornblende on the cleavages in some cases.

A sample of amphibolite (4013) collected near the bottom of the inner harbour consists of hypersthene ($2V = (\div)\ 67°$, $n\gamma = 1.707$), hornblende, plagioclase (50 $^0/_0$ an), ore and a little diopside. The latter mineral is only present as small, rounded inclusions in the hornblende. Hypersthene and hornblende are present in equal amounts as rather large grains. The hypersthene has occasionally inclusions of hornblende, but its porphyroblastic habit has disappeared. The hypersthene seems

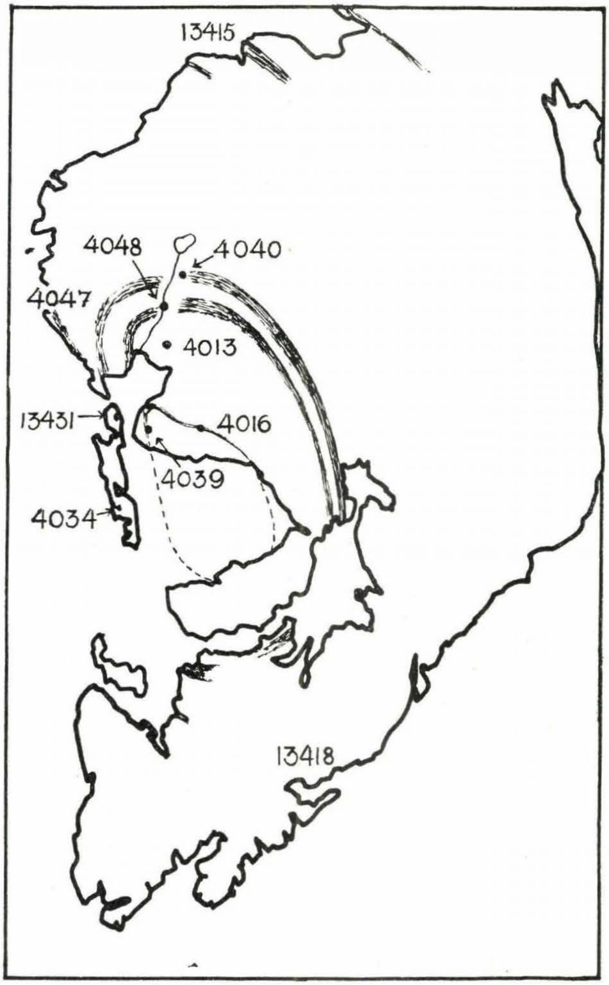

Fig. 10. Map of the Tovqussaq region showing the localities of the amphibolites described on page 20).

to be a true component of the rock being in equilibrium with the surroundings.

This is probably also the case in the amphibolite band (4016) just north of the outer harbour. This rock is composed of plagioclase (45—50 $^0/_0$ an), diopside, hornblende, hypersthene and ore, the latter occurring as rather large grains. The hypersthene has $2V = (\div)\ 57°$ and $n\gamma$ 1.715. Diopside aggregates are present but are penetrated by hornblende and plagioclase. The diopside is to some extent transformed into hornblende, a transformation which seems to follow the cleavages of the diopside in many cases.

In the southern part of the peninsula, near the outpost Atangmik a garnet-hypersthene-amphibolite (13418) occurs. It shows the following paragenesis: plagioclase (68 % an), hornblende, hypersthene ($n\gamma$ = 1.728), garnet and biotite. The hypersthene is present in much corroded grains clearly in process of replacement by hornblende and probably by garnet too, as the latter can have hypersthene inclusions of common orientation. Thus, garnet and hornblende seem to be secondary after hypersthene.

Hypersthene Amphibolite from the East Coast of Langø. (No. 13431 b).

The rock contains fairly large (0.2 centimeter), elongated, irregular grains of hornblende. Between these grains, the other minerals of the rock occur in an equigranular aggregate consisting of rather small (0.05 centimeter), equidimensional grains with rounded, somewhat irregular outlines. The minerals and their approximate quantities are: hornblende 40 %, hypersthene 22 %, plagioclase 19 %, diopside 18 % and ore 1 %.

The hornblende is of a slightly lighter colour than that of 13415. $2V = (\div)\ 86°$, $c : \gamma = 15°$, $n\gamma = 1.667$. The pleochroism: γ dark green, β yellowish-green, and α yellowish. It is fairly rich in ore inclusions and has more irregular outlines than the hornblende of 13415.

The diopside occurs in part as small, independent grains, in part intergrown with or enclosed in the hornblende. Traces of diopside aggregates can be observed. The optical data of the diopside are: $2V = (\div)$ 57°, $c : \gamma = 41°$, $n\beta = 1.694$, corresponding to ab. 34 % hedenbergite according to KENNEDY (1947).

Hypersthene is present partly in aggregates consisting of several small, rounded grains, partly as independent, slightly larger grains which may have inclusions of hornblende (plate 1, fig. 2) and more rarely of diopside. $2V$ is $(\div)$ 64°, $n\gamma = 1.708$.

Plagioclase occurs as irregular grains with alternating broad and thin lamellae (albite and pericline twins). The structure is normal zonal, and the composition varies from 65 to 58 % an.

The hypersthene seems to be a true component of the rock, all traces of porphyroblastic development have disappeared.

In the border zone towards the ultrabasic schlieren, the amphibolite has thin pegmatitic veins consisting of plagioclase, hypersthene and small amounts of quartz. A hornblende slightly darker than that of the amphibolite occurs associated with the hypersthene. The latter has often finely distributed ore pigmentation, $n\gamma$ is found to vary from 1.709 to 1.715.

The first value was found in hypersthenes of thin veins, the last-named in larger pegmatitic masses. 2 V was determined to be (÷) 66° close to the amphibolite.

Ultrabasic Schlieren in the Amphibolite (No. 13431 a).

The rock is equigranular with a grain size of about 0.1 centimeter. Most of its components occur as equidimensional grains of a more or less pronounced polygonal shape. The components are: hypersthene 40 %, hornblende 38 %, diopside 9 %, spinel and ore 11 %, and olivine 2 %. In addition a few very small plagioclase grains are seen (plate 2, fig. 3).

The hypersthene has a tendency to form aggregates consisting of several small, fairly polygonal grains. The cleavages are coated with magnetite dust and in addition the hypersthene contains a good deal of finely distributed pigment. 2 V = (÷) 84°, $n\beta$ = 1.686, $n\beta$ = 1.692, giving 15 and 19 % fs respectively according to POLDERVAART (1950). The pleochroism is well pronounced (see page 46).

The hornblende may occur as polygonal grains but is also present as parallely arranged elongated grains corresponding to the parallel stripes of hornblende in the surrounding amphibolite. It has, as the hypersthene, coatings of ore on the cleavages. The colour is fainter than in the amphibolite, the pleochroism is: γ green, β yellow-green, and α colourless). 2V = (÷) 88°, c : γ = 19°, and $n\gamma$ = 1.658.

Diopside is present in very few equidimensional grains bordering on both hypersthene and hornblende. The optical data are: 2V = (÷) 57°, c: γ = 41°, and $n\beta$ = 1.689, corresponding to 27 % hedenbergite.

A green pleonaste associated with magnetite occurs in the hypersthene grains as well as on the mutual hypersthene borders and on the borders between hypersthene and hornblende. In the latter mineral spinel is extremely rare. The spinel grains may be associated in small aggregates.

Olivine occurs as a few small, rounded grains, particularly between the grains of the hypersthene aggregates but also bordering on the hornblende. It has crusts of ore on the irregular fractures and is as a rule separated from the adjoining minerals by a zone of ore. The determination of the axial angle of the olivine by means of the universal stage was difficult, but a number of measurements gave values about 90°. $n\gamma$ was determined to be about 1.690. Both determinations correspond to ab. 10 % fayalite.

The boundaries between hypersthene and hornblende are usually well defined, but a pigmented zone may be seen here and there between the two minerals. Occasionally, the pigmentation appears as small rods at right angles to the boundaries.

Mr. H. PAULY, mag. scient. kindly examined the ores in polished samples. Magnetite is present, firstly, intimately intergrown with the spinel in some sort of graphic intergrowth, secondly as a few independent grains having thin spinel lamellae and more rarely ilmenite lamellae, and thirdly as the above mentioned coatings on the cleavages of hypersthene and hornblende. Spinel dust is also present there.

In addition larger grains composed of pyrrhotite, pentlandite, chalcopyrite and covellite are present. In one of these grains a mineral which might be sperrylite was observed at high magnification. These ore grains are partly altered into limonite.

The Border Relations Between Amphibolite and Ultrabasite at the East Coast of Langø.

On page 8 a brief survey of the conditions at the border between ultrabasite and amphibolite was given. Because of the difference in resistance to the weathering the border is usually marked by a furrow, but the border relations can occasionally, be studied in details.

Table 1. $n\gamma$ and the ferrosilite content of the hypersthenes in the border zone between 13431a and 13431b. Compare fig. 11.

Sample no.	$n\gamma$	% ferrosilite
1	1.695	22
2	1.698	24
3	1.700	26
4	1.703	29
5	1.705	30
6	1.701	27
7	1.702	27
8	1.705	30
9	1.698	24
10	1.698	24
11	1.710	34
12	1.697	24
13	1.697	24
14	1.709	34
15	1.704	29
16	1.699	25

Fig. 11 shows, in full size, a handspecimen with the borderzone. The left part is the amphibolite, the right the ultrabasite. The border is seen in the middle, the transition from amphibolite to ultrabasite covers one or two centimetres and is rather diffuse. Its most conspicuous feature is thin, light coloured secretions of the type described on page 22.

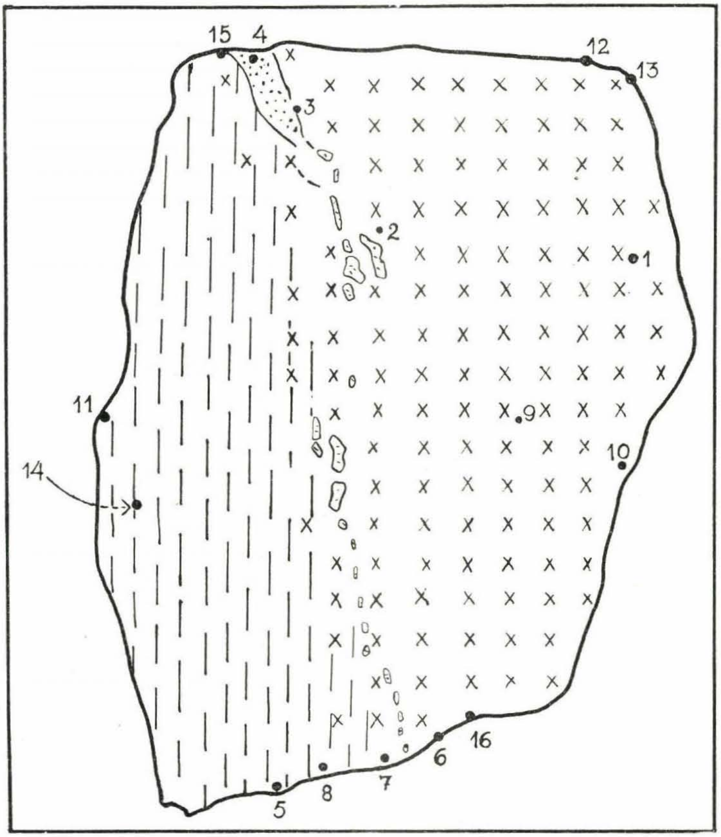

Fig. 11. Handspecimen showing the border between amphibolite and ultrabasic rock, the east coast of Langø. Crosses indicate ultrabasic rock, striation amphibolite. The dotted areas consist of pegmatitic material. The figures indicate the arrangement of the hypersthene samples examined (table 1). Sample no. 14 was taken on the underside of the specimen in a pegmatitic vein. Scale $^1/_1$.

The hypersthene has been examined across the border zone. The figures of fig. 11 indicate from where the hypersthene samples were taken. The corresponding refractive indices and ferrosilite contents (according to Poldervaart (1950)) are recorded in table 1.

The results of the examination show that also the mineralogical and chemical changes take place over one or two centimetres. These results will be further discussed on page 44.

Ultrabasic Rocks from Langø.

Samples of 13 different bodies of ultrabasic rocks, mainly from the west coast of Langø, but also from the eastern and central parts of the island, were examined in the laboratory.

Table 2. The Optical Properties and the Ferrosilite[1]) Contents of the Hypersthenes of the Tovqussaq Region.

	(÷) 2V	% fs	nγ	nβ	% fs
A. Hypersthenes of amphibolites:					
13415	49	42	1.719	..	42
13418	58	34	1.728	..	50
4040	59	33	1.714	..	38
4048	..	..	1.704	..	29
4013	67	25	1.707	..	32
4016	57	35	1.715	..	38
4034	..	..	1.708	..	33
13431 b	64	27	1.708	..	33
B. Hypersthenes of Ultrabasic rocks Langø:					
4033 West coast[2])	..	..	1.685	1.682	14
13427 —	..	..	1.690	..	17
4052 —	..	..	1.689	..	17
13428 —	..	..	1.689	1.685	17
13429 —	..	..	1.694	1.687	21
13430 —	..	..	1.694	1.686	21
13434 —	..	..	1.693	1.687	20
13433 —	84	15	1.693	..	20
13435 —	..	..	1.693	1.686	20
13437 —	84	15	1.693	1.687	20
13440 —	..	..	1.696	..	23
13431a East coast	84	15	1.692	1.686	19
13432 —	..	..	1.694	..	21
C. Hypersthenes of Pegmatites, Langø:					
13436	..	..	1.702	..	27
13434	70	23	1.702	..	27
13438	72	22	1.697	..	23
13432	..	..	1.715	..	38
13431 b	66	25	1.711	..	36
13440	..	..	1.710	..	34

[1]) According to POLDERVAART, 1950
[2]) The most southerly samples are named first, the most northerly last.

They are all, as 13437a, composed of hypersthene, hornblende, spinel, and magnetite. A few have a subordinate amount of diopside, but the paragenesis listed above is clearly the stable at the conditions prevailing under the formation of the ultrabasic rocks. As shown in table 2 the optical data of the hypersthenes of the examined rocks remain remarkably constant. This interesting fact will be dealt with on page 50.

Hornblende seems, in most samples, to grow at the expense of the hypersthene.

As mentioned on page 14 some of the ultrabasic bodies of the west coast of Langø have a well pronounced linear schistosity, even layered and "folded" rocks, in which hypersthene and hornblende are separated into layers, are present. These rocks may contain a considerable amount of biotite. As an example of these deformed rocks No. 13433 will be briefly described.

This rock is distinctly "layered" when examined in the hand-specimen. In thin section this structure is less pronounced as the hypersthene layers contain a good deal of hornblende.

Hornblende ($2V = (\div)\ 88°$, $c : \gamma = 19°$) occurs in grains as large as 0.2 centimeter arranged in parallel layers which are thinner than the hypersthene layers. The latter consist chiefly of small, rounded grains, although remnants of larger grains may be observed. $2V = (\div)\ 84°$, $n\gamma = 1.693$.

The hypersthene is clearly under transformation into hornblende, as well marginally, as along the cleavages. In addition small well-developed crystals of hornblende are observed as inclusions in the hypersthene (compare plate 7 fig. 13).

A red-brown biotite and plagioclase are present in small amounts.

Another ultrabasic rock (No. 4033) from the west coast of the island is almost exclusively composed of hypersthene porphyroblasts which vary in size from 0.5 to 1.0 centimeter. The porphyroblasts have a linear arrangement. The hypersthene has $n\gamma = 1.685$, corresponding to 14 % fs. This is the most magnesium-rich hypersthene observed on Langø so far.

Hornblende is confined to "the triangles of error" between the large hypersthene grains. The hypersthene is almost entirely free from inclusions of hornblende.

Spinel (+ magnetite) is present in considerable amount and occurs chiefly in elongated, irregular grains, which are parallely arranged. In adjoining hypersthene grains of different orientations, many spinel grains have the same common, parallel arrangement (plate 2, fig. 4). The latter is clearly independent of the crystallographic orientation of the enclosing hypersthene and seems rather to have a tectonical origin. The rare hornblende inclusions in the hypersthene may have the same parallel orientation.

The Transition from Ultrabasite to Amphibolite in one Inclusion in Gneiss, the West Coast of Langø (No. 13437).

a. The southern ultrabasic end: The rock is composed as follows: hypersthene 52 %, hornblende 36 %, spinel and ore 11 %, and plagioclase 1 %.

Hypersthene occurs both as irregular grains of as much as 0.3 centimetres (plate 3, fig. 5) and as aggregates of equidimensional grains with irregular outlines. The larger hypersthene grains and the aggregates are surrounded by and separated from one another by a network of hornblende. The latter is usually rather thin but may in places swell to aggregates as large as those composed of hypersthene. In that case the mutual hornblende boundaries are rectilinear while the boundaries against the hypersthene are lobed.

In some parts, the rock consists of an aggregate, the chief component of which is small grains of hypersthene. In addition hornblende and spinel occur, as well as small grains of plagioclase, against which the hypersthene may have a rim of hornblende.

The hypersthene has $2V = (\div)\ 84°$, $n\gamma = 1.693$, and $n\gamma = 1.687$. The optical data are almost identical with those of the hypersthene of 13431 a. The hypersthene has been partially transformed into hornblende along the cleavages and has somewhat ragged outlines on account of the hornblende of the network having grown into the hypersthene. Accordingly, the hornblende frequently contains inclusions of small, rounded grains of hypersthene generally with ore pigmentation towards the hornblende.

The hornblende has the same colour and pleochroism as that of 13431 a, but has $2V = (+)\ 88°$, $c: \gamma = 17°$, and $n\gamma = 1.664$.

Ore coatings on the cleavages of hypersthene and hornblende are very rarely seen.

The plagioclase has 65—70 $^0/_0$ an.

The spinel is intimately intergrown with magnetite as in 13431 a and occurs in rather corroded grains, but magnetite plays quantitatively a less important rôle in the intergrowths. It is present as rather thin threds.

The green spinel is especially concentrated in the outer part of the hypersthene grains and in the network of hornblende (table 3, fig. 5). In the latter the spinel grains have disintegrated and occur as accumulations of small spinel grains, in extreme cases as a pigmentation in the hornblende and apparently in a state of dissolution.

A comparison with 13431 a shows that the two rocks contain hypersthene of identical compositions, but it is evident that the hypersthene in 13437 a is in a process of replacement by hornblende. Neither diopside nor olivine are present in 13437 a (but the hypersthene has here small concentrations of ore which may be the last remains of olivine?).

Mr. H. PAULY has also examined the ores of 13437 a. In addition to the above mentioned should be noted that in 13437 a only traces of sulphide ores are present. At a magnification of 600 × small grains of chalcopyrite could be observed.

b. The ultrabasic rock in the central part of the inclusion: The rock is equigranular consisting of equidimensional, rounded grains of 0.05 centimeter. Hypersthene and hornblende are present in almost equal amounts, and some spinel also occurs.

In contradistinction to 13437a, this rock is more fine-grained but contains still a few irregular hypersthene grains of 0,2 centimeter. Most of the hypersthene present is found in aggregates. The larger amount of hornblende is caused by the formation of hornblende at the expense of hypersthene.

The amount of spinel is decreasing. It is found in rather corroded grains particularly in the outer parts of the hypersthene and hornblende grains.

c. The northern, amphibolitic part of the inclusion: Hypersthene and hornblende occur still in corresponding quantities, but the rock contains now about 20 % plagioclase. The amount of spinel is much reduced. The parallel arrangement of the hornblende grains gives the rock a linear schistosity.

All the minerals of the rock occur in rather irregular grains and the grain size is varying. Hypersthene with small inclusions of hornblende may reach the size of 0.15 centimeter. Hornblende may attain the same size, but the minerals occur generally as intimately intergrown grains of less than 0.1 centimeter.

Here, as in the two rocks described above it is possible to discern aggregates of hypersthene grains, but they are more intergrown with hornblende.

The axial angle of the hypersthene is still (÷) 84°, nγ = 1.697. The hornblende has 2V = (+) 85°, c : γ = 18°, and nγ = 1.666. The pleochroism is γ green, β yellow-green, and α faintly green.

The present rock differs in particular from the rocks described above in its content of plagioclase, irregular grains of this mineral being scattered all over the rock (plate 3, fig. 6). Repeated twinning, especially according to the albite law occurs. The anorthite content varies from 66 to 62 % in normal zoned grains. The border towards the hypersthene is sharp and without reaction rim.

Thus, the massive ultrabasic rock has changed into a plagioclase-carrying, hypersthene-amphibolitic rock with linear schistosity. In addition to the formation of plagioclase, the introduction of material from the surroundings has resulted in a gradual substitution of hornblende for hypersthene, in the varying composition of the hornblende, and in the material of the pleonaste having been digested by the hornblende.

Small, Partly Dissolved Inclusion of Amphibolite in the Gneiss from the West Coast of Langø. (No. 13438).

The inclusion is embedded in and penetrated by pegmatitic material. At the boundary between pegmatite and amphibolite a fine-grained, plagioclase-hypersthene aggregate occurs. The outermost part of the inclusion is coarse-grained and is deficient in hornblende.

The pegmatite consists chiefly of plagioclase (30 $^0/_0$ an) with alternating broad and narrow twin lamellae and a good deal of dark pigmentation. A minor quantity of quartz is present. Hypersthene occurs as slightly corroded grains of about 0.1 centimeter. It has ore-coatings on the cleavages and finely distributed pigmentation. $2V = (\div)\ 72°$, $n\gamma = 1.697$, corresponding to 23 $^0/_0$ fs. It is slightly uralitised along cracks. Rutile occurs as an accessory.

The content of hypersthene in the pegmatite increases towards the amphibolite. On the boundary an aggregate of plagioclase and hypersthene is found, the latter mineral occurs as smal, rounded grains, which are smaller than the hypersthene grains of the pegmatite.

The outer part of the inclusion contains a few small grains of biotite and hornblende. The red brown biotite is only found in a narrow outer zone in small quantities. Hornblende occurs as small grains of a dirty, brownish-green colour (darker than the hornblende of the amphibolite) and increases in amount towards the central part. The "transition" to this part is extremely abrupt. The narrow hornblende-biotite-carrying zone is sharply defined from the original amphibolite which contains fairly large grains of hornblende arranged parallel with the boundary.

The central part of the inclusion consists of hornblende and hypersthene, as well as a little plagioclase and diopside.

The hornblende is pleochroic with the colours: γ green, β olive-green, and α yellowish-green. $2V = (\div)\ 84°$, $c: \gamma = 17°$.

The hypersthene has $2V = (\div)\ 69°$, corresponding to 24 $^0/_0$ fs that is, slightly richer in iron that the hypersthene of the pegmatite which had 22 $^0/_0$ fs. It should be noted that the latter, contrary to the former, has coatings of ore on the cleavages.

The plagioclase has approx. 45 $^0/_0$ an.

In appearance and composition this inclusion resembles the amphibolite from the east coast of Langø and may be interpreted as the remains of one of the amphibolite bands which contained the ultrabasics from the west coast of the island.

Small, partly dissolved Inclusion of Ultrabasite in Pegmatite. (No. 13436).

The pegmatite consists chiefly of plagioclase (37 % an). There are also small amounts of quartz and the usual hypersthene of the pegmatites of this island ($n\gamma$ = 1.702).

On the boundary between pegmatite and ultrabasite there is a concentration of small hypersthene grains in aggregates together with plagioclase. The hypersthene is slightly uralitised. Next comes a zone with hypersthene grains of as much as $^1/_2$ centimeter, more or less replaced by plagioclase and biotite, the biotite is usually connected with the cleavages. The hypersthene has 2 V = (÷) 63°. The zone is intergrown with thin flakes of biotite arranged at right angles to the boundary and connected with the following zone. This consists of biotite flakes up to 0.3 centimeter long arranged parallel with the boundary, and of plagioclase.

Inside the mica zone and to some extent interwoven with it follows a hypersthene-rich zone with corroded hypersthene grains of 0,4 centimeter which have been displaced by biotite and, in particular by hornblende along the cleavages (plate 4, fig. 7). The substitution by plagioclase has not proceeded so far as in the outer zone. The axial angle of the hypersthene was found to vary from 66° to 76°. The optical sign is negative. In one grain with wavy extinction the core had 2 V = (÷) 76°, the rim 2 V = (÷) 67°. Most grains have, however, axial angles of about 68°, corresponding to 24 % fs.

The amount of biotite and plagioclase decreases towards the centre of the inclusion, where plagioclase only occurs as interstitial grains, biotite in a few small flakes. Hornblende is present either as irregular, independent grains or on the cleavages of the hypersthene. The hypersthene has 2 V = (÷) 70°, corresponding to 23 % fs.

In the central part of the inclusion, which was not examined in thin section, $n\beta$ of the hypersthene was found to be 1.693, corresponding to 24 % fs. Thus the content of fs in the hypersthene decreases from 27 % in the pegmatite, to 24 % in the central part of the inclusion. It should be noted, that the hypersthene of the ultrabasic body intersected by the pegmatite has $n\gamma$ = 1.693, which gives 20 % fs.

Fig. 7 shows the inclusion described as No. 13436.

Zone of Hornblende between Ultrabasite and Pegmatite. (No. 13434).

The pegmatite is composed of plagioclase (35 % an, dark pigmentation in linear arrangement), hypersthene (2 V = (÷) 70°), quartz, and small amounts of biotite and rutile. The hypersthene occurs as somewhat corroded, pigmented grains especially concentrated on the border between ultrabasite and pegmatite (plate 4, fig. 8).

Just inside the large hypersthene grains of the border, small flakes of a reddish-brown biotite are scattered over a zone of less than $^1/_2$ centimeter either among the minerals or associated with the hypersthene. The biotite zone runs a wavy course and its innermost part contains small grains of hornblende. Inside the biotite zone the biotite disappears and hornblende appears in fairly large quantities. The latter is of a dirty, brownish-green colour, $2V = (\div) 72°$, $c : \gamma = 18°$. The hornblende-rich zone contains a small amount of plagioclase with about 43 $^0/_0$ an.

Large grains of hypersthene lying across the biotite-, as well as across the hornblende-zone have close to the pegmatite (in the biotite zone) inclusions of biotite and are somewhat transformed along the cleavages. Towards the ultrabasite (in the hornblende zone) the hypersthene has inclusions of hornblende and looks fresher than in the biotite zone (plate 5, fig. 9).

The hornblende zone is almost one centimeter wide and has comparatively sharp borders against the hornblende-rich ultrabasic rock which still contains a little plagioclase in the outer parts. The plagioclase has fewer and broader twin lamellae than in the pegmatite and the anorthite content is 57 $^0/_0$.

The outer part of the ultrabasic body contains a few corroded hypersthene grains of 0.2 cm in a hornblende-hypersthene groundmass. The hypersthene seems, to a great extent, to be substituted by hornblende. The hypersthene has $2V = (\div) 76°$, the hornblende $2V = (+) 87°$, $c : \gamma = 17°$, pleochroism: γ green, β yellow-green, and α almost colourless.

The central part of the inclusion has hypersthene grains of one centimeter, partly transformed into hornblende. Moreover, it consists of hornblende and pleonaste, the latter in small quantities, especially in the network of hornblende between the hypersthene porphyroblasts. The hypersthene has $n\beta = 1.687$, and $n\gamma = 1.693$.

ON THE FORMATION OF THE ULTRABASIC BANDS IN THE AMPHIBOLITE

The amphibolite in which the ultrabasic rocks occur as bands is probably a para-amphibolite. All the minerals of the amphibolite are, however, secondary, that is to say formed during the metamorphism, and it is therefore difficult to determine the original mineralogical composition with certainty. Judging from the parageneses of the amphibolites described these rocks may in a previous state have been impure calcareous sediments. This conclusion is strengthened by a comparison with the amphibolite layers north of Tovqussaq which certainly had a sedimentary origin. The Tovqussaq amphibolites may be regarded as a southern continuation of the paragneiss series of the northern area (see page 59).

When the sedimentary origin of the amphibolite is taken for granted, the formation of the ultrabasic bands of the amphibolite seems rather difficult to explain in an orthodox way.

The ultrabasic rock cannot be considered as a magmatic secretion in the amphibolite; neither can the amphibolite be regarded as formed at the expense of an original, magmatic ultrabasic rock. Further, the mode of occurrence of the ultrabasite precludes the possibility of its being formed originally as an intrusion in the amphibolite.

Therefore, the ultrabasic rocks, just as the amphibolite, must have been formed by metamorphic processes. BARTH (1947, page 47) states for ultrabasic hypersthene-hornblende rocks (bahiaites) from Southern Norway that their formation is best explained by metamorphic differentiation. According to the description these rocks have a similar mode of occurrence and mineralogical and chemical composition as the ultrabasic rocks of Langø (see page 45). The writer shares BARTH's view and believes that the ultrabasic rocks described were formed *in situ* in the amphibolite by metamorphic differentiation combined with metasomatic processes (SØRENSEN, 1951 a).

In the following the processes leading to the formation of the ultrabasic rocks will be more fully discussed. It should be noted here that the chemical and mineralogical changes will be treated in the following chapter and that the whole problem of the origin of ultrabasic rocks in orogenic zones will be discussed on page 73.

The dome formation at Tovqussaq resulted in the existence of granulite facies conditions on the Tovqussaq peninsula and Langø. There has been a rise in temperature as indicated by the presence of hypersthene in gneisses and amphibolites. This is understandable as the Tov-

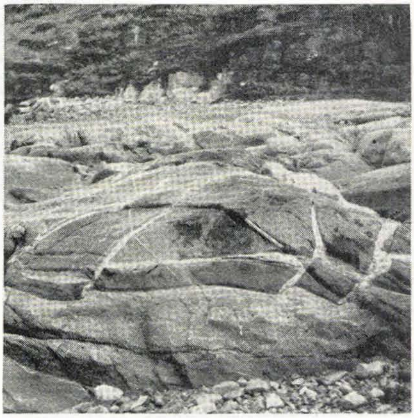

Fig. 12. Amphibolitic inclusion in gneiss, Pisigsarfik, Godthaab fiord. The gneiss is plastically deformed, the amphibolite is deformed by fractures parallel with and perpendicular to the strike of the layer. The fractures are filled with pegmatitic material. Note that the fractures parallel with the strike are older than the perpendicular fractures.

qussaq area is situated on the boundary between a northern amphibolite- and a southern granulite-facies region.

The structural development of the area, explained by BERTHELSEN (op. cit., page 568) as a result of an axial culmination in an underlying structure, took place under such P,T conditions that the quartz-feldspar rocks were deformed plastically. The amphibolite bands, being more rigid, were, during the deformation, rotated into their present concentric arrangement as a result of the doming.

The above mentioned difference in plasticity between amphibolite and gneiss (under proper P,T conditions) is familiar to all petrologists who have worked in pre-cambrian complexes. Fig. 12 shows a fine example from the Godthaab fiord, West Greenland.

It is convenient to discuss, in a more general way, the rôle played by the rigid bands of amphibolite in the deformation. It is a common experience for rigid layers to have a controlling influence on the deformation, a feature which may be studied in several localities along the west coast of Greenland (see for instance RAMBERG, 1948b, page 317).

The layers are rotated in such a way that the shearing forces are acting more or less parallel with their strike. The amphibolite resolves and transmits the forces into shearing- and direct stress and because of the parallelism between the forces and the strike of the amphibolite, the shearing component is predominant. Beautiful examples of this can be observed in the Godthaab fiord, where shear zones are conformable in the amphibolites, while similar zones in the surrounding gneiss may as well be of a cross-cutting nature.

Depending on the prevailing conditions the amphibolite may break in one or both of the following ways: 1) at right angles to the strike of the layer giving boudinage phenomenae or ac- or cross joints, 2) parallel with the strike along ab-surfaces. By the latter process the common banded amphibolites and some types of banded gneisses may be formed. The ab-zones of fracture are filled with material from the dispersed phase, which consolidates at places of low mechanical pressure at first giving rise to conformable replacement quartz veins, still with stripes of amphibolitic material. From this all transitions to banded gneisses and even homogeneous granitic rocks may be seen. The banded amphibolites may in some cases be formed by metamorphic differentiation because of the mechanical separation of the minerals according to their different gliding properties (W. Schmidt, 1932, page 183).

The above mentioned splitting up of the amphibolitic bands seems to be a common mode of deformation in the deeper parts of the orogenic zones. On Langø, however, the structural evolution followed a somewhat different course.

The gneissic rocks were plastically deformed, while the amphibolites behaved as rigid bodies, but, contrary to the above mentioned, the cohesional attraction of the amphibolite was not exceeded during the first stages of the deformation. Therefore, instead of fracture zones, potential shear zones were established. Because of the amphibolites resistance to the deformation, these zones may also be termed tension zones.

The kinetic energy required for the creation of these potential shear zones must, since no shearing took place, be converted into some other form of energy. The writer believes that this energy transformation is responsible for the formation of the ultrabasic rocks provided that the state of tension is maintained over a sufficiently long period.

In the following an attempt to explain the formation of the ultrabasic rocks will be made.

Let us consider a layer of amphibolite which, because of a difference in plasticity between the layer and its surroundings, has been arranged

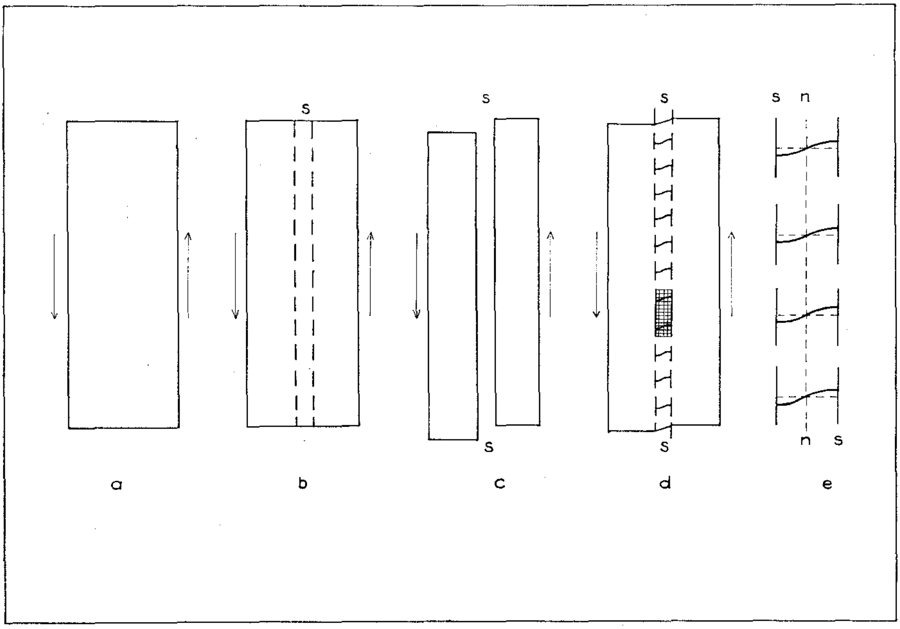

Fig. 13. Diagrammatic representation of the formation of shear zones and tension zones. a, b, c, and d see text. e: Enlarged part of the zone S—S showing state of no-strain (dotted lines) and state of tension (curved lines). n—n: neutral surface along which the shearing stress is at a maximum.

parallel with the shear forces in operation (fig. 13a). As mentioned on page 35 the layer will have a tendency to break parallel with the strike along the zone S—S of fig. 13b, the exact location of which is in the zone of least resistance under the given conditions[1]). S—S may be termed a potential shear zone. If the cohesional attraction of the rock is exceeded, displacement will take place as indicated in fig. 13c, that means, the kinetic energy is released in movement.

If the cohesional attraction of the amphibolite is not exceeded under the given conditions the deformation along the zone S—S may be elastic as suggested in fig. 13d. S—S may now be termed a tension zone, the physical conditions of which differ from those of the amphibolite outside S—S. The kinetic energy is in this case stored as elastic energy (potential energy of strain). As the amphibolite is not perfectly elastic a certain amount of heat is generated.

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[1]) The stress concentration, which is responsible for the formation of the zone S—S, is determined by the direction and size of the deforming forces, by the velocity of deformation, by the temperature, and by discontinuities in the rock. Additional information about stress concentration and stress distribution can be obtained from TIMOSHENKO (1941) and HAFNER (1951).

The amphibolite is a heterogeneous aggregate of different minerals, which may originally be very close to perfect hydrostatic equilibrium. "The arrangement and sizes of the crystals may be such that the stress in every crystal is the same hydrostatic pressure" (BIRCH and BANCROFT, 1938, page 65).

"For a given impermeable aggregate with definitely fixed crystals, there exists in general only one pressure (for a given temperature) at which the stresses on the individual crystals can be hydrostatic and equal to the pressure on the boundary, and it is of course possible that there will not be even one such pressure. At all other pressures the stresses on the crystals must depart more or less from hydrostatic" (BIRCH and BANCROFT, op. cit. page 66).

If the equilibrium is disturbed, for instance by establishment of the tension zones in the amphibolite, the stresses on the crystals are no longer hydrostatic. Recrystallisation (associated with metamorphic differentiation as discussed below) removes the non-hydrostatic strains and the rock will return to a state of hydrostatic pressure.

In the present case the less compressible ultrabasic rocks are formed in S—S in response to the change of pressure in that zone in such a way that the zone S—S is in hydrostatic equilibrium with the adjacent rocks.

Consider next the part of the amphibolite marked by cross-ruling in fig. 13d. As no movement takes place along S—S we may, for the sake of simplicity, assume, that the system in question does not give off work to the surroundings and that the total energy of the system is constant. The minerals of the amphibolite in the system considered were originally in mutual equilibrium. The equilibrium is disturbed by the establishment of the state of tension and a stabilizing irreversibel process sets in. As the total energy is constant there is, accordingly, a loss of work in the system in question i. e. a similar amount of heat is produced.

The statements above are valid for a system of constant total energy, where the reactions are caused by a change of the state of the system. Work supplied to the system from the outside as well as mechanical heat may of course also play an important rôle. We shall only mention frictional heat which may develop in several ways. A stretching of the amphibolite gives rise to internal friction which is probably concentrated in the shear zones. The resistance of the amphibolite towards deformation is another source of heat, the elastic energy being released as heat. Plastic flow along microscopical shear planes may be of importance especially in the later stages of the "tension period".

That the processes discussed can cause an increase in temperature seems to be beyond doubt. It might be objected, however, that the heat generated is dissipated by conduction so fast that it has no traceable influence in the tension zones. In this connection should be referred to the following points:

1) The thermal conductivity in the tension zone is most probably different from that in the surrounding amphibolite. It seems most reasonable to assume that the pressure is greatest in the tension zone. For this reason the conductivity may be lower in the tension zone than outside it, unfortunately experimental data is still lacking.

2) The "transition" from strained to unstrained rock is gradual, not abrupt (see page 49).

3) The rock in S—S and the amphibolite outside this zone have the same composition, but have been exposed to different physical conditions. Thus, there is a difference in chemical potential between S—S and the unstrained amphibolite. Accordingly, the energy released in S—S is gradually used in the stabilizing processes which continue until the minerals in S—S are in mutual equilibrium under the existing conditions and further are in equilibrium with the surroundings, in other words, until there is no potential difference between the systems in question.

To the statements above should be added that the temperature at which recrystallization of a given mineral will occur freely decreases with increasing stress (BUERGER and WASHKEN, 1947). The tension prevailing in S—S causes more ions than usual "to jump over the energy barrier" which accelerates the chemical and mineralogical changes in the zone (compare RAMBERG, 1946).

For the above-mentioned reasons it seems allowed to assume that the rock formed in S—S has a paragenesis corresponding to a higher temperature of formation than that of the unstrained amphibolite (see page 52).

Thus the ultrabasic rock is believed to be formed as a result of readjustment processes in response to changes in pressure and temperature in the most "active" (unstable) zones in amphibolites.

The main part of the amphibolite behaves as "passive (stable) masses in the tectonic stream"; the activity is confined to the tension zones (i. e. zones of stress concentration) in the stages of deformation discussed above. This is supported by the mineralogical evidence discussed in the following chapter. In addition we may refer to the common slicing of layers, where slices of the original, unaltered rock are found between zones of shearing.

It should be noted that differential laminar movement could have a similar result as the shear movements discussed above. Shearing produced by bending of the amphibolite layers may have played an important rôle in the formation of the Langø ultrabasics.

The conclusions of the preceding discussion are in close agreement with the results of the examination of the behaviour of deformed metals.

When a previously deformed polycrystalline metal is heated to a sufficiently high temperature, a recrystallized structure developes (primary recrystallization). Strains residing in the lattice after deformation are dissipated by reorientations having distinct crystallographic significance in relation to these strains.

"Under certain conditions, on prolonged annealing of the specimens, a type of grain coarsening occurs in which a small number of grains commence to grow at various points by devouring the neighboring fine-grained primary structure, which, in itself, exhibits a marked stability toward normal grain growth." (Rosi, Alexander, and Dube, 1952, page 189). This is the secondary recrystallization which is probably caused by the different boundary energies of the grains.

Chen and Mathewson, (1952) have examined the recrystallization of aluminum single crystals after plastic extension. They found a competition between nucleation and growth. In some cases the specimens recrystallized with the formation of a single crystal, in other cases polycrystals were formed ("polygonization"). The latter process seemed to be connected with deformation bands in the crystals. "Thus the process of recovery seems to be more closely associated with the intimate strain configuration than with the net value of strain" (op. cit. page 507).

The deformation of metals cannot be directly compared with the amphibolites discussed above. The metal samples have one component only and the experiments were carried out at low confining pressure. Conditions in the strained amphibolites are much more complicated. Nevertheless, the writer finds the experimental evidence strongly in favour of the hypothesis presented above. We shall not discuss the deformation of metals further here. More information can be obtained from the papers quoted above, which in addition have extensive lists of references.

The geological literature on the subject seems to be rather limited.

H. Backlund (1936, pag. 59) in his important discussion of the eclogites states that the eclogites were formed in a comparatively high level in the crust. The high density minerals of the eclogite were formed, not because of depth in the earth (great hydrostatic pressure), but by dynamic pressure which locally may reach higher values than the hydrostatic pressure. The rise of the migmatite front was responsible for the prevailing high temperature (op. cit. pag. 58).

H. EBERT (1936, pag. 74—75) in connection with BACKLUND's paper expressed the following view on the presence of high dynamic pressure in shallow depth in the crust:

"Um dies zu erklären, kann mann die Dampfdruckverhältnisse in der Erdkruste heranziehen. Allgemeine Überlegungen zeigen, dass in einer gewissen Tiefe die Dampfdrucke einen Maximalwert erreichen müssen, und zwar in der Zone der pegmatitischen Zustände. Die Berechnung aus Zustandsgleichungen — . . . . . — zeigt, dass auf diese Weise Drucke zustande kommen, die der Grössenordnung des reinen hydrostatischen Druck in erst weit grösseren Tiefen entsprechen. Damit diese Überdrucke tatsächlich wirksam werden können — theoretisch ist zunächst Volumenvergrösserung der fluiden phasen zu erwarten, bis der Dampfdruck dem normalen hydrostatischen Druck entsprecht — müssen über engbegrenzte Gebiete tektonische Spannungszustände herrschen, die den Druckausgleich verhindern. Das bedeutet, dass auch innerhalb einer typischen Eklogitzone nur ein geringer Prozentsatz der chemisch in Frage kommenden Gesteine wirklich in Eklogit verwandelt zu werden braucht."

Thus, BACKLUND was the first to combine deformation and the formation of massive rocks as eclogites in rigid rocks. EBERT showed that the formation was restricted to tension zones, and he, as well as BACKLUND, attributed the processes to high pressure.

J.W. AMBROSE (1936, pag. 257) has described an example of progressive kinetic metamorphism from Flinton, Manitoba. The field relations are such that neither proximity to an igneous mass nor depth of burial can be held to account for the observed advance in metamorphic rank of the various rocks. As the amount of movement in the higher rank zone (garnet-zone) is demonstrably greater than in the low rank zones (biotite- and chlorite-zones) AMBROSE proposes that the heat necessary for the metamorphism was developed mechanically by, and during, intimate internal shearing. He writes (op. cit. pag. 279): "Frictional heat must be generated on planes of shear when differential movements take place. If the shearing is restricted to that along faults, the relatively small amount of heat generated, though possibly of local importance, is obviously insufficient to cause any significant rise in the temperature of a large body of rocks. On the other hand, if the rocks are intimately sheared, although the amount of frictional heat developed along any given plane may be small, the total amount must be enormous. Provided it is generated faster than it can be conducted away, the temperature of the whole body of sheared rock must rise."

In 1944 J. L. DE LURY discussed the generation of magma by frictional heat, an idea already expressed by DANA. In the level of the crust under consideration the displacements are slow. "The rocks fail piecemeal in a long period of time when subjected to a stress difference small in comparison with the one required to produce instantaneous failure" (op. cit. pag. 119).

The movement even along extensive overthrust fault planes in shallow levels is so slow that frictional heat is removed by conduction and no magma is formed. But De Lury believes, however, "that projection of these same places into distant and deeper regions might well spell magma generation. Deformation is slow, even in critical periods, so that the thermal insulation provided by burial to depths of a few tens of kilometers becomes an important factor" (op. cit. pag. 125).

Turner and Verhoogen (1951, pag. 570) treat the problem of mechanically generated heat, but unfortunately in rather vague terms. They write that it can be shown that the rise of temperature is of the order of a few degrees, where the work of deformation, instead of being stored elastically in the rock, is dissipated as heat. No evidence confirming this statement is offered.

Further they reject the heat generated by friction on surfaces of rupture as a possible source of maintaining temperatures of regional metamorphism.

They admit the existence of pseudo-tachylite veins associated with mylonites. Now, frictional heat in these cases has only been effective in restricted zones, for which reason Turner and Verhoogen conclude that heat produced in this way must be of even less significance as a factor in regional metamorphism in which deformation is distributed through much greater thickness of rocks. "By Niggli, Harker, and others it has been argued that dissipation of mechanically generated heat by conduction, though a slow process, generally is efficient enough to nullify the influence of heat from such sources in regional metamorphism" (op. cit. pag. 571).

"Since it is thus unlikely that frictional heat contributes notably to metamorphic temperatures, and in so far as high nonhydrostatic stresses incidental to deformation are not essential to the crystallization of most metamorphic minerals, deformation cannot be regarded as the principal direct cause of regional metamorphism." (op. cit. pag. 571). They later acknowledge that deformation may exercise an all-important "catalytic" influence.

It sounds incredible that the writers of a textbook on petrology, written in 1951, should still be looking for "the possible existence and distribution of unexposed subjacent granitic or other igneous bodies" (op. cit. pag. 571) in explaining temperatures of metamorphism. The critical student of pre-cambrian areas knows that the majority of the granites there are not the cause, but, on the contrary, the end products of the metamorphism.

The comparison quoted above between mylonites and the main period of deformation is very unfortunate. As observed in several areas (e. g. BARTH, 1947, pag. 24, and BERTHELSEN, 1950, pag. 561) the mylonites belong to a late stage of the deformation, where the rocks, situated in shallow depths, behave as brittle bodies. The late faulting takes place at temperatures much lower than the temperatures prevailing during the main period. Pseudo-tachylytes indicate that mechanically-generated heat can fuse the rocks in restricted zones, before it is removed by conduction; but a comparison with the regions of regional metamorphism, where plastic flow may be promoted along microscopic shears, is definitely irrelevant.

On the preceding pages the importance of the mechanically-generated heat in metamorphism was briefly discussed. The writer believes that there is an intimate relationship between structural development and metamorphism, the metamorphism being intensified with increasing deformation (see pag. 53).

For this reason the writer accepts the main principles of AMBROSE's view. The mechanically-generated heat may play an important rôle in metamorphism, especially in shallow depths of the crust. The heat generated by the slow and piecemeal movement along the displacement surfaces, is probably dissipated by conduction, without affecting the place of formation, but steepening the temperature gradient in the crust instead. Where the movement is rapid enough thin bands of ultra-mylonites may be formed.

In the deeper levels of the crust, for instance in the granulite facies province of Tovqussaq, the deformation takes place in a somewhat different way. The rocks yield more plastic than at higher levels, perhaps still to some extent along microscopic shears. The rocks, however, have different plasticities. The most plastic rocks are deformed by flow (macro creep), the more rigid ones by fracture.

Under the physico-chemical conditions prevailing at Tovqussaq during the deformation the gneissic rocks were plastically deformed, the amphibolites behaved as rigid bodies.

On the east coast of Langø the cohesional attraction of the amphibolite was not exceeded during the first stages of the deformation. Stress of high order may, however, have been affecting the amphibolite during a long period. Potential shear zones (tension zones) developed and the writer believes that the change in physico-chemical conditions in these zones was responsible for the formation of the ultrabasic bands.

In a later stage of the deformation the strength of the amphibolite as well as that of the ultrabasite was overcome and cross joints, which were filled with pegmatitic material, were formed.

In zones of very intense shearing, as for instance the west coast of Langø, the amphibolite was fractured, rolled out in and partly digested by the gneiss. The thin bands of amphibolite may even be folded. The rigid ultrabasic bands yielded by fracture and formed boudinages (beads of pearls). The shearing acted parallel with the ultrabasic bands, viz. parallel with the potential shear zones. As there was, in the early stage of the deformation, a difference in plasticity between amphibolite and gneiss, the later stages were characterized by a similar difference between ultrabasite and amphibolite, the latter in this stage being deformed plastically with the gneiss.

Although the ultrabasics from the Godthaabs Fiord will be treated in a later chapter, it is appropriate here to note, that in the Godthaab Fiord displacements were released along the potential shear zones as the ultrabasics of the amphibolite bands are often found in contorted, chloritized mylonite zones.

The hypothesis outlined in this chapter cannot be substantiated at the moment. The writer hopes, however, that the evidence presented in this paper will be accepted, until a more quantitative treatment can be produced.

THE CHEMICAL AND MINERALOGICAL CHANGES ACCOMPANYING THE FORMATION OF THE ULTRABASIC ROCK

In the previous chapter a discussion of the structural evolution resulted in the formulation of a hypothesis according to which the ultrabasic rock was formed *in situ* in the amphibolite on the east coast of Langø. The chemical and mineralogical data discussed in the present chapter seem to support this statement.

The dome formation at Tovqussaq resulted in the existence of granulite facies conditions on the Tovqussaq peninsula. In response to the rise in temperature hypersthene appeared in the various rocks of the area.

The amphibolite layer from Eqaluk mentioned on page 19 is one of the most northerly layers in which a formation of hypersthene has been observed. It is evident from the examination of thin sections that hypersthene is the last formed mineral. Furthermore, a comparison of the mineral compositions, as they appear from the optical data, shows that material must have been added from the surroundings as well as removed from the amphibolite. Traces of the latter process may be the small grains of plagioclase, which seem to be especially connected with the hypersthene.

The rocks from the east coast of Langø, the amphibolite 13431 b and the ultrabasic rock 13431 a were chemically analysed by Dr. S. Palmqvist, Fagersta, Sweden. The results of the analyses are recorded in table 3.

The totals of the analyses are in excess of the limit usually allowed for the summation of rock analyses (namely 100.50 %). In addition, a comparison with the mineralogical compositions of the two rocks indicate that the H_2O values are much too low. Rocks containing about 40 % hornblende must contain more water than 0.06 % and 0.13 % respectively. A control determination on 13431 b gave 0.50 % H_2O (with $PbCrO_4$ as a flux), i. e. the total of the analysis exceeds 101 %. The analyses are certainly inaccurate, but as sufficient fresh rock material is not available

Table 3.
Analyst Dr. Sven Palmqvist.

	13431 b			13431 a		
	Weight %	Mol. no.	Number of cations in Standard cell	Weight %	Mol. no.	Number of cations in Standard cell
SiO_2	47.42	0.7894	46.1	44.40	0.7392	43.6
TiO_2	0.75	0.0093	0.5	0.80	0.0100	0.6
P_2O_5	0.02	0.0001	..	0.03	0.0002	..
Al_2O_3	13.38	0.1313	15.3	7.57	0.0743	8.7
Fe_2O_3	3.32	0.0208	2.4	6.01	0.0376	4.4
FeO	9.86	0.1373	8.0	8.28	0.1152	6.8
MgO	13.48	0.3344	19.5	23.43	0.5811	34.3
CaO	10.57	0.1885	11.0	8.33	0.1486	8.8
MnO	0.26	0.0037	0.2	0.26	0.0037	0.2
Na_2O	0.66	0.0106	1.2	0.60	0.0097	1.1
K_2O	0.51	0.0054	0.6	0.25	0.0027	0.3
NiO	0.07	0.0009	0.1	0.16	0.0021	0.1
Cr_2O_3	0.29	0.0019	0.2	0.50	0.0033	0.4
$H_2O\pm$	0.06	0.0033	0.4	0.13	0.0072	0.9
Total...	100.65	..	105.5	100.75	..	110.2

13431 b Mode:		13431 a Mode:	
hypersthene	22 %	hypersthene	40 %
hornblende	40 %	hornblende	38 %
diopside	18 %	diopside	9 %
plagioclase	19 %	spinel + magnetite	11 %
ore	1 %	olivine	2 %

at present, it has not been possible to undertake duplicate analyses. The following recalculation of the analyses should be taken with some reservation. The analyses are considered by the writer to illustrate the qualitative changes in chemical composition fairly well, especially as the inaccuracies are most probably to be found in the values for SiO_2, Al_2O_3, and MgO. The remaining components should be correct.

The best way of comparing the analyses seems to be a calculation of their "standard cells" (as suggested by Barth, 1949). The number of cations in the "standard cell" (i. e. the number corresponding to 160 oxygen ions) is tabulated in table 3.

A comparison shows that the amphibolite may have passed into the ultrabasic rock

By Adding	By Subtracting
0.1 ion of Ti	2.5 ions of Si
2.0 ions of $Fe^{\cdot\cdot\cdot}$	6.6 ions of Al
14.8 ions of Mg	1.2 ions of $Fe^{\cdot\cdot}$
	2.2 ions of Ca
0.2 ions of Cr	0.1 ion of Na
(0 5 ions of H)	0.3 ions of K
Total... 17.6 cations representing 37.1 valences.	Total... 12.9 cations representing 37.0 valences.

If the ultrabasic rock is formed *in situ* in the amphibolite, the latter rock having the same composition in the zone S—S (compare page 36) as elsewhere, it appears that the formation of the ultrabasic rock is accompanied by a considerable exchange of material.

Let us compare the chemical changes with the mineralogical evidence.

The hypersthene and hornblende of the ultrabasic rock (13431 a) are according to the optical data more Mg-rich than those of the amphibolite. The same applies to the diopside. In addition a great amount of spinel and a little Mg-rich olivine are present in 13431 a, which at the same time contains more hypersthene than 13431 b. Thus, the observed increase in Mg corresponds closely with the mineralogical data. The corresponding decrease in Si, Al, Ca and alkalis results in the disappearance of plagioclase and in the decreasing quantity of diopside (and hornblende).

The $Fe^{\cdot\cdot\cdot}/Fe^{\cdot\cdot}$ relation is most interesting. The added 2.0 ions of $Fe^{\cdot\cdot\cdot}$ are of the same magnitude as the subtracted 1.2 ions of $Fe^{\cdot\cdot}$ i. e. it appears as if the $Fe^{\cdot\cdot}$ released is oxidized and fixed as $Fe^{\cdot\cdot\cdot}$. This is in close agreement with the mineralogy of the ultrabasic rock as its hypersthene and hornblende have magnetite on the cleavages and are clouded with a fine ore pigmentation.

As to the TiO_2 of the rocks, the ore-microscopical examination proved ilmenite to be very rare. Thus, Ti must be accomodated in the silicates. As mentioned previously, the hypersthenes are strongly pleochroic with α clear red, β colourless to yellowish-brown, and γ faint green. This may be a result of a content of TiO_2 as described in a recent publication by QUENSEL (1951, page 248).

Cr and Ni follow as usually the Mg (see further page 66).

We may now discuss the formation of the ultrabasic rock in a more general way. In response to the change in pressure and temperature in the zone S—S (see page 36), the minerals of the ultrabasic rock were formed at the expense of the amphibolite minerals.

The change in pressure (probably an increase) may stabilize minerals, the vapour pressure of which increases least with increasing outer pressure (e. g. hypersthene and olivine, compare TURNER and VERHOOGEN, 1951, page 22). A result of the increase in temperature is the rise in the Mg/Fe ratios in the hypersthene and hornblende. A comparison with the ultrabasic rocks of the west coast of Langø indicates that the stable end-product of this transformation is a rock carrying hypersthene (of the same composition as that of 13431a) hornblende, pleonast, (and magnetite). For comparison one of these western rocks (13437a) was analysed by Dr. S. PALMQVIST (see table 4).

Table 4.
Analyst Dr. SVEN PALMQVIST.

	13431a	13437a
SiO_2	44.40 %	43.95 %
TiO_2	0.80 -	0.60 -
Al_2O_3	7.57 -	11.94 -
Fe_2O_3	6.01 -	6.59 -
FeO	8.28 -	8.78 -
MnO	0.26 -	0.28 -
MgO	23.43 -	21.98 -
CaO	8.33 -	4.88 -
Na_2O	0.60 -	0.68 -
K_2O	0.25 -	0.07 -
$H_2O\pm$	0.13 -	0.07 -
P_2O_5	0.03 -	0.03 -
NiO	0.16 -	0.10 -
Cr_2O_3	0.50 -	0.26 -
Total...	100.75 %	100.21 %
	Mode:	Mode:
	hypersthene ...... 40 %	hypersthene ...... 52 %
	hornblende .... 38 -	hornblende .... 36 -
	diopside .... 9 -	spinel + ore ...... 11 -
	spinel + ore ...... 11 -	plagioclase .... 1 -
	olivine .... 2 -	

The ultrabasic rock of the east coast (13431a) has apparently not reached a state of perfect equilibrium, we may say that the rock association was quenched. Thus, the $Fe^{\cdot\cdot}$-ions expelled from the hypersthene and hornblende were not totally removed, but were, on the contrary, oxidized and fixed as magnetite dust, especially on the cleavages of the minerals. This indicates that the diffusion processes at least partly followed the cleavages, which were characterized by low mechanical pressure.

A part of the Fe released reacted with the introduced Mg and formed spinel intimately intergrown with magnetite.

The small olivine grains which are so Mg-rich that they cannot be in equilibrium with the hypersthene, can most probably be regarded as formed, when the $Mg^{\cdot\cdot}$ introduced, because of the quenching, crystallized in the mineral with the highest velocity of formation. Thus, the olivine is a metastable mineral. (It is worth remembering that the diffusion processes, supposed to take place in the solid state, necessarily must proceed in the "intergranular-film" or through the lattices (see page 67). The magnetite dust and the olivine grains may therefore be described as "diffusion-channels").

Traces of the Si, Al, Ca, and alkalis removed from the zone of ultrabasite formation may be found in the pegmatitic schlieren and masses at the ends of the ultrabasic bands, and in the small plagioclase grains in the ultrabasic rock.

The diopside of 13431a occurs in a limited amount in corroded grains and it has not the clouding of ore-pigmentation which characterizes the hypersthene and hornblende. This indicates that the diopside was unable to take part in the stabilizing processes which in turn means that it is a primary amphibolitic mineral.

We may now consider two ways of formation of the ultrabasic bands, which were ignored in the previous chapter.

Firstly, the ultrabasics might be metamorphosed ophiolitic intrusions in the original, geosynclinal limestones. This possibility can be discarded because of the diopside of 13431a.

Secondly, the ultrabasic rock might be formed by recrystallization of an original Mg-rich layer of the amphibolite as a result of readjustment to increasing temperature and pressure. This possibility cannot be so easily rejected as the first-mentioned. The presence of diopside concretions in certain horizons in the amphibolites at Eqaluk indicates that such Mg-rich layers exist.

It is, however, very unlikely that hypersthenes of distinctly different compositions can be found in the same rock at the same P,T-conditions. We should at least, in that case, not expect that the hypersthenes of all the ultrabasic rocks examined on Langø had approximately the same composition. This fact speaks in favour of formation at certain P,T-conditions rather than in local Mg-rich bands of varying composition.

On the other hand, a discontinuity in the amphibolite (e.g. a horizon rich in diopside) may bring about the stress concentration which was responsible for the formation of the zone S—S of fig. 13. The writer has often observed that thin replacement quartz veins may follow layers containing diopside concretions which supports the statement above.

The diopside of 13431 a is slightly more Mg-rich than that of 13431b. As stated above it is unlikely that this difference is brought about by metamorphic differentiation (the possibility, that it is formed in this way, cannot be definitely excluded at the moment). It is more probable that the difference in composition between the diopsides of 13431 a and 13431 b is original.

Diopside occurs, as a metamorphic mineral, either in contact rocks (in a broad sense of the word) or in lime-silicate rocks. In the last-named rock type it is formed in the granulite-facies and in the amphibolite-facies. Examination of thin-sections proves that diopside is not in equilibrium with the adjoining minerals in lime-silicate rocks; it is always more or less replaced by hornblende. In accordance with this, the diopside concretions in amphibolites are, in most cases, separated from the surrounding rock by reaction rims of hypersthene, hornblende, or even biotite. Diopside may therefore be regarded as a metastable mineral which is formed, in an intermediate stage of the metasomatic transformation of a carbonate rock because of deficiency in Si, Al, and alkalis. Its composition is therefore determined by the primary chemical composition. This explains that the diopside of 13431a has a lower hedenbergite content than that of 13431 b.

As described on page 24 there is a gradual change in the composition of the hypersthene across the lateral border between amphibolite (13431 b) and ultrabasite (13431 a). This feature could be explained as a composition- or activity-gradient along which material migrated between the two rocks. It might also be considered as a result of conduction of heat from the tension zone towards the amphibolite (compare pp. 37 and 38). The writer, however, favours the idea that this gradual change in composition reflects the "transition" from strained to unstrained rock, i. e. the differences in composition are caused by different conditions of formation.

The last statement requires a discussion of the rôle of the orthorhombic pyroxene in metamorphism.

The hypersthenes of the Tovqussaq region (table 2) can be divided into three rather well defined groups, each of which is characterized by a fairly constant ferrosilite content.

This feature is most conspicuous in the hypersthenes of the ultrabasics, the fs-content of these having values about 20 % (14—24 %).

The amphibolites, in which hypersthene, according to the examination of thin-sections, is a true component, have pyroxenes with 32—38 % fs. (No. 4048 should not be included in these considerations as it has hypersthene-bearing layers alternating with hypersthene-free, probably as a result of metamorphic differentiation). The amphibolites of the peripheral parts of the Tovqussaq peninsula have more fs-rich hypersthenes.

Only a few pegmatites were examined. We may distinguish between two groups. The larger pegmatitic masses have hypersthene of about 36 % fs, i. e. the same as the amphibolite-hypersthenes. Towards the ultrabasic bodies the hypersthene has app. 27 % fs.

The above-mentioned features cannot be accidental. The fact, that rocks of different compositions (e. g. the ultrabasic rocks) nevertheless have hypersthenes of fairly constant compositions, indicates, that the composition of the hypersthene is determined within wide limits by the physical conditions rather than by the chemical bulk composition.

In a recent paper RAMBERG and DEVORE (1951) discussed the distribution of Fe^{++} and Mg^{++} in coexisting olivines and orthorhombic pyroxenes. They conclude that accurately measured entropy data of the system would enable us to calculate the temperature of formation from the distribution of Fe^{++} and Mg^{++} in the two minerals. They state that there is no relationship between temperature of formation and for instance Fe-content of the two minerals.

Unfortunately, RAMBERG and DEVORE do not discuss the mineralogical data on which their examination is based. It seems to the present writer that equilibrium between olivine and pyroxene in table 1 (op. cit. page 194) is the exception rather than the rule. Coexisting olivine and pyroxene belong frequently to two different generations of minerals.

The olivine-pyroxene relationship seems to be of limited value for the discussion of metamorphic rocks. First of all, RAMBERG and DEVORE's considerations are based on reactions in a closed system where the change in the physical conditions will cause an exchange of Mg^{++} and Fe^{++} between the two minerals until equilibrium is re-established. Most metamorphic processes take, however, place in an open system permitting introduction and removal of material, in which case the two minerals will be in equilibrium when they have attained the compositions corresponding to the given temperature (compare the isothermal planes in BOWEN and SCHAIRER, 1935). Another objection to RAMBERG and DEVORE's statements is, that olivine is no metamorphic mineral (with the exception of olivines of ultrabasic rocks and metastable forsterite of crystalline limestones). Thus, olivine cannot crystallize or recrystallize under common metamorphic conditions and olivine found in metamorphic rocks must therefore be remnants of primary igneous rocks or metamorphic ultrabasics. The olivine is not in equilibrium with the coexisting pyroxene (compare the dunite at Siorarssuit, West Greenland, where olivine of constant composition is found in contact with pyroxenes of at least three different compositions, SØRENSEN, 1951b).

The crystallization of a silicate melt is dependent on the temperature within the limits determined by the chemical bulk composition, i. e. at each stage of the crystallization, mineral phases are separated or resolved in such a way that the solid phases and the melt are in mutual equilibrium at the given temperature. In accordance with this the members of the solid-solution series in all rocks formed at a certain stage of the crystallization of a layered complex, have constant compositions, in spite of varying mineralogical composition of the rocks in question (see e. g. WAGER and DEER, 1939).

The metamorphic processes take place in an open system with free access to exchange of material. Chemical activity, not concentration is the driving force of diffusion. Material migrates from places of high, to places of low activity and the processes tend to eliminate differences in activity. A familiar large scale example is granitization. All rocks of an area are, irrespective of primary composition, transformed into granitic rocks, which have the least activity of all rocks in the level of the crust in question. In a recent paper RAMBERG (1951) has demonstrated that the endproduct in the granulite facies has quartz-dioritic composition, while the composition is grano-dioritic in amphibolite facies.

Let us return to the orthorhombic pyroxenes. The experimental work of BOWEN and SCHAIRER (1935) demonstrated that Mg-rich pyroxenes are stable at high temperature, Fe-rich at low. The isothermal planes of these writers show that coexisting olivine and hypersthene both become more and more Mg-rich at rising temperature in such a way that the Mg/Fe ratio is largest in the pyroxene.

We should expect that the Mg/Fe ratio in the hypersthene in metamorphic rocks increases with increasing temperature. It is natural to associate the change of composition observed in the hypersthene from the east coast of Langø (13431a) with a rise in temperature. We observed in this locality that Mg was introduced into and Fe expelled from the hypersthene. The same applies to the hornblende (and to the rock). This means that we have, in conformity with BOWEN and SCHAIRER's isothermal planes, a simultaneous increase of the Mg/Fe ratio in the hypersthene and in the hornblende.

Data are still scarce, but BARTH's examination (1950) of the bahiaites of Southern Norway indicates that the Mg/Fe ratio is larger in hornblende than in coexisting hypersthene.

Now the composition of the hornblende, being complicated, varies in no simple way with the temperature, but is probably to a certain extent determined by the bulk composition of the rock.

The hypersthene is, on the other hand, a pure Mg-Fe mineral which does absorb only subordinate amounts of other components (foreign components may still exist because of uncompleted diffusion processes). It is therefore reasonable to assume that the composition of the hypersthene of the open system is primarily determined by the physical conditions. Thus, hypersthene may be qualitatively used as a geological thermometer.

Consider the ultrabasic rocks of Langø. The temperature difference between the tension zone and the amphibolite has probably been small. If the temperature of the tension zones increases further, we may assume that the hypersthene becomes more Mg-rich; at the same time it increases in quantity. At a certain temperature, an olivine, which is in equilibrium with the hypersthene, appears and the end-product should be a dunitic rock. As will be mentioned below intermediate stages between the Langø rocks and dunites are actually found in other West-Greenlandish localities.

Thus, we may reach temperatures and pressures, corresponding to much deeper levels of the crust, in the tension zones. The high stress in the latter may, however, play an important rôle in lowering the temperature of crystallization of the minerals in question (see page 38).

The statements above should be taken with some reservation as it is extremely difficult to distinguish between effects of chemical composition and effects of P,T-conditions. Besides, it is not always possible to decide whether the minerals are stable, or not. But, the examination of the hypersthenes of the Tovqussaq region strongly suggests that the conclusion arrived at above is a possibility to take into account in petrological considerations.

The Retrograde Transformation on Langø.

The retrograde transformation has not progressed very far on the east coast of Langø, where the first stages of the ultrabasite formation may still be studied.

On the west coast, on the other hand, the transformation is very pronounced because of the more advanced tectonic deformation and the pegmatitic material (the dispersed phase) which has penetrated in between and into the ultrabasic masses. The transformation is "polar" since the northern ends of the masses are generally most highly transformed. This fact in connection with the almost total lack of low temperature minerals indicates that the retrograde transformation of the ultrabasic rocks was intimately connected with the deformation which took place under granulite facies conditions. The only traces of deformation at a later date are the mylonites, in part ultramylonites, described by A. BERTHELSEN (op. cit. pag. 561). The ultrabasic rocks, however, were not affected by transformation after the main period of deformation. Thus we have here an example of the dependence of metamorphism on deformation[1]).

As the deformation on the west coast of Langø shows a continuation of the processes which were in action on the east coast, the transformation might be described as progressive, but as the ultrabasic rocks are most important in the present discussion, and as the P,T conditions prevailing in the tension zones, in which the formation of the ultrabasic rocks took place, are not reached in the later stages of the deformation, the term "retrograde" seems to be justified.

The parageneses of the ultrabasic bodies of the west coast, which were stable in the tension zones, are obviously not in equilibrium when exposed to the observed change in the physico-chemical conditions. Nor is there equilibrium between the bodies and their surroundings.

As described on page 16 only the marginal and the "tectonically most exposed" parts of the ultrabasic masses were the subject of a more radical transformation. This may, as stated above, be taken as evidence of the connection between deformation and transformation. In addition it supports the view, obtained from field- and laboratory observations, that ionic diffusion in the solid state (replacement processes) was responsible for the transformation. The period of deformation was, although undoubtedly of long duration, nevertheless too short to permit a thorough alteration because of the slowness of the diffusion processes. Further, the marginal reaction zones may, to a certain extent, have protected the central parts of the ultrabasic bodies.

In the central parts of the ultrabasic masses, which were hardly reached by material from the surroundings, the retrograde processes consisted chiefly of a reorganization of material. Hornblende was formed at the expense of the hypersthene, and the components of the spinel entered into the hornblende. The magnetite also disappeared. Exam-

[1]) Static metamorphism (load metamorphism) is certainly much over-estimated in geological literature. The velocity of diffusion decreases with increasing confining pressure. Stress accelerates, on the contrary, the processes. (Compare J. A. W. BUGGE, 1945, p. 30, and H. RAMBERG, 1946). Therefore, metamorphism is probably intimately connected with deformation in the levels of the crust considered here. Static recrystallization may occur in the upper parts of the crust where vapours and solutions are of some importance.

ination of the hypersthenes of 13 different ultrabasic bodies and the example described on page 27 show that the composition of the hypersthene remained constant in the inner parts of the masses. Thus the original hypersthene is preserved.

The hornblende formation is most interesting. The scarce optical data may suggest that the original Al-poor hornblende was gradually transformed into a more Al-rich one. (Compare the analyses of table 4).

The inner parts of the ultrabasic masses may, taken as a whole, be regarded as a sort of "armed relics", as their parageneses correspond to a temperature higher than that prevailing during the deformation (as indicated by the marginal reaction rims).

In the marginal parts of the masses the transformation is more advanced as it is favoured by the tectonical processes and by the difference in chemical activity between the ultrabasic rocks and the surrounding mainly quartzo-feldspatic rocks. The gradient thereby formed is responsible for the lively exchange of material between the two rock types.

The surrounding rocks most often pegmatitic have in most cases developed a more fine-grained zone towards the ultrabasic rock. The composition of this border rock is fairly constant and it consists of andesine (35 $^0/_0$ An), a little quartz, and a hypersthene with a negative axial angle of about 70° and $n\gamma$ ab. 1.700. Biotite is found locally in rather large quantities, while hornblende only occurs as a fine-grained aggregate on the cleavages of the hypersthene.

It is worth noting that hornblende does not seem stable in quartz-feldspar rocks under the physico-chemical conditions in question. Consequently, the hornblende is dissolved in the outer parts of the ultrabasite, which explains the observed increase in An-content in the plagioclase from the pegmatite towards the ultrabasic rock (see page 32).

In the samples examined, a zone with fairly large hypersthene grains was observed between the pegmatite and the ultrabasite. There is also a concentration of hypersthene grains in the fine-grained border of the pegmatite. This hypersthene is secondary and is formed at the expense of the hornblende (and hypersthene) of the original rock. A similar example is described by RAMBERG (1948a, pp. 14 and 27) from Sukkertoppen, West Greenland, where hypersthene and plagioclase also are formed at the expense of the original minerals.

Inside the hypersthene zone follows a zone with plagioclase, hypersthene, biotite, and in some cases, near the ultrabasic rock, small grains of a secondary hornblende, darker than that of the unaltered rock.

Under the microscope the substitution of pegmatitic material for the original rock is quite evident. The hypersthene and the hornblende are replaced by plagioclase (and quartz), the released Mg and Fe are partly bound in biotite which may be associated with the cleavages of the hypersthene.

It is interesting that hornblende, although not being stable in the pegmatite, nevertheless occurs in small, secondary grains inside the outer biotite zone. A few fairly large hypersthene grains may have biotite on the cleavages at the pegmatite end, while they have hornblende towards the ultrabasic rock. This feature is probably connected with the composition gradient discussed in the following section.

As mentioned above there is a gradual transition in the composition of the hypersthenes in the border zone between ultrabasic rock and pegmatite. Even a zoned hypersthene, with a higher fs-content in the outer zone than in the core, was observed.

Now, the ultrabasic bodies, which were stable at higher temperature than that prevailing under the retrograde transformation, have lower chemical activity than the surrounding quartz-feldspar rocks; we may say that the ultrabasic rocks are not wanted in the rock association of the west coast of Langø. Consequently, material of the dispersed phase migrates towards the ultrabasic rock along activity gradients between the incompatible rocks (see Ramberg, 1944, page 107). The dispersed phase may in this case be compared with the leucocytes of the blood.

The activity gradient is reflected by the gradual change in composition across the border zone. The exchange of material takes place through the minerals of this narrow zone and as the system was evidently quenched in an early stage of the transformation, the appearance of the observed "composition gradient" is easily understood. It is not difficult to explain that the most An-rich plagioclase, the most fs-poor hypersthene, and hornblende (instead of biotite) occur close to the ultrabasic rock.

If the rate of diffusion was known under the P,T-conditions in question, we should on closer examination of the narrow border zone be able to get information about the length of the period during which the processes were active.

In addition to the above-mentioned facts concerning pegmatite formation by replacement, further evidence may be found in stripes of pigmentation in the plagioclase, the stripes may in some cases continue in corresponding pigmentation lines in the adjoining hypersthene. (Cf. also the example on page 16, where amphibolitic structure was observed in the pegmatite).

Accordingly, the so-called "space problem" is no problem here. The space in question is, since the processes take place in an open system, filled by replacement processes with the minerals stable at the given physico-chemical conditions.

As the chemical composition is not constant, the metamorphic processes should not be interpreted as reactions between simplified formulae for the compositions of the minerals, not even when it is a question of reaction rims between adjoining minerals.

ULTRABASIC ROCKS
FROM OTHER PRE-CAMBRIAN-AREAS

Ultrabasic rocks of various types have been described from most pre-cambrian areas. The ultrabasic rocks are, however, as a rule considered as being independent small intrusions and most descriptions pay little attention to the structure of the surroundings. Consequently it is difficult, on the basis of current literature, to extend the point of view expressed on the preceding pages to other occurrences.

Therefore this chapter presents a brief treatment of ultrabasic rocks from other areas studied by the writer. In the following chapter, which deals with the problem of the formation of ultrabasic rocks in orogenic zones, additional examples from the literature will be quoted.

1. South Harris, the Outer Hebrides.

In the summer of 1950 the writer with Mr. R. Phillips of Durham visited the area near Rodil, South Harris, which has been described by C. F. Davidson (1943).

The area is built up of alternate layers of paragneisses and metagabbros. A band of amphibolite stretching from the Rodil Hotel past the school has two wedge-shaped, conformable ultrabasic inclusions.

In a small quarry NNW of the Rodil Hotel (Davidson, op. cit. pag. 85) a fine-grained ultrabasic rock, which is very similar to 13431a from the east coast of Langø, occurs. It is composed of diopside, edenite, hypersthene ($(\div)\,2V = 79°$), olivine, magnetite, spinel, and serpentine. As on Langø olivine occurs as small grains but in slightly larger quantities.

The outer part of the inclusion is, at least in some places, transformed into a hornblenditic rock which has inclusions of unaltered ultrabasite. The border relations are, however, difficult to study because of the vegetation, but as small outcrops outside the ultrabasic rock consist of pegmatite, it seems reasonable to assume that pegmatitic material playes a rôle in the border between the inclusion and the surrounding amphibolite.

The occurrence is penetrated by pronounced crush zones, so much that it may be termed a crush-breccia. The rock is divided into angular fragments, which are bounded by slickensided surfaces.

The other ultrabasic body is found at the Rodil schoolhouse. It consists of diopside, hypersthene (2 V = (÷) 80° (DAVIDSON, page 85), and a small quantity of hornblende and mica. This occurrence is brecciated too, but the individual fragments have an outer zone of hornblende separated from the adjoining fragments by a thin layer of mica. The hornblende has obviously been formed at the expense of the ultrabasic rock and has inclusions of minerals belonging to the latter e. g. highly corroded grains of hypersthene (cf. DAVIDSON's amphibolite, op. cit. pag. 85).

The enclosing garnet-amphibolite consists of brown hornblende, diopside, garnet, plagioclase, a little biotite and quartz and, occasionally, hypersthene ((÷) 2 V = 60°, DAVIDSON, page 106). The garnet occurs in very irregular grains associated with diopside, hornblende, hypersthene and biotite.

Due east of Rodil Hotel hypersthene lumps which may reach a size of $^1/_2$ meter are found in a limited area of the amphibolite. Several lumps may be arranged in the same strike horizon. They are usually surrounded by light coloured schlieren and built up of hypersthene grains of up to 1 centimeter ((÷) 2 V varies from 59° to 72°, the grains have ore-coatings on the cleavages and are often clouded with a fine pigmentation), and have in addition garnet and diopside. In a few lumps hypersthene occurs in quantities of secondary importance, in which cases garnet, diopside and hornblende are the predominant minerals.

From the description above it will appear that there is much similarity between Tovqussaq and Harris, not only in the petrography of the ultrabasic rocks but also in their mode of occurrence. In both places the ultrabasites are conformably enclosed in continuous bands of amphibolite. There is then reason to believe that the rocks of Harris were formed in the same way as those of Langø. If so, the small lumps of hypersthene east of the Hotel (in the layer of amphibolite that encloses the ultrabasic bodies) may be explained as "ultrabasite embryos".

The retrograde transformation in Harris is more advanced than on Langø. The association of garnet-diopside-hornblende in the amphibolite shows that the rocks were transformed in the upper part of the amphibolite facies. The sporadic occurrence of hypersthene in the amphibolite renders it probable that the minerals in question were formed at the expense of hypersthene at falling temperature. This indicates that, originally, the rock has been a hypersthene-amphibolite which further emphasizes the similarity to Tovqussaq.

As the eclogites described by DAVIDSON contain as much as 10 $^0/_0$ hypersthene (($\div$ 2V varying from 70° to 80°, DAVIDSON, page 101), it cannot be precluded that, in accordance with the statements above, these eclogites represent transformation products of original hypersthene-carrying, ultrabasic rocks. The supposition is further confirmed by the diopsidic nature of the pyroxene of the eclogites (DAVIDSON, op. cit. page 99), and by the data given for the hypersthene of the rocks, as well as the lack of quartz in the eclogites.

2. The South Isortoq-Alangua Region, Sukkertoppen District, West Greenland.

It has already been mentioned on page 7 that ultrabasic rocks occur as inclusions in hornblende-gneisses in this district.

The hornblende-gneiss consists of plagioclase (30—50 $^0/_0$ An), hornblende (green, ($\div$) 2V varying from 60° to 70°, $c : \gamma$ = ca. 15°), brown biotite, quartz, and small quantities of apatite, magnetite and rutile. It has numerous light-coloured veins (often ptygmatically folded) but is otherwise fairly homogeneous. In places it contains dark, amphibolitic inclusions, and from its smooth transition to amphibolitic rocks it appears that it must have been formed at the expense of the latter. Examination under the microscope of the gneiss around the ultrabasic lenses shows, that the gneiss towards the latter is highly cataclastically deformed.

The amphibolites of the area vary from typical calc-silicate-rocks to common amphibolites. The first named have diopside ((+) 2V = 58°, $c : \gamma$ = 40°) as the predominant mineral, the other components are actinolitic hornblende and bytownitic plagioclase (74 $^0/_0$ An). A rock of this type occurs as a thick, continuous band on the west coast of Alángua (see map, fig. 1). The common amphibolite consists of green hornblende (($\div$) 2V = 71°, $c : \gamma$ = 18°), plagioclase (40 $^0/_0$ An) and subordinate amounts of diopside and quartz. Garnet may occur in all types, especially where the amphibolites have pegmatite-filled cross-joints. Inclusions of amphibolite in the pegmatites have their original orientation, thus the pegmatites are formed by replacement. The elements released by this process may be found in tourmaline, beryl and apatite in the pegmatite.

The hornblende-gneiss conformably encloses ultrabasic bodies with a diameter of as much as 100 metres. These bodies are always surrounded by pegmatites.

Petrographically, the ultrabasic rocks vary from types identical with the rock from the east coast of Langø to dunitic rocks.

The occurrence Ol.5 (cf. map fig. 14) consists of hypersthene (($\div$) 2V = 86, $n\gamma$ = 1.686), green hornblende (($\div$) 2V = 85°, $c : \gamma$ = 18°), pleonaste, magnetite, and a few corroded grains of diopside; in other words, a paragenesis closely corresponding to the Langø rocks (plate 5, fig. 10). This occurrence is enclosed in a gneiss with numerous amphibolitic inclusions.

North of Ol.5 is the occurrence Ol.3 which consists of hypersthene ((÷) $2V = 86°$), olivine in grains up to $^1/_2$ centimeter ((÷) $2V = 87°$), green hornblende ((÷) $2V = 87°$, $c : \gamma = 15°$), pleonaste and magnetite.

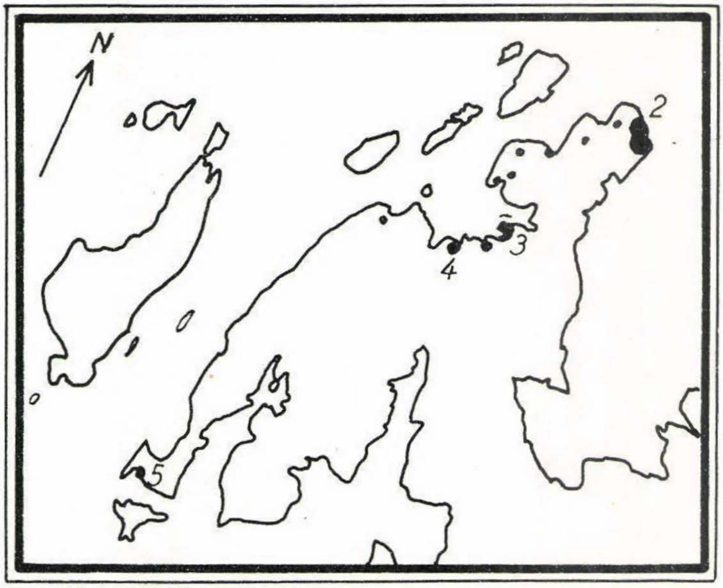

Fig. 14. The peninsula on which the occurrences Ol. 2, 3, 4, and 5 are situated (compare fig. 1).

Hornblende seems to be formed, at least partly, at the expense of the olivine. Hypersthene and olivine were probably formed in mutual equilibrium, judging from their compositions (plate 6, fig. 11).

The least disturbed part of the occurrence Ol.1 consists of olivine ((+) $2V = 87°$, enstatite ((+) $2V = 84°$), chromite (ab. 2 $^0/_0$), a little hornblende and serpentine, the latter especially in a few narrow displacement zones. The enstatite was, at least partly, formed at the expense of the olivine. In conformity with this, the enstatite has more fs than the olivine fa.

The undisturbed part of the occurrence in Amitsuarssoralak shows conditions similar to Ol.1, but the olivine has (+) $2V = 89°$ and the orthorhombic pyroxene (+) $2V = 86°$ and $n\gamma = 1.680$ (plate 6, fig. 12). Chemical analyses of the mentioned rocks from Ol.1 and Amitsuarssoralak are given by Sørensen (1951b).

Fig. 15. Pegmatite penetrating the southern part of Ol.5.

Fig. 16. Detail of fig. 14. Note the partly dissolved ultrabasic inclusions in the pegmatite and the large mica flakes of the latter.

The ultrabasic bodies bear marks of a more or less pronounced deformation. The transformation is polar as on the west coast of Langø, intense at one end of the lenses (in the direction N 20—40° E, parallel with the fold axis of the area), while the other end has apparently been in a sheltered position during the deformation processes.

Fig. 17. Displacement zones in the northern end of Ol. 3. Note the large hypersthene grains of the ultrabasic rock.

The northern part of Ol.5 is of a massive appearance, while the southern part is broken up into angular fragments by a pegmatite containing partly dissolved inclusions of ultrabasic rock and clearly formed by replacement (figs. 15 & 16). Under the miscroscope the pegmatite appears highly cataclastically deformed.

The northern, massive part of Ol.3 is intersected by zones filled with serpentine-anthophyllite along which a slight displacement has taken place (fig. 17). The southern part, on the other hand, is foliated or even folded (fig. 18).

In Amitsuarssoralak, the southern part of the ultrabasic body is a massive olivine-bronzite-rock, while the northern part is a foliated bronzitite with bronzite prisms (up to 20 centimeters) indicating the direction of the axis of folding of the area. The formation of this rock took

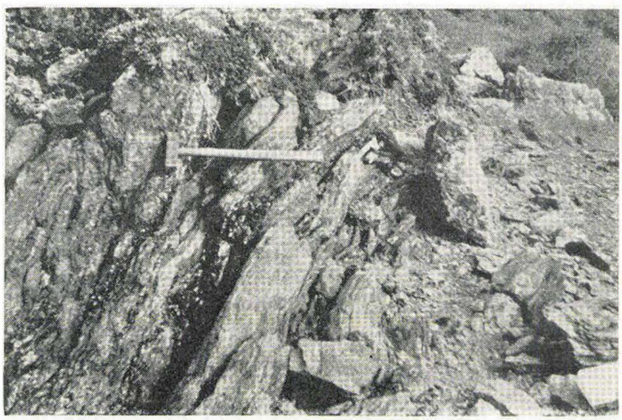

Fig. 18. Schistose ultrabasic rock at the southern end of Ol. 3.

probably place in an early phase of the deformation. In accordance with this view the bronzite seems to have almost the same composition as that of the olivine-bronzite rock ((+) $2V = 88°$, $n\gamma = 1.684$). At a later stage of the deformation, with lower temperature and less stress, hornblende begins to replace the bronzite, as the latter poikilitically encloses numerous small, euhedrale hornblende grains the longitudinal direction of which is at right angles to the foliation of the rock (plate 7, fig. 13). The largest bronzite prisms have only hornblende inclusions in their outer parts.

At Ol.1, conditions are more complicated. The occurrence is divided into numerous large and small lumps. The southernmost of these is the above mentioned massive olivine-enstatite rock which in its outer parts is a monomineralic enstatite rock. The northernmost lumps are hypersthene-hornblende rocks similar to the rock from the northern part of the Amitsuarssoralak occurrence. The orthorhombic pyroxene in this rock has $n\beta = 1.682$ and is thus more fs-rich than the pyroxene of the southern lump.

To the north-west, the rocks of Ol. 1 are "rolled out" in the gneiss and are partly digested by the latter so that a hybrid, hypersthene-carrying rock is formed. Small lumps of the original hypersthene-hornblende rock may still be found "swimming" in the hybrid rock, just as primary, large hypersthene grains, of the same appearance as in the original rock, are scattered all over the hybrid zone. These large grains are much corroded and may be divided into several small parts. As these have a common optical orientation and are still arranged as the hypersthene of the original rock, the transformation must have taken place by replacement in the solid state (see plate 7 fig. 14). The hybrid rock consists of plagioclase (60 $^0/_0$ An), hypersthene of the same composition as the hypersthene of the hypersthene-hornblende rock, hornblende, biotite and quartz. A similar rock has been observed as an inclusion in the hornblende-gneiss south of the Amitsuarssoralak occurrence. In contradistinction to the rock mentioned above this contains anthophyllite prisms, and its plagioclase has 85 $^0/_0$ An. Its hypersthene is identical with that of the above-mentioned hybrid rock. The final product of the described transformation is a rock of plagioclase, biotite, and quartz. This can, however, be studied best in the much transformed occurrence Ol.2.

The principal rock at Ol. 2 is composed of hypersthene and hornblende. The axial angle of the hypersthene is (÷) 82°. In the outer parts of the occurrence and in small inclusions of ultrabasite in the surrounding gneisses the incipient transformation can be studied. Light-coloured schlieren of plagioclase are seen penetrating the rock between the large hypersthene porphyroblasts of the ultrabasic rock. Hypersthene is replaced by plagioclase (ab. 70 $^0/_0$ An) and biotite in the way described on page 31 from the west coast of Langø. But while hypersthene (and not hornblende) was found in the quartzo-feldspatic rocks of Langø, quite the opposite is the case here. The plagioclase schlieren contain grains of hornblende with well developed crystallographic outlines (plate 8 fig. 15), while hypersthene is not observed. This difference may easily be explained by the fact that the transformation took place in granulite facies on Langø, while in the present case it took place in amphibolite facies, where hypersthene is not stable.

Hybrid rocks like those described from Ol. 1 and also certain types from Ol. 2 have highly corroded grains of hypersthene in quartzo-feldspatic rocks (plate 8, fig. 16). The hypersthene is, however, obviously not in equilibrium with the surroundings in these cases.

The light-coloured schlieren mark the first stages of the transformation. Dispersed pegmatitic material enters the cracks in the ultrabasic rock and supplants the ultrabasic material. These first stages are followed by a more thorough transformation which may be studied especially in the outer parts of the ultrabasic body and in small inclusions of ultrabasic affinity in the surroundings (fig. 19). Thus, the southernmost part of the body is a plagioclase (32 °/₀ An)-biotite-quartz rock which has lost all ultrabasic characteristics. Only the great amount of biotite discloses its ultrabasic origin, the biotite representing released ferro-magnesian material not removed by migration.

In addition to the above mentioned light schlieren the ultrabasic rock of Ol. 2 in places contains "druses" chiefly of plagioclase, quartz and biotite. The plagioclase here has the same composition as in the surrounding pegmatite, i. e. 25—35 °/₀ An. The druses must therefore be regarded as pegmatitic material which consolidated without having

Fig. 19. Partly dissolved ultrabasic body in hybrid rock immediately to the north of Ol. 2.

reacted with the minerals of the original rock. The possibility of the druses being formed by secretion of acid material as a result of metamorphic differentiation, is rendered improbable by the trend of the transformation.

As stated above all ultrabasic bodies of this area are separated from the gneiss by pegmatites. In the least transformed parts of some of the occurrences traces of an outer zone may be observed, in which the hypersthene contains more fs than in the central parts. In other words, we have here the remains of a high temperature border facies. As the surrounding rocks are formed in amphibolite facies, the hypersthene cannot exist in contact with the border pegmatites. Accordingly, the ultrabasics are separated from the latter by a reaction zone consisting of hornblende and/or biotite, more rarely of anthophyllite. In one instance the boundary rock consists of biotite, garnet, plagioclase and quartz.

As on Langø, the retrograde transformation is most advanced in the marginal parts and along cracks in the ultrabasic masses. The central parts are in many cases not reached by the processes in question. The excess of the material migrating towards the ultrabasic rocks consolidated in the pegmatites.

It is worth noting that a biotite zone may be observed in some occurrences in the unaltered rock inside the outer reactionzone. This is seen at best at Ol. 1 (fig. 20) where a zone in the original olivine-enstatite rock, a short distance inside the outer monomineralic pyroxenite zone, has large flakes of mica, which obviously replace the olivine and the enstatite (plate 17, fig. 9). This zone probably marks the front of the migrating material received from the surroundings.

Fig. 20. Mica zone (in the center of the photograph) between bronzite-rock to the left and olivine-enstatite rock to the right. Southern end of Ol. 1.

Low temperature transformation products have hardly been observed. The rocks carrying olivine, however, contain a little serpentine, talc and carbonate which are especially associated with the displacement zones.

The ultrabasic rocks of this area show conformity with the west coast of Langø on the following points:—

1) constant association with hornblende-gneisses formed at the expense of para-amphibolites,
2) the occurrence Ol. 5 is petrographically identical with the rocks of Langø,
3) the ultrabasic bodies show signs of polar deformation.

Hence it seems natural to assume that these rocks were formed through processes of a similar kind as those which gave rise to the formation of the ultrabasic rocks of Langø.

The structural evolution in connection with the dome formation was made responsible for the formation of the ultrabasics of Langø. In the area dealt with in this chapter the formation of the rocks discussed may be associated with high stress in connection with the intense folding of the area.

On Langø, where the processes may be regarded as a "laboratory experiment" demonstrating the course of the formation of the ultrabasic rocks, the processes were on a small scale. In the South-Isortoq-Alángua region greater forces were at work. The ultrabasic masses reached a greater magnitude, and higher temperature (in connection with stress of high order) resulted in the formation of olivine.

It is most interesting to note that the hypersthene-olivine-pleonaste-hornblende rock of Ol.3 is an intermediate step between the rocks of Langø and Ol.5 on the one hand and the rocks of Ol.1 and Amitsuarssoralak on the other. This is in close agreement with the statements presented on page 52, especially as the coexisting olivines and hypersthenes of Ol.3 and Amitsuarssoralak have compositions which, according to the isothermal planes of BOWEN and SCHAIRER, indicate that the minerals were close to mutual equilibrium (compare BOWEN and SCHAIRER, op. cit. p. 213).

That chromite is present only in the olivine rocks is easily explained. Thus WAGER and MITTCHEL (1951, page 183) state that "the emphatic preference of chromium for pyroxene rather than olivine is clear". This means that Cr in the hypersthene-rich rocks is present in the crystal structure of the pyroxene. When olivine substitutes the pyroxene, the Cr released consolidates as chromite. When olivine is transformed into hypersthene the opposite is true concerning Ni which enters olivine more easily than pyroxene. The Ni released is concentrated in pentlandite-bearing pyrrhotite associated with the secondary hypersthene.

The examples described on page 63 and page 62 from Ol.2 and Amitsuarssoralak indicate that the retrograde transformation of the ultrabasic rocks with large hypersthene porphyroblasts proceeds from the grain boundaries. This supports WEGMANN's view that the material migrates in the "intergranular film" (WEGMANN, 1935).

This is easily understood by a consideration of the border relations of the minerals.

"A particle at the boundary between two phases 1 and 2 is simultaneously subjected to forces exerted on the one side by particles of phase 1 and on the other side by particles of phase 2. If the kind or arrangement of particles is different in the two phases, these two forces will be unequal. To maintain equilibrium, it is therefore necessary that particles of both phases in the immediate vicinity of the boundary should become rearranged in such a manner as to balance the forces exerted between them" (Turner and Verhoogen, 1951, page 397).

Thus the minerals are "surrounded by disorder zones" stable at given P,T-conditions. If the P,T-conditions are changed this surface phase arrangement is no longer stable. The particles will be rearranged in such a way that equilibrium is reestablished. The nuclei of corona minerals may be formed in this way. Besides it is possible that the unstable surface phases may act as migration channels (i. e. intergranular films).

3. The Angmagssalik District, East Greenland.

Through the kindness of Professor L. R. Wager the writer, when visiting Durham in the spring of 1950, was given the opportunity of examining rock specimens and thin-sections of ultrabasic rocks from the Angmagssalik district and of studying the field notes which describe the mode of occurrence of these rocks. A preliminary description of the rocks has been published by Professor Wager (1934, page 13).

The ultrabasic rocks occur as inclusions in grey hornblende-gneiss which also contains bands of amphibolite and highly transformed sediments. The amphibolite may be broken up into boudins, and here and there the gneiss has incorporated material from the more or less dissolved parts of amphibolite.

The ultrabasic masses may be as much as 2—300 metres across and are most often seen as small masses arranged in the strike direction of the gneiss. The rocks have often hypersthene porphyroblasts and consist moreover of hornblende, biotite, chlorite, calcite, ore and serpentine. Some rocks contain olivine which may be enclosed in the hypersthene grains.

The ultrabasic inclusions are much transformed along the borders and fissures and may contain biotite, hornblende, anthophyllite, and chlorite there.

It will be seen that there is on many points a likeness to the West Greenland rocks mentioned in the preceding paragraph, but in contradistinction to the rocks from the Isortoq-Alángua region, the ultrabasic rocks from Angmagssalik contain a large quantity of chlorite and serpentine. This shows that the retrograde transformation took place at a low temperature perhaps under conditions corresponding to epidote-amphibolite-facies.

4. The Godthaab Fiord Area, West Greenland.

As mentioned on page 7 the area south of Tovqussaq is formed under granulite facies conditions. Further south, in the Godthaab fiord area, it is succeeded by a new amphibolite facies complex. The latter was examined in the summer of 1951 by Dr. H. RAMBERG, Mrs. M. L. RAMBERG and the writer. So far, the rocks of the complex have not been examined in the laboratory, but, nevertheless, some features of importance for the problems discussed in the present paper will be briefly and preliminarily described in the following.

The complex comprises a series of various gneissic rocks including thick layers of amphibolite and garnet-sillimanite-micaschists. The two last named may be intimately interbanded giving the impression of a highly metamorphosed sedimentary series. In addition, quartz-rich, in part sparagmitic, rocks are occasionally seen, but in subordinate amount. In one locality a rock, which may be interpreted as a highly deformed conglomerate, was found.

The complex is intensely folded, the fold axes (NE—SW) being parallel with the axes in the Isortoq-Alángua area. The underlying structure, which was probably responsible for the formation of the Tovqussaq dome, may be traced in NW—SE-directed structural elements. In this connection it should be noted, that the study of aerial photographs has revealed a dome-structure in the country between Tovqussaq and the Godthaab fiord probably situated in the borderzone between a western granulite- and an eastern amphibolite-facies area.

Numerous occurrences of ultrabasic rocks were visited during the field work. They vary in size from the very small to masses several hundred metres across and comprise peridotites with all transitions through anthophyllite-hornblende rocks to serpentinites and soapstones. They occur in part as bands in amphibolite, in part as conformable lenses in gneisses, which in almost all cases could be proved to be formed at the expense of amphibolites. The amphibolitic layers in the mica schists have also occasionally bands of ultrabasic rocks.

The thick amphibolite layer, which runs along the east coasts of the two islands Sermitsiaq and Bjørneø, has conformable, up to 50 metres thick mylonite zones. The amphibolite is, in these zones, recrystallized into a green hornblende-chlorite-biotite rock which, especially in the western part of the amphibolite, encloses boudinages of ultrabasic rocks. Great masses of pegmatite, in part associated with cross-joints, are locally present in this western zone. Pegmatites of this category in the south-eastern part of Sermitsiaq contain a huge quantity of tourmaline and muscovite, the first-named often in "suns" of large crystals.

The ultrabasic masses in the mylonites consist in places of a more or less serpentinized dunitic rock, the olivine of which has ab. 6 % fayalite. Olivine is the predominant mineral in limited areas, but most often the olivine grains are more or less corroded and rolled out in a serpentine-anthophyllite-hornblende rock formed at the expense of the dunite (plate 9, fig. 18), probably at the same time as the amphibolite in the mylonite zones recrystallized into the above mentioned green rock. It is a remarkable feature that the corroded olivine grains are very fresh and without traces of translation lamellae.

The shores of the sound, which separates Sermitsiaq and Bjørneø, provide two very instructive sections across the strike. The large eastern amphibolite layer apparently continued further west than it is seen now, but it has here been digested by the gneiss. The latter

Fig. 21. Small scale folding. South coast of Bjørneø.

contains in places a great amount of amphibolitic inclusions. The green mylonite zones may also be found as well as their ultrabasic inclusions, both rock types being more or less attacked by the granitization.

In the middle part of the islands (on the shores of the above mentioned sound) conditions are most complicated. The granitic gneiss, pegmatites, green chlorite rocks from the shear zones, and highly transformed, mica-rich ultrabasic rocks are intimately interfolded in very small folds (fig. 21). The main axis of folding of the area — NE—SW — plunges steeply towards the NE as well as towards the SW. In addition a cross-folding axis — NW—SE — with south-eastern plunge is observed[1]). The cross-axis corresponds with the direction of trend lines of the supposed underlying structure, which are probably also responsible for the formation of the sound in question.

The small scale folding mentioned above was followed by a granitization in such a way that the intimately folded rocks were partly

[1]) A similar small-scale folding has been described by N. Edelman, 1949, page 31.

or wholly replaced by an almost massive granite. All stages of transition exist from folded rocks to granite with faint traces of, or without folding structures. In an intermediate stage fragments of small-folded rock are enclosed in granite. As there is no parallelism between the linear structures of the different inclusions because of the small scale folding, we have conditions which would formerly have been described as an intrusion breccia. It seems to be a common feature that a period of high plasticity precedes granitization.

West of the folded complex, granitic gneisses, in part hornblende-bearing, begin to appear again, still containing ultrabasic inclusions.

Fig. 22. Amphibolite layer with ultrabasic bands in its western part (to the left). Immediately to the west of the amphibolite, the ultrabasic zone mentioned in the text can be seen. West coast of Bjørneø.

Then follow large masses of peridotitic rocks which consist of olivine, bronzite, hornblende and spinel. They are enclosed in a hornblende gneiss which contains remnants of amphibolite as well as chloritic mylonite zones. This makes it probable that we have the site of a former immense shear zone here.

The peridotites are situated on the eastern limb of a large anticline which has a "core filling" of a homogeneous and fairly massive granite.

In the western part of the anticline a peridotite complex corresponding to the above mentioned occurs. There are numerous large bodies of peridotite in the western part of the two islands mentioned.

On the peninsula north of the south-western point of Bjørneø olivine-bronzite-hornblende-spinel rocks are present as bands in a layer of amphibolite. Two bands, the maximum thicknesses of which are about 5 meters, are separated from one another by ab. 2 meters of amphibolite. They can be followed along the strike of the amphibolite for more than 100 meters. Then, they wedge out, but, after an interruption, ultrabasic rocks appear again in the same strike horizon. Immediately to the west of the amphibolite, elongated masses of ultrabasic rocks occur in one and the same strike horizon in banded gneiss. The thickness of this interrupted ultrabasic layer is 30 meters (fig. 22).

These western occurrences may be compared with the conditions in the large eastern amphibolite layer previous to the formation of the green mylonite zones. The ultrabasic bands are further comparable with the ultrabasics of Langø.

Ultrabasic rocks are a prominent feature of the geology of the Godthaab fiord. To the above mentioned may be added the occurrences of bands of tough serpentinites in amphibolite on the west coast of Storø and the north-east point of Bjørneø on the east slope of the Qornoq mountain. Finally, numerous bodies of serpentinites and talc rocks are present in the eastern highly granitized area of the fiord.

The border zones of the ultrabasic bodies are most interesting, especially where the rocks are brecciated in a late stage of the deformation. Beautiful large-scale reaction rims are then developed. Most often the outer zone consists of biotite; inside it follows a light green hornblende zone and then a zone with long anthophyllite prisms arranged at right angles to the border. The ultrabasic fragments may be quite fresh, but are in many cases altered into serpentine, and talc rocks.

The ultrabasic masses are, as in the other areas described, separated from the surrounding gneiss by pegmatites.

Large flakes of molybdenite occur in some ultrabasic masses in the outer green hornblende zone or in the biotite-hornblende border zones of penetrating pegmatites. This feature is most satisfactorily explained by assuming that the original minerals of the ultrabasics had Mo in the crystal lattices, while the secondary minerals were unable to accomodate Mo. The latter was released and fixed in the mineral molybdenite. It is worth noting that light coloured, conformable replacement veins in the amphibolite also contain molybdenite in a similar way.

As to the age of the ultrabasics we can state that they are definitely older than the granitization. The latter process was progressive, i. e. it was preceded by a period during which the rocks were brittle and a deformation by fracture was predominant. Displacement zones penetrate all the rocks of the complex, also the ultrabasics. (In later stages of the granitization the displacement zones recrystallized in such a way that they now appear as "granite dykes" which correspond closely to similar dykes in Finland (see EDELMAN, 1949, page 33)).

As the granitization progressed, the plasticity of the rocks increased, the maximum of plasticity being reached in the small scale folding described on page 69. This folding was succeeded by the granitic replacement processes in connection with which the ultrabasics were transformed into serpentine, and talc rocks.

The mylonite zones, in which the ultrabasic rocks are occasionally enclosed, are apparently older than the "granite dykes". It seems therefore to be most reasonable to assume that these zones were formed in the latest stages of the deformation which was responsible for the formation of the ultrabasic rocks. Thus, there may be a considerable space of time between the formation of the ultrabasic rocks and the granitization.

The field work is, however, not yet concluded. The solution of the problems concerning the age of the ultrabasic rocks cannot be given until we have a more clear picture of the "geometrical orientation" of the ultrabasics, than we have at present.

In conclusion we may state that the large ultrabasic masses of this area indicate that we have had a zone of strong deformation here (compare page 70), a view which is supported by the "shape" of the Godthaab fiord.

The mode of occurrence of the ultrabasic rocks in the Godthaab fiord supports, as far as the writer can see, the conclusions arrived at in Tovqussaq. Certainly, the supporters of BOWEN's hypothesis, which states that the peridotites of orogenic regions were intruded as crystal aggregates along zones of dislocation, may apply a similar view on the present area. The writer finds, on the contrary, that the Godthaab fiord occurrences may be interpreted as a link between the Tovqussaq rocks and the great peridotite masses e.g. of the alpine regions. There seems to be a logical evolution from the small scale features at Tovqussaq to the large peridotitic masses of the west coast of Bjørneø. This evolution is, however, the subject of the following chapter.

ULTRABASIC ROCKS IN OROGENIC ZONES

Ultrabasic rocks of peridotitic affinity belong to the most conspicuous features of orogenic zones of all ages. In this chapter these rocks will be dealt with in brief[1]).

At first a general outline of the characteristics of the rocks in question will be given in the following 11 paragraphs:—

1) The peridotitic rocks are limited to orogenic zones and occur in the folding chains of all ages all over the world.

2) They occur as steeply inclined sheets or as lenticular bodies, in both cases typically enclosed conformably in the surrounding more or less deformed rocks. Several lenses may be found in one and the same strike horizon. Transgressive borders towards the enclosing rocks occasionally occur.

3) The size of the individual bodies varies from a few meters to several kilometers along the strike.

4) The rocks are most often associated with geosynclinal sediments and volcanics.

5) In rare cases only, association with other intrusive rocks can be demonstrated. The ultrabasics occur usually as independent masses.

6) The ultrabasics were formed during the first stages of orogenesis and are older than the granite batholiths of the mountain chains. The ultrabasic masses may be highly deformed during the later stages of deformation.

7) The rocks in question are often situated along zones of dislocation and strong deformation.

8) The rocks are characterized by simple mineralogical composition. Rocks of this type melt at extremely high temperatures (over 1500° C).

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[1]) We exclude from this discussion peridotites of layered complexes, and the so-called peridotite dykes. The peridotite dykes in Skye have been satisfactorily explained by Bowen (1928). Other occurrences as for instance the peridotite dykes of Eastern North America (Kemp and Ross (1907), Matson (1905), and Sosman (1938)) are of a more problematic nature.

9) The contact zones observed for instance around gabbro masses are never seen around the peridotites, the rocks enclosing the latter are metamorphosed only to a slight extent by the peridotite. The temperature of emplacement of the ultrabasics must for this reason have been within the ranges of metamorphism.

10) No lavas of peridotitic composition are known.

11) In many cases, the ultrabasic rocks are in the same metamorphic facies as the surroundings. The primary minerals may be partly or totally transformed into secondary during the retrograde metamorphism connected with the later stages of the deformation.

The formation of the rocks discussed provides one of the most puzzling problems of petrology.

In the first decades of this century the ultrabasics and their enclosing gneissic rocks were regarded as differentiates of one and the same magma, as for instance ESKOLA did in his classical paper on the eclogites of Norway (1921, page 19). The peridotites of the younger mountain chains were believed to have consolidated from melts of their own composition. These ideas are inconsistent with the high melting temperatures of the peridotites and the slight metamorphism around masses of these rocks. For the same reasons VOGT's view (1924) that ultrabasic melts were formed by settling and remelting of the minerals first crystallized from a basaltic magma, was abandoned a long time ago. Moreover, VOGT's view is not in agreement with BOWEN's reaction-principle.

H. H. HESS (1938) derived peridotites by differential fusion of a peridotitic substratum. The intrusion is believed to take place during the first stages of folding, when the geosynclinal sediments are buckled downwards. The intrusion should be facilitated by a large content of water in the original melt. First of all it seems very unlikely that peridotites could be transported from a presumed peridotitic substratum to their present level. Besides, the assumption is not in agreement with recent experimental data.

BOWEN and TUTTLE (1949) have demonstrated that a water rich peridotitic magma can exist only at high temperatures (above 1000° C). They therefore state (op. cit. page 455) that "the possibility of the formation of dunites, serpentinites, and peridotites from such supposed magma intruded at low temperature is definitely excluded".

BOWEN as early as in 1915 suggested that dunite was never liquid as such but was formed by accumulation of early olivine crystals from a complex magma. The mass of olivine crystals might be intruded into other rocks as a solid or substantially solid mass. BOWEN and TUTTLE support this hypothesis in the paper quoted above. They believe that dunitic and related material can be intruded in a completely crystalline state under certain conditions of crustal deformation. That peridotites are found in zones of dislocation and strong deformation is taken by BOWEN to be in favour of the hypothesis. The "cold intrusion" is believed to be facilitated by interstitial liquid, translation lamellae in olivine and pyroxene, and maybe by beginning serpentinization. This mechanism may be particularly effective in producing pervasive serpentinization. "A still-standing mass of dunite or peridotite might not be as readily serpentinized through and through" (op. cit. page 456).

It has been argued that the common feature of lack of large amounts of basic rocks formed at the same time as the ultrabasics and complementary to the latter contradicts BOWEN's view. TURNER and VERHOOGEN (1951, page 247) have overcome this difficulty by stating that the spilitic geosynclinal basalts and the peridotites represent complementary differentiates from the same magma, the time interval between the eruption and intrusion of the two rock types might reflect the difference in mobility between the two "magmas".

BOWEN's hypothesis is probably the one at present favoured by most petrologists. It explains the lack of a contact zone around the peridotites and this mechanism of intrusion may be effective in the formation of peridotites, but, on the other hand, some occurrences of ultrabasics seem not to be in agreement with BOWEN's view.

If the peridotites were intruded as crystalline masses, the large massive and homogeneous bodies of ultrabasics without traces of stress and cataclasis should be the exception rather than the rule. The central part of the masses are very often undeformed, while the marginal parts may be highly deformed. This deformation is in most cases of a late date. The objection, that the primary cataclasis might be healed by the original interstitial liquid, or by solutions or vapour seaving through the cracks, can be rejected. The minerals of the peridotites (olivine and pyroxene) cannot crystallize at the low temperatures prevailing in the level of the earth crust in question.

The connection between the peridotites and shear zones does not necessarily imply that the peridotites intruded along the shear zones. The opposite relationship is also possible as the rigid peridotite masses certainly have a directing influence during the deformation.

BOWEN and TUTTLE (1949, page 459) have further advocated the view that dunite may be formed along cracks in pyroxenite and pyroxenite along cracks in dunite because of hydrothermal or pneumatolytic activities. BARTH (1950, page 627) has recently explained some bahiaitic rocks from Southern Norway as formed in this way. Another example are the "veins" of bronzitite in dunite at Siorarsuit, West Greenland (SØRENSEN, 1951b, page 65). This principle seems only to be of importance in restricted zones and cannot explain the formation of larger masses of rocks, the minerals of which indicate a higher temperature of formation than the temperature prevailing in the surroundings, although it must be admitted that the solutions or vapours in operation may cause a local heating of the rocks along the cracks.

J. AVIAS (1949) has explained the peridotites of New Caledonia as the products of metasomatic processes. We shall return later to this very interesting paper.

It should be mentioned at this point that large stratified lopolithes as Bushveld and Sudbury have recently been interpreted as being formed in a metasomatic way.

Thus, it appears from the preceding brief discussion of the peridotites that the orthodox magmatic explanation is not very satisfactory and that there is a tendency to interpret the rocks in question as the products of hydrothermal or metasomatic activities[1]. In the following a contribution to the discussion of the peridotite problem will be given.

As stated on page 72 there is a logical "evolution" from the thin ultrabasic bands of Tovqussaq to the large peridotitic masses in the Godthaab fiord area. In both cases the ultrabasics may be formed in zones of tension in rigid layers. The same may be the case with the huge bodies of peridotites of the orogenic zones of later date.

In support of this idea should be mentioned the difference in folding style between the lower and upper parts of the orogenic zones.

In the lower part (mainly WEGMANN's *Unterbau*) as for instance the pre-cambrian of West Greenland the deformation is more or less plastic. The comparatively high temperatures are normally insufficient to cause fusion at the prevailing pressure. The temperature gradients are generally rather smooth. The heat generated by the metamorphic processes seems not to give rise to pronounced differences in temperature at neighbouring places, as indicated for instance by the gradual transition from "country rock" to granite formed at the expense of the first named. Exceptions to this may be found where there is a difference in potentials e. g. where we have the postulated tension zones.

In the upper parts of the orogenic zones (WEGMANN's *Oberbau*), where the rocks usually are more brittle, the deformation is more cataclastic. Overfoldings and overthrusts are a prominant feature contrary to the "smooth" folds of the Unterbau. The unconsolidated sediments are folded rather easily, while the more competent layers may be folded

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[1]) PERRIN and ROUBAULT (e. g. 1949, page 375) have recently suggested that all granular rocks were formed by metasomatic processes. BARTH (1952 pp. 184 and 226) favours also a metasomatic explanation.

by bending or shear. The temperature is of a low order, but frictional heat plays a great rôle. The temperature gradients are therefore steeper in the upper parts of the orogens than in the lower, and as the pressure is low, granitic melts may be formed under favourable conditions. Heat may be accumulated locally, partly as a result of friction, partly introduced with the granitizing "solutions" (cf. P. Misch, 1949, page 242). In addition local discontinuities may be of importance, preventing migration of material and equalization of differences in temperature and pressure. Therefore, granites formed by granitization may have sharp contacts in this level of the crust as demonstrated by Misch (op. cit. page 379).

If the genetical relationship between tension zones and ultrabasics in pre-cambrian areas be accepted, it is a short step to apply the same view on the upper parts of the orogenic zones. The local accumulation of heat seems to favour this idea.

As pointed out in several papers, for instance Bowen and Tuttle (1949, page 458) the peridotites and serpentinites are situated along zones of intense deformation. This has been demonstrated by Hess (1948) for island arcs in the Pacific i. e. in the North Pacific. Similarly the serpentine belt of Australia is associated with a great overthrust. As each movement along the fault- or thrust-zones is preceded by a more or less prolonged period of elastic deformation and tension, it seems to the writer that conditions favouring the formation of ultrabasic rocks may arise locally.

Hess (op. cit. page 432) states for ultrabasic rocks in the East Asiatic mountain belts, that they occur in two belts about 50 miles on either side of the original location of the tectogene axis and less commonly in the area between the two belts. "They are intruded during the first great deformation of the belt, presumably during buckling of the crust, and later deformations of the same belt are not accompanied by intrusions of peridotites. Thus location of the peridotite belt and dating its intrusion locate the old tectogene axis and date the initiation of the deformation of that zone". By this Hess has located and dated three different zones of deformation along the west coast of the Pacific.

Comparing these features with for instance Tovqussaq we observe that the formation of the ultrabasics in both areas took place during the first stages of deformation. The ultrabasics of Tovqussaq were formed in tension zones or potential zones of deformation in rigid bands of amphibolite. The pacific rocks were formed in an immense deformation zone between two huge rigid masses, namely the continent and the bottom of the Pacific. As suggested by Hess (op. cit.) the peridotites are associated with island arcs, which in turn are probably associated with the thrust planes between Asia and the Pacific.

The writer therefore tentatively suggests that the Pacific peridotites were formed in the first "tension-stage" of deformation in shear zones of continental dimensions by processes similar to those believed to be responsible for the formation of the ultrabasic rocks at Tovqussaq. The large size of the first named may be due to the large scale deformation taking place along the shear zones.

A similar view may be taken for peridotites at other places on the earth.

The ideas outlined above concerning the peridotite formation seems to the writer to be in close agreement with the 11 paragraphs at the beginning of this chapter. By this it should not be understood that the writer does not believe in magmatic peridotites. On the contrary it may be possible that some peridotites are separated from their magmatic associates during the deformation because of a difference in mobility.

It should be emphasized that although the proper conditions for ultrabasite formation were obtained only in amphibolites in the pre-cambrian of West Greenland, ultrabasic rocks may be formed in other rocks in the higher levels of the crust.

A few additional examples from the literature may illustrate the view expressed above.

E. MIKKOLA and TH. G. SAHAMA (1936, page 357) have described ultrabasic rocks closely resembling the West Greenlandish rocks from the region SW of the "granulite series" in Finnish Lapland. This region forms, from a tectonical point of view, a transition from the granulite to the tectonically quite different rocks further SW. In this region, too, a connection between amphibolite and ultrabasics seems present.

TUOMINEN and MIKKOLA (1950) in their re-interpretation of the classical Orijärvi area in Southern Finland associate the Mg-Fe concentration of the anthophyllite rocks with the folding of the area and assume that the thick competent beds glided along the relatively thin clayey beds, which were thus subjected to strong penetrative movements and to flowage towards the hinges of the folds. The rock gradually recrystallized and simultaneously the constituents in excess emigrated, thus causing an enrichment in Mg and Fe within the residue. They state (op. cit. page 91) that similar rocks will be formed, if instead of clayey rocks, say, feldspar rocks are subjected to a corresponding metamorphosis. Thus, the Orijärvi region is another example on the tectonical origin of the ultrabasic rocks. It should be mentioned that WEG-

MANN already in 1931 suggested a tectonic origin of the anthophyllite rocks of Orijärvi.

That ultrabasic rocks may be formed at the expense of calc-silicate rocks is for instance seen at Siorarssuit, West Greenland (SØRENSEN, 1951b, page 65). The dunite at this locality encloses and replaces lumps of a diopside-hornblende-rock. Corroded grains of the latter are scattered all over the dunitic rock.

Another example is found in rocks of caledonian age in Sunmøre, Western Norway (T. GJELSVIK, 1951). Along the coast of this area dunitic bodies are associated with zones of lime-silicate-gneisses and crystalline limestones. In the interior part of the area the dunites seem to follow a sparagmitic zone.

A. LACROIX in his classical papers on the granite contacts in the Pyrenees described the association limestone, ultrabasic rocks. The granites have penetrated and metamorphosed sediments of silurian, devonian and permo-carboniferous age. The granites enclose more or less metamorphosed sediments, which are still found in their original orientation. The inclusions of limestone are surrounded, not by the normal granite but by hornblende-granite or diorite. Where the limestone inclusions are broken into fragments during the deformation, more basic rocks are found, i. e. norites, hornblendites, and hornblende-peridotites. The acid and ultrabasic rocks are connected by a gradual transition, which occasionally takes place within few meters, but normally within "plusieurs centaines de metres" (1898, page 60). The granite and the diorite have gneissic structures. The ultrabasic rocks are found in small masses as inclusions in these gneisses.

LACROIX believes that the ultrabasic rocks were formed at places where the limestone bands in the granite were broken into numerous small fragments so that the granite at one time could assimilate a great amount of Ca- and Mg-bearing material. Although the tectonic conditions of the area is not described, a "tectonic formation" of the ultrabasics seems to be applicable, when the occurrences in the Pyrenees are compared with the rocks described in this paper. In that case the formation of the ultrabasic rocks in question may be connected with potential cross- or tension-joints. The bahiaites, described by BARTH (1950) occuring in cross-joints in norite, might be explained in a similar way.

J. AVIAS (1949) in a very interesting paper discussed the formation of the peridotites and serpentinites of New Caledonia. The peridotites here are surrounded by a serpentinite zone. The "contact metamorphism" in the country rock is slight, the transition between serpentinite and country rock being gradual. Volcanites comprise 80 % of the country rock. Inclusions of country rock in the serpentinite have preserved their orientation. Country rock structures may be found in the serpentinite, indicating that the latter is a sort of pseudomorph after the former.

The peridotite masses are heterogeneous, occasionally containing even granitic rocks. The layering in the peridotite corresponds with the original layering of the sediments.

AVIAS believes that the peridotite and serpentinite were formed by metasomatic processes. The serpentinite is derived from the original volcanic rocks by metamorphic serpentinisation at low temperature (below 500° C). If the primary rock was Al-rich, chlorite-rocks were formed. At higher temperature (and pressure) olivine and ortho-rhombic pyroxene were formed by progressive metamorphism at the expense of serpentine. In case of Al-rich rocks, gabbro, or even granite were formed. Transgressive borders may be the result of changes in volume accompanying the various transformations or difference in plasticity under the folding.

The volcanic rocks of the mountain belts are mainly of andesitic composition and contain usually a large amount of glass. These rocks are unstable and react very easily when exposed to metamorphism.

AVIAS's interpretation of the peridotites of New Caledonia is, as far as the writer knows, the only "transformistic" study of alpine peridotites published so far.[1])

The peridotites of Blashke Island and related areas in South-Eastern Alaska, described by M. S. WATSON jr. (1951) may be interpreted in analogy with AVIAS' view.

The occurrence in Blashke Island has a central dunitic core, surrounded by an olivine-pyroxenite ring, which in turn is surrounded by a zone comprising, adjacent to the pyroxenite, olivine gabbro, and adjacent to the country rock, hornblende gabbro. The country rocks consisting of various sedimentary and volcanic rocks are transformed into hornfels around the gabbro, the metamorphism decreasing away from the latter. The borders between the rocks of the complex and between the complex and the surroundings are irregular, steep, and transgressive. The complex is situated in the Kashevaraf anticlinorium. The minerals of the complex show a change in composition from the central part outwards from Mg-rich, Fe-poor to Fe-rich, Mg-poor.

WATSON explains the conditions by assuming the intrusion of an

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[1]) Other papers regarding ultrabasic rocks of orogenic zones as products of metamorphic processes are: R. W. v. BEMMELEN, On the origin of igneous rocks in Indonesia. Geologie en Mijnbouw, Juli, 1950, p. 218; M. ROUBAULT, Sur la nature metamorphique des serpentines de la Kabylie de Collo (Algérie). C. R. Acad. Sciences. t. 232, p. 2032—33, 1951, and R. PERRIN and M. ROUBAULT, Les idées nouvelles en pétrographie et l'étude du métamorphisme alpin. Schw. Min. Petr. Mitt. Bd. 31, 1951 p. 575.

ultrabasic magma. A temperature-composition gradient between magma and surroundings should be responsible for the rock sequence. The gabbro should be a result of contact metamorphism.

The ultrabasic masses of the area are situated near the roof-zones of granite batholithes, and they are formed earlier than and above the granite. According to WATSON the ultrabasics and granites are common products of a process, not of a common magma.

The writer agrees with WATSON in the last mentioned statement, but does not believe in the existence of the supposed ultrabasic magma in the level of the earth crust in question. An interpretation in analogy with AVIAS' treatment of the rocks from New Caledonia seems much more promising.

PROUD and OSBORNE (1952) discuss some ultrabasic rocks from Woodsreef, New South Wales.

An ultrabasic body consisting of harzburgite and serpentinite is situated along a fault zone. The areas on either side of the fault are characterized by different styles of deformation. The fault zone actually forms the western border of the ultrabasic body. PROUD and OSBORNE (op. cit. page 23) state that harzburgite was injected into a master fault of gravity origin opened up along the line now occupied by the western margin of the serpentine belt. Conditions were those of tensional control with periods of hydrostatic character. A second dunitic phase was structurally controlled by horizontal shearing stress. Although the writers of the quoted paper regard the ultrabasics as magmatic rocks, a metamorphic interpretation seems to be at least just as satisfactory. As to the serpentine fibers PROUD and OSBORNE believe that they were formed in the appropriate stress environment, involving torsion and concomitant tension.

More examples could be quoted, but unfortunately most descriptions pay little attention to the tectonic environment of the ultrabasics. A discussion of peridotites described as independent intrusions must therefore be very subjective and of limited value.

In conclusion the writer wishes to draw the attention to the possible "tectonic origin" of the ultrabasics of orogenic zones and hopes that petrologists working in other areas will test in the field the working hypothesis expressed in this paper.

ON THE FORMATION OF ECLOGITES

As mentioned on page 39 BACKLUND in an important paper (1936) has given a metamorphic, kinetic explanation of the formation of eclogites. He has in the caledonian rocks of Northern Sweden and East Greenland demonstrated the transformation of geosynclinal basalt into eclogite, a transformation which goes over amphibolites and garnet amphibolites.

He assumes that eclogite has been formed at high dynamic pressure (stress) and high temperature. The appearance of the high density minerals of the eclogite does not depend on depth of formation but on the dynamic pressure which in effect may exceed the hydrostatic.

EBERTH (1936, page 74) showed that the eclogite formation took place in limited zones in the amphibolite.

Thus, the eclogite is in a sense a "tectonite" (BACKLUND: op. cit. page 60).

BACKLUND's view may probably be extended to most occurrences of eclogites. At this point the eclogites of France (Y. BRIERE, 1920) and Tirol (L. HEZNER, 1903) should be mentioned. Both authors emphasize the association amphibolite-eclogite, the eclogites often occuring as bands or lenses in amphibolite or garnet amphibolite. In the caledonian rocks of Sunmøre, Norway (P. ESKOLA, 1921, and T. GJELSVIK, 1951) eclogites are present as conformable lenses in migmatites. T. GJELSVIK (op. cit. page 23) states that the eclogites are connected on the one side with garnet-bearing olivine gabbros, on the other side with garnet-rich lime-silicate gneisses. Gradational types between the different rocks are found. In addition rocks of eclogitic affinity (Cr-diopside instead of omphasite) occur as schlieren in dunite (ESKOLA, op. cit. page 51). As suggested by BACKLUND (op. cit. page 58) the eclogites of this area may also be explained in analogy with the above.

The explanation of the formation of ultrabasic rocks of peridotitic affinity put forward in the present paper is in close agreement with BACKLUND's explanation of the formation of eclogites. But, while BACKLUND believes high dynamic pressure to be most important, the present writer would prefer to formulate the explanation in such a way that a difference in pressure (tension or stress) starts the process while the higher temperature in the tension zones determines the appearance of the rocks formed in these zones. The texture and parageneses of the rocks in question support the view that static conditions prevailed in the zones mentioned.

The paragenesis garnet-pyroxene may have come into existence in two ways. Firstly, as mentioned on page 59 by regressive metamorphism of an original hypersthene-carrying rock. Secondly, eclogites may be formed by progressive metamorphism as demonstrated by BACKLUND.

The paragenesis garnet-pyroxene indicates P,T-conditions corresponding to the upper part of the amphibolite-facies, the presence of hypersthene shows P,T-conditions corresponding to granulite-facies.

We may therefore establish two new high-pressure- or stress-sub-facies, i. e. eclogite sub-facies in amphibolite facies and peridotite sub-facies in granulite-facies. When olivine is an important component of the ultrabasics, the latter are formed at temperatures exceeding those generally prevailing in granulite facies. We are here dealing with the highest temperatures reached in metamorphism (with the possible exception of certain types of mylonites).

Since the association garnet-pyroxene is not stable in granulite facies, eclogites are only found in rocks belonging to this facies as products of retrograde processes. On the other hand, peridotitic rocks may be formed in tension zones in all levels of the crust in which tension is a phenomena of importance.

KORJINSKY seems to have advanced a similar unorthodox view as early as in 1937.

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Appendix.

The volume of the "Medd. Dansk Geol. Forening" which is in print now (Bd. 12. pt. 2) contains two papers of interest for the present study, namely, A. NOE-NYGAARD and A. BERTHELSEN: Structural geology of a high-metamorphic gneiss complex, with a general discussion on related problems (compare p. 76 of the present paper), and H. SØRENSEN: Further studies on ultrabasic rocks in Sukkertoppen District, West Greenland, which gives a detailed description of the occurrence Ol.2 (see p. 63).

Færdig fra trykkeriet den 14. marts 1953.

PLATES

Plate 1.

Fig. 1. Hypersthene amphibolite, Ekaluq. No. 13415. 30 ×, 1 nic. Hypersthene porphyroblast with hornblende inclusions. Top left: diopside enclosed in hornblende. The white grains are plagioclase. (CHR. HALKIER phot.).

Fig. 2. Hypersthene amphibolite, the east coast of Langø. No. 13431b, 30 ×, 1 nic. In the middle: diopside, hornblende, and plagioclase. To the right: hypersthene with hornblende inclusions. (CHR. HALKIER phot.).

Fig. 1.

Fig. 2.

Plate 2.

Fig. 3. Ultrabasic rock. No. 13431 a. The east coast of Langø. 30 ×, 1 nic. The rock is composed of hypersthene (predominant), hornblende, diopside (two grains close to the centre between the two basal cleavages of hornblende), olivine (to the right), and spinel + magnetite. Note the dark cleavages and the finely distributed pigmentation in the hypersthene and in the hornblende. (Chr. Halkier phot.).

Fig. 4. Ultrabasic rock. No. 4033. The west coast of Langø. 40 ×, 1 nic. The photograph shows that the spinel grains (black) have approximately the same parallel orientations in three adjoining grains of hypersthene. (B. Mauritz phot.).

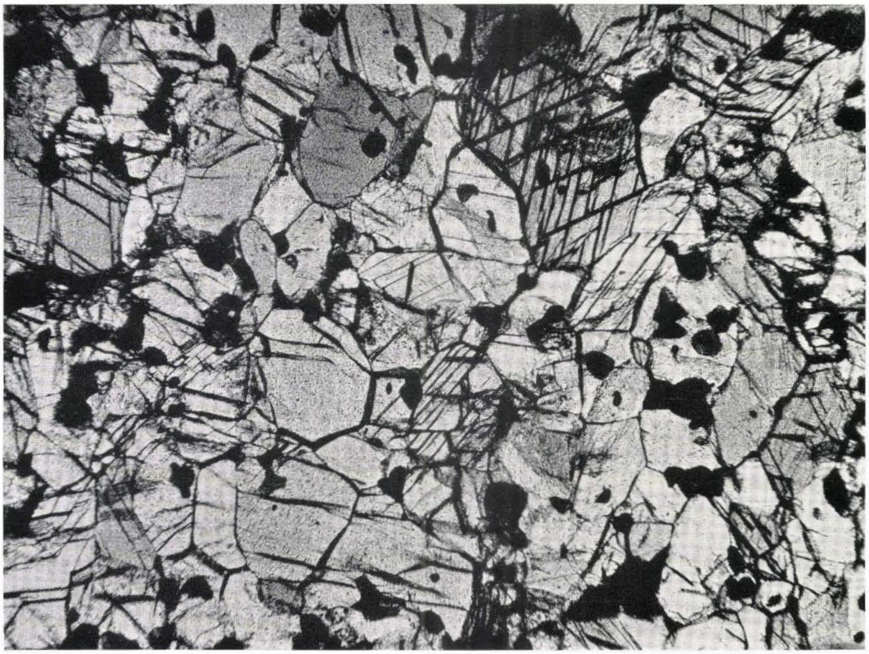

Fig. 3.

Fig. 4.

Plate 3.

Fig. 5. Ultrabasic rock. No. 13437 a. The west coast of Langø. 30 ×, 1 nic. Hypersthene grain surrounded by hornblende network with concentrations of spinel. (CHR. HALKIER phot.).

Fig. 6. Ultrabasic rock. No. 13437 c. The west coast of Langø. 30 ×, 1 nic. The rock consists of hypersthene (partly penetrated by hornblende), hornblende, and plagioclase. (CHR. HALKIER phot.).

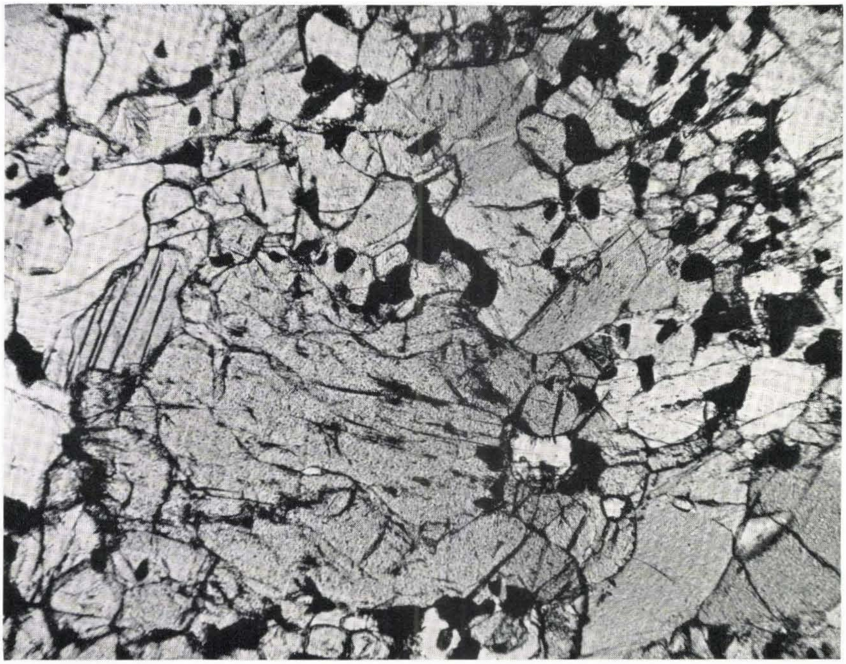

Fig. 5.

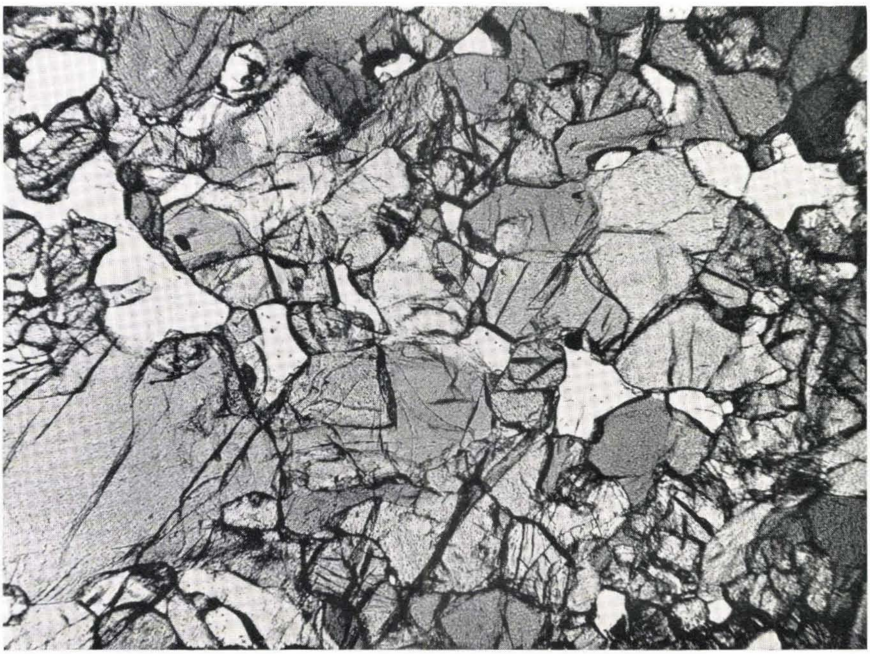

Fig. 6.

Plate 4.

Fig. 7. Partly dissolved ultrabasic inclusion in pegmatite. The west coast of Langø. No. 13436. 30 ×, 1 nic. The large hypersthene grain is partly replaced by plagioclase and hornblende. Note the areas with amphibole cleavage in the hypersthene. (Chr. Halkier phot.).

Fig. 8. Pegmatite. No. 13434. West coast of Langø. 30 ×, 1 nic. The photograph shows corroded grains of hypersthene enclosed in plagioclase. Note the lines of pigmentation in the plagioclase. (Chr. Halkier phot.).

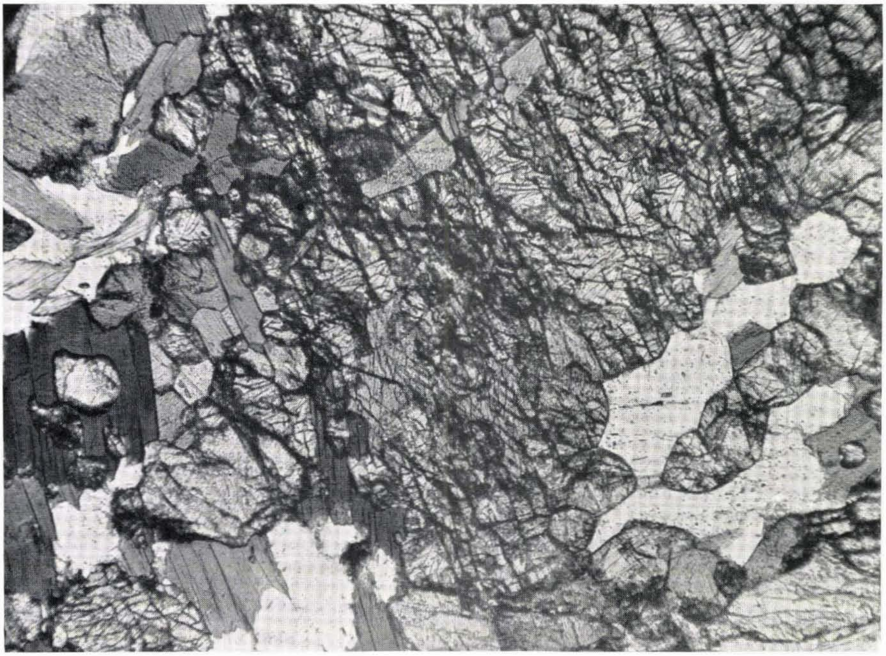

Fig. 7.

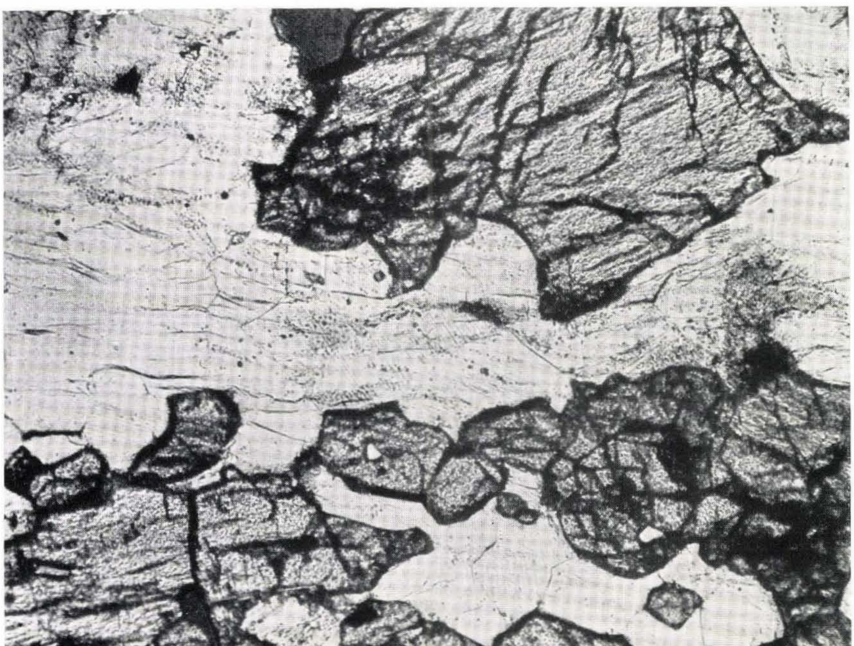

Fig. 8.

Plate 5.

Fig. 9. Border between ultrabasite and pegmatite. No. 13434. The west coast of Langø. 30 ×, 1 nic. The large hypersthene grain has inclusions of biotite towards the pegmatite (top left), hornblende towards the ultrabasite (center). (CHR. HALKIER, phot.).

Fig. 10. Ultrabasic rock. No. 13299. Ol. 5. 30 ×, 1 nic. Large hypersthene grain with inclusions of hornblende and hornblende along the cleavages. The groundmass consists of hornblende, hypersthene, spinel, and diopside. (CHR. HALKIER phot.).

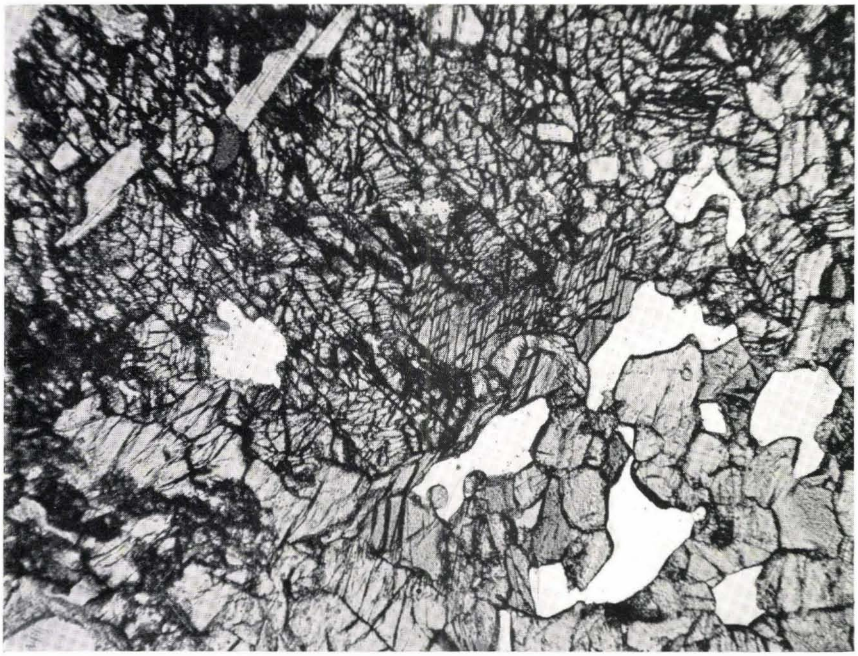

Fig. 9.

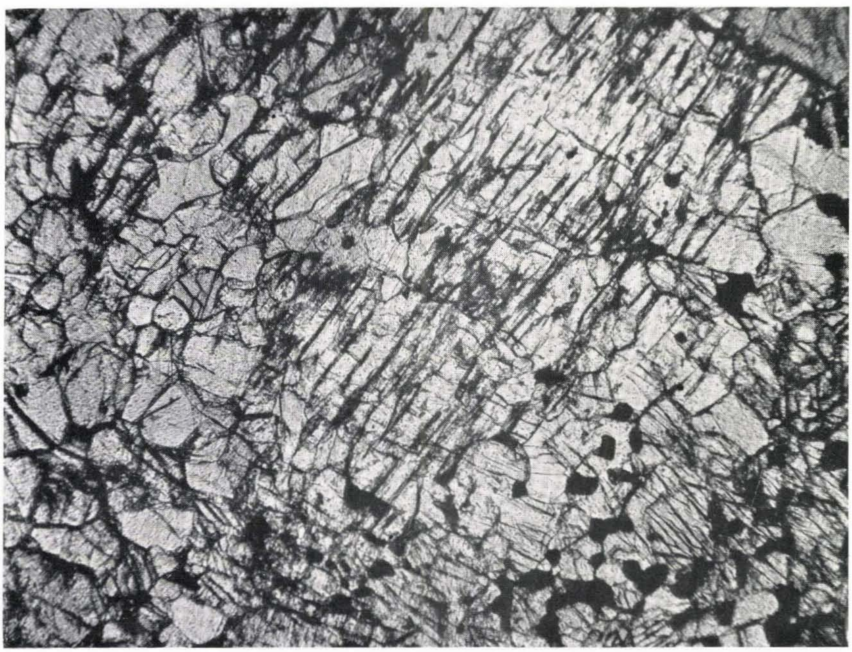

Fig. 10.

Plate 6.

Fig. 11. Ultrabasic rock. No. 13517. Ol. 3. 30 ×, 1 nic. Two hypersthene grains (to the right and to the left) separated from one another by hornblende. Olivine is seen top right. The black grains are spinel. (Chr. Halkier phot.).

Fig. 12. Ultrabasic rock. No. 13485. Amitsuarssoralak. 30 ×, 1 nic. Olivine to the right, bronzite to the left. (Chr. Halkier phot.).

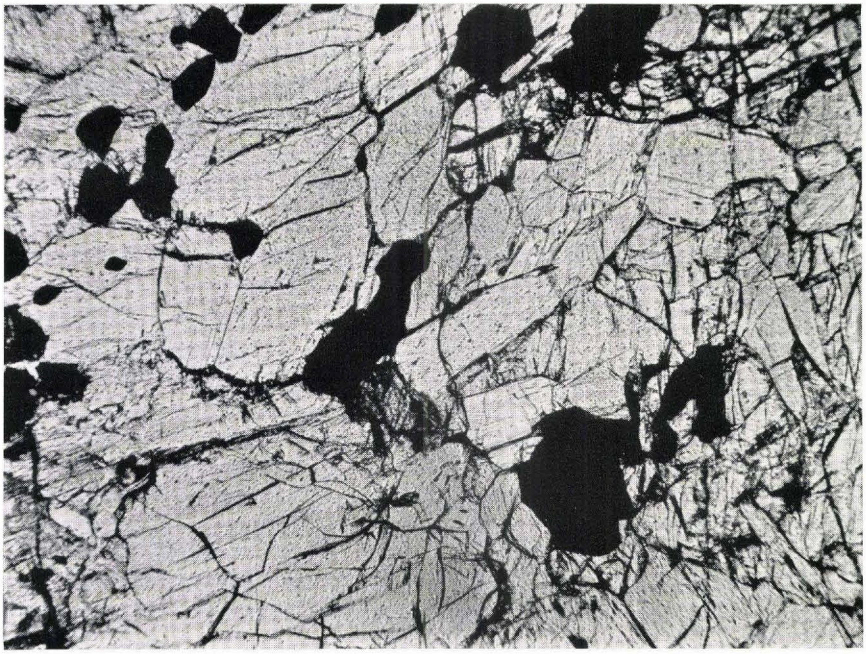

Fig. 11.

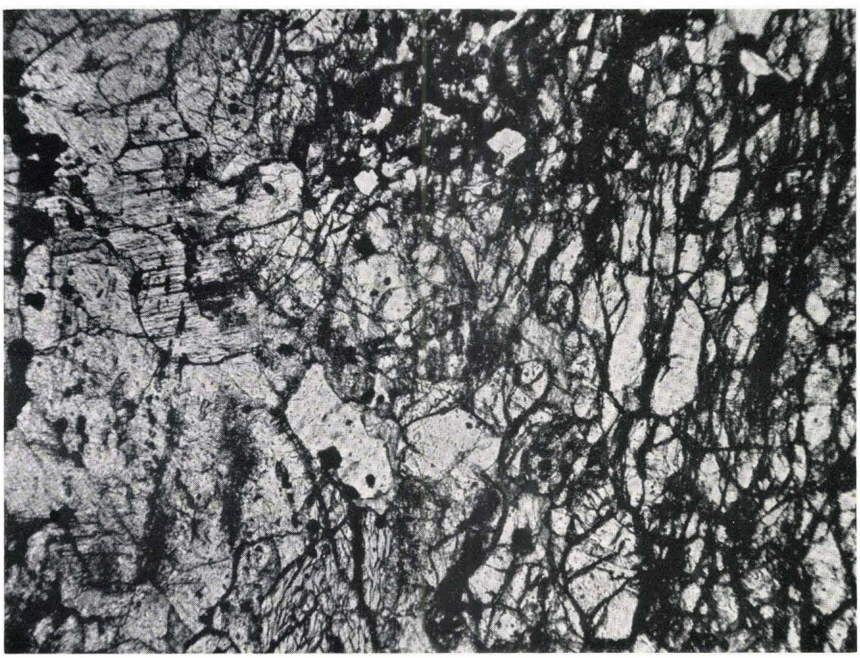

Fig. 12.

Plate 7.

Fig. 13. Ultrabasic rock. No. 13488. Amitsuarssoralak, 30 ×, 1 nic. Bronzite porphyroblast with poikilitic inclusions of hornblende. Note the crystallographic outlines of many of the hornblende grains. (Chr. Halkier phot.).

Fig. 14. Hybrid rock. No. 13535. Ol. 1. 30 ×, 1 nic. The photograph shows a hypersthene porphyroblast partly replaced by plagioclase. Hornblende is seen to the right, below the centre, and to the left. (Chr. Halkier phot.).

Fig. 13.

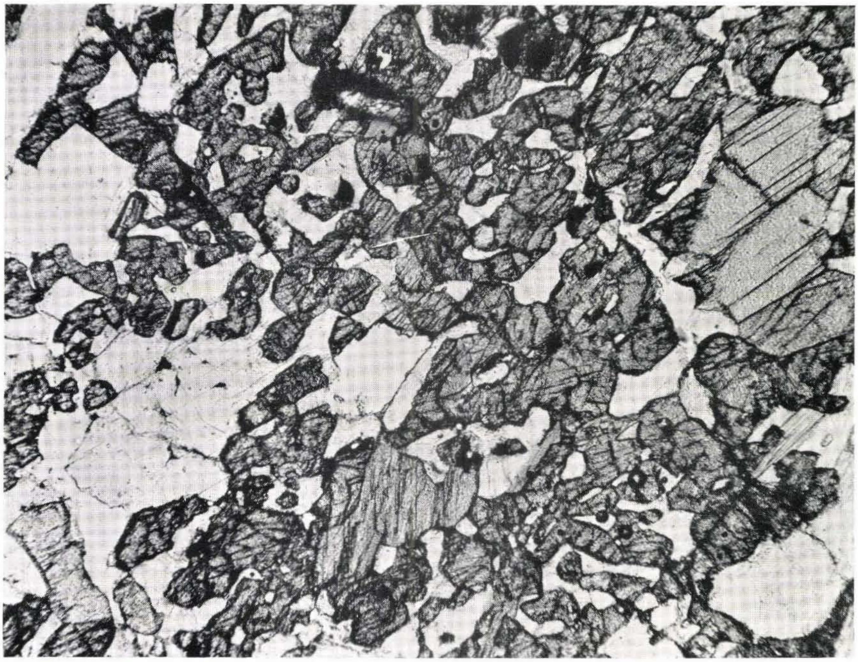

Fig. 14.

Plate 8.

Fig. 15. Central part of ultrabasic inclusion in gneiss. No. 13508. Ol. 2. 30 ×, 1 nic. Replacement vein (composed of plagioclase and hornblende) between the original hypersthene porphyroblasts (seen to the right, top and bottom). (CHR. HALKIER phot.).

Fig. 16. Outer part of the same inclusion as fig. 15. No. 13507. Ol. 2. 30 ×, 1 nic. Hypersthene porphyroblast partly replaced by plagioclase and biotite. To the left: hybrid rock composed of plagioclase, hornblende, hypersthene, and biotite. (CHR. HALKIER phot.).

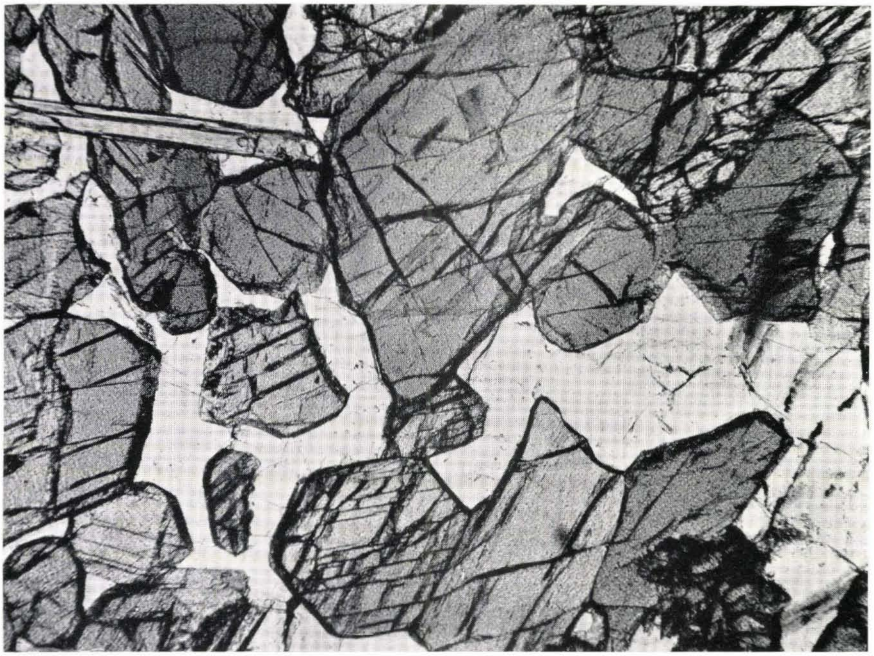

Fig. 15.

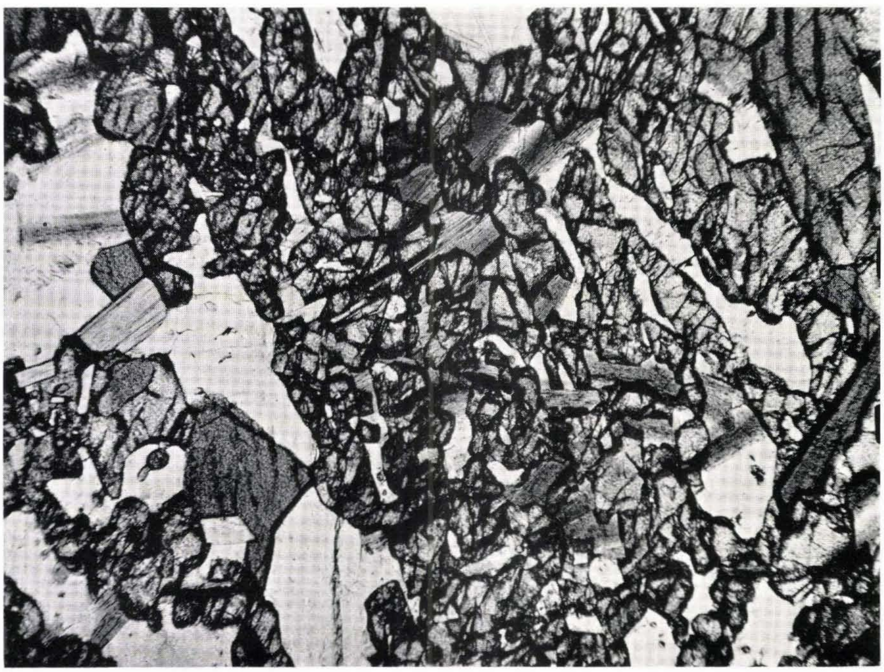

Fig. 16.

Plate 9.

Fig. 17. Mica zone in ultrabasic rock. No. 13542. Ol. 1. 30 ×, 1 nic. Centre: Olivine, enstatite, and chromite. To the right and to the left: flakes of mica with inclusions of olivine, enstatite, and chromite. (CHR. HALKIER phot.).

Fig. 18. Ultrabasic rock in mylonite. No. 13957. South-east coast of Sermitsiaq, Godthaab fiord. 40 ×, 1 nic. Grains of olivine in a matrix of hornblende, anthophyllite, and serpentine. (B. MAURITZ phot.).

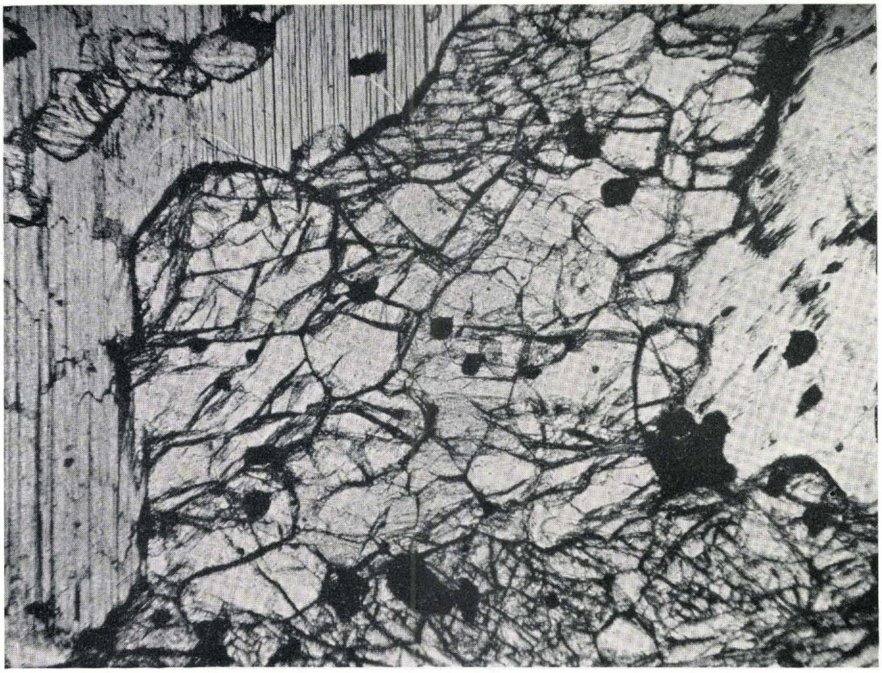

Fig. 17.

Fig. 18.