GRØNLANDS GEOLOGISKE UNDERSØGELSE
BULLETIN No. 25.

STRUCTURAL STUDIES IN THE PRE-CAMBRIAN OF WESTERN GREENLAND

II
GEOLOGY OF TOVQUSSAP NUNÂ

BY

ASGER BERTHELSEN

WITH 84 FIGURES IN THE TEXT
AND 4 PLATES

С РУССКИМ РЕЗЮМЕ́

Reprinted from
Meddelelser om Grønland Bd. 123, Nr. 1

KØBENHAVN
BIANCO LUNOS BOGTRYKKERI A/S
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Denne afhandling er af det matematisk-naturvidenskabelige fakultet ved Københavns universitet antaget til offentlig at forsvares for den filosofiske doktorgrad.

København, den 9. september 1960.

H. H. Ussing,
h. a. dec.

CONTENTS

Abstract

This work deals with the structural analysis of a small area, Tovqussap nunâ, in the southern part of the Sukkertoppen district, West Greenland (fig. 1). This unusually well-exposed area is built up of pre-Cambrian granulite and amphibolite facies rocks. After an introduction, in which an outline of the mapping method is given and the petrographic nomenclature discussed, about a hundred selected rock specimens are described. This is followed by a detailed account of the geological map (Plates 1 and 2). A special rock division, the diorites, is given a more thorough treatment on account of their considerable chronological significance.

The structural analysis begins with an elucidation of the geometric relations in eastern Tovqussap nunâ. With the help of Wulff net stereograms, structural contour maps and profiles, an analysis is carried out on this area's complicated structures. This analysis has revealed three superimposed fold phases. The experience thus obtained was used in the analysis of central and western Tovqussap nunâ. Through the results of the analysis it is reckoned that the high-metamorphic Tovqussaq rocks belong to an approximately thousand metres thick supracrustal series comprising five major concordant stratigraphic units (Pl. 4). This series was first folded by flat-lying isoclinal folds with roughly NNW axes (the Midterhøj phase, see table 1 and fig. 78). This folding is assumed from analogies to have taken place during an early stage of the metamorphic development (slate/schist facies). The structures belonging to this first phase were later refolded on approximately ENE axes in large recumbent folds with amplitudes which can exceed 10 km. This fold phase (the Smalledal phase) was probably accompanied by amphibolite facies metamorphism.

Refolding of the large western nappe during the last stage of the Smalledal phase is thought to have created the conditions for a later diapirism. This diapirism expressed itself by a transport of plastic material towards the top of the dome within the single rock units (the gneisses). The diapirism seemingly occurred during a change in the tectonic movement pattern leading to renewed refolding, this time on SE to S plunging axes under granulite facies conditions. The structures which were developed during this renewed refolding (the Pâkitsoq phase) show a distinct dependence on the form of the earlier structures. At the end of the Pâkitsoq phase a sub-phase (the Langø sub-phase) can be distinguished, during which steep foliation planes were superimposed on the earlier formed folds. The formation of the latekinematic diorites can be shown to have taken place in the antiformal hinge zones of pyribolite layers where movement along steep foliation planes has facilitated the ascent of salic material migrating from the underlying gneiss of the fold cores. The postkinematic diorite dykes and aplites, which also are regarded as having been formed metasomatically, represent an intermediate period during which the Tovqussaq district was subjected to tensional stresses. The originally planar aplites make it possible to distinguish yet another movement phase, which locally has left a strong imprint

on the detailed structural picture, though it has only brought about small changes in the form of the major structures. This posthumous phase took place under amphibolite facies conditions, and led to a retrograde metamorphism of the older granulite facies rocks. Where the hypersthene gneisses were only subjected to weak penetrative movements they were made over to hornblende-biotite-bearing 'purple gneisses', whereas where penetrative movements were more intense, they recrystallised into light coloured biotite gneisses. The granulites on Langø and Tugdlerúnarssuit were also formed during the posthumous phase, on account of a local higher Mg/Fe ratio which permitted the preservation (or formation anew) of these rocks' present granulite facies mineral assemblage in spite of PT conditions which otherwise would have led to the formation of typical amphibolite facies minerals (as in the light coloured biotite gneisses).

This retrograde metamorphism resulted also in local granitisation due to a redistribution of the mobile granite material under the prevailing tectonic conditions. Towards the close of the posthumous phase para- to postcrystalline mylonites and faults were developed. These are believed to represent the last traces of orogenic phenomena in the area. Later followed the intrusion of two groups of basic dykes, separated in time by a regional system of NE-trending wrench faults. This cratogenic stage of the area's evolution has been dealt with separately in an earlier publication (Berthelsen and Bridgwater, 1960).

The structural evolution under consideration here is interpreted as belonging to a single geological cycle. The different fold phases are taken to represent successive structural events in which the rocks of the area have been involved during their passage through different tectonic levels under the mountain chain. This kinematic analysis enables one to follow the Tovqussaq rocks through an early sinking stage and a later rising stage. The deformation of the Tovqussaq district probably took place during the Ketilidian orogeny which was established by E. Wegmann in SW. Greenland.

In the closing chapter the progressive metamorphism is discussed first, after which the mineral facies position of the Tovqussaq rocks is dealt with. Furthermore the results of some trace element analyses are discussed. The author is of the opinion that the present content of trace elements in the rocks may only partly reflect their original distribution in the parent rocks. Even Zr, which is usually regarded as a very stable element during metamorphism, is thought to have undergone a real redistribution. A high content of Ni and Cr in certain calc-silicate rocks is considered to bear witness of a migration of these elements from the surrounding gneisses during skarn-metasomatism. With regard to the origin of the ultrabasic rocks, the author disputes H. Sørensen's theory that they were segregated from hypersthene amphibolites (H. Sørensen, 1953). The ultrabasic composition of these rocks is considered by the author to date from the geosynclinal stage, where they could have been either siliceous dolomites or, perhaps more likely, ultrabasic ophiolites of extrusive origin. The basic rocks (pyribolites and gabbro-anorthosites) could have formed from extrusives (or early formed sills) or marly and limy sediments. Their present relatively high content of Ni and Cr indicates however that the former is the more probable. As the basic rocks are quantitatively important the enclosing gneisses cannot have provided Ni and Cr in sufficient quantity to account for their present concentration in the pyribolites and gabbro-anorthosites without a migration of these elements over extraordinarily long distances. Since the gneisses and granulites in Tovqussaq and in the surrounding areas show a genetic connection to definite sedimentogenous rocks, it is considered that the quartz dioritic to granitic rocks are metamorphosed and metasomatised greywackes and schists.

In connection with this petrogenetic discussion the author has introduced some general considerations, amongst others concerning what role anatectic processes have played during granulite and amphibolite facies metamorphism. By a comparison of the structures in non-metamorphic sediments with those in high-metamorphic Greenland gneisses, it is concluded that the introduction of anatectic processes to account for the complicated gneiss structures is redundant. The discussion concludes with some remarks on the formation of dome structures in gneiss areas in the light of the results from the kinematic analysis of the Tovqussaq dome.

I. INTRODUCTION

The Structural Analysis

The aim of this work is to demonstrate what may be obtained by applying "the structural analysis" in the study of a well exposed region built up of highly metamorphosed rocks. It is the author's belief that an increasing number of carefully performed analyses of such regions may help to establish a collection of 'styles tectonique' which in its turn may render a more exact interpretation of less well exposed areas possible.

"The structural analysis", as developed by E. WEGMANN from the heritage of the Alpine school of tectonics and SEDERHOLM's classical studies, has become an almost independent branch of geology.

In a few words, "the structural analysis" may be defined as a s t u d y o f t h e g e o l o g i c a l m a t e r i a l v i s u a l i z e d i n t h e d i m e n s i o n s o f s p a c e a n d t i m e. Its foundation is empirical. It serves as a new line of attack on regions where neither the classical methods of stratigraphy nor the modern ideas of petrology make a thorough study possible. Its field of application thus covers the domain of metamorphic rocks. Its goal is to open up new ways of thinking in geological science.

Its methods are many and variable. One of its fundamental principles being to use only open systems of classification and to use multiple working hypotheses, descriptions of its "way of application" can only be made on theoretical grounds or by giving concrete examples of its use in particular cases. In this way, the merits of the structural analysis have been discussed repeatedly by WEGMANN.

Since the methods, each time they are applied to a new region, should be refined according to the particular requirements of this, the "savoir faire" of the structural analyst must grow (more or less) out of his personal experience.

The author has striven to present the work in a fashion which should make it possible for the reader to follow the analytical procedure through its various phases from the descriptive into the interpretative stage. Doing so he wants to apologize for some unavoidable repetitions and somewhat lengthy descriptions of observations with postponement of the conclusions. Even if a complete objective presentation in this manner can hardly be achieved, because some interpretations—intended or not—may infiltrate at an early stage, (see page 14), this way of presentation is thought to be the most honest and the most rewarding for the reader interested in the methods.

ASGER BERTHELSEN
Geological Survey of Greeenland,
Geological Institute of the University, Copenhagen

History and Development of Research

K. L. GIESECKE, who in the years from 1806 to 1813 carried out mineralogical studies along the west coast of Greenland, visited Tovqussap nunâ in 1807 on his journey from Godthåb to North Greenland, fig. 1. His notes from this voyage are cited here in English translation: "We passed thus Niakok [Fiskefjord] and the camping place Atangmik and rounded Nesanguak (Kalotten). This passage is dangerous because the mainland faces the open sea.

The rocks along this stretch are reddish granite and gneiss in which syenite (hornblende gneiss) and occasionally hornblende schists form important layers. On the mainland a steep mountain, Tikarnak [Tovqussaq Mt.] rises high above the others". (GIESECKE, 1878, p. 45).

No other geologists visited the area until 11 years ago when Dr. HANS RAMBERG, whom the present author at that time served as an assistant, entrusted him with the preliminary mapping of what appeared to be a dome structure at Tovqussaq. Working from a base camp at Gammel Lejrskar, the author mapped out the central portions of the peninsula and the existence of an overturned dome structure was verified. During this work a hand drawn enlargement of a map in 1:200.000 was used for mapping. The geological sketch map accompanying the preliminary report published in 1950 was prepared from aerial photos after the return to Copenhagen. Mr. H. SØRENSEN—also in the season of 1949—studied some ultrabasic rocks at Langø. His results of this study together with those obtained from similar occurrences within the surrounding district, were published in 1953.

When the author in 1953 returned to the Tovqussaq area, maps in 1 : 20.000 had become available. This season, however, was spent largely on reconnaissance of the surrounding regions, from which outline maps and photo-mosaics on a scale of 1 : 40.000 had been prepared. The area north of Tovqussaq was surveyed in the pleasant company of Mr. HENNING SØRENSEN. We also worked together further to the south, between Godthåb and Færingehavn. The Fiskefjord region was surveyed by the author.

In 1954, the author spent one and a half months on detailed mapping of the Tovqussaq peninsula. Unfortunately bad weather delayed the

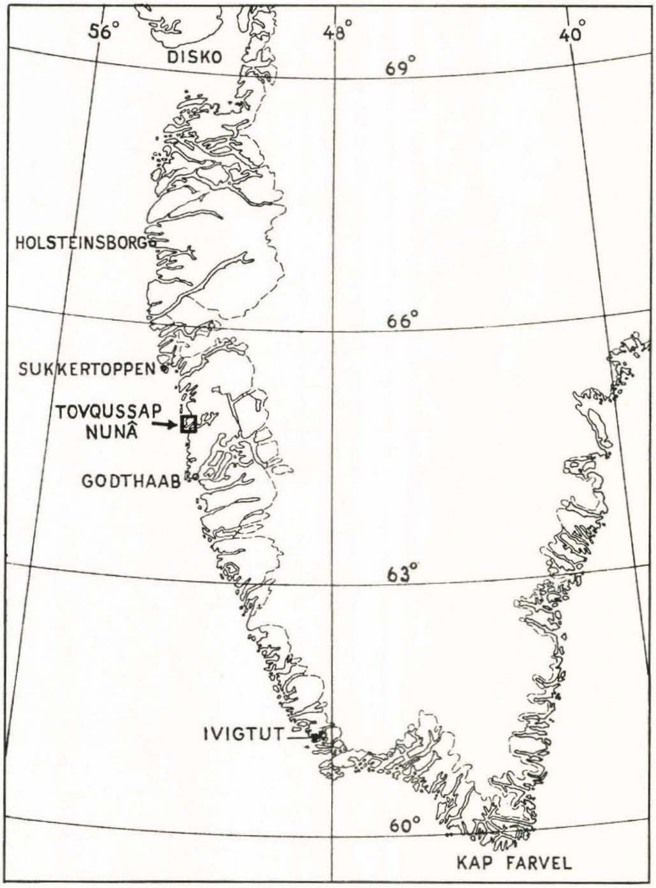

Fig. 1. Key map showing the location of Tovqussap nunâ, West Greenland.

programme and the southernmost and eastern part of the area could only be touched on.

The results from 1949, 1953 and 1954 were utilised in the compilation of an unpublished prize disertation in January, 1956. Abstracts of this work were presented at the Nordic geological winter meeting in Oslo in January, 1956, and at the assembly of the Geologischen Vereinigung at Göttingen in March, 1956. (Berthelsen, 1957).

When, in 1956, the author got the opportunity to spend a fortnight at Tovqussaq, he was able to map the southern and eastern parts. During this season's work the composite nature of the Krebsesø antiform was realised, and the need of a revision of the older mapping to the west and north became apparent. Since the author was very occupied with other work in south-west Greenland, this revision could not be undertaken before the end of the season in 1957, when, in ten days of concentrated work, he succeeded in finishing the map — as presented in plates 1 and 2.

Although another visit to this very interesting region would be highly desirable to check up various details and problems, the author has abandoned the idea of a return since leave from Greenland during the summer of 1959 offered him an opportunity to write up his results from Tovqussaq, an opportunity which may not recur in the coming few years.

The present work summarizes all results from the author's field work in Tovqussap nunâ from 1949 to 1957 and his laboratory work. The regional geology of the southern Sukkertoppen district will be described in a forthcoming publication (part III).

A separate account of the post-orogenic basic dykes and their relation to some major faults has been prepared by Mr. D. BRIDGWATER and the author (BERTHELSEN and BRIDGWATER, 1960).

Acknowledgements

The author is grateful to the board of GGU for having sanctioned and sponsored his continued research on the Tovqussaq area even at times when this work interfered with other duties. The author is in great debt to his former teacher, professor, dr. phil. ARNE NOE-NYGAARD, whose moral and professional support and never-failing interest in the author's research work have stimulated him to its fulfilment. His introduction to the principles of the structural analysis the author owes to professor, dr. phil. E. WEGMANN. He cannot emphasise too much the value of the innumerable hours which professor WEGMANN so unselfishly spent with him during his stay at Neuchâtel in the spring of 1953, to his subsequent work in Greenland.

It is also a pleasant duty for the author to extend his best thanks to all persons who in one way or another helped him in Greenland or in Copenhagen; to the skippers, Messrs. SANDER NIELSEN, MAURENTIUS POULSEN and ÅGE HANSEN; to his former chief, professor, Dr. HANS RAMBERG; to his colleagues and field companions, Messrs. ERLING BONDESEN mag. scient., NIELS HENRIKSEN stud. mag., STIG BAK JENSEN cand. mag., SIMON LÆGAARD cand. mag., HARRY MICHEELSEN mag. scient. and HENNING SØRENSEN, lecturer, mag. scient; to Mr. and Mrs. MAULE FREDERIKSEN and Mr. HJALMAR ANDERSEN for the great hospitality we received in their homes at Tovqussaq; to Mr. HARRY MICHEELSEN for refractive index measurements; to Mr. IB SØRENSEN cand. polyt. et lic. techn., for trace element determinations; to Mr. RØHLING for his very skilful draughting of the geological map (Plates 1 and 2); to Mr. POVEL POVELSEN, photographer at GGU, for enlargements of photos for reproduction; to Messrs. C. A. Jensen, German Petersen and H. Valentin for preparation of thin sections.

The preparation of the maps (Pl. 1 and 2) accompanying this paper has been made possible through the courtesy of the Geodetic Institute, Copenhagen.

Mr. C. Pulvertaft kindly corrected the English manuscript. The text has been ably typed by Miss E. Boye.

II. DISCUSSION OF METHODS AND NOMENCLATURE

The litho-structural mapping

The essential base for the structural analysis is a geological map. The more accurate and detailed the mapping the more accurate are the results of the analysis. Having had the opportunity of working in an exceedingly well exposed region with good and reliable topographic maps and excellent aerial photographs available, all the prerequisites for the preparation of a reliable geological map were fulfilled, and the author hopes that his efforts have not been in vain.

Mapping of highly metamorphosed or even ultra-metamorphosed rocks is, however, no simple matter. In an area characterised by more or less continuous exposures, the wealth of details seen in the field is so overwhelming that the selection of the feature to be mapped may greatly trouble the conscience of the geologist. To overcome these problems the author has used the procedure of mapping, remapping and revising the map yet again and of selecting out smaller and smaller regions for more and more detailed study. Thus a general impression of the relative importance of different kinds of features and the interrelation between the various orders of size of these, was achieved before the final compilation of the map was undertaken. In this selection, a subjective element may have been brought in at an early stage of the work, but this seems hard to avoid when rocks of composite and often obscure origin are dealt with.

During the actual process of mapping litho-structural divisions were used, a technique which presupposes good exposures, or great experience. The litho-structural mapping was facilitated by the occurrence of several distinct marker horizons, namely the pyribolites shown in the geological map. By a continuous record of the course of these layers, a primary division into major structural units was obtained. The marker horizons were easily detectable in most places from their basic composition, but in places the continuation of these marker horizons could only be traced by step-by-step mapping along the strike of the rocks. The disappearance of the general lithological difference between the marker horizon and its surroundings was moreover often accompanied by intense smallfolding. In such cases only very good exposures allow a continuous tracing of the horizon. By this technique some remarkable structures were revealed, which doubtless would have been neglected if ordinary traverse mapping had been applied, or if only scattered outcrops could have been studied.

The field behaviour of the marker horizons furnish good examples of the doubtful value of a purely lithological (petrographical) classification as a guide to mapping. Although lithology and structure are congruent in many cases, there are some very important exceptions. Border lines drawn according to an arbitrary lithological rock classification may thus lead to a false impression of the structures.

The litho-structural mapping pays due regard to the lithology but strives first of all to establish a reliable record of the structures. Doing so, it also includes observations of small scale structures and their relation to major structures. Through a structural approach many petrological problems may be viewed from new angles, since changes in composition can be correlated with changing structural situations.

The litho-structural mapping can be performed without any preexisting knowledge of the origin or nature of the structural features mapped. Such a knowledge will gradually grow out of the experience gained as the mapping proceeds.

When rocks of composite origin or rapidly changing facies are met with, the structural mapping enables such rocks to be classified roughly into major litho-structural units. It becomes possible thereby to distinguish between local structurally controlled lithological variation and more fundamental differences in composition (the term fundamental is used instead of "primary").

The geological map is presented in two sheets, one indicating the lithological variations found within the structural frame, and another which shows the final interpretation of the latter.

The structural map largely contains the material used to perform the structural analysis. The scarcity of structural readings is primarily caused by the rather homogeneous nature of many of the rocks. Moreover, readings were taken principally in the crucial areas (such as hinge zones). In simpler areas, where the readings are more scattered, their value is however greater, since they are more representative. When mapping in such areas, a constant "open eye" was kept for possible variations of the small scale features, but readings were only noted when sensible changes occurred. The representative nature of these readings should be born in mind, when at a later stage they are used in the geometrical analysis. This again is one of the advantages of well exposed regions. The litho-structural mapping has shown that the Tovqussaq rocks, in their present highly metamorphic state, form a conformable succession which has been folded and refolded in a very complex manner. The metamorphic and kinetic transformations have obliterated almost all primary structures and it is only possible to trace the evolution back to a certain stage.

All the rocks shown on the map of Pl. 1 (excluding the diorites) have passed through a progressive metamorphism and several phases of movement before they attained their maximum metamorphic state in granulite facies. Structures developed during the progressive metamorphism can thus be shown to have been refolded under granulite facies metamorphism. Towards the decline of the deformation, but still under granulite facies conditions, certain pyroxene-bearing diorites were formed —largely at the expense of pyribolitic rocks. The dioritic rocks help to separate posthumous movements which occurred at a still later stage, when amphibolite facies conditions prevailed.

In order to analyse the structural and metamorphic evolution outlined above, we have to treat the rock succession as a conformable one, but this does not necessarily mean that it originally possessed this character. Unravelling the evolution backwards, step by step, we reach a certain stage beyond which we only can guess. Is the conformity of the succession, which we can trace right back to this point, due to a still older kinetic influence or does it represent the primary bedding? Although important in itself, this question has no bearing on the validity of the analysis of subsequent superimposed deformations and can thus be left unanswered for the moment.

When, therefore, in the following, expressions like layers, bands, sequences and successions are used, it should be remembered that they do not involve any interpretation as to the origin of these layers, bands, etc., but are purely used as descriptive terms for the stratiform lithological units revealed from the mapping. They permit an analysis of structures developed later. Thus they bring out the existence of old recumbent folds, a dome structure and still younger superimposed structures. From this analysis it may appear that what was regarded as a 'layer' on the map, in reality shows up to be a repeated series. It is not until the end of the analysis that a series, in a stratigraphical sense, can be established — if then the origin of the layering is known.

This procedure has influenced the naming of some of the important litho-structural units. What at an advanced stage of the analysis showed up to be one and the same layer had in some cases been given different names in the field. Although in some ways inconsistent, this system of naming has been retained in the final maps, since it reflects the trend in the progress of the analysis. The names given to the different layers are indicated on the structural map, Plate 2.

On the use of the terms Granulite and Granofels

In his first attempts to write up the present paper, the author ran into serious trouble concerning the nomenclature to be used for the Tovqussaq rocks. From his search to find an appropriate system of classification for high- or ultra-metamorphosed rocks it soon became apparent that no such one exists. This is particularly the case with the granulite facies rocks, many of which have been described and named as if they were magmatic rocks. Almost every school of petrology has its own system, but unfortunately many identically spelt terms are used in different meanings within these systems. This inconsistency becomes the more apparent when the author is compelled to write in a foreign language in order to be understood by his colleagues.

Thus, at present, there are French, English, German and mineral facies granulites.

In French, granulite is a fine grained, granular, acid granite containing muscovite and accessories such as topaz, apatite and tourmaline (TERMIER and TERMIER, 1956, p. 256, and JOHANNSEN, 1958, I, p. 215).

In English, granulite is a metamorphic rock, with granulose texture and composed of even-sized granular minerals (e.g. feldspars, pyroxenes and garnets). Parallel or banded structure is due either to the presence of streaks or lenticles of non-granular quartz, or to the alternation of bands in which different minerals predominate (RICE, 1953; HOLMES, 1920).

In German, granulite in its typical development is a metamorphic, rather fine to extremely fine grained, garnet-bearing rock of white or light reddish or yellowish colour. The rock consists mainly of feldspar, quartz and scattered garnets. Biotite, tourmaline, rutile, sillimanite, kyanite and green spinel (hercynite) may form minor constituents, while apatite, zircon and ore are sparse accessories. The feldspar is microperthitic potash feldspar and antiperthitic oligoclase (including mesoperthite; for this term see MICHOT, 1951, p. 270). The texture is characterised by thin layers or bands of xenoblastic feldspar and quartz which alternate with layers or bands made up of plate-like quartz grains (ROSENBUSCH, 1898, and SCHEUMANN, 1954). SCHEUMANN calls this texture leptynitic, but since this usage of the term 'leptynitic' does not agree with its common meaning in French, the author would prefer to speak of a Plättung texture.

In the classic granulite region, the Granulite Gebierge of Saxony, these granulites sensu stricto are intimately associated and interbanded with pyroxene-bearing metabasites, ultrabasites, para-rocks and eclogites. This rock association was first described by LEHMANN (1884) as the granulite series. LEHMANN used the name pyroxene-granulite to describe the metabasic plagioclase-pyroxene-bearing rocks, although he was well aware that this was an unfortunate use of the term granulite as these pyroxenic rocks are granoblastic and do not show Plättung texture. "It seems necessary to give these dark rocks another name" (translated from German, as cited by SCHEUMANN, 1954, p. 103). LEHMANN's classification, however, has been left largely unchanged by later students of this classical area, even if Scheumann still points to the lack of an appropriate name for these plagioclase-pyroxene rocks, which name would also cover the "basic charnockites" of QUENSEL (SCHEUMANN, 1954, p. 103).

The manner in which granulite gradually lost its original meaning, being used for all rock types within the "granulite series", is quite similar to the modification which HOLLAND's term charnockite has undergone. Being originally coined for the hypersthene-bearing granitic member of the "charnockite series", it is now generally used for intermediate and basic or even ultrabasic "charnockites" as well. In consequence charnockite sensu stricto is now referred to as acid charnockite.

With the introduction of the mineral facies classification (ESKOLA, 1920; BARTH, CORRENS and ESKOLA, 1939), the use of the term granulite has grown still more confusing. Thus, in his paper on the Finnish granulites of Lapland, ESKOLA states: "All the rocks of granulite facies are granulites" (ESKOLA, 1952, p. 142). As Plättung texture and extreme fine grain seem to be exceptions rather than common features in granulite facies complexes, the incorporation of "granulite" in the nomenclature of the facies classification invalidates even the former extended use of the term (as in the "granulite series"), because now it may be applied to rock associations which do not even include granulite s. s.

The wide acceptance and use of the facies classification makes a complete abolition of the term granulite impossible here, but the term may be restricted to include only granulite s. s., this rock being regarded as a typomorphic rock of (at least some) granulite facies complexes. This attitude has also been taken by HANS RAMBERG in his textbook on metamorphic and metasomatic rocks (RAMBERG, 1952). No one would contemplate calling all rock types found within an amphibolite facies complex amphibolites. Amphibolite (i. e. a plagioclase-hornblende metamorphite) is a typomorphic rock of this facies—and no more.

In the ever increasing literature on charnockites, the English textural term granulite has become widely used to describe even-grained rocks such as pyroxene-granulites. (These rocks would actually be identical to German pyroxene granulites).

The conclusion arrived at from the above, that the use of granulite should be restricted to granulite s. s. and to granulite facies (meaning a mineral facies, where granulite s. s. is a typomorphic rock), prohibits, however, any further use of the term even in a textural sense.

R. Goldsmith (1959) has recently suggested a new term, granofels, which might fill the space left by dropping granulite in the English sense. Goldsmith defines granofels as a medium- to coarse-grained, granoblastic, metamorphic rock without, or with only indistinct, foliation or lineation. Unfortunately granofels is a linguistic hybrid, which most probably will not appeal to continental geologists. It is nevertheless a very useful term, especially because it includes rocks with some kind of preferred orientation. In this respect it seems more suitable than the discarded "textural granulite".

In the author's experience, even-grained rocks, devoid of any preferred orientation, are rare or of rather restricted occurrence in regionally metamorphosed complexes.

Most rocks which at first sight appear to be granulose often betray some sort of preferred orientation (usually a lineation) when studied more closely or if really granulose, they pass rapidly into lineated varieties. Thus in the Tovqussaq region most rocks are B- or S-tectonites in the sense of Sander.

As no linguistically more suitable way out of this problem is available and fifty years have passed without any other appellations the author has chosen to use R. Goldsmith's term, granofels. A purely textural classification, however, in several cases may be difficult to apply and so the term has only been used where appropriate compositional rock names were lacking.

The classification which works best in the litho-structural mapping is one which is based primarily on the mineral content of the rocks and thus also takes the chemical composition into account.

For the basic rock types, the lack of appropriate rock names is evident. "Pyroxene-granulite", in which the pyroxenes dominate over hornblende but which carries the same plagioclase as an amphibolite, has earlier been described as pyroxene-amphibolite (e.g. Parras, 1958). This is, however, a somewhat risky extension of the term amphibolite. To call such rocks diopside, hypersthene, plagioclase granofelses would also be rather clumsy.

Although well aware of the doubtful outcome of introducing new rock names, the author feels compelled to make use of some new terms which, he hopes, may largely replace the unfortunate term "basic granulite" (or pyroxene-granulite) in future.

Towards a better nomenclature for "Basic Granulites"

In many regions amphibolite is connected with pure pyroxene-plagioclase rocks by a transitional series of rocks. We will consider the case where hornblende gives way to ortho- as well as clino-pyroxenes.

Amphibolite containing these two pyroxenes in quantities less than hornblende, may be termed pyroxene-amphibolite, but when the content of pyroxene increases further, the resulting rocks are not amphibolites, as these should be composed essentially of hornblende and plagioclase. With the complete disappearance of hornblende a rock results which hitherto has been described as pyroxene-plagioclase granulite. Now, the term granulite having been abandoned, the lack of a new name for this rock is evident. It might be substituted by pyroxene-plagioclase grano-

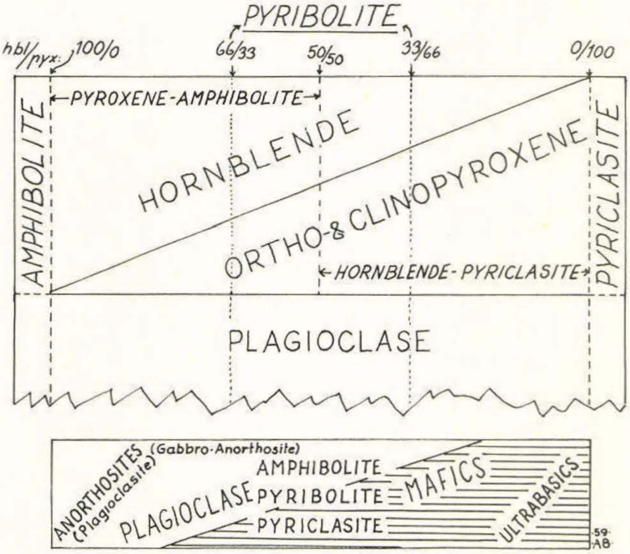

Fig. 2. Diagram showing the possible variation in mineral composition of pyribolite and pyriclasitic rocks.

fels, but a congruent name to amphibolite, i.e. a name which by definition takes only the mineral content – and not the textures – into account, would be desirable. Here the author would suggest "pyriclasite", which is constructed from pyr (pyroxene) and clasite (plagioclase), to describe rocks composed of two pyroxenes and plagioclase (pyriclasite is favoured instead of pyroclasite, because of the resemblance of this latter to pyroclastics). In consequence, the rocks which contain pyroxenes in excess of hornblende could be called hornblende-pyriclasites. The transition would thus be covered by the succession: amphibolite, pyroxene-amphibolite, hornblende-pyriclasite and pyriclasite, fig. 2.

This system of classification, however, may cause some difficulties when applied to rocks containing the two pyroxenes in nearly same amount as hornblende. To overcome this difficulty, such rocks are tentatively given a separate name, pyribolite, which is constructed from pyribole, a term used by Johannsen as a collective name for hornblende and pyroxene. The position of the term pyribolite is thus analogous to that of adamellite, which lies within the border zone between granite and granodiorite.

Pyribolite will then comprise the transitional rocks which show a hbl./pyrox. ratio varying from 66/33 to 33/66, fig. 2. Although a classification in which the nomenclature overlaps may cause some confusion if misused, it seems preferable because hornblende-pyroxene-plagioclase rocks often show a certain variation within themselves due to banding. The suggested classification renders a description of such banded rocks possible without the use of specific terms for the individual bands. The majority of the Tovqussaq hornblende-pyroxene-plagioclase rocks may thus be described as pyribolites, because they generally take a position intermediate between amphibolites and pyriclasites.

Fig. 2 shows diagrammatically the possible variations in mineral content of the rocks discussed above. The content of plagioclase in relation to content of mafics is also variable. With increasing content of feldspar, the rocks pass into anorthosites (plagioclasites), while they grade into ultrabasic rocks with disappearance of the light constituent. No strict limits of the mafic/felsic ratio are given for the application of the terms pyribolite and pyriclasite except that they should correspond to those set for rocks of mela-dioritic to gabbro-noritic composition. In consequence the plagioclase may vary from andesine to bytownite with an average of about An 50. It should be remembered that the presence of orthopyroxene is essential for the use of the terms pyribolite and pyriclasite. This requisite is deemed necessary if conflicts with the facies classification should not arise. It also serves to exclude the entry of various calc-silicate rocks into the suggested system. Scapolite, garnet and biotite may enter as varietal constituents in the rock transition discussed. Where wanted, the varietal mineral names may be used as prefixes, e.g. garnet-pyribolite.

The prefixes clino- or ortho- may be used to designate varieties of pyribolites or pyriclasites where one of the pyroxenes predominates to such an extent that the other is almost excluded.

The suggested terms fit into the mineral facies classification, but have also been defined in a manner which should prevent their contamination with the nomenclature of the latter.

It is hoped that the new terms may help to make future petrographic description of "basic granulites" more stringent. The suggested nomenclature is flexible and easily applicable, not requiring calculations of norms but only modal estimates. If a purely textural basis of classification is preferred, the new terms may also be used descriptively, as for example in pyribolitic schist.

On the use of the terms Gneiss and Charnockite

With the acid to intermediate rocks similar problems do not exist, as these rocks generally can be described as gneisses. Doing so, the author uses the following definition for gneiss: a metamorphosed, medium to coarse grained rock in which (or part of which) quartz-feldspar predominates and which on a mesoscopic scale shows banded, veined, streaky, smallfolded or any similar structures (due to compositional and textural variations) and/or parallel arrangement of the dark minerals in some or all of the rocks. Preferred orientation of the light minerals may exist as well, but is not used as a criterion in the definition.

Thus defined, gneiss is a broad field term, which may include schistose or granofelsic components where these are involved in the structures mentioned above. The variation in composition between the different components (e. g. the bands) may be great, but the bulk composition within the area of an outcrop should be that of a quartz-feldspar-bearing rock of leucocratic to melanocratic composition. Handspecimens of gneiss may thus show rocks with foliation (closed, interrupted or dispersed), lineation or pure granofelsic texture. The gneiss type which is made up of only one component with interrupted or dispersed foliation is called homogeneous gneiss.

In the definition above, the grain size was allowed to vary from medium to coarse grained. When the grain is finer but still visible with the unaided eye, the author prefers to speak of fine grained gneisses. Scandinavian geologists often use the term leptitic gneiss for such rocks which are transitional between leptite and gneiss. But, as SEDERHOLM remarks, the term leptite is a "cul de sac" for all fine grained metamorphic rocks of dubious origin.

Even if only approximate modal estimates have been made during the petrographic study of the Tovqussaq rocks it has been possible to classify them after JOHANNSEN's descriptive system. Gneisses of quartz-syenitic, alaskitic, granitic, granodioritic and quartz-dioritic composition will be described. Such a division is only meant to describe the variation in chemical composition. Actually the terms in most cases only apply in this sense, because the feldspars in the Tovqussaq rocks are antiperthitic and micro-perthitic types generally not found in "normal" quartz-syenites, alaskites etc.

In fact some of the Tovqussaq rocks show greater affinities to the charnockite suite. SUBRAMANIAM has recently revised the type area and the nomenclature of charnockites. According to a preliminary report (SUBRAMANIAM, 1959) he suggests the term "charnockite suite" be restricted to what previously was embraced under "acid charnockite", the remaining intermediate, basic, and ultrabasic divisions now being regarded as alien to the proper charnockitic rocks. Thus redefined the charnockite suite embraces:

1. Charnockite sensu stricto (hypersthene granite, birkremite),
2. Enderbite,
3. Hypersthene quartz syenite,
4. Alaskite.

Charnockite sensu stricto has been redefined since the type specimen, JOB CHARNOCK's tombstone, was found to contain small grains of red garnet. Charnockite sensu stricto is now a hypersthene-quartz-feldspar-rock with or without garnet, characterised by greenish blue feldspar, and greenish blue quartz, the dominant feldspar being a microperthite (SUBRAMANIAM, 1959, p. 328).

Interestingly enough, the enderbitic member was found to be quantitatively the most important within the type area near Madras.

The revision of the charnockite nomenclature is useful, because charnockite now is restricted to hypersthene-bearing quartzo-feldspathic rocks. Too wide an extension of a rock term has always lead to its misuse. However, the use of quartz-syenite and alaskite is somewhat unfortunate if a classification taking into account the high-temperature perthites found in rocks of the charnockite suite should be strived at. This was more or less fulfilled by the former extended use of "charnockite".

SUBRAMANIAM claims that the charnockite suite is of magmatic origin. The present author is not completely convinced that the type area, with its poverty of outcrops, allows such a conclusion to be drawn. From structural arguments a metamorphic origin for these rocks may just as well be claimed. (See also WILSON, 1957).

When the term charnockite has not been applied in the description of the Tovqussaq rocks it is primarily because charnockite sensu stricto has not been found within the region. It may formerly have existed, in which case it has been completely altered to biotite gneisses during a subsequent phase of metamorphism under amphibolite facies conditions.

A complete revision of the nomenclature of the quartzo-feldspathic rocks with high temperature perthites is still needed. Material for such a revision has not, however, been at hand for the present author.

Some dioritic and the ultrabasic rocks of the Tovqussaq peninsula have been described using the available terms for similar igneous rocks. In spite of occasional preferred orientation, these rocks have such a plutonic appearance that this procedure seems justified.

Textural Considerations

During his study of the Tovqussaq rocks, the author has made ample use of the textural nomenclature developed by F. Becke (1904) in connection with his classical work in the eastern Alps.

In order to enable a detailed description of the granoblastic gneisses and granofelses to be made, a few additional terms and prefixes have been brought into use. First of all, a further distinction in the relative grain size seems necessary. Rocks composed of mineral grains of similar, or nearly similar, size are called eu-granoblastic, while rocks which show varying grain size but which cannot be called porphyroblastic, are de-

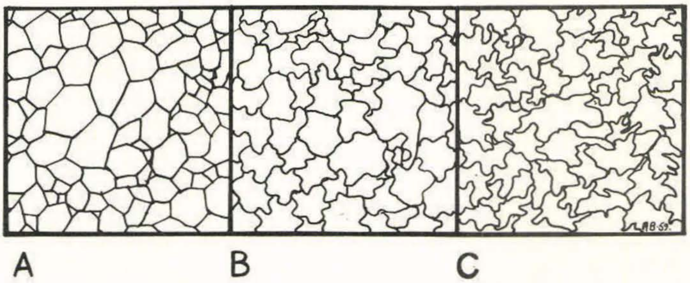

Fig. 3. Division of granoblastic textures.

scribed as hemi-granoblastic. The prefixes eu- and hemi- (real- or half way-the same size) have been chosen to avoid confusion with Becke's terms homoeoblastic and heteroblastic, where the homoeoblastic textures comprise granoblastic, lepidoblastic, and nematoblastic variants. The new terms eu-granoblastic and hemi-granoblastic are thus subdivisions of Becke's granoblastic, which again is one of several kinds of homoeoblastic tecture.

A further descriptive division of eu- and hemi-granoblastic textures is based on the design or pattern shown by the outlines of the individual grains, fig. 3. Where grains border on each other with more or less rectilinear mutual borders, the resulting texture is called saccharoidal (eu- or hemi-granoblastic), fig. 3 A. Where the grain border has a curved or sinuous course, i.e. when the grains are lobate, the texture is called interlobate (eu- or hemi-granoblastic), fig. 3 B. Where neighbouring grains interfere and interlock in a manner suggestive of pseudopodia given off from an amoeba, and when the texture—in spite of these irregularities—still has preserved its granoblastic appearance, the expression amoeboid (eu- or hemi-granoblastic) is used, fig. 3 C. With the loss of the granoblastic texture, diablastic textures appear.

Regarding grain size, rocks with grains smaller than 1 mm (but still visible with the naked eye) are called fine grained. Medium grained rocks have a grain size between 1 mm and 1 cm while coarse grained rocks have grains larger than 1 cm (in diameter).

III. PETROGRAPHIC DESCRIPTION AND CLASSIFICATION OF THE TOVQUSSAQ ROCKS

A description and discussion of the structural evolution of a high-metamorphic to ultra-metamorphic area like the Tovqussaq region presupposes a knowledge of the petrography. The composition of the material involved must be known. In most geological descriptions from such areas the greatest attention has been paid to petrographic and petrogenetic problems, while structural questions are left more or less aside (often because of the insufficiency of outcrops). The rocks are classified mainly on petrographical and mineralogical criteria. The author has chosen the other way round, using mainly structural criteria as a basis for his mapping, but in order to make the presentation of the results intelligible for the petrographic-minded readers, an attempt has been made to present a petrographical division as well as a structural classification. The petrography, therefore, will be treated twice. First, the petrographic types are described and detailed descriptions of about 100 thin sections of selected samples are given to supply as exact as possible an illustration of the variations amongst these: great importance is attached here to the transitional types. Next the petrographic types and their variations will be correlated with the field observations made on the outcrops and their bearing on the area of the geological map discussed. Due regard should thus be paid to changing "l'ordre de grandeur" (E. WEGMANN, 1951). A summary with petrogenetic conclusions and discussions is given finally in conjunction with the results of the structural analysis.

The only mineral which has had a fairly thorough treatment during the petrographic studies is plagioclase. The An content has been determined either by means of extinction angles in sections normal to (010) as well as (001) (in favourable sections or by aid of the universal stage) or by RI measurements. The latter were kindly performed by Mr. H. MICHEELSEN, mag. scient. Some untwinned oligoclases and andesines were determined simply by comparison of the refractive indices of the plagioclase with that of the neighbouring quartz.

Mr. MICHEELSEN has also carried out some RI determinations of hypersthenes and has determined $c \wedge Z$ and 2 V in diopsides from an eclogitic rock. Determinations of $c \wedge Z$ and $2V$ in some hornblendes and diopsides from various basic rocks were carried out by the author.

In the following description the customary petrological division into acid, intermediate, basic, ultrabasic and "other" rocks is used.

A. The acid division

This division comprises rocks with compositions varying from alaskitic to quartz-dioritic.

Most of the gneisses, except for the highly mixed or composite types, belong to this group. Among the more acid gneisses three main types may be distinguished.

a) Hypersthene-bearing gneisses,
b) Purple gneisses,
c) Light coloured biotite gneisses.

All three types belong, however, to a transitional series and may pass gradually into each other. It should be noted that the term gneiss here is used in a broad sense, and that the rocks of this group may appear granofelsic as well as gneissic in handspecimens.

The hypersthene-bearing gneisses

These usually show the dark greenish to brownish colour and the blue quartz typical of granulite facies rocks. They are composed of plagioclase, quartz and pyroxene. Apatite and ore form common accessories. Sphene is totally absent. The plagioclase generally varies in composition from about 30 to 40 % An, and is usually well twinned and often antiperthitic. The content of quartz is rather variable. A transparent to pale green diopside occurs in several slides in an intergrowth with hypersthene or forming thin rims around this mineral. Uralitic hornblende (bluish green) and yellow to brown biotite are seen in many slides. They seem to take the place of the pyroxenes. Orthoclase and microcline have been noticed in only subordinate amounts (antiperthite excepted).

The following thin section descriptions contain more detailed information about these rocks.

35872. Homogeneous quartz-dioritic granofels, SSE of Snirkelsø.

A homogeneous medium grained rock with a somewhat interlobate, hemigranoblastic texture. The minerals are antiperthitic andesine, quartz, hypersthene, a little diopside, microcline and hornblende together with accessory apatite and ore. The total content of mafic minerals is rather small. The antiperthitic texture seen in the plagioclase is very fine to almost submicroscopic. The microcline is found as scarce small, interstitial grains with irregular extinction. The hypersthene, which is pleochroic from pinkish to greenish, is rimmed by greenish diopside, which may have been altered later into a bluish green hornblende.

4029 C. Quartz-dioritic hypersthene gneiss, top of Tovqussaq Mt.

A fine to medium grained rock of greenish brown colour. In the handspecimen a faint lineation of the dark minerals may be discerned. When seen under the microscope, the texture appears interlobate hemi-granoblastic. The minerals are antiperthitic plagioclase (32—38 % An), quartz, hypersthene, diopside, a little biotite and accessory apatite and ore. The plagioclase may show reverse zoning. It contains antiperthite partly as very fine (almost submicroscopic) inclusions and partly as coarser strings and patches. Except for a few zircon needles, the quartz seems devoid of inclusions. It shows lobate to amoeboid outlines and its grain size is highly variable. The faintly pleochroic hypersthene forms irregular grains and may be intergrown with diopside. In some grains, hypersthene is seen to be replaced by a yellow to brown biotite which has grown along the cleavage planes of the pyroxene. Biotite may also be associated with ore grains. A few small, interstitial grains of potash feldspar were also observed.

4029 A. Pegmatitic type associated with sp. 4029 C (see above).

A medium to coarse grained, greenish, leucocratic rock with blue quartz. The texture is highly interlobate and granoblastic. The minerals are quartz, antiperthitic plagioclase, and a little microcline. Quartz occurs in granulated amoeboid grains. The plagioclase (about 30 % An) contains broad strings and patches of antiperthite. Microcline is seen as grid twinned interstitial grains.

19246. Banded hypersthene gneiss, Hulebugt.

A purple to greenish, medium grained gneiss with more fine grained amphibolitic bands (about 2 cm thick). When seen under the microscope the texture appears as saccharoidal granoblastic with lobate quartz. The minerals are plagioclase (finely antiperthitic), quartz, hypersthene, some hornblende and biotite, accessory apatite and ore. The well twinned plagioclase is a slightly normal zoned andesine which shows an almost submicroscopic antiperthitic texture. The quartz forms larger somewhat undulating grains and small bleb-like grains. The pleochroic hypersthene forms irregular grains and granular aggregates. It is partly replaced by a greenish hornblende. The brown biotite is relatively rare. Hornblende is dominant in the mafic bands.

4094 B. Pyroxene-bearing band in purple gneiss, Lejrsø.

An almost fine grained, grey, poorly foliated rock with a hemi-granoblastic texture. The minerals are plagioclase, some quartz and diopside, hypersthene, biotite and hornblende, plus apatite and ore. The total content of mafics exceeds that of quartz. The plagioclase is an sodic andesine (35—40 % An). The xenoblastic diopside and hypersthene are partly uralitised or replaced by brown biotite. Diopside is seen as a rim on some hypersthene grains. Quartz shows lobate outlines and seems to be a late mineral, possibly introduced from outside. No antiperthite or microcline were detectable within the slide, although the specimen was taken in the immediate neighbourhood of an alaskitic rock to be described later (specimen 4094 A).

19274. Quartz-dioritic hypersthene gneiss, core of Pâkitsoq-antiform, north coast of Tovqussap nunâ.

A medium grained, brownish rock with a weakly developed lineation. The minerals are plagioclase (35—40 % An), quartz, biotite, hypersthene, hornblende and apatite. The texture is dominated by the saccharoidal granoblastic plagioclase, while quartz forms grains of lobate outline and varying size. A few small more or less rectangular inclusions of potash feldspar occur in the well twinned plagioclase, but otherwise no traces of antiperthite are seen. The biotite is pleochroic from yellowish to brown and shows lepidoblastic outlines. Corroded relics and small grains of pink pleochroic hypersthene are associated with green hornblende or biotite.

18219. Hypersthene gneiss, northern Langø.

A greenish brown gneiss with fine grained darker bands alternating with medium grained leucocratic bands containing blue quartz. The minerals are quartz (predominates in the light bands), plagioclase (26—27 % An), hypersthene, diopside, biotite, and hornblende plus accessory apatite and ore. The texture is dominated partly by the parallel arrangement of quartz and partly by the eu-granoblastic plagioclase. In the mafic-bearing band a reticulate texture with mafic spongy aggregates is developed. Quartz shows lobate or amoeboid outlines, even where forming orientated elongate grains. Undulate extinction is prominent. The plagioclase shows faint traces of a very finely antiperthitic texture. The faint pinkish hypersthene may form an intergrowth with diopside. Both are rimmed with or in the state of being replaced by hornblende or yellow-brown biotite.

Transition from hypersthene gneiss to purple gneiss

From the foregoing description it is seen that typical hypersthene-bearing rocks completely devoid of hydroxyl-bearing minerals are rare. In most specimens, the pyroxenes are in a state of being substituted by hornblende or biotite. When this transition proceeds further, other mineralogical and textural features appear and the microscopical appearance of the rock changes as well. The following three descriptions have been selected to illustrate this transition. The specimens were collected from the Frame Layer and the core of the Tovqussaq dome.

19243. Hypersthene-bearing quartz-dioritic gneiss, north-east coast of Qaersup ilua.

A medium grained, greyish yellow rock with a faint foliation marked by the arrangement of the dark minerals. Seen in thin section, the texture is slightly interlobate, hemi-granoblastic. The content of the dark minerals is estimated as 5—10 %. The rock is composed of plagioclase, quartz, some hornblende, hypersthene, biotite, diopside and accessory apatite, zircon and ore, besides a little secondary sericite. The plagioclase is faintly antiperthitic and shows an An content of about 30 %. Quartz forms an important constituent and occurs as lobate grains and small rounded blebs in the plagioclase. Hornblende forms xenoblastic grains with irregular to lobate outlines and replaces hypersthene parallel to its cleavage traces. The hornblende itself may be replaced by biotite in parallel intergrowth. The hornblende is pleochroic, X: pale greenish yellow, Y: pale green and Z: faint brownish green. The hypersthene is pleochroic from pale pink to transparent greenish. The transparent diopside is associated with hypersthene and partly rims this mineral. Biotite is pleochroic from yellowish to dark brown and forms lepidoblastic grains and unorientated flakes. Apatite is a rather common accessory whereas zircon only occurs in a few grains enclosed in hornblende or biotite. The ore is associated with the dark minerals.

13417 A. Hypersthene-bearing quartz-dioritic gneiss, north coast of Qaersup ilua.

A fine to medium grained, greyish green rock with an interlobate, hemi-granoblastic texture. The minerals are plagioclase, quartz, hypersthene, diopside, biotite, hornblende and accessory apatite and ore. The dark minerals constitute less than 5 % of the rock. The plagioclase is somewhat antiperthitic and contains about 25 % An. Quartz is a prominent constituent as lobate grains and smaller rounded blebs. The pinkish pleochroic hypersthene and the faint greenish diopside form irregular xenoblastic to corroded grains. In some cases a rim of bluish green hornblende is seen on the hypersthene, which also may be replaced by a yellowish to dark brown biotite. The biotite also forms flakes associated with ore.

13417 B. Purplish quartz-dioritic biotite gneiss, north coast of Qaersup ilua.

A light coloured purple rock of medium grain. The colour is due to the purple feldspar and the blue quartz. The texture is extremely interlobate, hemi-granoblastic. The minerals are plagioclase (antiperthitic), quartz, biotite and a little hornblende, some microcline and accessory apatite and ore. The dark minerals amount to less than 5 %. The antiperthitic plagioclase shows an An content of 20—25 %. The abundant quartz forms lobate to almost amoeboid grains and smaller rounded blebs. The larger grains show undulate extinction. The biotite is pleochroic from yellow/orange-yellow to dark brown/black. It forms seemingly unorientated lepidoblasts. In one place the biotite is seen to replace a strongly corroded to almost skeletal grain of bluish green hornblende. Microcline occurs very sparsely and interstitially.

Comparing the descriptions of specimens 19243, 13417 A and 13417 B, it becomes evident that the An content decreases along with increasing potash content (as reflected by antiperthite, microcline and biotite). At the same time, the texture grows less orientated and becomes more interlobate. The pyroxenes in their turn are completely converted into biotite. Parallel to this transition a complete change in rock colour takes place from greenish into purplish tinges.

The purple gneiss

This is a very spectacular rock because of its unusual colours which may include purple or even violet, depending on the relative proportions of purple feldspar and bluish quartz. The purple gneisses seem to be slightly more potassic than the hypersthene gneisses. They always carry a deeply coloured biotite, and a bluish green hornblende may also be present. The antiperthitic plagioclase generally varies within the range 25—35 % An. Interstitial potash feldspar is not uncommon. The accessories are the same as those found in the hypersthene-gneisses and sphene is still absent. While the hypersthene-gneisses showed textures varying between saccharoidal and interlobate, the purple gneisses generally have developed more interlobate or even amoeboid textures.

The two specimens described below originate from areas where a typical development of the purple gneiss can be seen over large stretches.

19235. Homogeneous purple gneiss, north coast of Tovqussap nunâ, about 1 km NE of Kangeq.

The handspecimen shows a purple, medium grained rock. The texture is distinctly interlobate, hemi-granoblastic. A faint parallel arrangement of elongated quartz and feldspar crystals may be perceived. The minerals are quartz, antiperthitic plagioclase and small amounts of biotite and hornblende besides accessory apatite and ore. The lobate quartz grains are somewhat sheared, which may also be the case with the plagioclase. The latter shows a finely developed antiperthitic texture which partly seems submicroscopic. The An content is about 30 %. The yellowish to dark grey biotite and the bluish green hornblende are scarce.

14976 A and B. Purple gneiss, E coast of Tovqussaq peninsula, NE of Dioritnæs.

The two specimens represent fine to medium grained, and medium to almost coarse grained varieties of the purple gneisses found at this locality. The most fine grained rock is poorly foliated and very leucocratic, while the more coarse grained is richer in quartz and biotite and better foliated. The texture of both rocks is interlobate, hemi-granoblastic—when the lepidoblastic biotite is disregarded. The plagioclase varies from an antiperthitic, basic oligoclase in the more fine grained rock to a less antiperthitic acid andesine in the coarser type. Microcline in small amounts is associated with the oligoclase, while absent in the quartz-rich rock. Where forming larger lepidoblastic grains, the biotite attains a dark brown colour. A few grains of bluish green hornblende and some ore were noticed in both rocks. Zircon was seen in the coarse grained, quartz-rich type, while apatite was absent in both types.

Transition from purple gneiss to light coloured biotite gneiss

The transition from the purple gneiss into the light coloured biotite gneiss may be illustrated by describing a series of samples collected at the coast north of Skiverne. Of six samples, four have been sectioned. First, the observations made on handspecimens are given:

Sp. 19276 A is a medium-fine grained purple granofels. Hypersthene may be discerned in places as a relic core in uralite. Otherwise, biotite is the most common dark mineral. Sp. 19276 B is slightly more coarse grained and shows interrupted foliation. The larger grains of blue quartz cause the more violet colour. Sp. 19277 is more light coloured, medium to fine grained and somewhat foliated. Biotite forms small scattered but parallel orientated flakes. Purple feldspar and blue quartz may still be seen, but most of the felsic minerals are distinctly more light coloured. Sp. 19278 represents a fine to medium grained rock with scattered but less orientated biotite. The colour is light grey with a very faint violet tinge. Pink thin veins traverse the specimen, which also contains larger (up to 1 cm) porphyroblasts of microcline microperthite. Sp. 19279 is a very light grey, almost fine grained rock with elongated rod-like quartz and scattered but orientated biotite. Some small red speckles, presumably due to weathering of small ore grains, are seen.

Under the microscope:

19276 A is seen to be interlobate, hemi-granoblastic, and the biotite flakes show a wavy orientation. The minerals are antiperthitic plagioclase (over 60 %), quartz, microcline, biotite, uralite, plus apatite and ore. The untwinned plagioclase seems to be an oligoclase. It contains antiperthite of two kinds, one almost submicroscopic and the other coarser, both being of string or hair type. The amounts of microline (interstitial) and biotite are about equal. The latter mineral is yellow to greenish brown. In one place it is seen to be intergrown with uralite—a pale bluish green hornblende with scarce relics of pyroxene. Quartz and ore also participate in this intergrowth.

In slide nr. 19277, the texture is amoeboid, hemi-granoblastic. The minerals are feldspar, quartz (more abundant than in 19276 A), biotite and accessory apatite and ore. The feldspar comprises oligoclase and microcline microperthite. The latter seems to be concentrated in certain bands parallel to the foliation. It is so filled with strings that it approaches a mesoperthitic composition. Where bordering on plagioclase, it seemingly corrodes the latter, which contains irregular rods and protrusions given off from the potash feldspar. The plagioclase is antiperthitic of the coarse string type, but may within one and the same grain grade into the above-mentioned diablastic plagioclase with intergrowth of potash feldspar. Where not hidden by antiperthitic or diablastic textures, closely spaced albite and pericline twins are seen in the oligoclase. The biotite is of the same type as found in sp. 19276.

Summarizing, it may be said that the content of quartz and potash feldspar seems to be greater in 19277 than in 19276. (This may, however, be due to the tendency of these minerals to concentrate in certain bands). A complete exsolution and rearrangement of the potash feldspar is seen in slide 19277, where the "diablastic plagioclase" is found. The elongated shape of the associated quartz may point towards the paratectonic nature of this process.

Sp. 19278 shows a still more advanced stage of this development. Of the former plagioclase, only diablastic grains are left, the other feldspar being microcline microperthite partly of porphyroblastic development. Comparing the quartz in the three sections, it is seen that in the purple gneiss, the quartz contains liquid inclusions (or gas bubbles), while in the next stage (19277) it becomes rutilated. In sp. 19278 inclusions of zircon, and a red-brown mineral (tourmaline) appear as well. The full development of the inclusion-rich quartz is met with in 19279. This rock shows a hemi-granoblastic texture which seems to have been formed by recrystallisation of a partly cataclased rock. The orientation of the biotite (now of a more greenish type) indicates that this cataclasis occurred along younger S-planes which cut the old foliation obliquely and caused its plication. This movement and the succeeding recrystallisation have evidently resulted in a new rearrangement of the plagioclase and potash feldspar. The former is largely reconstituted as an oligoclase which contains only relics of antiperthite. That we actually have relics is shown by the occurrence of the string type as well as a type which is so irregular that it must have developed from the diablastic intergrowth described above. The microcline in its turn forms separate grains with well developed cross hatching and hardly shows perthite texture.

Another important feature which may be noticed in this rock is the first appearance of sphene. In the earlier stages, the biotite seemingly has been capable of incorporating the titanium released through the break-down of the hypersthene. The change in the colour of the biotite from brown to greenish, however, can hardly be explained in this way, since in other rocks described below, greenish biotite occurs abundantly without the presence of sphene.

The deformation mentioned above, leading to the almost complete separation of the two feldspar phases, has left fewer traces in the relic mineral associations of the purplish rock type, but that it all the same has attacked this rock is evident from the undulate quartz and occasionally bent plagioclase twins.

Apart from these specimens, which enable one to establish an almost continuous record of the transition from the purple gneiss into the light coloured biotite gneiss, the study of other specimens has brought forward a wide variety of "evolutionary glimpses", which fit well into the ideas developed above. Some of these rocks which show greatest affinity to the purple gneisses will be described now. Others, which resemble more the light coloured biotite gneiss type, will be described after this.

Transitional types related to the purple gneiss

4027. Biotite/garnet-bearing grey to violet gneiss, south of the Tovqussaq mountain.

A greenish grey to violet, fairly foliated biotite gneiss of medium grain. Small deep brownish red garnets are seen in one part of the handspecimen, but do not occur in the slide, where the texture is seen to be slightly interlobate, hemi-granoblastic, with seams of parallel orientated dark minerals. The constituents are a faintly antiperthitic plagioclase (25—26 % An), quartz, hornblende, biotite, epidote and some microcline besides accessory apatite and ore. The yellow-green to bluish green hornblende is partly replaced by a greenish biotite, which in this case shows diablastic intergrowth with quartz. Faintly yellow epidote occurs quite abundantly in hypidioblastic grains associated with biotite. The occurrence of epidote (seemingly as a stable mineral) within this rock is remarkable.

4038. Pale violet biotite gneiss, Langø.

A pale violet, foliated rock of medium grain. The texture is dominated by the bands of granulated quartz and others of hemi-granoblastic plagioclase with associated lobate to amoeboid quartz. Faintly antiperthitic plagioclase (20–22 % An) and quartz occur in almost equal amounts. The scarce, greenish brown biotite forms interrupted foliae. Microcline occurs in very small quantities as interstitial fillings. A few relics of a bluish green hornblende are also found. Apatite and ore form scarce accessories.

19136. Pale purple biotite gneiss, Langø.

A light coloured pale purplish rock with pink veins carrying blue quartz. The sparse biotite is arranged parallel to the banding or veins. The grain size is fine to medium. The texture is highly interlobate, almost gneissose with elongated quartz grains. In thin section, the pink veins or bands are seen to correspond to stained stripes, which are somewhat richer in microcline than the surrounding areas. The minerals are feldspar, quartz, a little biotite and accessory ore. The feldspar is partly oligoclase (23 % An) and partly microline. Mesoperthitic and antiperthitic textures are found, although much of the plagioclase is free of inclusions. The interstitial relation of the microcline is still discernable, but is less clear than in slide 19136 (see above). The biotite is greenish and may show diablastic intergrowth with quartz. A single relic of the usual hornblende was found.

4018. Light coloured biotite gneiss, 1st Intermediate Layer, north coast of Inderhavn.

A light coloured grey to pink gneiss with medium to almost fine grain size. Streaks of biotite and ore indicate a weak foliation. Otherwise the texture is granoblastic to cataclastic with a faint banding. The minerals are antiperthitic plagioclase, quartz, potash feldspar, biotite and accessory apatite and ore. The calcic oligoclase is often untwinned. Where twinning is seen, the lamellae have been bent. The extinction is also undulose. Quartz forms granulated lobate to amoeboid grains and smaller irregular individuals. The potash feldspar occurs partly as fine to string-like antiperthite and partly as interstitial or separate grains. Where the latter border on plagioclase, microdiablastic intergrowths between the two feldspars have been observed. Mesoperthite has also been noticed. The biotite is pleochroic from yellowish brown to dark greenish brown. The lepidoblastic grains, which may be bent, are intimately associated with the ore grains. In the leucocratic bands a dark pigment is seen along cracks and fractures.

Among the transitional stages between the typical purple gneiss and the light coloured gneiss, some remarkable rocks, the feldspars of which are almost exclusively mesoperthite, have been met with. These rocks are generally very leucocratic and show quartz-syenitic to alaskitic compositions; because of their local and restricted field occurrence, they are not considered as "types".

4019. Purple to pink quartz-syenitic granofels, west of Sorthat.

A medium grained, saccharoidal, purple to pink rock, which is extremely leucocratic. The minerals are microperthite (almost mesoperthite), quartz, and very scarce biotite and accessory apatite and ore. The granoblastic feldspar constitutes more than two-thirds of the rock. Quartz occurs in lobate to amoeboid grains of varying size. They may be granulated and show undulating extinction. The biotite is deep yellowish brown.

4094 A. Light coloured to faint purple, medium grained, alaskitic granofels (interbanded with hypersthene-gneiss, see 4094 B), west of Lejrsø.

The rock is very homogeneous although a faint orientation may be seen in the arrangement of larger plate-like quartz grains. The minerals are quartz, microcline, plagioclase and very scarce biotite with allanite.

When seen under the microscope, an extraordinary microdiablastic texture is revealed. This texture seemingly has developed from a granoblastic mosaic, traces of which still may be seen in some relic areas within the slide. Quartz occurs as large lobate to almost amoeboid grains showing undulating extinction and as smaller myrmekitic individuals or blebs. The microcline shows irregular extinction but grid twinning has not been observed. It occurs partly interstitially and partly in diablastic intergrowths with plagioclase. This latter mineral is only faintly twinned and seems to be an oligoclase. Together with quartz, it shows relics of a former granoblastic texture in certain areas. In this position it is antiperthitic. When studied with the aid of the highest magnification a very high content of the finely distributed antiperthite was seen in addition to coarser patches of potash feldspar. In the plagioclase from the diablastic parts, the fine antiperthite is not seen and the coarse type is not common. The potash feldspar seems almost completely rearranged forming diablastic intergrowth with plagioclase. Grains have been observed where coarse antiperthitic texture gradually changes into an almost graphic arrangement of the two feldspar phases across the crystal. One of these grains even showed relic zoning within the plagioclase.

The light coloured biotite gneisses

These rocks form a very distinct type characterised by their finer grain and light grey to almost white colour. The composition is generally granodioritic to granitic. The minerals are oligoclase (20—25 % An), microcline, quartz (rich in tiny inclusions) and a more greenish biotite. Sphene has so far only been found in two of the specimens (19279, see above, and 19238, see below). Apatite and ore are common accessories. This rock type evidently represents the end member of the transitional series of gneisses dealt with here and it is connected to it by a large number of variations.

4041. Light coloured biotite gneiss, 2nd Intermediate Layer, between Sorthat and Ankerbugtelv.

A light coloured, medium to fine grained biotite gneiss which shows parallel arrangement of the scattered biotite flakes and partly lineated quartz. Otherwise, the rock is very homogeneous. The texture of the light minerals is interlobate, hemi-granoblastic and has seemingly been developed by recrystallisation of a cataclastic texture. The minerals are feldspar, quartz, a little biotite and accessory apatite and ore. (The minerals found as tiny inclusions in quartz are not listed as accessories in any of the descriptions). The feldspar comprises oligoclase (ca. 23 $^0/_0$ An) and a somewhat microperthitic microcline. The former predominates slightly over microcline, which forms interstitial to granular grains. In one place it is also seen to replace the plagioclase in a perthitic fashion. The biotite is pleochroic from greenish to black. Quartz is filled with tiny inclusions similar to those described below from 4020.

4020. Light coloured biotite gneiss, 2nd Intermediate Layer, SW of Enehøj.

A light coloured, almost fine grained biotite gneiss with a slight dispersed foliation. The texture is hemi-granoblastic. The minerals are plagioclase, quartz, microcline and biotite together with ore and apatite. The quartz tends towards lobate outlines, but otherwise more granular grains dominate among the light minerals. The plagioclase is an oligoclase (ca. 23 $^0/_0$ An). The interstitial microcline contains a few thin microperthitic strings. Biotite is pleochroic from yellowish green to very dark olive.

Quartz contains several kinds of tiny inclusions. Slender needles with a very high refractive index are thought to be rutile. They occur all over the grains in certain preferred directions. Small prismatic to rounded yellowish grains seemingly consist of zircon. Their relief is higher than that of apatite and they are length-slow. Rounded to elongate grains, which in a few cases show the typical basal section of tourmaline and which are but slightly pleochroic in purple to red colours, are all considered to be tourmaline. They are not very common. Finally, the quartz contains plenty of small irregular inclusions which seem to be of liquid (? or gas bubbles). They are concentrated in the central parts of the grains and are usually absent next to cracks and fissures.

4075. Light coloured biotite gneiss, east of Ankerbugtelv, 1st Intermediate Layer.

The handspecimens show a light-grey to somewhat pinkish, medium to almost fine grained rock. Rod-like light brown quartz grains are macroscopically visible. When viewed with a hand lens or under the microscope, the arrangement of the biotite indicates the existence of an old foliation (S_1) which has been shear-folded into micro-folds by movements along partly recrystallised S-planes (S_2). These latter are only microscopically visible. The S_1-planes are indicated by wavy rows of undeformed biotite flakes. The younger shear planes (S_2) are shown by zones of intense mortar texture and, where these traverse relic cataclasts of quartz, by parallel limonite-coated fractures. Where they continue into preserved larger plagioclase grains, the younger fractures are accompanied by saussuritisation.

The minerals are quartz, microcline, plagioclase, biotite and accessory apatite and ore. A division into an older mineral assemblage corresponding to the development of the S_1-planes and a younger mineral association created along with the formation of the later S_2-planes is evident.

To the older assemblage belong the elongated larger quartz individuals, an antiperthitic to mesoperthitic andesine, and apatite. The apatite occurs in relatively large crystals, which locally are cut by the S_2-planes.

The younger association undoubtedly has been formed by mechanical breaking down and recrystallisation of the older. During the recrystallisation of the "crushed" old highly antiperthitic plagioclase, the potash component separated out to form interstitial individual crystals of grid twinned microcline. The recrystallised plagioclase is an oligoclase. The recrystallisation clearly outlasted the deformation, as the cataclastic texture in most places has been changed into a blasto-cataclastic mosaic where lobate and interlocking grain-borders predominate. At the same time, the biotite seemingly has undergone a reconstitution—judging from its present unsheared habit. It is pleochroic from yellowish to dark greenish brown.

19238. Light coloured biotite gneiss, coast of Gule Hav.

A light grey, medium grained rock of granitic composition. In hand specimen, a dispersed foliation is shown by the parallel arrangement of the small biotite flakes. The minerals are quartz, microcline, oligoclase, biotite, muscovite, epidote (allanite), apatite, sphene and ore. The composition is thus granitic, since microcline dominates over plagioclase. The texture is hemi-granoblastic. Microcline shows cross hatching and forms granoblastic to almost porphyroblastic grains, which latter replace the plagioclase. Myrmekite is rare, but noticeable. The plagioclase is somewhat saussuritised and twinning is only seen on a few grains. It may show more sodic reaction rims towards microcline. The quartz forms large lobate grains with undulating extinction and small bleb-like individuals which seem undeformed. The biotite is strongly pleochroic from yellowish green to very dark grey-brown. Muscovite occurs as stumpy flakes seemingly secondary to biotite. Epidote of a yellow-green colour forms small irregular grains associated with biotite. Sphene is found as rims on the ore grains.

4010 A. Light coloured, slightly banded, biotite gneiss, East of Kosakfjeld.

A rather light coloured, medium grained gneiss with more coarse leucocratic bands alternating with more fine grained biotite-bearing bands. In the section the texture is hemi-granoblastic with lobate to amoeboid outlines. The minerals are antiperthitic plagioclase, quartz, microcline, biotite and accessory allanite and ore. The quartz shows the usual inclusions and undulating extinction. The plagioclase (about 23 $^0/_0$ An) contains rather coarse antiperthitic inclusions. Microcline forms small interstitial grains. Biotite is yellow-green to greenish brown and contains small grains of a yellow-brown allanite.

4010 B. Coarse grained granite pegmatite, associated with specimen 4010 A, described above.

The pink rock contains quartz, microcline and plagioclase. The texture is interlobate, hemi-granoblastic. Quartz shows the same minute inclusions as in the adjacent gneiss. The microcline is somewhat microperthitic. The plagioclase is commonly severely altered, but seems to be an oligoclase with an extinction angle reaching 10°.

4049 A. Light coloured biotite gneiss, west of Strøget.

A fine to medium grained, light grey biotite gneiss of almost quartz-dioritic composition. The foliation is of the dispersed type with scattered biotite flakes arranged parallel. The texture is hemi-granoblastic with slightly lobate outlines of the individual grains. The minerals are plagioclase (22—25 % An), quartz, microcline, biotite and very scarce zircon and ore. The oligoclase is just a little antiperthitic. Quartz tends to develop amoeboid outlines. Microcline occurs interstitially. Some grains show irregular extinction, others grid twinning. The biotite is somewhat greenish yellow to brown.

4049 B. A more quartz-rich, leucocratic type of medium grain.

The texture is characterised by smeared-out thin layers of quartz. The oligoclase-andesine is finely antiperthitic and shows bent twins. The quartz layers seem to have been developed by "squeezing out" of former large, elongated amoeboid quartz grains into granulated bands with prominent undulating extinction.

19302. Light coloured biotite gneiss, 1st Intermediate Layer, Ørenæs.

The medium to almost fine grained light grey rock is relatively well foliated and lineated with rod-like quartz grains. The texture is cataclastic to blasto-cataclastic with relic islands of medium grain size. Mortar-textured seams and zones are parallel to the rows of undeformed lepidoblastic biotite. The following minerals have been observed: quartz, plagioclase (partly developed as mesoperthite), microcline and biotite plus accessory apatite and ore.

A mosaic of irregular to lobate quartz grains indicates an original larger grain size. The blasto-cataclastic small quartz grains have recrystallised into interlocking or bleb-like grains. Plagioclase is developed both as mesoperthite and seemingly younger non-perthitic grains which may border on microcline. The interstitial microcline seems to be in optical continuation with the potassic component of the mesoperthite. In thel atter mode of occurrence, the potash feldspar occasionally shows the typical grid twinning of microcline. Where microcline occurs in the more fine grained blasto-cataclastic seams, it looses its interstitial habit. In a single case, it was seen that the potash feldspar strings in a grain of mesoperthite were bent and showed wavy extinction. The deformation giving rise to the crushing of the rock may thus have outlasted or been later than the formation of the mesoperthite. The biotite is yellowish to dark greenish brown. Apatite is found in a few large idioblastic crystals and as small rounded grains.

Transitional types related to the light coloured biotite gneiss

Specimens 35220 and 35221 show great analogies in the histories of their textural development. They undoubtedly represent two different evolutional stages of what was once one and the same gneiss. On this assumption, the study of the two specimens brings forward some highly interesting aspects as to the origin of the microtextures in the feldspars.

35820. Banded biotite gneiss, NW coast of Qaersup ilua, 1 metre east of the Interior Pyribolite.

The gneiss is made up of more medium grained, purple to pink, leucocratic bands which alternate with more fine grained grey bands containing biotite. The thickness of the layers varies from half a cm to about 2 cm. The reddish colour of the most coarse grained band is seen under the microscope to be due to finely distributed pigmentation of the feldspars with iron oxide. The two different components of the gneiss will be dealt with separately in the following.

The medium grained rock shows a composite texture. It consists of plagioclase, quartz, some microcline and small amounts of biotite, ore and apatite. With uncrossed nicols large lobate to amoeboid grains of quartz with an average length of 5 mm are seen. With crossed nicols each of these grains is seen to be made up of many smaller grains with mutual sutured to lobate contacts. These small grains also show undulating extinction. Quartz occurs moreover as small lobate grains or blebs in blasto-cataclastic areas between larger plagioclase grains. We thus have evidence of an original more coarse grained texture, which has been reduced in grain size by cataclasis followed by recrystallisation. The plagioclase is slightly antiperthitic (ca. 23 % An). Microcline is interstitial in its development and occurs preferably in the blasto-cataclastic seams and patches.

The fine grained rock shows a hemi-granoblastic texture with curved to lobate grain outlines. The arrangement of the lepidoblastic biotite seems to indicate a shearfolding of an old S-surface parallel to the banding. Some biotites show the direction of the younger shear planes which cut the banding obliquely. In a few preserved biotite arches no strain shadows can be seen in the biotite flakes. The development of the shear planes and the following reconstitution of the biotite may correspond to the granulation and partial healing (sutured textures) of the large quartz grains described above from the neighbouring band.

The minerals of the fine grained band are plagioclase (ca. 20 % An), quartz, brownish biotite and accessory ore and apatite. The microcline is interstitial and shows irregular extinction or well developed grid twinning.

35821. Banded biotite gneiss, 7 m east of 35820, NW coast of Qaersup ilua.

A banded gneiss with light coloured grey thin bands and pink to reddish bands. As in sp. 35220 the reddish colour is caused by staining of the feldspar. The two types of bands can also be separated by their grain size, the light grey being the most fine grained.

Texturally the two components are very similar to those just described from specimen 35220. Traces of shearfolding are seen in the more fine grained bands and granulated formerly large quartz grains are found in the more coarse grained bands. Mineralogically, there are however some pronounced differences between the two specimens, 35220 and 35221.

In 35221, the most coarse grained (pink) rock consists of quartz, plagioclase, microcline and very few dark minerals (biotite and ore). The microcline evidently replaces plagioclase and is found as patchy areas within the larger grains of this latter mineral. It also forms irregular grains and more granular aggregates. A few microperthitic strings may be observed in the microcline.

The most fine grained band consists of feldspar, quartz, biotite and accessory apatite and ore. The feldspar comprises microcline and oligoclase (ca. 25 % An). In places the two minerals occur independently, but microperthitic mixtures dominate most larger feldspar grains. All transitions from a plagioclase with a high content of antiperthite, via mesoperthite to microcline microperthite may be seen. In some of the more or less mesoperthitic grains, the outer part of the grain is almost exclusively made up of plagioclase, in others of microcline. In bulk potassic feldspar seems to dominate over plagioclase.

In both specimens, the quartz is rich in inclusions (rutile, zircon, tourmaline, liquid inclusions). The fine grained part of sp. 35220 shows, however, quartz which is poorer in inclusions.

Since the appearance of the tiny inclusions in the quartz seemingly accompanies the other changes described from the transition from the purple gneiss into the light coloured biotite gneiss, specimen 35220 undoubtedly represents an earlier stage than sp. 35221 in this transition. When the similarity in the sequences in textural development of the two specimens is kept in mind, it may be concluded that within the most fine grained bands the mesoperthite has been formed by replacement of a primary plagioclase. The latter in this case did not show an antiperthitic texture which could provide preferred "paths of invasion" during the entry of the potash feldspar.

Granulites sensu stricto

In composition the granulites bear some analogies to the alaskitic rock (see, for example, sp. 4094 A, described above) in carrying mesoperthite as an important feldspar, but they are distinguished from the alaskitic rocks by the presence of red-brown garnet and the Plättung texture.

Among the various samples collected, two which were less foliated were chosen for slicing because they were regarded as the most likely to supply information on the origin of this rock type.

4037 and 19140. Almost granoblastic variations of granulite, Langø.

White to pinkish rocks of fine to medium grain and consisting of feldspar, quartz, scattered garnets and small quantities of biotite, sillimanite and ore. In specimen 19140, the quartz is lineated, while in sp. 4037 bands (1—2 mm thick) of quartz have been developed so that a true Plättung texture is approached. The garnets may also be dragged out parallel to the quartz bands in this rock. In the thin section, the texture is dominated by the extremely lobate, inequidimensional quartz and feldspar. The feldspars are a highly microperthitic microcline (mesoperthite) and diablastic to normal antiperthitic plagioclase (25—27 % An). All microtextural transitions from pure oligoclase to pure microcline are in fact met with. The garnet is faint pink and may be surrounded by a corona of greenish biotite which occasionally contains slender needles of sillimanite.

A specimen of the typical granulite was also sectioned, and it will be described below.

19264, A. B. and C. Granulite, Langø.

A white to yellowish fine grained rock with sparsely scattered deep red garnets (1 mm to 1 cm). The plate-like to rod-like quartz is grey-brown, and glimmering sericite and sillimanite may be seen on cleavage surfaces. The minerals are feldspars, quartz, some garnet, biotite, muscovite, sillimanite and ore. The feldspar comprises microcline, oligoclase (ca. An 20 %), and mesoperthite. Relics of a coarse diablastic intergrowth between oligoclase and microcline are also found. Microcline is somewhat microperthitic and is for the most part interstitial although more granoblastic in the most fine grained bands. Orientated needles of sillimanite occur associated with "smears" of muscovite, which is partly replacing a greenish biotite. Coronas of green biotite are observed around the pink garnets.

In this specimen the Plättung texture with banding due to different grain size, elongated quartz etc., is typically developed. It has evidently been produced by movements causing a grinding down of the original coarser grain and a kinetic-metamorphic differentiation into monomineralic (quartz) bands. The movement has also lead to a partial "splitting up" of the mixed feldspar into two separated phases, the oligoclase and the microcline.

Regarding the feldspars, the granulitic rocks show great analogies to the light coloured biotite gneisses.

B. The intermediate division

This division includes several dioritic and some anorthositic (andesinitic) rocks. Quantitatively, the dioritic rocks are not particularly important, but they form a very characteristic member of the rock association at Tovqussaq. Moreover from a chronological point of view they are a very important group of rocks. The various occurrences will therefore be described and discussed in detail later on, while here only some general comments on their petrography are given. The diorite occurrence at Dioritnæs has been made the subject of a more detailed petrographic study, the results of which, however, will be presented best in continuity with the general description of the occurrence (see page 125).

The dioritic rocks at Tovqussaq all seem to have carried pyroxenes, but in places they have suffered from retrograde alteration, whereby hornblende or biotite diorites have been produced. These alterations are very similar to those observed within the gneisses (hypersthene gneiss to purple gneiss). The following brief description of the diorites, therefore, deals mainly with their primary mineral assemblage.

The pyroxene diorites are usually medium grained rocks which show greenish grey or brownish colours. Their composition is generally rather leucocratic. The well twinned plagioclase shows an An content varying from 30 to 45 $^0/_0$. It may be slightly antiperthitic, but whether this is a primary feature or not is uncertain. The pyroxenes usually include both orthopyroxene and clinopyroxene. The orthopyroxene seems to be a hypersthene (19293: 36 $^0/_0$ Fs and 14971: 33—37 $^0/_0$ Fs, acc. to RI determinations by H. Micheelsen). A brown biotite may be primary in some of the diorites, but biotite developed during the retrograde alteration also occurs. The hornblende, which usually is uralitic, seems to vary in composition from a common hornblende to Mg-hastingsite (see fig. 4). A pale greenish blue actinolitic amphibole may locally replace the more strongly coloured hornblende. The accessories include apatite, zircon and ore. Quartz, when present, seems to have originated from retrograde metamorphism.

19137. Leucocratic biotite diorite. Conformable diorite layer within the western part of the Great Pyribolite, north coast of Langø.

A medium to almost coarse grained grey greenish rock with a foliation. The texture is slightly interlobate, hemi-granoblastic, but mortar seams traverse the slide. The minerals are plagioclase (25—27 $^0/_0$ An), a little biotite, relics of hypersthene and accessory apatite, ore and zircon. The biotite is red brownish and seems to have formed at the expense of a hypersthene.

19248. Hypersthene-bearing diorite. Western part of 2nd Intermediate Layer, north coast of the Påkitsoq fjord.

A greenish grey, medium grained rock with a poorly developed foliation. The texture is hemi-granoblastic. The minerals are andesine, hypersthene, hornblende, biotite, quartz, very little potash feldspar and accessory apatite, ore and zircon. A little interstitial calcite may also be noticed. The pleochroic pinkish hypersthene, which shows rounded to lobate outlines, is partly altered into a pale green hornblende or into brownish biotite. The quartz seems always associated with such hornblende or biotite.

19141. Pyroxene-bearing diorite, west of Granulitsø, Langø.

A greenish brown, medium grained rock with slight lineation. The texture is saccharoidal to interlobate, hemi-granoblastic. The minerals are faintly antiperthitic plagioclase (33—39 $^0/_0$ An), brownish green hornblende, hypersthene, brown biotite, pale green diopside, a little quartz, apatite and ore. Interstitial calcite may also occur.

19145. Pyroxene-bearing diorite, south of Nordnor, Langø.

A medium grained grey to reddish brown rock with a prominent planar and a less well developed linear orientation of the mafics. The reddish tinge is due to the flesh-coloured feldspar. The texture is crystalloblastic. The minerals are plagioclase, hornblende, hypersthene, diopside, biotite and accessory zircon and ore. The hemi-granoblastic plagioclase is an andesine (42—45 $^0/_0$ An) which in the larger grains may show a faint normal zoning. The hornblende is pleochroic from light yellowish green (X) to brownish green (Y) and grass green (Z). It is generally densely packed with minute needles and inclusions of ore (schiller texture). This hornblende seems to have been formed by replacement of a faint greenish diopside. The pinkish hypersthene forms xenoblastic grains partly rimmed by diopside or is found as partly digested relics within larger grains of diopside. Where diopside with relics of ortho-pyroxene has been uralitised, the hornblende in direct contact with hypersthene is bluer than usual. The biotite is pleochroic from yellow-brown to deep brown or nearly black. It occurs in separate lepidoblasts and as replacement products of green hornblende. All mafics show preferred orientation with their c axis more or less parallel to the macroscopically visible lineation. The only accessories noticed in the three sections are zircon (enclosed by hornblende) and ore.

4042 and 13416. Hornblende diorite. Kosakfjeld, north of Inderhavnen.

A medium grained, yellow-brown to red-brown, lineated rock which contains the following minerals: plagioclase (sp. 4042: 36—38 $^0/_0$ An, sp. 13416: 40—45 $^0/_0$ An), hornblende, biotite and accessory apatite and ore. The texture is granoblastic, but the individual grains often show mortar seams or slightly sutured outlines. There are two sorts of amphibole, the normal brownish green hornblende and a pale greenish blue type which judging from its smaller extinction angle (about 16°) may be actino-litic. The latter is seen to replace the more deeply coloured variety. The biotite is olive greenish and, like the hornblende, forms quite irregular grains.

19655. Biotite diorite cut by a granitic aplite, island west of Gule Hav.

The hand specimen shows a medium to coarse grained yellow brownish diorite which is spotted with aggregates of biotite. It is cut (with a clean and sharp contact) by a fine to medium grained granitic aplite of light grey colour. Seen under the microscope, the diorite is hemi-granoblastic with partly recrystallised mortar seams between the grains. It is leucocratic and contains plagioclase, biotite and quartz. No accessories were noticed within the slide. The faintly zoned plagioclase shows an

An content from 30 to 40 %. The biotite forms aggregate flakes which show diablastic intergrowth with quartz, and small scattered scales. It is pleochroic from yellowish to green. Quartz may also be found in a few separate grains. The aplite shows an aplitic texture with studded microcline porphyroblasts. Thin mortar seams may be seen between the grains. The minerals are quartz, microcline, plagioclase, biotite and accessory apatite, zircon and ore. The plagioclase is an antiperthitic oligoclase, which in contrast to the plagioclase of the diorite shows albite—Carlsbad twins. The microcline is grid twinned and contains rounded blebs of quartz. The biotite is pleochroic from yellow-green to greenish brown, and forms irregular, bent and strained flakes. Apatite and zircon are both associated with the biotite. The contact is sharp and clearly indicated by the abrupt change in texture. Where the microcline of the aplite borders on the plagioclase of the diorite, myrmekite is developed close to the contact. Moreover, the normally andesinic plagioclase of the diorite becomes slightly more sodic towards the contact, and contains, within a zone parallel to this, antiperthitic inclusions of potash feldspar. This zone with antiperthite is developed in grains bordering the entire length of the contact and is most probably formed by replacement of plagioclase by potash feldspar introduced from the aplite. Following the same line of reasoning the antiperthitic plagioclase of the aplite may also have been formed by replacement processes during introduction of the microcline. Carried to the extreme, the conclusion could be drawn that the aplite originally was a dioritic aplite, which has suffered from an intense granitisation.

19144. Hypersthene-bearing diorite aplite, isthmus east of Nordnor, Langø.

A fine grained greenish grey rock with a faint lineation produced by the dark minerals. The texture appears almost wholly saccharoidal eu-granoblastic. The minerals are plagioclase (31—34 % An), olive green hornblende, hypersthene, diopside, a little biotite and apatite and ore. The xenoblastic hornblende often corrodes or includes the pyroxenes.

19146. Biotite diorite aplite, south of Nordnor, Langø.

A fine grained rather dense looking dark grey rock. The texture is cataclastic, but has obviously developed from an original saccharoidal eu-granoblastic texture. The minerals are plagioclase, biotite, quartz and accessory apatite and ore. The plagioclase is a calcic oligoclase which shows irregular extinction and bent twin lamellae. The greenish brown biotite may also be contorted. The quartz occurs as small granules together with the biotite in between the larger plagioclase cataclasts, whereby a sort of flaser texture is developed.

The dioritic rocks from the Dioritnæs occurrence are described on page 130.

C. The basic division

This division comprises two major groups of rocks, the pyribolites and the gabbro-anorthosites.

The pyribolites

The pyribolites and the associated pyroxene-amphibolites build up the thick marker horizons mentioned earlier. They also form an important component in many of the mixed gneisses. Some of these rocks have been studied previously by H. Sørensen (1953) who described them as hypersthene-amphibolites. According to the classification suggested by the present author, most of them should now be called pyribolites.

The limits of the hornblende: pyroxene ratio allowed for in pyribolite are rarely exceeded in any rocks of this group. The relative proportion of mafics to feldspar, however, varies rather from specimen to specimen —or from band to band within a single specimen—but in general plagioclase constitutes from a quarter to a half of the rock. The pyribolites and the pyroxene amphibolites are medium to fine grained, greenish to brownish, dark rocks, which at first glance appear structureless, but most of them reveal preferred orientation (generally a lineation) when studied more closely. Banding is seen in some types, whereas other types are more homogeneous. The minerals are ortho- and clinopyroxene, plagioclase and hornblende, scarce biotite and accessory apatite and ore. Garnet is a rare constituent.

The twinned plagioclase is generally a labradorite, but may vary from 40—75 % An. Zoning, normal or reverse, occurs in some specimens. The crystals are water-clear.

The hypersthene is pleochroic from pink to greenish. According to H. SØRENSEN, the Fs content varies from 29—50 %, the highest iron content being reached in a rock where a late formation of garnet has taken place.

The diopside is usually faintly greenish. For this mineral H. SØRENSEN has determined 2V : 57,c $\wedge$ Z : 41°, ny : 1,694 in one slide and 2V : 57, ny : 1,694 in another which correspond to about 30 % hedenbergite (TRÖGER, 1956, p. 63). The author has measured c $\wedge$ Z: 42° and 2V: 58 approx. (in 19306), which gives a hedenbergite % about 35.

The hornblende was studied in two samples by H. SØRENSEN, who classified it as an actinolitic hornblende (H. SØRENSEN, 1953, p. 20). Measuring maximum extinction angles, the present author realised that a great variation in the c $\wedge$ Z occurred in different slides. As a closer study of the hornblendes was desirable in order to define more precisely the difference between the hornblendes of the pyribolites and the gabbro-anorthosites, measurements of 2V and c $\wedge$ Z were performed on the universal stage.

The results of this study are shown in fig. 4 where c $\wedge$ Z is plotted against 2V. The variations of these two optical properties for some known hornblende series are also indicated (TRÖGER, 1956). The most striking result is the extreme variation exhibited by different grains within one and the same slide. The fact that three different samples show quite similar variations proves that these are not fortuitous. The data obtained on hornblendes from a gabbro-anorthosite also group themselves in an analogous way. Another striking feature is the extremely small c $\wedge$ Z found in some hornblendes which thus are pseudo-orthorhombic.

All the hornblendes are length-slow and show a normal but slightly varying pleochroism from X: faint yellowish brown, Y: greenish brown,

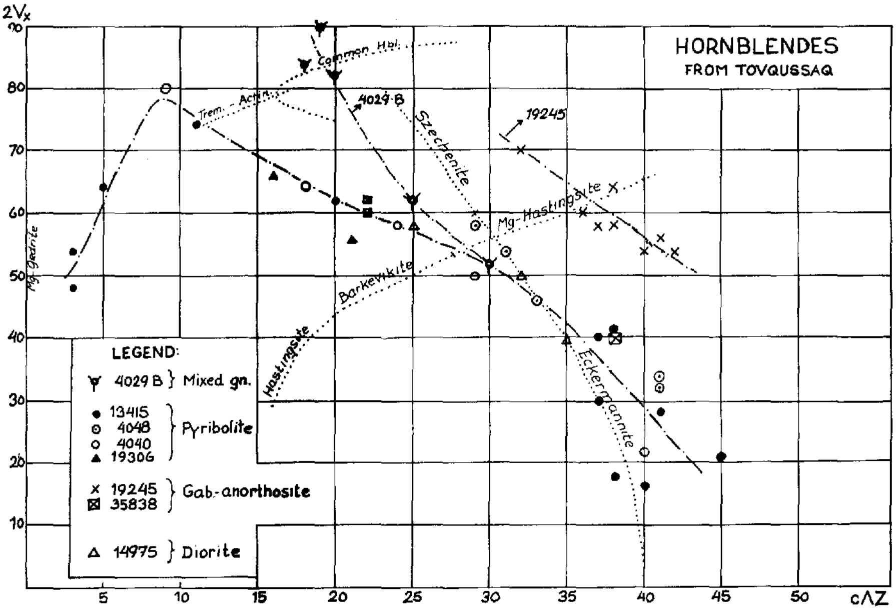

Fig. 4. Diagram showing the observed variation in the optical constants (c ∧ Z and 2V) in the Tovqussaq hornblendes.

Z: brownish green to X: yellowish green, Y: olive and Z: green or almost bluish green. The latter type of pleochroism seems confined to the grains yielding the largest values for c ∧ Z.

Undoubtedly a thorough study of the problem just touched on here would give interesting results, but such a study falls outside the scope to the present work.

The data shown in fig. 4 indicate that the hornblendes of the thick pyribolite layers are rather rich in Mg, since the variational trend-line intersects the Mg-hastingsites, the tremolite end of the actinolite-tremolite series, and since the pseudo-orthorhombic variety approaches values which should be expected from a pure Mg-gedrite. The varieties with larger c ∧ Z but smaller 2V could be interpreted as soda-rich species developed from Mg-hastingsites. This assumption would also explain the local occurrence of bluish rims on the hornblende where the adjacent plagioclase shows inverse zoning, the soda expelled from the plagioclase having been absorbed by the hornblende. Various other changes may naturally also accompany the shift in colours.

The few readings from a pyribolite (4029 B), which occurs as a component of the gneiss, indicate that here there is a transition from more normal hornblendes to Mg-hastingsite.

Biotite occurs only in small quantities in the pyribolites. It is usually brownish to red-brown and forms independent small flakes or has grown on grains of ore.

A chemical analysis of a melanocratic pyribolite from Langø has been published by H. SØRENSEN (1953, p. 45). The specimen analysed contained only 19 % plagioclase and cannot be regarded as representative of the pyribolites. In the following descriptions, some textural relations are dealt with. The thick marker horizons are given special names (see plate 2).

4016. The Interior Pyribolite, north of Navlen.

A dark greenish brown, fine to medium grained, slightly banded rock with a saccharoidal, almost hemi-granoblastic texture. The minerals are plagioclase, diopside, hornblende, hypersthene and accessory biotite, apatite and ore. According to H. SØRENSEN the An content of the plagioclase is 40—45 %. The diopside is faint greenish, while hypersthene is pleochroic from pale pink to greenish (Fs 38 % according to H. SØRENSEN). The brownish green hornblende shows concave borders against the pyroxenes. A few small grains of a red-brown biotite have grown on some ore grains.

4048. The Great Pyribolite, west of Ankerbugtdalen.

A slightly banded dark greenish medium to fine grained rock with an almost nematoblastic texture, due to a prominent lineation of the hornblende. The minerals are hornblende, plagioclase (about 50 % An), diopside, hypersthene and ore. Both pyroxenes appear as rather rounded grains, while hornblende tends to form elongated prismatic grains. The banding is caused by variations in the content of plagioclase and the predominance of one or the other of the pyroxenes. The hypersthene contains 29 % Fs approx. (H. SØRENSEN) and thus strictly speaking should be called a bronzite.

4040. The Pas Pyribolite, east of Ankerbugtdalen.

A dark, fine to medium grained rock. The minerals are plagioclase (53 % An, according to H. Sørensen), greenish diopside, hornblende, hypersthene and some apatite and ore. The pink to greenish hypersthene may form poikiloblastic grains, which only are detectable because the separate parts extinguish as a unit. H. SØRENSEN (1953, p. 20) studying the same slide, concluded that the hypersthene was the last-formed mineral.

19247. The Great Pyribolite, small point on the north-west coast of Pâkitsoq.

The dark greenish rock is fine to medium grained (1 mm—5 mm) with some more coarse grained bands. It is finely banded and shows a well developed lineation. Under the microscope the texture is seen to be granoblastic to nematoblastic. The minerals are plagioclase, hypersthene, diopside, hornblende and accessory ore. The hypersthene may be almost 1 cm across with irregular outlines and inclusions of saccharoidal plagioclase. The pleochroism is hardly noticeable. The plagioclase is a labradorite (ca. 55 % An). The yellow-green to brownish green hornblende seems concentrated in certain bands where it is associated with small granular hypersthene grains. Diopside is transparent and also seems to contribute to the banding by its uneven distribution in the rock.

19299. The Great Pyribolite, south coast of Pâkitsoq fjord.

A dark greenish-grey, rather fine grained rock with a weak lineation. The texture is saccharoidal granoblastic to somewhat nematoblastic. The minerals are plagioclase (about 70 % An), pleochroic hypersthene, hornblende and faint greenish diopside besides accessory apatite. In some grains, the plagioclase shows reverse zoning.

19273. The Pâkitsoq Pyribolite, south of the Eqaluk fjord.

A dark grey, lineated fine grained rock with a granoblastic to nematoblastic texture. The minerals are plagioclase (ca. 60 % An), diopside, hornblende, hypersthene and the accessories apatite and ore. The poikiloblastic diopside does not disturb the general granoblastic mosaic. In a few cases, hypersthene also seems to form large more spongy grains since disconnected neighbouring grains extinguish together. The plagioclase may show reverse zoning.

13415. Mela-pyribolite, from the Pâkitsoq Pyribolite, south coast Eqaluk, collected by H. Sørensen.

The specimen represents a very dark green, fine grained band. The minerals found in the thin section are hornblende, diopside, hypersthene and small quantities of plagioclase. This latter mineral, however, occurs more abundantly in other parts of the layer. The texture is granoblastic to poikiloblastic due to the occurrence of up to 5 mm large spongy hypersthene grains which include granular hornblende, plagioclase and diopside. The hornblende shows concave outlines against the pyroxenes. The pleochroism and the optical data of the hornblende have been dealt with above (see also fig. 4). The An content of the plagioclase is 45—50 % according to H. Sørensen, who also found that the pink hypersthene contains 42 % Fs. (H. Sørensen, 1953, p. 19—20).

13431 B. The Great Pyribolite, east coast, northern Langø, collected by H. Sørensen.

A banded rock, with brownish to black, fine grained layers separated by more feldspar-rich medium grained bands. The minerals of the dark layers are hornblende (40 %), hypersthene (22 %), plagioclase (19 %), diopside (18 %) and ore (1 %), according to H. Sørensen. The texture is granoblastic to almost nematoblastic, due to preferred orientation of the hornblende. There is also a slight tendency to an elongation of aggregates of saccharoidal plagioclase and the stumpy prismatic diopside. The hornblende shows the usual concave border against pyroxene. A few larger xenoblasts may surround diopside and hypersthene. The pleochroic hypersthene contains 33 % Fs, the faint greenish diopside about 30 % hedenbergite and the normal zoned plagioclase 58—65 % An, (all determinations by H. Sørensen).

19306. Pyribolite layer, Tugdlerúnarssuit, south of Langø.

A dark greenish medium to almost fine grained rock with slight lineation. In a section at right angles to the lineation, the texture appears saccharoidal granoblastic. The minerals are pale greenish diopside, plagioclase (ca. 65 % An), hornblende and pinkish hypersthene besides the accessories ore and apatite. A little secondary calcite was also noted. The diopside shows $c \wedge Z$: 42° and 2 V: 58, approx. which according to Tröger would indicate a hedenbergite content of about 35 %.

The following description deals with a pyribolite which forms part of the mixed gneiss. The associated gneiss is described above under the hypersthene gneisses (4029 A and C).

4029 B. Pyribolite, from mixed gneiss at the top of the Tovqussaq Mt.

A greenish grey, fine to medium grained rock with a distinct foliation. The texture is granoblastic to nematoblastic. The minerals are plagioclase, hornblende, diopside, hypersthene and accessory apatite and ore. The plagioclase is a labradorite-bytownite which shows irregular zoning. Among the pyroxenes, which predominate over hornblende, the faint greenish diopside is far the most common. Ore coatings on the cleavage planes of hornblende were noticed. The hornblende shows nematoblastic development with concave borders against pyroxene.

The following sample, which previously has been described by H. SØRENSEN as a hypersthene amphibolite, retains this name according to the new system of classification, since in this rock, hornblende is sensibly more prominent than pyroxene.

13418. Garnet-bearing hypersthene amphibolite, north-east of Atangmik collected by H SØRENSEN.

The texture is granoblastic to porphyroblastic. The minerals are plagioclase (55—70 % An), hornblende, garnet, hypersthene, biotite and ore. The plagioclase shows strong reverse zoning. The xenoblastic hornblende showed a maximum extinction angle of 41° when measured under the ordinary microscope (no basal sections suitable for universal stage work are found in the slide). The large extinction angle (nearly c $\wedge$ Z) and a relatively small 2V justifies the assumption that this hornblende corresponds to the sodic Mg-hastingsites found in other slides. The garnet is pink and completely isotropic. Hypersthene is pleochroic from pink to greenish and was estimated as 50 % Fs by H. SØRENSEN. The pale yellowish to dark red-brown biotite occurs in bent flakes with irregular extinction. Garnet contains idioblastic biotite flakes and small grains of hypersthene. Larger hypersthene grains may show bent cleavage planes. In this slide, the hornblende exhibits convex outlines against hypersthene.

The gabbro-anorthosites

The term gabbro-anorthosite is used here to designate a group of rocks which consist essentially of calcic plagioclase and some mafics (diopside, hornblende, biotite) together with accessory apatite and ore. The field characters of these rocks will be described later on; here it should only be mentioned that the gabbro-anorthosites show great variation in composition and structure. The plagioclase content may vary from 50 to 100 %, but usually the mafics constitute about 10—30 % of the rock. Most gabbro-anorthosites are medium to fine grained and show a typical blue-grey colour due to the feldspar.

The An content of the plagioclase generally varies between 45 and 85 %, and great variations may be found within one and the same slide. Zoning is also common. It may be normal, reversed or quite irregular in relation to the grain outlines. Twinning is generally better developed than in the plagioclase of the pyribolites.

In some thin sections, the plagioclase grains, when studied under the highest magnification exhibit a peculiar almost submicroscopic texture which to some degree is reminiscent of very fine spindle or hair perthite texture. The extreme small size of the spindles and their abundance makes their identification by the aid of the microscope impossible. Possibly this unusual texture is caused by unmixing of a very calcic and less calcic plagioclase. Similar textures have also been observed in some pyribolitic rocks, where the plagioclase shows strong zoning. It is hoped that this very interesting mineralogical problem, although now left aside since it falls outside the scope of the present paper, may be treated properly in the near future.

The diopside is colourless to faint greenish and may be completely replaced by hornblende. In slide 35838, diopside shows $c \wedge Z$: 41° and 2V: 56 (ca. 20 % hedenbergite).

The hornblende shows some unusual optical properties, which may help to distinguish the gabbro-anorthosites from plagioclase-rich diopside-amphibolites. The hornblendes of the gabbro-anorthosites are generally pleochroic from X: yellowish (faint brownish) to Y: green and Z: bluish green or greenish blue. In sp. 19245, which does not carry diopside, $c \wedge Z$ varies from 32—42° and the 2Vx from 70—54. In sp. 35838, which carries diopside and uralitic hornblende; the latter showed $c \wedge Z$ from 22–38° and 2Vx from 62–40; the smallest $c \wedge Z$ being obtained from a uralitic rim on diopside. Hornblende is usually nematoblastically developed.

Biotite occurs in slender flakes, which are pleochroic from yellowish to olive brown or dark brown.

Apatite is scarce, but forms relatively large grains. The ore is generally associated with the hornblende.

Scapolite, epidote and sericite are rather common secondary minerals in the gabbro-anorthosites. Scapolite may also in a few instances be regarded as a "primary" constituent.

The following descriptions contain more information about these rocks.

19262. Gabbro-anorthosite, east of the Pas Pyribolite, Gl. Lejrskar.

A dark medium to almost fine grained rock with a prominent foliation and stripes and lenticles of pure anorthosite. Under the microscope the texture is seen to be saccharoidal to nematoblastic. The minerals are plagioclase (about 50 %), hornblende, biotite and accessory apatite and ore. The plagioclase is well twinned, and may show zoning with cores reaching 70 % An, which value may again be reached in the outermost parts of the grain. In general the composition varies between 55—70 % An. In some grains, secondary sericite is found in the central parts of the individuals. The bluish green hornblende is nematoblastic. The yellowish to olive green biotite is mostly arranged parallel, but more haphazardly orientated flakes are also seen. Only a single, but large, grain of apatite was seen. Ore is associated with the hornblende.

4097. Gabbro-anorthosite, Verdens Begyndelse, north of Hestenæs.

A blue-grey rock with some banding. The grain size is about 1 mm. The texture is saccharoidal eu-granoblastic. The minerals are plagioclase (ca. 95 %), blue-green hornblende, apatite and ore. In addition the following secondary products occur; scapolite, epidote, sericite and biotite. The plagioclase contains ca. 47 % An and may locally show traces of a fine antiperthitic texture. The scapolitisation has followed certain bands, where diablastic intergrowth between scapolite and plagioclase may be seen. Here, scapolite is also associated with epidote, which latter, together with an orange brown biotite, seems to replace hornblende. Sericite occurs as tiny scales in the central part of several plagioclase grains.

19245. Gabbro-anorthosite, north-east coast Qaersup ilua, east of Interior Pyribolite.

A medium to fine grained rock with a flaser-like foliation, which in the hand-specimen is seen to be cut off discordantly by the surrounding quartz-rich purple gneiss. The minerals are plagioclase (about 80 % of the rock), hornblende and biotite. It is the hornblende which causes the foliation. The very pale brown to darker olive biotite occurs as slender randomly orientated flakes. The plagioclase, which is a zoned labradorite (68—54 % An), shows a saccharoidal hemi-granoblastic texture. The hornblende is pleochroic from pale greenish brown to bluish green. It may show a thin rim of greenish biotite.

19234 A and B. Gabbro-anorthosite, north-east of Kangeq.

A medium to fine grained rock with a sort of augen structure, where the "augen" are formed by aggregates of fine grained plagioclase. The mafics, hornblende, some diopside and scarce biotite, constitute about one third of the rock. The texture is dominated by the saccharoidal hemi-granoblastic plagioclase, which is a sodic bytownite (70—82 % An) with more calcic cores or rims. The bluish green hornblende is somewhat xenoblastic and replaces the faint greenish diopside, which latter is found as smaller irregular to rounded grains often surrounded by hornblende. A few flakes of yellow-brown biotite and ore grains are associated with hornblende. Apatite is rare.

4050. Gabbro-anorthosite, south of Strøget.

The handspecimen shows a medium to fine grained blue-grey to purple rock with schlieren of dark minerals between drawn-out lenses consisting of granular feldspar. The minerals are plagioclase (over 90 % in handspecimen), diopside, hornblende, biotite, ore and secondary scapolite. The plagioclase is mainly labradorite but due to zoning the composition varies from calcic andesine to calcic bytownite. Scarce inclusions of antiperthite were noticed. The diopside is somewhat greenish and occurs in irregular to corroded grains partly replaced by hornblende. The hornblende, which is more deeply coloured than usual, in its turn is in one place seen to be replaced by yellowish to dark brown biotite, which otherwise forms small randomly orientated flakes. Scapolite enters the eu-granoblastic texture (of the plagioclase) to a small extent, but may also be secondary since it occurs in radial aggregates on the walls of cracks and fissures in the rock. A single amoeboid grain of quartz was seen enclosed by granoblastic scapolite.

35838. Gabbro-anorthosite, north of Lejrsø.

This rock is now medium grained consisting of plagioclase and mafics. The plagioclase forms granular aggregates up to 5—10 cm long which from their shape and distribution can be seen to be former megacrysts densely packed in a mafic matrix. The plagioclase megacrysts are bluish grey with lighter rims. The grain size within the granulated plagioclase crystals varies from 1 to 2 mm, while the mafic parts are slightly more coarse grained. Under the microscope they are seen to contain small scattered grains of hornblende and diopside. The texture is saccharoidal and almost eu-granoblastic. The well twinned plagioclase shows strong zoning, which may be reversed, normal or irregular. The extinction angle in sections $\perp$ (010) and (001) varies from 39—43 (An 68—76 %). H. Micheelsen determined the An content as 57 % from refractive indices measurements, but the sample used may have come from the peripheral part of a relic large crystal, while the section is cut from the central part of such a crystal. The hornblende is pleochroic from X; greenish yellow to Y: dark green and Z: bluish green (for $c \wedge Z$ and 2V consult fig. 4). It evidently formed at the expense of the greenish faintly pleochroic diopside, which may contain schiller inclusions. A few ore grains are also found in the saccharoidal plagioclases.

D. The ultrabasic division

The petrography of several hornblende-bearing bronzitites from Langø has been described in great detail by H. SØRENSEN (1953). Since then, a great variety of ultrabasic rock types have been found on the Tovqussaq peninsula. The present author has placed this material at the disposal of Mr. SØRENSEN, who in the future hopes to work it up. In order not to interfere with this research, only a few types of ultrabasic rocks are described here merely to give an impression of the variation displayed by the rocks of this division. Only one occurrence will be described in more detail. At this locality eclogitic rocks are intimately associated with more "ordinary" ultrabasites.

Following JOHANNSEN's classification, the ultrabasic rocks at Tovqussaq comprise olivinites, peridotites and pyroxenites. Descriptions are given below of representatives from all these three families. As indicated in the legend of the geological map, dunitic types may also occur, but they have not so far been identified in the rock collection. When not otherwise specified, the samples originate from the central parts of ultrabasic "bodies" (of varying size) found within the thick layers of pyribolite described above.

Olivinitic to pyroxenitic types

a. Olivinites

19281. Hornblende-bearing bronzite-olivinite, east coast of the western peninsula, Kangerdlugssuit.

A fine grained yellow-green to greyish rock with a pronounced lineation of the darker green hornblende. Olivine (2V ca. 90, i. e. about 10—15 % Fa) makes up slightly more than the half of the rock. The other minerals are a faintly pleochroic bronzite (2V slightly less than 90, opt. neg.) and a weakly pleochroic, pale greenish hornblende (max. ext. angle 25°, 2V nearly 90) in addition to considerable amounts of ore and small quantities of a dark green spinel. The texture of the olivine is granular, but the hornblende is nematoblastically developed. More or less parallel antigorite fissures intersect the lineation obliquely.

b. Peridotites

19311. Hornblende-bearing lherzolite, Great Pyribolite, NW of Ankerbugt.

A yellow-green to grey, almost fine grained rock with a faint lineation. The texture is granular to nematoblastic. The minerals are olivine (approx. 30 %), transparent to faint greenish hornblende, colourless diopside, weakly pleochroic orthopyroxene and ore.

19263. Saxonite, Little Pyribolite, SE of Tovqussaq Mt.

A very fine grained dark yellowish green rock of massive appearance. The minerals are orthopyroxene, olivine, secondary antigorite, and a few grains of a pale brownish amphibole and ore. The antigorite, which makes up nearly half of the rock, replaces olivine and orthopyroxene along cracks and partings. The olivine shows 2V 90 approx. (i. e. 10—15 % Fa).

c. Pyroxenites

19258. Hornblende-bearing bronzitite, Little Pyribolite, Gl. Lejrskar.

An almost fine grained dark brownish rock with a eu-granoblastic texture. The minerals are bronzite (55 %), hornblende (38 %), diopside (2 %), olivine (1 %) and ore (4 %), (percentages only estimated). Bronzite (negative 2V nearly 90) is somewhat pleochroic and forms granular grains set in a matrix of hornblende which shows concave outlines against the former mineral. The hornblende is pleochroic from pale yellowish brown to pale green. Diopside shows a more transparent bluish green tinge.

Associated eclogitic rocks

About half a km north of Stjernesø (central Tovqussap nunâ), a large ultrabasic body, which occurs in a layer of pyribolite, is cut into two halves by a NW-striking fault. In the southern part of this occurrence an interesting profile can be studied, fig. 5. The yellow-green olivine-bearing ultrabasite is seen here to underly a fine grained dark green mafic rock, which in the field was designated diopsidfels. The two rocks were obviously folded together. The fold axis shows the same orientation as the axes measured in the surrounding rocks. Within the underlying ultrabasite, a marked lineation is developed parallel to the fold axis.

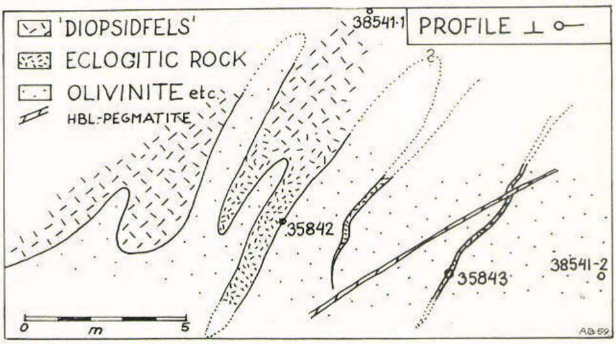

Fig. 5. Schematic profile through an ultrabasic body and associated eclogitic rocks. The numbers show the location of the samples described.

Where the overlying diopsidfels has been folded down into the ultrabasic mass as tightly squeezed synclines, the diopsidfels changes into an eclogitic rock. This passage seems to be gradual. The narrow eclogitic synclines could have been taken for dykes or veins, if the fold pattern had not been recognised.

The greater part of the ultrabasic mass is made up of yellow-brown to yellow-green olivinite. The microscopic characteristics of a sample collected a few metres to the west of the profile of fig. 5 are as follows:

35841/2. Amphibole-olivinite. The minerals are olivine, which constitutes about two thirds of the rock, and an almost colourless hornblende. The granular olivine shows a 2V 90 approx. (10—15 % Fa). The very pale grey to pale brownish grey hornblende is nematoblastically developed and shows idioblastic basal sections. The max. ext. angle reaches 32° and 2V is fairly large (—). No accessories were seen in the slide, which, however, is traversed by thin antigorite-filled fissures.

Where bordering on the eclogitic rocks, the olivinite changes into a pyroxenite of medium grain and green colour. This variety shows the following microscopical features.

35843. Pyroxenitic border facies against eclogitic rock.

The granoblastic textured rock consists of enstatite, diopside and accessory apatite. The enstatite is very faint brownish grey, shows a rather small 2V over Z (= c), is opt. pos. and has a RI less than diopside. This latter mineral shows a moderate 2V and is opt. neg. Apatite is found as rather large, usually rounded grains.

35841/1. Diopsidfels and eclogitic rocks (35842 and 35843).

The dark green, fine grained diopsidfels, which overlies the olivinitic mass, consists, from a rough estimate, of 60 % diopside, 25 % orthopyroxene, 7 % hornblende and 7 % plagioclase, besides which there are very small quantities of ore. The texture is hemi-granoblastic. Diopside is faintly greyish green and a few grains contain small ore grains in the central parts of the crystals. Judging from the moderate 2V (—), the orthopyroxene seems to be a hypersthene. It is weakly pleochroic from almost transparent to pinkish. The hornblende is pleochroic from yellow green to rather deep brownish green. The plagioclase (approx. 45 % An) occurs mainly as small irregular interstitial grains and patches.

The paragenesis of this rock seems to indicate that it is alien to the olivinitic rock described above, but that it may be genetically related to the pyribolites described earlier under the basic division. The hypersthenic composition of the orthopyroxene, the deep pleochroism of the hornblende and the presence of plagioclase are all features which may support this idea. In consequence, the diopsidfels may be classified as a hornblende-bearing melanocratic pyriclasite, which is directly related to the surrounding pyribolite.

In the field, the eclogitic rocks stand out as narrow dyke-like bands with a rather light colour on the weathered surface. Garnet and diopside are the two main constituents. The colour of the garnet varies from deep red to yellow-brown, while diopside usually is apple-greenish. In one specimen, the contact between "eclogite" and pyroxenite is sharp, in another a gradual passage takes place within a few centimetres. The garnet may be finely granular or form up to 2 cm large poikiloblasts. The most yellow-brown garnet was found in a specimen which contains scattered grains of interstitial clean calcite.

Under the microscope, the eclogitic rocks (35842 and 35843) are seen to be composed of garnet, diopside, smaller quantities of an epidote mineral, sphene and also apatite and ore. The larger diopside grains (2—3 mm) appear somewhat corroded in shape whereas the smaller (1 mm) are hypidioblastic and may show crystal faces. The garnet forms granular aggregates with a grain size just below 1 mm. Rather large but irregular grains of sphene occur scattered throughout the slides. They are pleochroic from pink to red-brown. Smaller irregular grains of epidote are found in garnet or diopside. The diopside contains about 5—10 % hedenbergite according to RI and 2V determinations most kindly carried out by Mr. H. Micheelsen ($n_y = 1{,}678$, $n_x = 1{,}696$ ($\pm$ 0,001), 2V average $56\frac{1}{2}° \pm 1\frac{1}{2}$).
Close to the border against the pyroxenite (sp. 35843), a few small grains of plagioclase may also be seen and the epidote mineral becomes more abundant. In the transition rock, enstatite is also found enclosed by the garnet aggregates, in addition

to diopside. Epidote is also found in the pyroxenite immediately bordering on the eclogitic rock.

As long as the composition of the garnets in the eclogitic rocks has not been determined, an exact classification of this rock cannot be made. It can only be said that the field relations point towards a metamorphic origin of the rock and that, in spite of its close association with the olivinitic rocks, it cannot be compared with the eclogitic bands in the olivine rocks in Norway described by ESKOSLA (1921, pp. 51—54). The eclogitic rocks from Tovqussaq, however, show great similarities to some transformed amphibolitic inclusions of eclogitic appearance in peridotite, described by H. SØRENSEN from Northern Norway (H. SØRENSEN, 1955 a, p. 93 and 1955 b).

E. Quartzites and calc-silicate rocks

Siliceous types

Pure quartzites are not known from Tovqussaq, but some calc-silicate-bearing types, which generally carry disseminated ore minerals, have been recorded. Mineralogically, these rocks are closely related to the more pure calc-silicate or skarn rocks found.

The quartzitic rocks are generally interbanded with the pyribolite and pyroxene-amphibolitic rocks described above.

18226. Calc-silicate-bearing quartzite, from a rust zone in the western part of the Great Pyribolite, north coast Langø. Collected by H. SØRENSEN.

A grey to greenish grey, fine grained rock with a rust coloured surface due to weathering of disseminated pyrrhotite and pyrite. The minerals are quartz (very predominant), garnet, diopside, scapolite, calcite, plagioclase, antigorite, phlogopite and accessory apatite, zircon and ore. The rock shows an almost true Plättung texture with bands from a mm to half a cm in size due to variations in texture and composition. Quartz forms hemi-granoblastic layers, where the grains show sutured contacts, and thin bands with elongated plate-like grains. Undulating extinction is prominent. The remaining minerals form parallel seams. Aggregates of calcite are seen to be replaced by phlogopite, garnet and diopside, and seemingly also scapolite. The orange-brown garnets form irregular intergrowths with diopside, and may partly rim this latter mineral or replace it along cracks. Occasionally a thin rim of garnet surrounds scapolite in contact with calcite. In this case a narrow corona of antigorite is found on garnet where adjacent to calcite. Antigorite is also seen to replace diopside along cracks and in other places it may have been formed at the expense of phlogopite. The diopside is bluish green and somewhat pleochroic. Schiller inclusions are prominent. The plagioclase is a labradorite, which generally is replaced by scapolite. Diablastic intergrowth between these two minerals may be seen. The scapolite shows a fine (100) cleavage, which in many cases is parallel to the banding of the rock. Most scapolite grains are very rich in tiny needle-shaped inclusions which are arranged parallel to the cleavage. The centres of the grains contain these inclusions in so great an amount that the scapolite becomes bluish grey. Zircon occurs in a few rounded grains which are uniaxial positive; this makes it possible to distinguish this brownish mineral from sphene.

19652. Siliceous calc-silicate with magnetite, west of Trehøje.

A dark greenish, fine to medium grained rock with foliation and a diffuse banding. The texture is "gneissic" with irregular elongated magnetite grains, and a less prominent preferred orientation of the diopside and quartz. The minerals are quartz, magnetite and diopside, (diopside and magnetite in nearly equal proportions and together constituting just over half of the rock), plagioclase and scarce apatite. Quartz shows undulating extinction. The diopside is transparent to faint greenish. Zoning has been noticed, with a rather hedenbergitic core and a diopsidic rim ($c \wedge Z$: 46° and 40° respectively). Plagioclase, with a refractive index well above quartz, occurs in a few small grains and as narrow rims around the magnetite. A few small rounded grains of a peculiar pale brown apatite may also be seen. They show a refractive index below diopside and are uniaxial negative.

Calc-silicate and skarn rocks

Pure calc-silicate rocks are found 1) as individual layers, 2) associated with the pyribolites described above, and 3) as enclaves in the gneiss. They consist of diopside, calcic plagioclase, garnet, scapolite and occasional quartz. Relic calcite may also occur. The diopside is strongly coloured in bluish green tinges and filled with schiller inclusions. The orange-brown colour of the garnet and the paragenesis suggest that it is rich in Ca. Where scapolite is primary, it is filled with tiny needles, as in sp. 18226 described above. Diablastic reaction textures are prominent in some of these rocks.

18221. Finely banded calc-silicate granofels, forming an independent layer in the gneiss, northern Langø; collected by H. Sørensen.

A finely banded, medium to fine grained greenish grey dark rock with a saccharoidal granoblastic texture. The grain size varies from layer to layer according to the relative proportions of the dark and light minerals. The rock contains plagioclase (ca. 70 % An), diopside, scapolite, and sphene together with apatite and ore. Some bands consist almost exclusively of granoblastic diopside, others mainly of plagioclase with small diopside grains in an almost reticulate arrangement. In one particular layer a concentration of small rounded to irregular grains of sphene is found. Bluish grey, inclusion-rich granular scapolite occurs side by side with plagioclase in one part of the section. The diopside is green to bluish green and filled with schiller inclusions. The $c \wedge Z$ varies from 44 to 48°, and 2V from 60 to 64, which correspond to a hedenbergite content of 70—100 %.

The two samples described below originate from skarn enclaves in gneiss.

35858. Skarn enclave in gneiss, NE of Enehøj.

A medium to fine grained rock consisting of plagioclase, garnet, diopside, scapolite and quartz (the latter mineral is only seen in the handspecimen). Garnet and plagioclase form a diablastic intergrowth where the garnet possibly is poikiloblastic. The well twinned basic plagioclase (about An 70 %) is granular and seems to have formed by granulation of former larger grains, as neighbouring grains in some areas show almost similar orientation. The garnet is red-brown in section and the diopside is pleochroic from yellow-green to bluish green. This strongly coloured diopside contains also fine schiller inclusions. Diopside and scapolite seem to be associated and do not enter the diablastic intergrowth illustrated by the plagioclase and the garnet. The scapolite contains plenty of tiny inclusions which seemingly cause its grey colour.

In the hand specimen, the quartz is seen to be associated with the dark coloured diopside.

35832. Skarn lens in gneiss, south of Great Pyribolite, east coast of Qagssiarssuk.

The large handspecimen shows a medium to very coarse grained rock, which contains dark bluish grey plagioclase, deep red-brown garnet, dark green diopside, clear quartz, relic rather dark grey calcite and greasy green scapolite. Quartz, garnet, diopside and scapolite seem associated with the relic calcite. In thin section the texture is seen to be hemi-granoblastic to diablastic. The plagioclase is a bytownite (ca. An 80 %), which is partly scapolitised. Scapolite also forms independent saccharoidal grains, which are clear and free of inclusions. The garnet is yellow to orange-brown, while diopside is bluish green and faintly pleochroic. Quartz occurs in rather large somewhat amoeboid grains with undulate extinction. Where adjacent to scapolite, the quartz is fringed by a thin rim of garnet or diopside. The garnet shows sieve texture where replacing plagioclase.

F. Various schists

Biotite-pyriclasite schists and their derivates

Among the various schistose rocks found, a particular type, pyriclasitic schist, deserves a more detailed description. These pyriclasitic rocks generally contain some biotite besides diopside, plagioclase and hypersthene. It is mainly the biotite which causes the schistosity.

This rock type is found as enclaves in gneiss and granulite. In the latter mode of occurrence, the pyriclasites seem to have been enriched in Mg, as may be inferred from the appearance of cummingtonite and cordierite.

Below, a pyriclasite enclosed in a gneiss which also carries enclaves of gabbro-anorthosite, will first be described. Then follows a description of the cummingtonite- and cordierite-bearing types found within the granulite on Langø. These rock types were not mappable on a scale of 1 : 20.000, but they are indicated on the detailed map of Langø, fig. 12.

35845. Diopside-rich schist.

A dark greenish brown schistose rock of medium grain and with lenticles and patches of larger grain and less preferred orientation. The minerals are diopside (about half of the rock), plagioclase, biotite, hypersthene and accessory apatite. Apart from the biotite, which is generally lepidoblastic, and the apatite all the remaining minerals form a diablastic texture. The yellowish to red-brown biotite also enters this texture to some extent, since in places it shows lobate outlines against the pyroxenes. It is then associated with plagioclase, both minerals together forming "pseudopodian protrusions". The plagioclase is a calcic andesine (about 45 % An). Diopside is faint greenish to transparent; the hypersthene is faint pinkish. Apatite forms relatively large rounded grains.

35817. Cummingtonite-cordierite-bearing schist, Sydtangen, Langø.

A rather light coloured greenish grey rock which is medium to almost coarse grained and which exhibits an irregular schistosity. The minerals are labradorite, diopside, hypersthene, cummingtonite, quartz, biotite and cordierite plus accessory garnet and apatite. The texture is lepidoblastic to poikiloblastic. The plagioclase forms well twinned grains which may be somewhat antiperthitic, but the presumed potash feldspar is very hard to distinguish from very small myrmekitic quartz "worms". Quartz occurs in a few larger, elongated grains, as more granular grains, and in myrmekitic intergrowth with plagioclase. Even the larger grains show irregular outlines and give off small pseudopodiae which invade the plagioclase. It is easily imagined that bleb- or worm-like grains of quartz in plagioclase originate from such "offshoots" from below or above the plane of the section. On the whole, the impression is given that quartz is replacing plagioclase. The faintly brownish pink hypersthene forms larger, irregular and corroded grains, which show thin exolution lamellae. It is rimmed by xenoblastic to poikiloblastic diopside or replaced by cummingtonite. This latter mineral is faintly pleochroic in yellowish brown tinges, shows a 2V approaching 90 and is optically positive. Schiller structure and relic cross fractures may be seen in the cummingtonite formed from hypersthene. The lepidoblastic biotite is pleochroic from yellowish to orange-brown. Cordierite shows penetration twins. The garnet is colourless.

35661. Cummingtonite-cordierite-bearing schist, south of Granulitsø, northern Langø.

A very similar, but more lineated and schistose rock consisting of labradorite, biotite, hypersthene, diopside, cummingtonite, quartz, cordierite, a little scapolite and accessory zircon. The quartz shows the same myrmekitic intergrowth with plagioclase as in 35817, and the mafic minerals correspond closely to those described above. Scapolite forms individual granular grains.

Garnet- and/or hypersthene-bearing biotite schists

On Langø some relics of biotite schists which may carry garnet and/or hypersthene are found (see map, fig. 12). As will be seen from the following descriptions these rocks reveal some interesting micro-tectonic features.

37097. Hypersthene–biotite–plagioclase schist (collected by H. Ramberg), Sydtangen, Langø.

In the slide irregular porphyroblasts (1 cm) of hypersthene are seen in addition to plagioclase (ca. 30 $^0/_0$ An), biotite, and accessory zircon. The texture is dominated by the incipient cataclasis, by which all the minerals are affected. The large hypersthene grains are broken up and show irregular extinction. Plagioclase shows bent twin lamellae and mortar textured or sutured grain borders. Large biotite flakes form aggregates with irregular extinction. The zircon is found in the hypersthene and the biotite.

35818. Garnet-biotite schist with black hypersthene, Sydtangen, Langø.

A dark coloured schist with light, thin quartzo–feldspathic veins. Porphyroblasts (from half to 1 cm) of deep red garnet and black hypersthene are seen in the hand-specimen. Under the microscope the section is seen to comprise one part consisting of biotite and garnet and another quartzo-feldspathic with hypersthene and biotite. Accessory zircon is found in both parts.

In the schistose biotite-rich part, contorted and bent lepidoblasts of a faint yellowish to deeper brown biotite form the "groundmass" in which somewhat cataclased porphyroblasts of pinkish grey garnet are set. The biotite is strained except for some small idioblasts which have grown within 'competent' grains of garnet.

In the quartzo-feldspathic part of the section, large grains of a highly pleochroic hypersthene are seen. Its colour changes from intense pink to somewhat bluish green. Although the thin slide is slightly thicker than normal, this type of pleochroism can be said to be unique among the hypersthenes so far seen in the Tovqussaq rocks.

The same applies to its black colour in the handspecimen. The hypersthene has formerly formed larger grains, which now are broken up and separated by seams of biotite. Almost biotitised relics are also found in the smeared-out biotite aggregates. Quartz shows undulating extinction and the plagioclase highly contorted twin lamellae. Felty seams of saussurite may separate the plagioclase grains. Biotite with irregular extinction also occurs between the grains of the felsic minerals. The refractive indices of the plagioclase are equal to or slightly larger than that of the quartz, suggesting an An content of about 40 %.

35816. Partly gneissified mica schist, Sydtangen, Langø.

A foliated to schistose, medium to coarse grained biotite-rich rock with turbid greenish feldspar and transparent blue quartz. The thin section is cut from a feldspar-rich part of the rock and shows plagioclase (antiperthitic oligoclase), quartz and a yellow to brown biotite. The texture is hemi-granoblastic in the quartzo-feldspathic parts while lepidoblastic where biotite predominates. The rock is slightly cataclased with incipient saussuritisation of the feldspar and quartz showing undulating extinction. A few grains of interstitial calcite were noticed.

In the extreme northeastern part of the Tovqussaq peninsula, there occurs a mappable layer of garnet-bearing biotite schists which occasionally carries sillimanite. Such Al-rich rocks are otherwise rare in Tovqussaq. Some of the garnet-biotite bearing rocks can be proved to have been formed by retrograde metamorphism of the pyribolite (cf. sp. 13418 described under the basic division). Most unfortunately, the sample collected from the proper garnet-bearing schists east of Kangerdlugssuit cannot be said to be typical since it is veined by quartzo-feldspathic material and carries hypersthene and not sillimanite. Recalling that the field determination of the sillimanite is beyond any doubt, the author presents the following description of the hypersthene-bearing variety collected.

19283. Garnet-biotite schist, east of Kangerdlugssuit, south coast Eqaluk.

When weathered, a strongly stained, medium grained schistose rock with rust-brown colours. The texture is hemi-granoblastic with bands or seams rich in lepidoblastic biotite and others with elongated, slightly amoeboid quartz grains with prominent undulating extinction. The contorted biotite and the occasional bent twin lamellae in the plagioclase also indicate a postcrystalline deformation. The minerals are sodic andesine (ca. 30—35 % An), quartz, biotite, garnet, a few relics of hypersthene, accessory zircon and ore. The biotite is pleochroic from yellowish to dark red-brown and may contain rounded small grains of zircon surrounded by pleochroic rims. Zircon also occurs enclosed by quartz, feldspar or garnet. The garnet is pale pinkish under the microscope and shows idioblastic (in the smaller grains) to almost poikiloblastic development. The garnet seems to be older than the deformation which influenced the biotite. The hypersthene is strongly pleochroic from greenish to pink and is found in a few relic grains, which are replaced by a more yellow to olive biotite along cracks and cleavages.

IV. DESCRIPTION TO THE GEOLOGICAL MAP

Morphology

Tovqussap nunâ is situated in the southern part of the Sukkertoppen district, West Greenland (see key map fig. 1). It forms a peninsula (ca. 15 × 9 km), which to the north is bordered by the Eqaluk fjord and to the south and south-east by the Angmagssivik-Sangmissoq fjord. On the west coast, which faces the Davis Strait, a natural harbour, Inderhavnen, is found on the lee side of Hestenæsset and Langø. South of Inderhavnen, a broad deep bay, Qaersup ilua, is found. A narrow and crooked fjord, Pâkitsoq, has been cut into the country south of Qaersup ilua. On the south coast of the Tovqussaq peninsula, a Greenlandic settlement, Atangmik, is situated. In 1948 a fishing station was established at Langø. In 1954 it was rebuilt into a whaling station, which, however, was abandoned in 1959.

Tovqussaq is really the name of a 524 m high mountain north of Qaersup ilua, fig. 6. Due to its characteristic form (a broad dome with a small summit cone) it serves as an important landmark for sailors. The northern part of Tovqussap nunâ (which means the Tovqussaq-country) is formed by rather high (400—550 m) mountains and belongs morphologically to the highland north of the Eqaluk fjord. The mountains in northern Tovqussap nunâ are usually smooth and rounded and often dome-shaped. A similar morphological development has also been found within certain areas of the northern Sukkertoppen district. Rather deep-cut and steep-sided valleys separate the more softly sculptured mountains. This feature explains why the northern part of Tovqussap nunâ, particularly when seen from the sea, appears as an important highland in spite of its moderate altitudes.

Morphologically, the southern part of Tovqussap nunâ belongs to a 'strandflat' which is typically developed further to the south, in the archipelago of outer Fiskefjord and along the west coast of Nordlandet, fig. 6.

The region between Atangmik and the Pâkitsoq fjord is characterised by relatively low hills (100—150 m) separated by mainly NW-trending often swampy valleys. The low hills and ridges usually show a very steep SW flank (lee side during the glaciation) and more gently sloping NE faces.

Fig. 6. View towards south and south-east from the 250 m hill south of Sorthat. To the left, the Tovqussaq Mt. with its characteristic summit cone. The land south of Qaersup ilua is seen to form the northern extension of the lowland of Nordlandet (Photo: GGU, A.B.).

Nearly all valleys of the Tovqussaq peninsula are fault-controlled. Thus in the northern part, two NE-striking faults have given rise to two important systems of valleys, the Ankerbugtdal-Lejrsø-Breddal and the Gammel Lejrskar-Stjernesø-Krebsesø systems, which both have served as convenient approaches. Irdalen and the Pâkitsoqelv valley are strike valleys.

Lakes are quite numerous, but never very large.

The mountains of Tovqussap nunâ are generally brownish in colour, which is only partly caused by the rock colours, since these are often subdued by a partial, superficial cover of brown and black lichens. Along the coasts, on the west-facing slopes, and above an altitude of about 200 m, the rocks are generally exceedingly well exposed.

From the end of September to the beginning of June, the country is normally so snow covered that field work is rendered impossible. In some years the ice on the highest lakes does not break up before late in July or the beginning of August.

In order to systematise the description of the geological map, this has been divided into several litho-structural units, which in turn build up the various macro-structures; the latter will be dealt with later on. The litho-structural units comprise the layers and successions established from the mapping.

The description starts within the central parts of the Tovqussaq dome, which has been described already (Berthelsen, 1950), and is gradually extended to the remaining parts of the peninsula.

The core rocks of the Tovqussaq dome

The central part of the Tovqussaq dome structure (Berthelsen, 1950) is made up of a rather homogeneous 'core rock' of greenish brown to purple colour and mainly quartz-dioritic composition. This rock passes gradually into the pink to light coloured, granodioritic to granitic types, which separate the 'core rock' from the enclosing pyribolite layer, the Interior Pyribolite. The poverty of structures within the 'core rock' is obviously due to an advanced homogenisation. This may be observed in the field from the occurrence of more or less nebulitic structures which are preserved as relics in certain places. When seen under favourable light a large scale banding, conformable to the dome structure indicated by the Interior Pyribolite, may also be observed in the mountains north of the deep bay, Qaersup ilua, which has been eroded out in the dome core. On a much smaller scale, the same orientation is shown by traces of an old foliation and, just west of Navlen, by a narrow band of rusty biotite-rich schist (traced over more than 30 m). This band, as well as the foliation, are folded. Within the rather leucocratic core rock, the old foliation has in places been thrown into shear folds. Due to the low plunge of the axes in these shear folds, the outcrop pattern shows very "spitzwinklige Falten". The formation of the microcline-bearing rocks, which surround the central dome core like an onion shell, is most probably related to this shear folding. The following observations support this assumption: 1) microcline schlieren show rodding or lineation parallel to the axes of these shear folds on the small point south of Qaersup ilua, 2) in other places, the granitisation is seen to have out lasted the shear movements and the microclinisation proceeds along irregular mutually intersecting planes. Undeformed biotite-microcline-bearing pegmatites may also be found. On the mentioned point, south of Qaersup ilua, it may also be seen that the granodioritic or granitic types are developed by granitisation of quartz-dioritic types which latter are here found as small relic areas within the pink microclinised rocks.

The shearing seems to have involved preferentially the peripheral parts of the core rock.

Thus it is possible, within the scale of an outcrop, to trace the same phases of movement and transformations as were observed on a microscopic scale. The transition from the pyroxene-bearing core rock (13417, A) into the biotite-bearing purple granofels (13417, B) was characterised by a change towards more amoeboid textures, which seem to indicate that the recrystallisation was accompanied by only feeble intergranular movements. In the samples from the outer part of the dome core (35820 and 35821) the shear folding of the old foliation could be traced within the finer-grained bands. The post-tectonic reconstitution of the biotite was pointed out as well as the introduction of potash feldspar.

A characteristic section within the northern part of the dome core, beginning at the Interior Pyribolite on the NW coast of Qaersup ilua will now be described. Here the Interior Pyribolite is well banded and may contain smaller ultrabasic or calc-silicate inclusions and lenticles. It shows a gradual passage into the light coloured core rock. The dioritic

Fig. 7. Banded granitic gneisses of the core layer north of Qaersup ilua (Photo: GGU, A.B.).

transition zone is about one metre broad and contains up to half a metre large ultrabasic inclusions. The dioritic transition zone, therefore, has possibly been formed at the expense of the amphibolite. Within a distance of about 100 metres from this contact, the core rock is rich in microcline and may be crossed by irregular pink veins and zones of impregnation. Finely banded types are also met with (fig. 7), and nebulitic light shear folding was noticed in other places (fig. 8). Further east, the pinkish rocks pass gradually into the purple granofelses or gneisses, which locally contain greenish brown areas with relic pyroxene.

Apart from biotitised amphibolite bands and a few (up to 4 m broad) lenticles of ultrabasic rocks, no other foreign rock types have been met with in the dome core.

The Interior Pyribolite

This pyribolite circumscribes the core of the Tovqussaq dome. As mentioned above, it is banded and contains ultrabasic inclusions where seen in the coastal section NW of Qaersup ilua. The purple gneiss of the

Fig. 8. Shear folds in the core rocks of the Tovqussaq dome, north coast of Qaersup ilua (Photo: GGU, A.B.).

western Frame Layer passes into a hypersthene-bearing gneiss with abundant pyribolite inclusions of elongated shape close to the basic layer. Further to the north, the Interior Pyribolite thickens due to intense folding on a mesoscopic scale. The folds are open drag folds with NNW plunging axes and eastern 'Vergenz'. In the turn-round (northern hinge zone of the dome) the well banded pyribolite is folded around rather steep NNE plunging axes, causing the strike to change rapidly to E—W. On the NW-facing slopes to the east, it can be seen how the dip decreases with increasing altitude. A prominent rust zone, with intense orange to yellow colours on the weathered surfaces, is also seen here. The ore content is only small. The widening out of the pyribolite on the 200 m spur is determined principally by the low dip. The layer forms here the upper edge of a steep escarpment. Small folds are commonly encountered. They all seem to be drag folds with more or less NE-plunging axes. The direction of relative tectonic transport[1]) is towards SE on a mesoscopic scale. Following the pyribolite further to the SE, it can be observed that the

[1]) Relative tectonic transport in this paper always refers to the direction of movement shown by the upper layers relative to the lower.

axial plunge swings into E and ESE. The mesoscopic folds become almost isoclinal and indicate now a relative tectonic transport towards S and SSW. Similar small folds may even be observed along the coast within the core rocks underlying the pyribolite, which here consists of two layers separated by a narrow gneiss band. Conformable ultrabasic bands may also be noticed in the pyribolite. The lowermost pyribolite layer can be followed NW-wards into the above mentioned escarpment south of the 200 m spur, but it then dies out within the strongly granitised core rocks.

South of Qaersup ilua, the Interior Pyribolite reappears on the northern slopes of Akuliaruserssuaq. To the west, it is strongly folded around SSE plunging axes. Further to the east it becomes partly gneissified and changes into a banded amphibolitic rock.

The Frame Layer of the Tovqussaq dome

The thick succession of more or less banded gneisses and various other rocks which outcrops between the Interior Pyribolite and the Little Pyribolite of the Tovqussaq dome has been given the name "the Frame Layer." It is composed mainly of quartz-dioritic gneisses which display all transitions between common hypersthene-bearing (4029, C and A, 19246, 19243) to purple biotite-bearing types, but light coloured biotite gneisses (4010, A) may also be developed. Garnets have been found locally within all three gneiss types (e.g. 4027). The Frame Layer gneisses are commonly mixed or migmatitic and contain in certain zones abundant basic material as boudins, enclaves, schlieren or bands, which may be in all stages of gneissification. Agmatitic structures are also developed in some horizons, but have only been indicated on the lithological map where very prominent. The basic material comprises mainly pyribolite (4029, B) and its derivates, but boudins and inclusions of gabbro-anorthositic rocks (19245) occur as well and are in fact characteristic for this lithological unit.

On the broad point between Qaersup ilua and Inderhavnen, a pyribolitic band is found on the west coast (partly under water). It is separated from the eastern hypersthene gneiss by a narrow layer of purple gneiss. Further east a broader purple gneiss layer occurs. It passes into a biotite-hypersthene gneiss with thin concordant basic bands and schlieren which show a prominent drag folding. The relative tectonic transport indicated by these folds and by a row of rotated pyribolite boudins is similar to that deduced from the folds in the Interior Pyribolite close by (see above).

North of the hinge zone of the Interior Pyribolite, purple to light coloured gneisses with partly biotitised amphibolitic bands and inclusions are found. Small folds with steeply NNW-plunging axes are common around the remains of some barracks. Boudins of ultrabasite (hornblende-hypersthene-diopside rock) with a slight clockwise rotation have also been noticed within the purple gneiss. The light coloured gneiss immediately NW of the pyribolite contains amphibolitic bands, streaks and inclusions and has obviously been developed from a rock similar to the hypersthene gneiss rich in basic material further to the south. Eastwards the light coloured gneiss becomes more homogeneous and exhibits a prominent sheeting. Towards the SE it passes gradually into the purple gneiss which here overlies the Interior Pyribolite. This purple gneiss occasionally contains gabbro-anorthositic boudins and elongated inclusions, which show a flaser-like foliation.

The purple gneiss layer may also be traced south of Qaersup ilua, here again being separated from the Interior Pyribolite by a layer of hypersthene gneiss (in part dioritic) rich in basic bands and schlieren. The southern parts of Akuliaruserssuaq are wholly built up of the Frame Layer. South of the purple gneiss layer, a fairly well banded mainly hypersthene gneiss follows. It contains up to half a metre thick agmatised bands of more basic material—largely amphibolitic. It is succeeded by a better banded still hypersthene-bearing gneiss with occasional gabbro-anorthositic inclusions. Along the south coast of the peninsula, a slightly banded and generally smallfolded rock is found. North of Sarfaq silardleq partly digested amphibolitic inclusions show recumbent small folds, which indicate a relative tectonic transport towards east, the axes plunging to S. On the eastern shore of the peninsula, an opposite direction of relative transport can be deduced from folds with an amplitude of more than one metre. Gabbro-anorthositic inclusions are not uncommon along this coast. On the north coast of Pâkitsoq, a purple gneiss layer is seen to the east of the smallfolded gneiss. It is separated from the Little Pyribolite by a less broad layer of hypersthene gneiss.

North of Krumryggen and around the Tovqussaq Mt. (524 m) the Frame Layer shows a more complex development and the subdivision into different gneiss layers, which could be carried out in the southern area, is not practicable here. The gneisses are generally very mixed and often agmatitic so strike and dip readings are difficult to obtain or have to be taken from the general orientation over larger areas. The gneiss which encloses the more basic inclusions has often developed a more pegmatitic facies. Between Gammel Lejrskar and Ankerbugtelven the Frame Layer has been exposed to a strong granitisation with the formation of more homogeneous light coloured biotite gneisses. In places the microclinisation is seen to be a rather late event, since it has followed existing, mutually intersecting fractures. Pink microcline-bearing pegmatites may also be developed (4010, B).

The Little Pyribolite

This pyribolite layer separates the Frame Layer from the 1st Intermediate Layer throughout the Tovqussaq dome structure. It has been traced from Ankerbugten north of the Tovqussaq Mt. to Pâkitsoq and Ørenæs, the remaining part of its course being covered by the sea. North of Ankerbugten, where outcropping within a faulted wedge, it is folded around a nearly vertical axis. In the section at Gammel Lejrskar, it is seen to enclose metre broad ultrabasic lenticles (19258). ESE of the Tovqussaq Mt. the Little Pyribolite causes a local but strong magnetic disturbance, but no magnetic minerals could be found in the outcrops. The dioritic dyke close by (to be described separately later on) does not exhibit a similar disturbing effect, and it must be assumed that the pyribolite contains some hidden magnetic bands (compare page 104). SE of the Tovqussaq Mt., saxonitic inclusions (19263) are found in the pyribolite.

On Ørenæs, the structural relations are highly complicated, but apart from a local replacement to form diorite on the NE point of the said peninsula, the Little Pyribolite seems normally developed.

The 1st Intermediate Layer

The 1st Intermediate Layer also forms part of the Tovqussaq dome. It has been traced from Ankerbugten over the northern flank of the Tovqussaq Mt. to Pâkitsoq and the root of Ørenæs. Like the Little Pyribolite, its remaining extension passes below the sea, through Sarfaq silardleq, around Akuliaruserssuaq, east of Langø to Inderhavnen. Along its periphery, the 1st Intermediate Layer is bordered by the Great Pyribolite, which also circumscribes the dome structure, and which, due to its occurrence along the east coast of Langø, allows one to reconstruct the course of the inner layer as outlined above. The 1st Intermediate layer, like the 2nd, is generally developed as a light coloured biotite gneiss, (4075, 19302) with local transitions into purplish types (4018). Its contacts with the enclosing pyribolites are sharp and considerable movement seems to have occurred along the contact planes. At Gammel Lejrskar, a purple facies, with foliated quartz, borders the Great Pyribolite. This "Einregelung" of the quartz would appear to correspond to the shear movements mentioned in the description of sample 4075, where an older plicated foliation and younger movements accompanied by an important recrystallisation could be discerned. The sample from the light biotite gneiss south of Ørenæs (19302) supplied information on a very similar recrystallisation. The accompanying deformation can be directly matched here with shear parallel to the contact of the gneiss layer, since it has given rise to a mesoscopically visible flaser-like splitting up and joining.

The Great Pyribolite

This thick layer of pyribolite circumscribes the Tovqussaq dome with a nearly ideal oval outcrop pattern. The pyribolite (4048, 19247, 19299, 13431) contains intercalations of pyroxene diorite, thin calc-silicate-bearing quartzitic layers (18226) and abundant ultrabasic inclusions of all sizes (and sometimes also thin conformable ultrabasic bands). The pyribolite itself may be banded or quite homogeneous. Calc-silicate schlieren, pockets and bands are not uncommon and have been found in places to contain calcite. In other words, we are concerned more with a pyribolitic "series" than a layer.

Although these intercalations have not been mapped continuously, the available information seems to indicate that the dioritic layers, for example, have a very wide extension within the part of the Great Pyribolite layer exposed around the dome. Along the outer periphery (in relation to the dome) a characteristic diorite layer with elongated inclusions of basic and ultrabasic rocks has been found within the Great Pyribolite 1) on the east coast of Langø, 2) west of Ankerbugten, and 3) in the section in Gammel Lejrskar; it was not observed on the north coast of Pâkitsoq nor south of Ørenæs. In this eastern and southern area, however, a dioritic layer has been found close to the innermost contact of the Great Pyribolite. The distribution of the rust-zones seems much more irregular and less persistent. As regards the ultrabasics the observed material is unfortunately too scarce to allow any detailed description of their distribution throughout the Great Pyribolite to be made. In places, as for example on the eastern slopes of Ankerbugtdalen, a row of inclusions of ultrabasites is seen close to the southern contact of the 1st Intermediate Layer, but a little further to the west (west of Ankerbugtelven) large peridotitic lenses (19311) are more irregularly distributed due to strong folding of the pyribolite around axes with a moderate westerly plunge. The amplitude of the folds is around 100 m. In the cliffs on the west coast of Ankerbugten, more regularly arranged ultrabasic bodies are found again.

Where the Great Pyribolite reoccurs on the north-east coast of Langø, it contains conformable bands (1 m thick approx.) of hornblende-bearing bronzitite. This locality has been described by H. Sørensen in connection with his study of the ultrabasic rocks of Langø (H. Sørensen, 1953). Since then, the present author has had an opportunity to revisit the locality and carry out a more detailed study of the field relations of these ultrabasic rocks. H. Sørensen (1953, p. 8) mentions that the ultrabasic bands can be followed for some 50 metres, but then die out — passing into normal looking amphibolite (now termed pyribolite). However the re-examination brought out that the apparent wedge shape

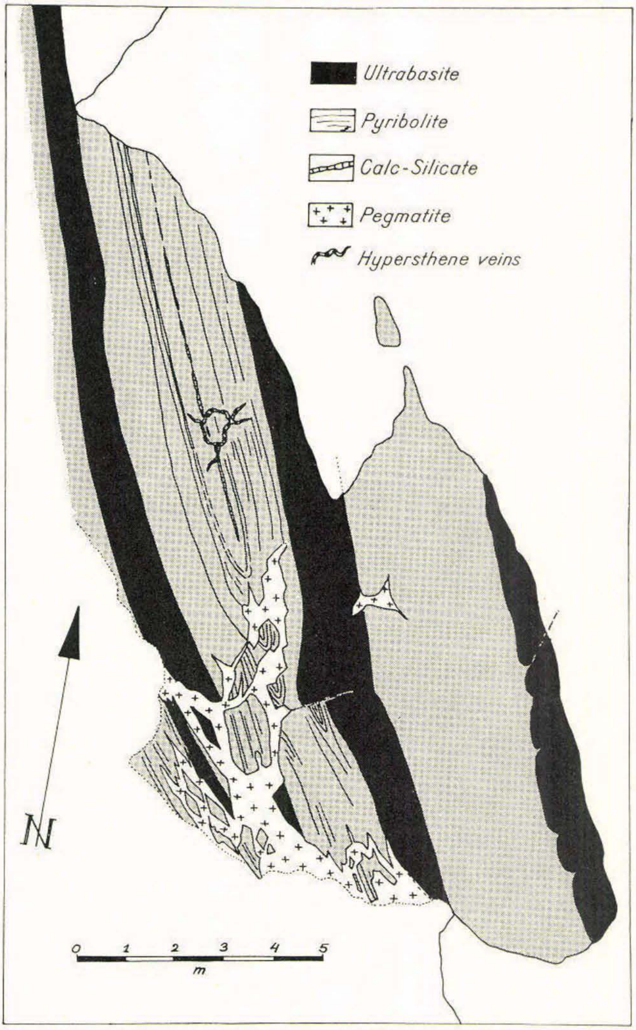

Fig. 9. Sketch map showing isoclinally folded bands and layers of ultrabasic rocks in the Great Pyribolite, north-east coast, Langø.

of the ultrabasic bands is caused by their being folded with the enclosing rocks into tightly compressed isoclinal structures. The hornblende bronzitites and the banded pyribolite with its calc-silicate intercalations all show a remarkable conformity and even very narrow bands may be traced through the fold structures, figs. 9 and 10. The ultrabasic bands show a certain tendency to thicken in the hinge zones and to become thinner or disappear on the sheared flanks (fig. 10). Still more convincing examples of these features will be described on a later page. It may also be noted that the ultrabasic rock in one place (fig. 9) seems to have been more competent than the pyribolite, while at a nearby locality, a synformal core of pyribolite surrounded by ultrabasite shows an initial break-

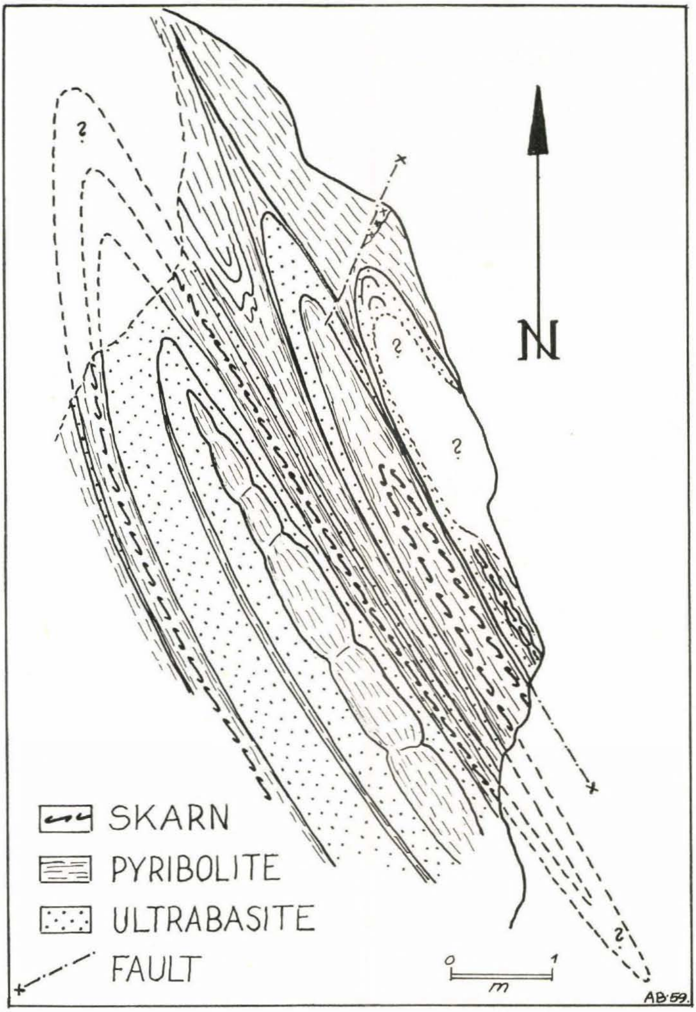

Fig. 10. Sketch map of isoclinal folds in the Great Pyribolite just north of the locality pictured in fig. 9.

ing up into boudins (fig. 10). Small differences in texture or composition may explain this change in relative competence in different parts of seemingly contemporary structures.

Within the banded pyribolite conformable thin pyroxene-plagioclase-bearing pegmatites are not uncommon. Basic segregation pegmatites may also be found in places (fig. 9). Along the east coast of northern Langø, thin cross-cutting biotite-microcline-bearing pegmatites with a steep SSW dip and slickensided contacts are abundant. Close to a small fault, a larger light coloured pegmatite was also seen (fig. 9.) This pegmatite is remarkable since it contains displaced inclusions of the enclosing rocks.

The 2nd Intermediate Layer

The 2nd Intermediate Layer comprises the rocks enclosed between the Great Pyribolite and the Pas Pyribolite. On Langø and north of Qaersup ilua, this division is obvious, while in the southern part of the Tovqussaq dome, the structural relations are more complicated.

In the northern part of the dome, which here is overturned to the north, the 2nd Intermediate Layer is developed mainly as a light coloured biotite gneiss of very uniform appearance (4020, 4041). It encloses a few conformable layers of pyribolite. Around the Ankerbugtdal, some remarkable hinge zones have been mapped out within these basic layers. The axes plunge gently to the east. The northernmost largest hinge zone is cut and displaced by the Ankerbugtdal fault in such a manner that it is exposed now on both sides of the fault. The southernmost hinge zone is seemingly less disturbed by the fault. As mentioned by the author (Berthelsen, 1950), the light coloured biotite gneiss exhibits a very prominent sheeting in this area. The individual sheets are usually from one to over 2 m thick. The joints dip slightly towards the central parts of the gneiss layer, steepening somewhat towards the contacts with the enclosing thick pyribolite layers. A similar sheeting has also been observed north-east of Krumryggen. Further south, the western half of the 2nd Intermediate Layer is developed as a gneissose diorite, while the eastern part retains its granodioritic to granitic composition.

On the western flank of the dome, south of Kolbesø and on Langø, the 2nd Intermediate Layer shows a less uniform development. Intercalations of pyribolite and calc-silicate rocks are found, and on Langø the light coloured gneiss in places gives way to granulite s. s. The region south of Kolbesø is rather poorly exposed. Detailed mapping (fig. 11) has shown, however, that the bordering thick pyribolite layers can be traced continuously south to the coast of Inderhavnen. The intervening gneiss layer contains abundant amphibolitic and pyribolitic bands and carries occasional ultrabasic inclusions. All transitions from hypersthene-bearing via purple to light coloured gneisses are met with. The poverty of outcrops and the many cross-cutting faults make mapping of the minor basic bands impossible.

On Langø, which was mapped also on a scale of about 1 : 4.000, the better exposed intercalations within the 2nd Intermediate Layer can be traced for considerable distances (fig. 12). Here they comprise thin layers of pyribolite, pyroxene amphibolite, calc-silicate rocks and mica-schits. The enclosing gneiss is in part developed as granulite s. s. (4037, 19140, 19264), but purple gneiss, with transitions into either hypersthene-bearing (18219) or banded, pink to light coloured biotite-bearing types (19136), forms the major part of the main layer. North of Nordnoret, these tran-

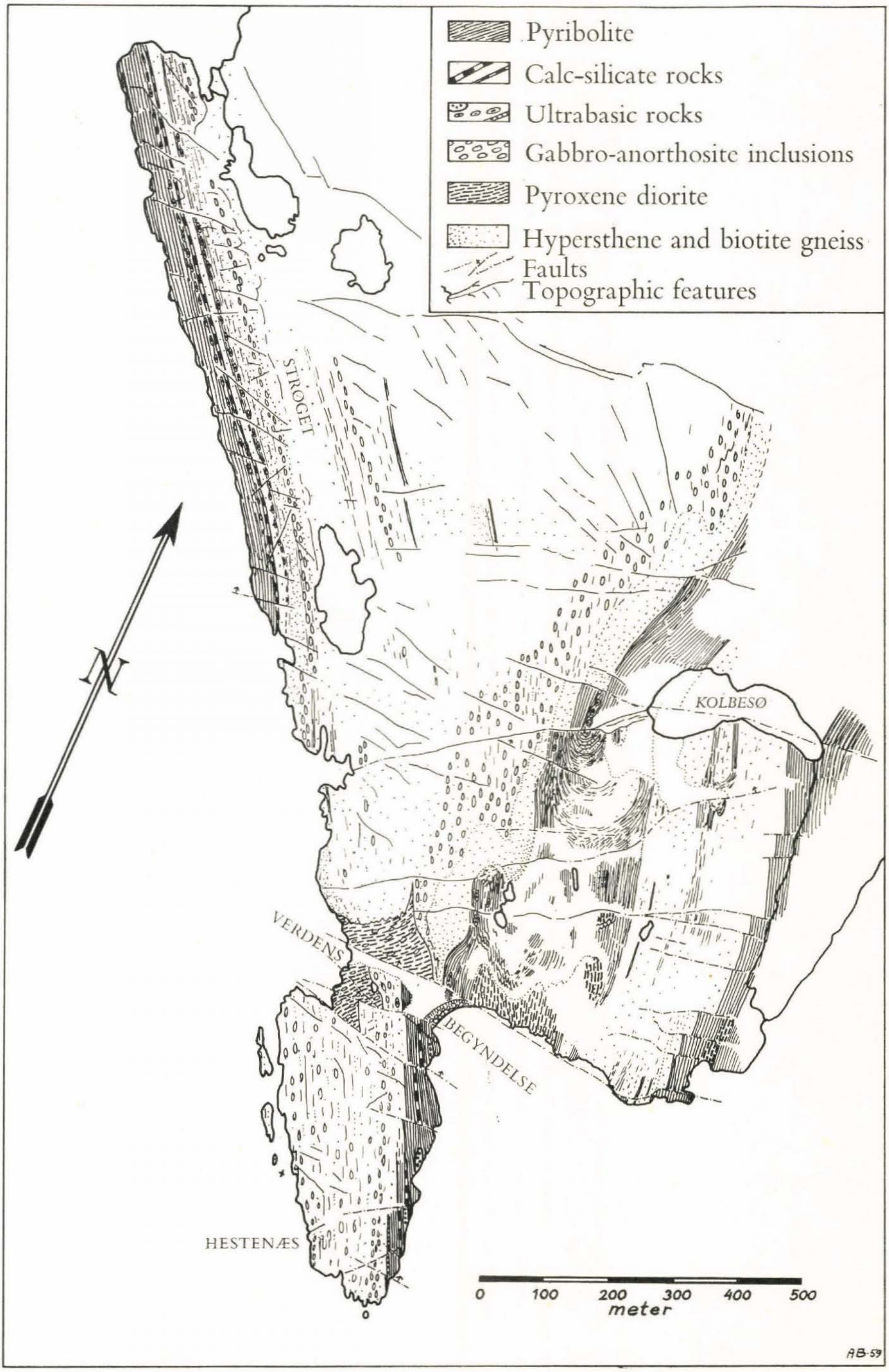

Fig. 11. Detailed geological map of the Hestenæs-Kolbesø area.

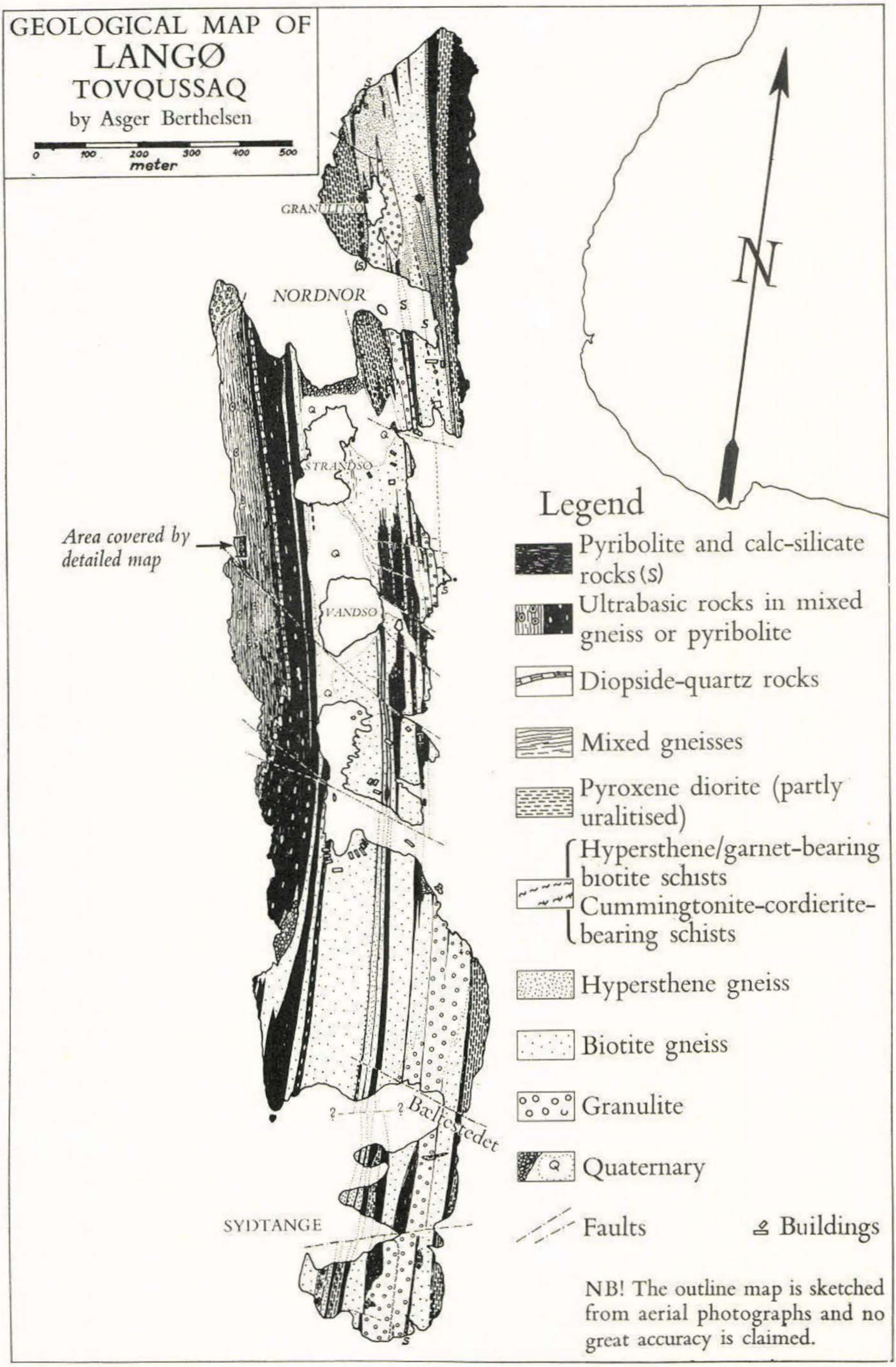

Fig. 12. Detailed geological map of Langø.

Fig. 13. Partly migmatised garnet-biotite schists, Sydtangen, Langø (Photo: GGU, A.B.).

sitions have also been noticed taking place along the strike direction. The granulite shows a similar transgressive development in relation to the strike, although it occurs mainly within a zone east of the many thin pyribolitic layers which extend from Sydtangen to the eastern shores of Vandsøen. North-east of this lake these layers pass into a hinge zone. Taking the universal SSE to SE plunging axes of Langø into account, this would mean that the pyribolitic intercalations occupy an isoclinal synform. Although this structure is not very apparent from the map, its existence has been proved in the field by observations on the small folds. These also bring out that the granulite east of the said structure should correspond to the rather homogeneous purple gneisses south of Vandsøen.

The structural relations on Northern Langø are seemingly still more complicated. The granulite, which is well developed south of Granulitsøen, is succeeded by hypersthene gneiss to the north of the lake due to the interfolding of these two rocks. A similar synformal structure was observed in a disrupted band of calc-silicates, which has been traced west and

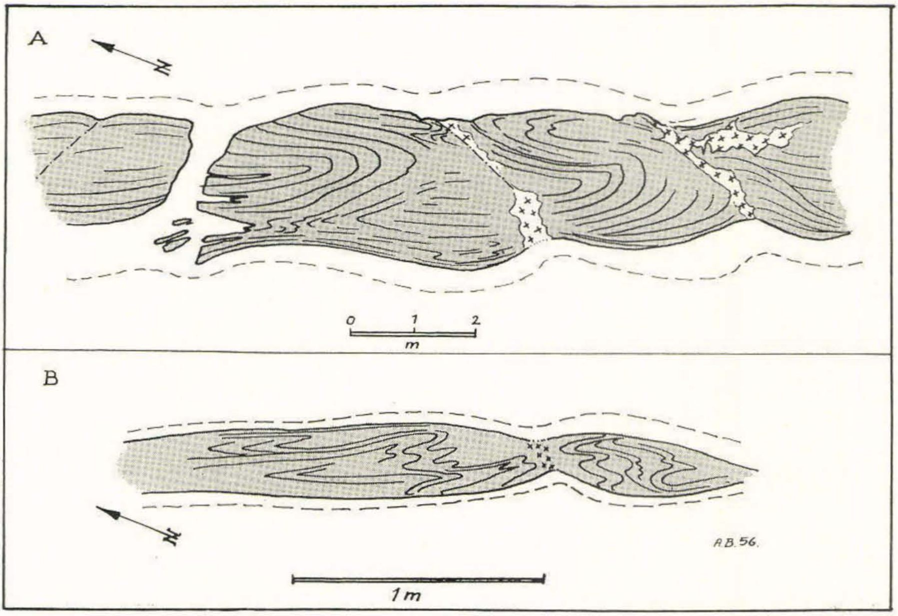

Fig. 14. Boudins of smallfolded calc-silicate rocks, north of Nordnor, Langø.

north of Granulitsø, and which seemingly reoccurs east of the granulite on the northern shore of Nordnoret. When due regard is taken of displacement caused by later faulting, the present synform at Granulitsø can be directly matched with the isoclinal synform shown by the pyribolite bands north-east of Vandsøen. The irregular border of the granulite south-east of Granulitsøen is most probably due to a later refolding of the isoclinal fold. These considerations lead us finally to the conclusion that the dioritic rocks (19141, 19145) around Nordnoret occupy an antiform.

Amongst the various rock types found on Langø, a few deserve special mention here. On Sydtangen, a narrow layer of hypersthene-garnet-bearing biotite schist (37097, 35818, 35816) occurs within the granulite, fig. 13. In places it may be seen how the schist, traced along its foliation, passes into granulite. Rows of garnet indicate the continuation of the schist-foliae. The granulite also contains boudins of pyriclasitic schists, which may be altered into cummingtonite-cordierite-bearing schists (35817, 35661). The boudins, which are about 1 m long, show structures indicative of internal movements within the disrupted layer. Very similar boudinage structure has been described by K. Coe (1959) from some siltstones in West Cork, Ireland.

Although, as described above, the granulite may be seen in places to have been developed by granitisation of mica schists, its origin is far from simple. In other places, garnets have been formed in otherwise normal purple gneisses and even within a dioritic dyke, where this latter cuts the granulite.

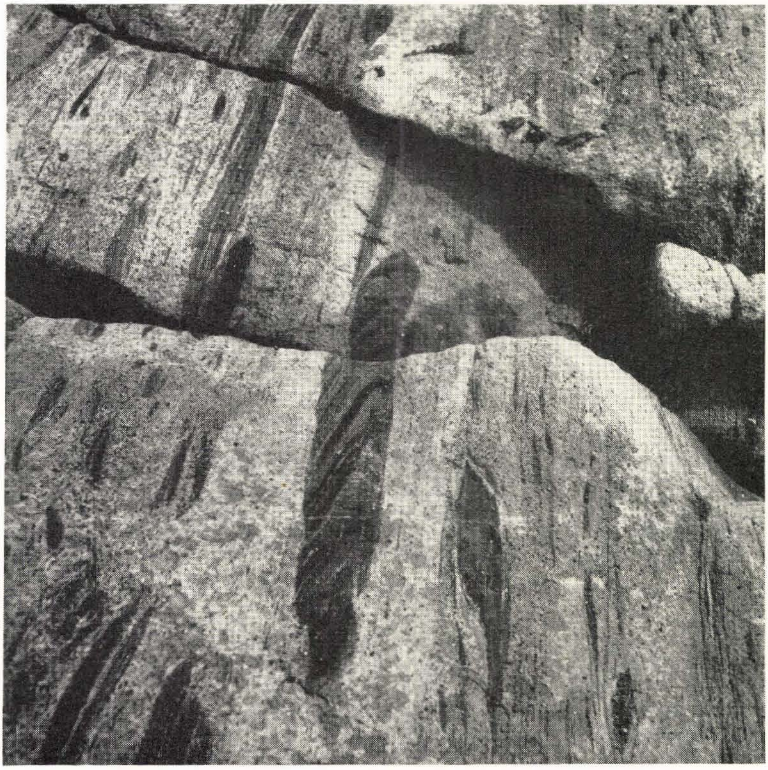

Fig. 15. Basic inclusions in hypersthene gneiss, north-eastern Langø (Photo: GGU, A.B.).

Another rock type which is typically developed on Langø is the banded calc-silicate (18221). This is indicated with the same symbol as the pyribolite, but has in addition been marked with an S on the map fig. 12. These rocks are finely banded and occur either as smallfolded layers or as boudins within gneiss or granulite. North of Nordnoret, the westernmost is strongly smallfolded. The axes of these folds show a SE plunge similar to that observed in the surrounding rocks. South of Nordnoret the same layer has been traced right to the coast east of Strandsø. The easternmost layer on the map fig. 12 forms boudins, which, however, show internal smallfolding. As may be seen from fig. 14 this folding is obviously older than the formation of the boudinage structures. In a thinner band of calc-silicate which is found about one metre further to the east an initial stage in the formation of the boudins can be observed. The smallfolding is here so well preserved that its origin as drag folding can be ascertained. The axes are again parallel to those found in the enclosing rocks. These sketches present definite evidence for the interpretation previously given by the author for this particular structure (Berthelsen, 1957, p. 180 and fig. 6). The petrographic study has left

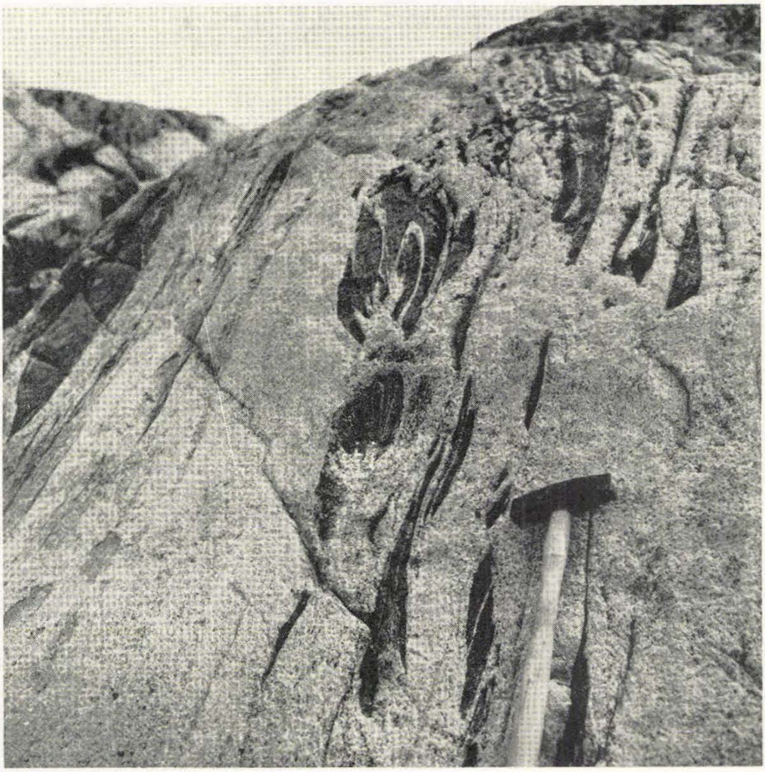

Fig. 16. Basic inclusions in hypersthene gneiss, eastern Langø (Photo: GGU, A.B.).

little doubt that here we are dealing with high grade calc-silicate rocks of sedimentary origin. From the field-observations, we can say with certainty that at an early stage of the metamorphic evolution the original sediment behaved as an incompetent layer and was thrown into drag folds. At a subsequent stage in the metamorphic evolution, the calc-silicate rocks gained in competence relative to the enclosing gneisses and became boudiné. These movements were accompanied by formation of pegmatites. Where the disruption was complete, a sort of flow occurred within the boudins and the asymmetrical style of the original drag folds was more or less destroyed. As seen from fig. 14 A, the breaking up into boudins may have been controlled by diagonal S-planes rather than cross-fractures. Very similar boudins with internal fold structures have been described recently by K. Coe from West Cork, Ireland. In the Irish example, however, the folding is due to slumping within an unconsolidated bed of sandstone, which later on, when lithified and enclosed by siltstone, was boudiné. On Langø a similar mechanism can hardly be invoked, since the folds shown by the calc-silicate rocks are homoaxial with those observed in the surrounding rocks. Here the change in relative competence should probably be explained as the result of the conversion of the original sediment into a calc-silicate rock, or, since the change is relative, by a change in the physical properties of the enclosing rocks.

On the southern part of Sydtangen, a smallfolded calc-silicate layer separates granulite from purple biotite gneiss. From a structural point of view, this layer corresponds to the thick, western layer at Nordnoret.

Fig. 17. Ultrabasic rocks accumulating in small hinge zones, eastern Langø.

Finally, the narrow layer of hypersthene gneiss, which on northern Langø occurs just west of the Great Pyribolite, should be mentioned. It contains rather abundant elongated inclusions and streaks of pyribolitic or ultrabasic material. Smallfolding may be seen in nebulitic structures or within the enclaves. Fig. 15 and 16 show typical examples of the folding in the basic enclaves. The smallfolds are closed or isoclinal and it is particularly their hinge zones which have survived the gneissification. The two larger enclaves in the centre of fig. 15 are thus both relics of a formerly coherent hinge zone. The plastic behaviour of the ultrabasic rocks during this folding—even down to the smallest scale—is shown by fig. 17 where the black ultrabasic band shows thickening in the closures and migration from the flanks of the smallfolds. A quartz vein was at the same time disrupted.

Tugdlerúnarssuit and the archipelago of Qagssiarssuk

The rock association of the 2nd Intermediate Layer on Langø, reappears on eastern Tugdlerúnarssuit, where granulitic rocks are exposed in two areas separated by a layer of pyribolite (19306). On the north coast, this layer contains a median band of diorite, in the same manner as in the pyribolite found on the western parts of Sydtangen, Langø. To the south-east, the whole layer is replaced by rusty diorite. On Tugdlerúnarssuit the garnets of the granulite have often been altered into biotite which forms greenish grey spots in the light coloured rock. Where the granulite has been mylonitised, as on the eastern point, these spots have been dragged out and the rock has a streaky appearance. On the three small islands south of Tugdlerúnarssuit a strongly mylonitic light coloured biotite gneiss with conformable microcline schlieren is found. It contains a few small (20—40 cm) inclusions of calc-silicate and thin grey dykes which are folded but cut by undeformed microcline-bearing pegmatites. This mylonitic gneiss undoubtedly represents a strongly sheared and recrystallised equivalent of the granulite of Tugdlerúnarssuit and Langø. It is structurally enclosed between two layers of pyribolite, a western and a south-eastern. The western one can be traced northwards into the rather thick pyribolite layer which on Tugdlerúnarssuit separates the granulite from the western hypersthene-bearing gneisses with pyribolitic, ultrabasic and gabbro-anorthositic inclusions. This pyribolite layer can be followed north through Langø, Hestenæsset and Kolbesø into the northern thick pyribolite layer of the dome, the Pas Pyribolite. The latter everywhere separates a gabbro-anorthosite-bearing gneiss from the purple to light coloured gneisses and granulites of the 2nd Intermediate Layer.

As indicated by the change in strike within the mylonitic gneiss of the three small islands mentioned, this pyribolite layer continues under sea into the narrow wedge of pyribolite exposed on the east coast of the point south-west of Tugdlerúnarssuit. The hypersthene gneisses of western Tugdlerúnarssuit just reach the same coast. Here they are strongly folded and the smallfolds indicate that they form the closure of a southward-plunging antiform. The wedge shape of the overlying pyribolite seems to be due to its pinching out on the western flank of this structure. The hinge zone has also been recognised within the hypersthene gneisses of western Tugdlerúnarssuit, which will be described later on.

On the point south-west of Tugdlerúnarssuit, a mylonitised light coloured biotite gneiss overlies the partly pinched-out Pas Pyribolite. The gneiss in its turn is overlain by a thick pyribolite layer which forms the cliffs of the south coast of Qagssiarssuk and then curves south towards Blindtarmen lake. Where it runs over the point west of the bay it decreases in thickness and gives way locally to garnet-bearing gneisses.

The author has correlated this pyribolite layer with the Pas Pyribolite, since in this southern region it always separates a mixed, mainly hypersthene-bearing gneiss from rocks which lithologically as well as structurally can be traced directly into the 2nd Intermediate Layer as recognised on eastern Tugdlerúnarssuit. In order to make the Pas Pyribolite appear as the thick pyribolite layer further to the south on the point mentioned a tight synform with a sea-covered hinge zone west of Tugdlerúnarssuit has to be postulated.

The area between Qaĝssiarssuk and Pâkitsoq

This area, south of the Great Pyribolite and north of the Pas Pyribolite, contains some very complex geology. It is dissected by many often swampy valleys which are fault-determined. Retrograde alterations along these faults have caused a bleaching of the various rocks so that their identification in the field is greatly hampered. Shear accompanying the faulting may also have changed or even obliterated the older structures. The original rock assemblage and its structure can only be studied on the exposures in the scattered hills. The representation given in the present map is based largely on one night's work when favourable illumination due to a low sun in the west made possible a tracing of the structures and layers from one hillock to another; at this time all confusing features within the valleys were hidden in shadow. A few hours mapping thus enabled the author to group all the detailed observations made previously into a definite pattern.

In the western part of the area a more or less dioritised or gneissified pyribolite layer (indicated as ordinary pyribolite on the map of pl. 1) can be traced from the small point on the east coast of Qagssiarssuk over the hills first towards north-east and then curving towards north-west. In fact it forms a hinge zone in a partly recumbent structure, since the axis plunges to the south. The core of this structure is made up of a larger body of almost totally uralitised diorite with a few amphibolitic bands or inclusions. A bleached gneiss separates the diorite from the gneissified pyribolite. West of the latter a banded to streaky gneiss, varying from hypersthene-bearing to purple gneiss with discontinuous bands of garnet-biotite-rich gneiss, is exposed. West of Qáqârssúnguit (140 m), this gneiss contains a zone rich in basic material (pyribolite and amphibolite) which here is intensively smallfolded on southward-plunging axes. Towards the north-east this zone fades out, but possibly can be traced into the dioritic rock on the east coast of Qagssiarssuk just south of the Great Pyribolite, where skarn rocks (35832) are found. North and north-east of Qáqârssúnguit, smallfolding is also present in the gneisses which here have developed a strong elongation and lineation parallel to the fold axis, fig. 18. The smallfolding of the lithological banding can actually only be observed on cross joint surfaces while the trace of the

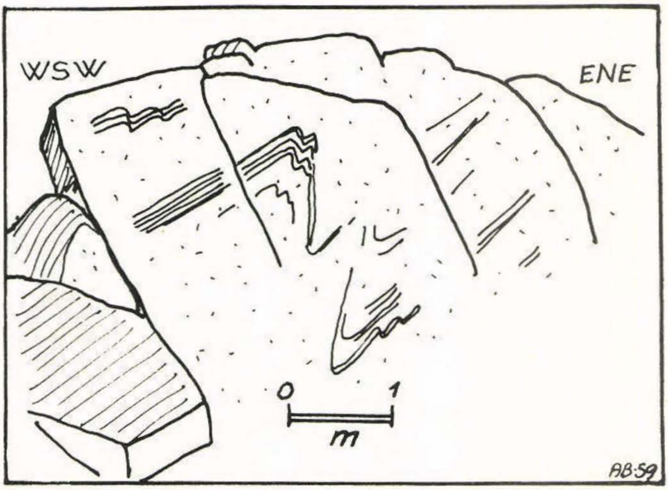

Fig. 18. Asymmetric smallfolds in gneisses (2nd Int. Layer), east of Qagssiarssuk.

elongation on more horizontal exposures simulates that of a nearly vertical foliation.

The recumbent structure just described is undoubtedly directly connected with the antiform found on western Tugdlerúnarssuit and south of this island. The dioritic core can thus be interpreted as having been developed from the median pyribolite of eastern Tugdlerúnarssuit. South-west of Pâkitsoq, an analogous structure is found within the rocks of the 2nd Int. Layer. The pyribolite, which forms the 80 m high hill east of Ørenæs, is thus replaced by diorite along its periphery in the hinge zone. A quartz-dioritic gneiss with small basic and ultrabasic inclusions occupies the core of the structure shown by the pyribolite.

The Pas Pyribolite and its immediate northern surroundings

This layer circumscribes the Tovqussaq dome, but it has a much more complex configuration on the map than any of the other layers so far described. Some of these complications have just been discussed in connection with the description of the lithology and structures of the

2nd Intermediate Layer. The Pas Pyribolite will now be described beginning in the northern part of the dome and continuing anticlockwise around this structure.

Much of what has been said about the lithology of the Great Pyribolite applies also to the Pas Pyribolite, although the latter in several localities shows a somewhat different development. Immediately east of Ankerbugtdalen (4040), it appears as a normal banded pyribolite. On Sorthat (307 m) it is smallfolded close to its northern contact. The axes of these smallfolds trend E—W and are horizontal. The folds indicate a relative tectonic transport towards north. Further to the west the Pas layer appears more homogeneous and slightly more coarse grained. A well developed mineral lineation plunges between 25 and 30° to SSE, and does not lie in the plane of the pyribolite layer (as far as this can be determined from contour construction), but is inclined to this.

Thus it seems that this lineation originated along with a recrystallisation of the pyribolite and its direction would appear to have been controlled by movements which are younger than the formation of the overturned dome. The influence of this younger deformation is also manifest around Kolbesø, where the Pas Pyribolite is strongly folded and in part transformed into diorite. This area was mapped in greater detail in order to find out whether the Pas Pyribolite could really be traced through these complications or not. As may be seen from fig. 11, which shows the outcome of this detailed study, there can be no doubt that the Pas layer extends south on to the east coast of Hestenæsset. The general foliation, however, strikes NW, i.e. obliquely to the trend of the large scale lithological layering. In spite of the rather poor exposures in this area, in some northward-facing sections along the younger faults it could be ascertained that smallfolds developed by differential shear movements along the present foliation planes had been superimposed on older broad folds. Apart from larger lithological variations, local relics of mesoscopic banding help to establish the existence of the older and larger folds within the pyribolite. These structural relations will be discussed at greater length in connection with the geometrical analysis. Due to the changes which took place along with the shear movements, the pyribolite is more or less gneissified, particularly along its western contact. Inclusions of ultrabasic rocks are rather common south of Kolbesø, and a mappable band of calc-silicates has been met with at various localities along the western contact of the pyribolite. On Hestenæsset, this band helps to determine the displacements caused by the younger dextral faults. Within the area shown in fig. 11 the calc-silicate band is bordered by a purple biotite gneiss to the west. This gneiss layer separates the pyribolite and the calc-silicate rocks from a zone rich in gabbro-anorthosite inclusions. This lithological succession helps to

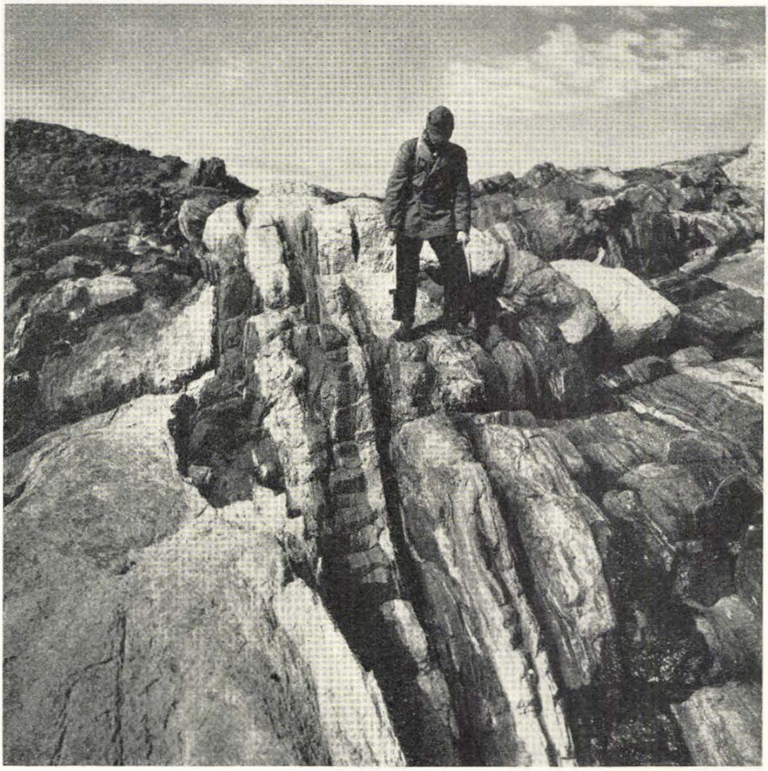

Fig. 19. Banded succession in the Pas Pyribolite, west coast of Langø (Photo: GGU, A.B.).

identify the Pas Pyribolite on the west coast of Langø. West of the 2nd Intermediate Layer there occurs here a rather thick layer of pyribolite, which, along its contact against the gneiss, contains a few narrow layers of purple to violet gneiss (4038). To the west it enclose a thin (1—2 m) layer of quartz-diopside rock, which may be banded and which in its southern extension shows prominent smallfolding. West of this calc-silicate bearing layer, there follows a rather broad belt of highly mixed rocks, comprising bands of pyribolite, amphibolitic gneiss, ultrabasic rocks, garnet-amphibolite and leucocratic purple granofelses, fig. 19. On the point west of Nordnoret, this succession, which is represented here by a mixed purple gneiss, is separated by a low-dipping mylonite from bleached gneiss with enclaves of gabbro-anorthosite on the north-western small point.

Thus on Langø the same rock succession is found as on Hestenæsset, although the western gabbro-anorthosite-bearing gneiss is in tectonic contact with the Pas Pyribolite. As previously mentioned, this rock succession may be traced south on Tugdlerúnarssuit.

Fig. 20. Boudinage structures in ultrabasic rocks occupying hinge zones. Sketched from a photograph taken near the west coast, Langø.

H. Sørensen has described some ultrabasic hornblende-hypersthene rocks from the banded and streaky amphibolitic gneisses of the west coast of Langø (H. Sørensen, 1953). According to Sørensen the ultrabasic rocks form lenticles comformably enclosed by gneiss, which latter is regarded as having been formed at the expense of amphibolite. He mentions that the lenticles often diminish in size towards north, a feature which he has also observed within the inclusion-rich dioritic layer of the Great Pyribolite on the east coast of Langø. He ascribes this feature to the formation of boudinage structures from original wedge-shaped ultrabasic bands. The present author's studies of different localities on Langø have convinced him that the initial wedge shape is only an apparent one. The predominant decrease in size towards north, which can be observed where isoclinally folded bands have later on been boudiné, is explained simply by the fact that the ultrabasic rocks have accumulated

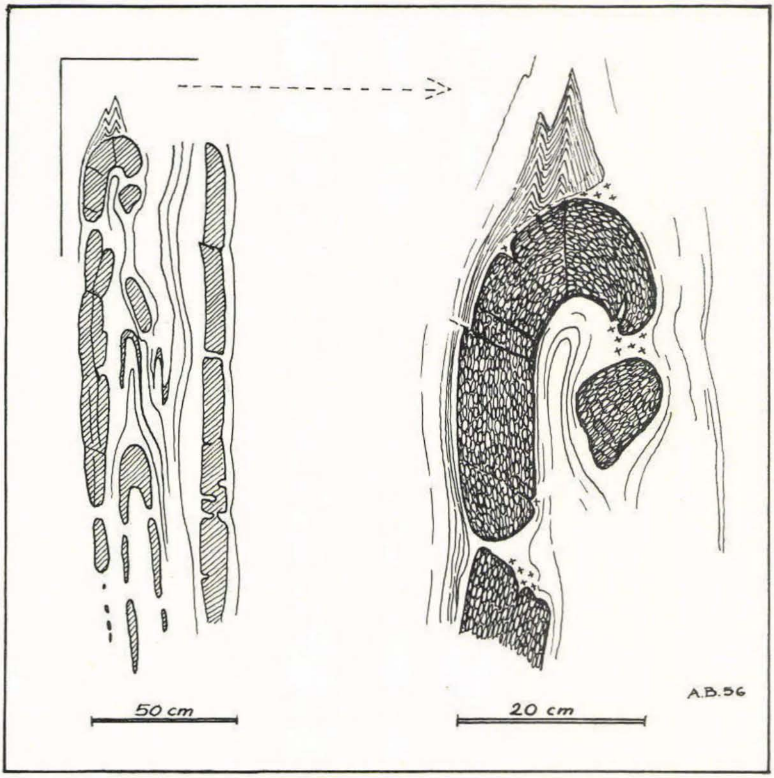

Fig. 21. Small scale structures of ultrabasic rocks in the Pas Pyribolite, western Langø.

preferentially within the antiformal hinge zones. The general south-east axial plunge then accounts for the "polarity" (to use H. Sørensen's expression) seen within the more or less horizontal exposures. It should be noted that decrease in size of the boudins or inclusions has also been observed when going from north to south; in these cases, the ultrabasic rocks have evidently been concentrated in the synformal and not antiformal hinge zones.

In fig. 20, a typical example of boudinage structure within an antiformally arranged ultrabasic band is shown. Another example is shown in fig. 21 where the breaking up of the ultrabasic band is less advanced. As may be seen in the sketch of the antiformal hinge zone the elongated grains of rhombic pyroxene, which largely build up the ultrabasite, in their direction follow the outer shape of the band, but have obviously been deformed in connection with the formation of the boudinage structure. The gneiss seems to have yielded plastically, while the amphibolitic band on the flank of the anticline has been displaced by a small fault in a similar manner to the ultrabasic band.

In fig. 22 several examples are shown of the behaviour of the ultrabasic rocks during deformation. They have all been sketched from occur-

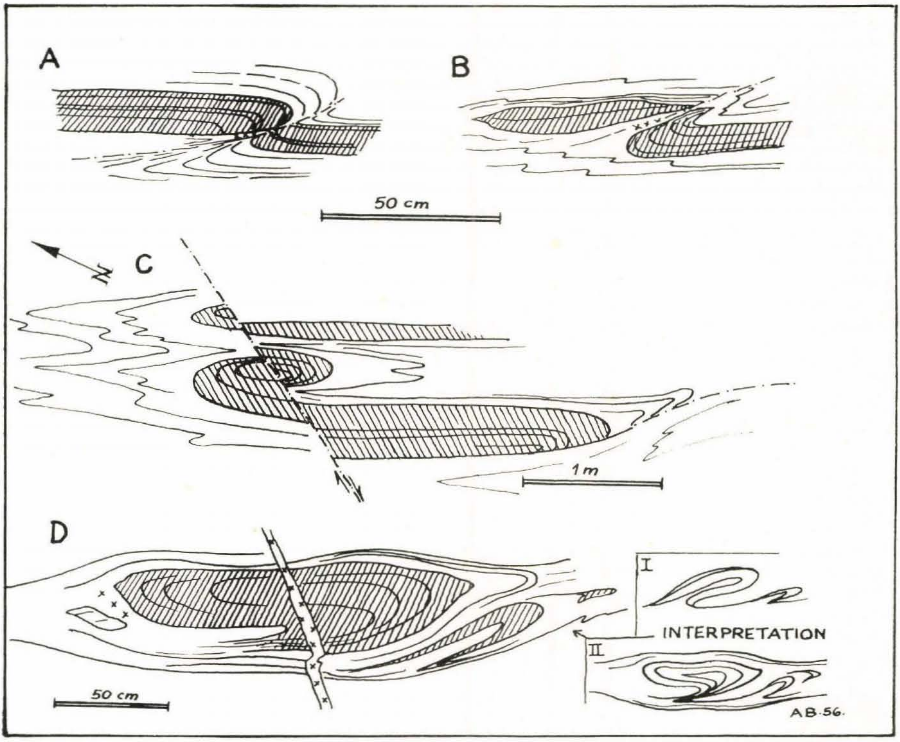

Fig. 22. Sketches illustrating the structural behaviour of ultrabasic rocks, Langø.

rences within the western Pas Pyribolite, except for the lowermost example, which was found along the east coast of Langø. Example C in this figure shows how an ultrabasic band has been isoclinally folded. Such structures may at first sight resemble lenticles or boudins, especially when one of the fold flanks is sheared off. This resemblance is still greater in example D, where the ultrabasite occupies the core of a detached hinge zone. This complex fold can best be imagined to have developed in a two stage folding; first two thickened and dragged hinge zones were formed and then the largest of these was refolded along with changing movements within the surrounding rocks. The examples A and B seemingly represent a deformational stage where the ultrabasic rocks had gained in relative competence. This is particularly evident from figure B, where the ultrabasic band has also been disrupted. The development of pegmatite along the shear fractures may also be mentioned (cf. the pegmatite between the ultrabasic fragments in fig. 21).

In order to see whether the conclusions drawn from these small scale features are also valid on larger scale, an area of about 50 × 40 m where ultrabasic rocks are abundant was mapped on a scale of 1 : 100. The ultrabasic rocks, which comprise hornblende-bearing hypersthenite and hypersthene-bearing hornblendites, are associated here with garnet

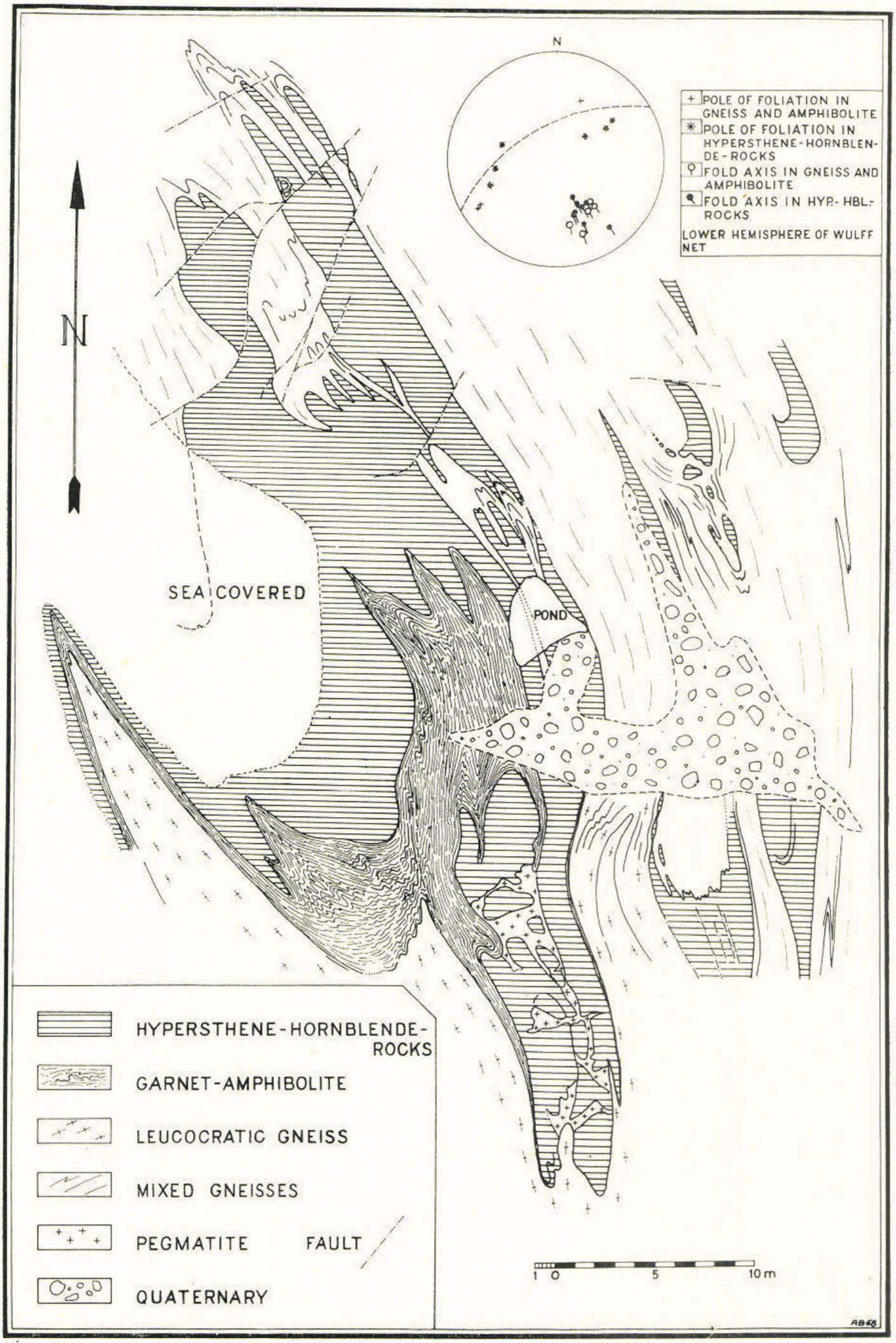

Fig. 23. Sketch map of a small area within the Pas Pyribolite, western Langø. For exact location see geological map of Langø, fig. 12.

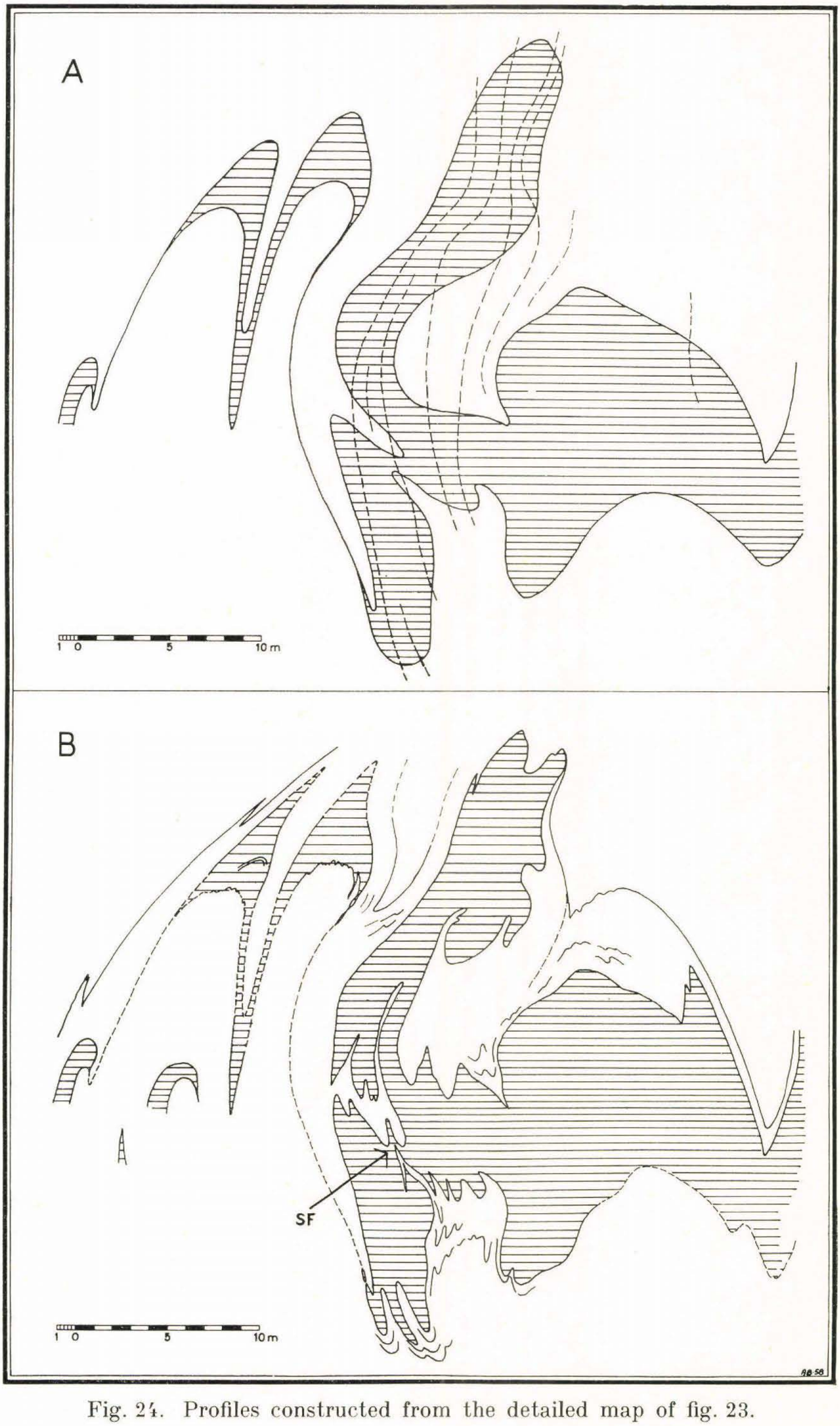

Fig. 24. Profiles constructed from the detailed map of fig. 23.

amphibolite and mixed gneiss with bands and schlieren of pyribolite and amphibolite. Very leucocratic purple granofelses also occur. The map resulting from this study is shown in fig. 23, and the location of the map area is indicated on the geological map of Langø.

The resulting outcrop pattern is a very irregular one. In the southern part of the map, the ultrabasite is partly agmatised by pegmatites, but the original structures are still discernable. The area does not present a flat surface, but is slightly rocky, so when standing at one place no general impression of the structures can be obtained, and the ultrabasic rocks appear from a hasty inspection to form lens-shaped inclusions. Only when these are traced in detail does a complex fold pattern show up. Apart from the general impression formed during the mapping, the readings plotted in the stereogram of the figure indicate that the ultrabasic rocks have been folded conformably with the enclosing rocks. Assuming this, a profile at right angles to the fold axis was constructed. This construction did not include corrections for the small variations in altitude. The correction for the displacements along the small faults was made after the projection into the profile was performed. The errors involved hereby do not, however, have much influence on the final profile (fig. 24, B), since the vertical component of the fault movements is rather insignificant. In the profile of fig. 24, B, there are several indications of a shearfolding which has been superimposed on older and larger folds. For example, the smallfolds marked by SF in fig. 24, B, give a clue for the understanding of the structures, since they indicate the existence of shearfolding. In fig. 24, A, a profile is shown which is the result of an attempt at eliminating the effect of the later shearfolding. The antiformal axial planes are indicated as broken lines so that a comparison with the lower profile can be made. By this procedure, we arrive at a 'style tectonique' which is directly comparable to that observed on a mesoscopic scale. The style is characterised by the highly plastic behaviour of the ultrabasic layer. It would be interesting to extend this analysis to a still larger area in order to see whether the apparent conformity between the ultrabasite and the enclosing rocks persisted.

This detailed analysis also shows that in this region the lithological layering is not as steep as would be imagined from most of the field readings. This is also shown by the folded banded skarn horizon and the gneiss layers further to the south (see fig. 12).

The course of the Pas Pyribolite on Tugdlerúnarssuit and on the south coast of Qagssiarssuk has already been described. Between Qagssiarssuk and Blindtarmen, the pyribolite increases in thickness. Since it mainly occupies a broad valley, and is only exposed on some ridges, the ultrabasic rocks, which were mapped close to the coast, could not be traced further south. The Blindtarmen lake has been eroded out in an important antiformal hinge zone which makes the Pas Pyribolite reappear with a northeasterly strike on the 140 m high mountain, Qáqârssúnguit. This hinge zone is dissected by several later faults striking ESE and showing a dextral displacement.

NE of the long and narrow fault-determined lake, the pyribolite passes more or less gradually along the strike into gneissified pyribolite and finally into a mixed gneiss which contains only slightly more basic material than the mixed gneisses south of the Pas layer. Where passing two smaller lakes, the Pas layer is characterised furthermore by the presence of several ultrabasic inclusions and more or less irregular bodies of pyroxene diorite. Still further to the north-east, the layer is sheared out along a marked zone of movement.

On the north coast of the Pâkitsoq fjord, the Pas Pyribolite reappears east of the 2nd Intermediate Layer. The pyribolite layer is also somewhat gneissified here, i.e. it in fact forms a layer of more or less dioritic gneiss with abundant bands, schlieren and inclusions of pyribolite. Further to the north, however, it attains its normal banded or massive appearance. It is joined here, on its eastern border, by a thin layer of pyribolite, which on the coast of Pâkitsoq was separated from the Pas layer by an intervening wedge-shaped layer of hypersthene-bearing or purple gneiss.

South of Enehøj and just north of Gammel Lejrskar, this thin pyribolite layer again splits off from the Pas Pyribolite.

In the triangular fault block, south of Gammel Lejrskar, the Pas Pyribolite is affected by small scale and large scale folding. Just north of the two parallel NE-striking faults a larger antiform and a neighbouring synform are found. The axes of these folds plunge gently towards NNW within the hinge zones, but further north assume a horizontal position. This explains the reappearance of the pyribolite as a small inlier within the light coloured gneiss of the synformal core. In the antiform, the axis takes up a SSE plunge before it finally, north of Gammel Lejrskar, plunges rather steeply to NNW. Within this region, the pyribolite is thrown into 1—2 m large drag folds and it contains abundant inclusions of ultrabasic rocks.

From Gammel Lejrskar to Ankerbugtdal, the Pas Pyribolite describes a slight arc. Where it has been studied in detail, (close to the 374 m Lake) the contact with the 2nd Intermediate Layer is knife sharp and seemingly tectonic, since the adjoining gneiss contains platy drawn-out quartz grains. Further west a narrow layer splits off from the main pyribolite and enters the synformal structure within the 2nd gneiss layer described above.

Within the northern part of the Tovqussaq dome structure, two additional litho-structural units have been mapped. From Verdens Begyndelse, north of Sorthat, across the Ankerbugtdal to Enehøj, the

Pas Pyribolite is bordered by a narrow gneiss layer which separates it from a layer very rich in enclaves or bands of gabbro-anorthosite. In this northern part, these layers still belong to the dome structure. They are best developed between Sorthat and Enehøj. To the west as well as the east, on Hestenæsset and SE of Enehøj respectively, the gabbro-anorthositic layer seems to join the Pas Pyribolite. Around and east of the Ankerbugt valley, the intervening gneiss layer is developed as hypersthene gneiss, but further east it passes into a banded light coloured gneiss. West of the Ankerbugtdal, the northern 'gabbro-anorthositic' layer is separated from the northern light coloured to pink, homogeneous rocks by a narrow band of smallfolded and banded pyribolite, which seemingly belongs to the gabbro-anorthosite layer. West of Enehøj, this latter layer is nearly 90 $^0/_0$ gabbro-anorthosite, which here is intensely smallfolded. Where this layer is involved in the synformal and antiformal structures west of Vinkelsø, the gabbro-anorthosite occurs as more scattered inclusions in the gneisses. Structurally the inclusion-rich zone joins up with the narrow pyribolite layer, which south of Vinkelsø was found to split off from the Pas Pyribolite.

An analogous development may be observed west of Kolbesø where the gabbro-anorthosite-bearing layer has been involved in the folding which also affected the Pas Pyribolite here (cf. fig. 11 and the text above).

The western surroundings of the Tovqussaq dome

Under this heading the region including the country south of Lille Ekkodal to Hestenæs and from western Tugdlerúnarssuit to Atangmik in the extreme south will be described. For the northernmost part of this region the reader is referred to fig. 11, which is a reduction from a map on a scale of about 1 : 4000.

In this map a fan-shaped arrangement may be seen in the strike of the lithological units. It seems probable that this outcrop pattern is caused by a fold, and the mapping also showed that, going from west to east, the lithological succession is more or less repeated. Along the coast, west of the broad strike valley, Strøget, an important pyribolite layer occurs.

At its eastern contact it is underlain by a narrow band of calc-silicate rocks. Close to these, the pyribolite contains a parallel row of ultrabasic lenticles, which seemingly once formed a conformable band (possibly isoclinally folded). East of the calc-silicate band there occurs a hypersthene-bearing gneiss layer, which to the south rapidly changes into light coloured rocks (4049, A and B). Then mixed gneisses, which run parallel to the contact with the above-mentioned gneiss layer and which contain ultrabasic rocks, are met with. Towards the northern point these

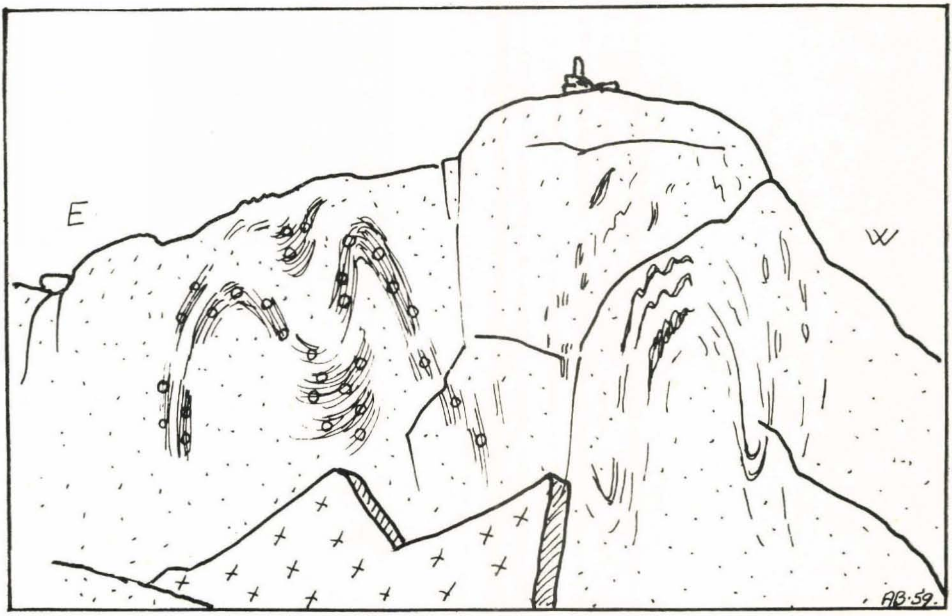

Fig. 25. Smallfolded mixed gneisses, central Hestenæs. The open circles indicate garnet amphibolite. In the front is a sheared pegmatite.

gneisses pass into slightly gneissified pyribolite. They are generally hypersthene-bearing, but change to the east, where they contain abundant gabbro-anorthositic enclaves (4050), into more purplish gneisses, which latter grade into the more light coloured rocks underlying the strike valley, Strøget.

Although not completely identical in detail, the rock succession just described shows a great resemblance to that described above from the east coast of Hestenæsset and west of Kolbesø. The intervening area contains mostly mixed gneiss, with discontinuous basic bands and occasional enclaves of gabbro-anorthosite. Purple gneiss types predominate, but quartz-syenitic granofelses (4019) and more rusty coloured hornblende-hypersthene-bearing types are also met with. On Hestenæsset schlieren of pyribolite and garnet amphibolite were found on the west side of the point, where folds with an amplitude of 1—2 m are developed, fig. 25. The mesoscopic drag folds in the cliffs of the west coast indicate that they are on the western flank of the antiform. The sketches shown in fig. 26 were made in the central part of Hestenæsset. In A of this figure it may be seen how gabbro-anorthosite with thin intercalations of gneiss and pyribolite has been flexural-slip folded. A small thrust cuts off the crest of the antiform. Part B of the same figure shows a row of slightly clockwise rotated boudins in biotite-hypersthene gneiss. The three southernmost boudins consist of gabbro-anorthosite with

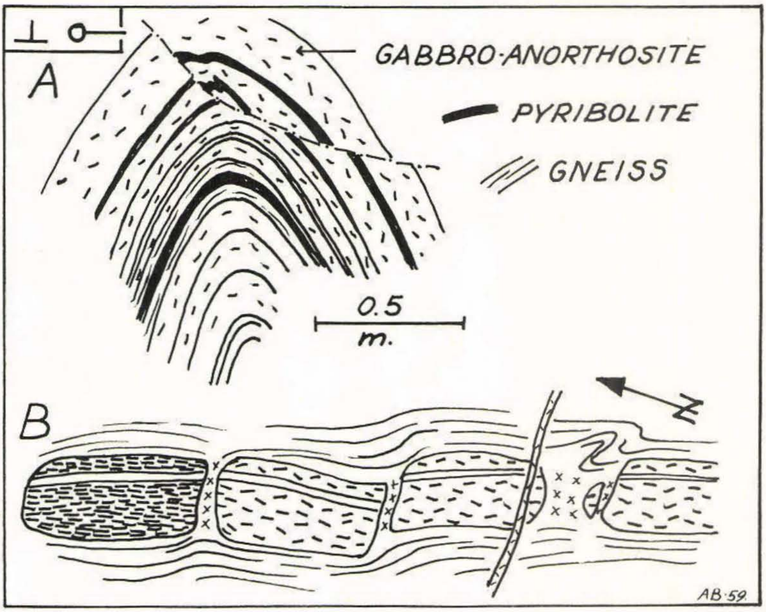

Fig. 26. Small scale structures in gabbro-anorthosite, Hestenæsset.

a narrow band of almost pure anorthosite. This band is also found in the northern boudin which otherwise is made up of a melanocratic diopside amphibolite with certain affinities to the gabbro-anorthosite. Pegmatites separate the boudins and an undeformed vein cuts the lot. The movements which caused the rotation of the boudins have also left their trace in a small drag fold. The relative tectonic transport indicated by these structures fits well into the kinematic picture of the large antiform over Hestenæsset.

On western Tugdlerúnarssuit the southern continuation of the same antiform is visible in the arrangement of the inclusions and schlieren in the quartz-dioritic hypersthene gneiss, which latter may in places attain an almost dioritic composition.

The enclaves and schlieren consist of pyribolite, biotite-pyribolite, gabbro-anorthosite and less commonly of ultrabasite. On more or less vertical walls, where the weathering has given the inclusions a slight negative relief, their arrangement reveals a gently plicated antiformal crest, fig. 27.

Within these inclusions transitions from skarn-amphibolite into gabbro-anorthosite have been noticed. Fig. 28 illustrates another interesting feature seen at this locality. Within a larger enclave of foliated gabbro-anorthosite, two conformably arranged boudins of almost pure anorthosite are found, surrounded by a thin zone mainly composed of diopside with a little biotite and plagioclase. This observation is im-

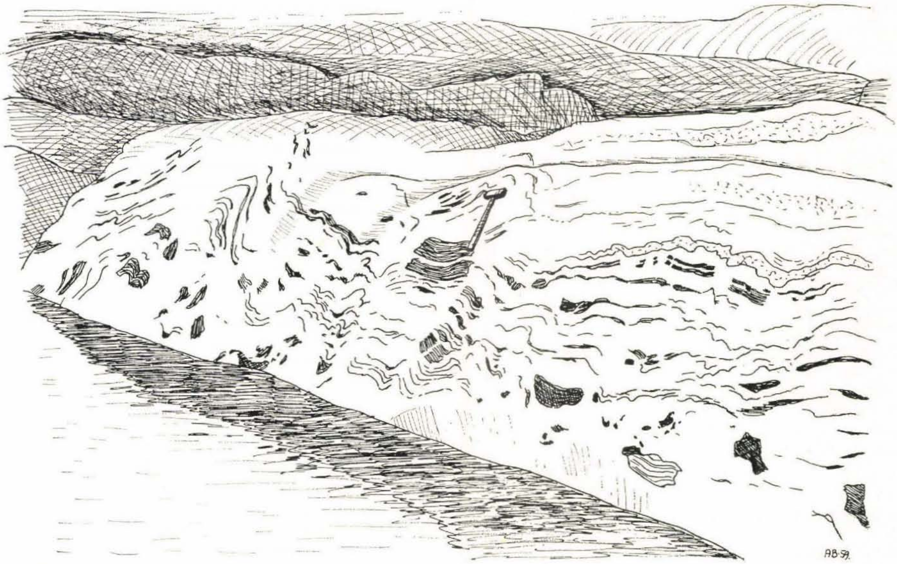

Fig. 27. Small scale structures in gabbro-anorthosite-bearing gneisses, northern point of Tugdlerúnarssuit. Drawn from a colour photograph.

portant, since it demonstrates that the gabbro-anorthosite, before it was broken up into the present large blocks, had been exposed to a still older phase of deformation during which not only the foliation, but also internal boudinage structure were developed.

In the smaller enclaves and schlieren of gabbro-anorthosite, hornblende and biotite generally constitute the mafic minerals.

Fig. 29 shows another example from western Tugdlerúnarssuit. Although the smallfolded pyribolite and amphibolitic bands still indicate the presence of the antiformal crest, the tectonic style has radically changed. The larger amplitude may be explained by the control of the thicker competent basic bands and as such is not unusual, but the occurrence of recrystallised finer grained zones of former movement which cut through the fold pattern is a new feature. Where gabbro-anorthositic enclaves have been affected by the younger movement, they may now be rotated and displaced. These dyke-like zones presumably represent recrystallised and plastically moved old mylonites or shear zones. Rather similar features have also been observed west of Kolbesøen within the purple gneisses of the central part of the same antiform.

The low land between Qagsiarssuk and Atangmik appears rather monotonous in composition and structure. East and south of Blindtarmen, a rather large scale banding within the gneisses shows that the acute hinge zone of the Pas Pyribolite around Blindtarmen may be traced in the gneisses towards the extreme south-western point of the

Tovqussaq peninsula. The same impression was obtained from the study of the coasts. On a smaller scale the gneisses are, however, rather heterogeneous. Darker bands and schlieren are common, as well as lenticular or more irregular enclaves of basic and ultrabasic rocks. Smallfolding is very prominent in the gneisses surrounding the Pas Pyribolite, where this makes a sharp bend around Blindtarmen. Readings of fold axes, mineral lineation and elongation of inclusions are often more easy to

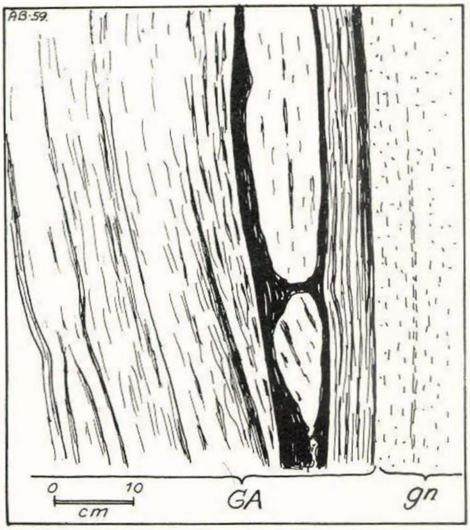

Fig. 28. Enclave of gabbro-anorthosite with internal boudinage structure, northwestern Tugdlerúnarssuit.

obtain than measurements of the lithological banding. Where these gneisses have been visited, they were not found to contain gabbro-anorthositic inclusions, but possibly a more intensive search could prove their existence. Along the west coast, the gneisses are universally hypersthene-bearing, where not affected by retrograde alterations due to mylonitisation or faulting. West and south of Blindtarmen, transitions into purple, occasionally garnet-bearing types are met with. The above-mentioned 110 m hill is composed of somewhat smallfolded hypersthene gneiss which otherwise is rather homogeneous. On the north coast of the small Atangmik fjord, more or less banded hypersthene gneiss is found.

The area north of the Tovqussaq dome and south of Smalledal

The area discussed here extends from Kalotbugt and Gule Hav across the Ankerbugtdal to Gammel Lejrskar. From the field work it is evident that the geology of this area is very complex. The total absence

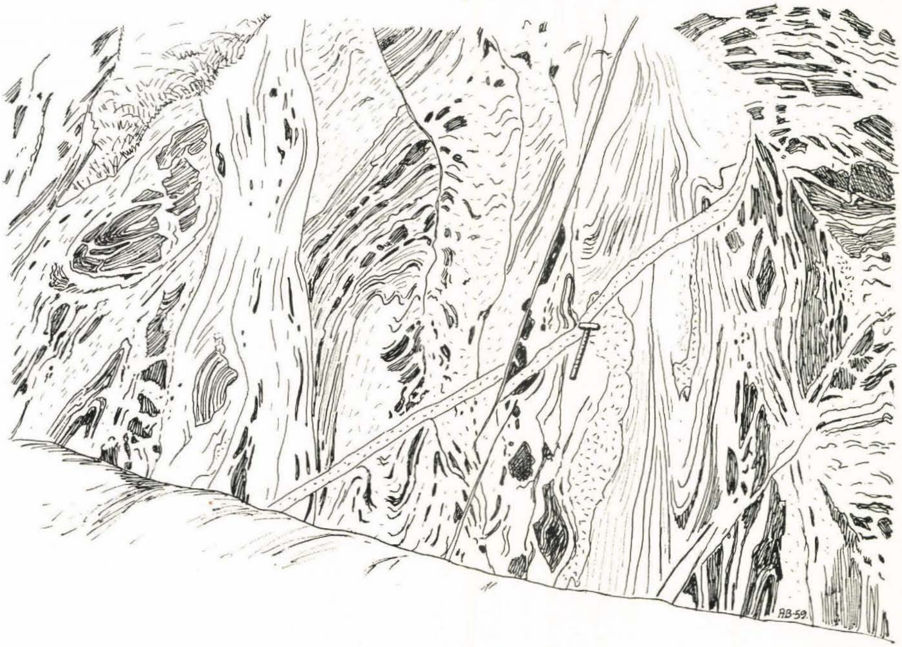

Fig. 29. Small scale structures within the hinge zone of the Western Antiform, western Tugdlerúnarssuit. Drawn from a colour photograph.

of any reliable marker horizons makes it very difficult to present a litho-structural map of this region. Within larger areas, such as Skraaland, the rocks are so homogenised that no structural readings can be taken over a stretch of more than one kilometre. In other places, 'layers' more rich in basic and ultrabasic inclusions can be traced for some distance. Two such inclusion-rich zones have been traced continuously from Store Ekkodal right to Ankerbugtdal. Otherwise the basis for the mapping is mainly petrographical, and of doubtful value from a structural point of view since the granodioritic and granitic rocks have been developed in connection with the last phase of movement and hence do not necessarily reflect the older-formed litho-structural units in their present distribution on the map.

The shores between Kalotbugt and Gule Hav are mainly built up of yellow-brown to rusty coloured dioritic rocks, which also form the western part of the small island west of Gule Hav. The eastern part of the island is made up of mainly hypersthene-bearing banded gneisses, in which small scale double folds are seen. The youngest axes plunge at 20° to the NNW (341°).

On the slopes east of Gule Hav, very light coloured granitic rocks (19238) are exposed up to the fault in Smalledal. Partly agmatised amphibolitic bands are enclosed by the granitic gneiss, which here also

Fig. 30. Agmatitic gneisses, northern wall of Store Ekkodal (Photo: GGU, A.B.).

contains abundant and metre-large biotite-bearing pegmatites with crystals of microcline up to 20 cm across. Eastwards the granitised rocks pass gradually into a more mixed gneiss of the purple type. On the point just north of Kalotbugt, the diorite on the extreme tip borders directly on hypersthene-bearing or purple gneiss with a band rich in enclaves of gabbro-anorthosite.

On the northern wall of Store Ekkodal an illustrative nearly vertical profile through the agmatitic zone north of Skraaland can be seen, fig. 30. On this photograph smaller dark areas represent ultrabasic rocks, while the less dark patches are more or less transformed basic rocks. In the centre these rocks appear slightly more light coloured due to granitisation. Light coloured, irregular in part, pegmatitic veins traverse this association. The photograph shows a profile some 25 m high. South of Store Ekkodal, this agmatitic zone seems to form a hinge zone with an axial plunge to the SSE. Further north-east the agmatite fades out within the granitised roks of Skraaland. These gneisses correspond closely to those described from the Intermediate Layers in the northern part of the dome, although possibly they carry slightly more microcline. They can be traced eastwards just across the Ankerbugtdal, but further on more heterogeneous rocks are met with. They are light purplish streaky to agmatitic gneisses. In the transition zone nebulitic agmatites are seen.

The northern mixed gneiss band south of Smalledal is generally less agmatitic than the one seen in fig. 30; it rarely contains ultrabasic

Fig. 31. Agmatitic gneiss, south coast of Kalotbugt. The internal structures of the basic enclaves show an older fold structure which is only slightly influenced by post-replacive movements. Drawn from a colour photograph.

inclusions and in places may be well banded. However, where occurring on the south shore of Kalotbugt, it is strongly agmatised with partly digested inclusions, see fig. 31. None of these two 'layers' can be traced with certainty east of Ankerbugtdal. North-west of Enehøj a zone with small ultrabasic inclusions is found, but otherwise it can only be said that most of the gneisses are more or less agmatitic. This applies particularly to the region north and east of Enehøj. The darker enclaves generally consist of biotite-amphibolite, but may also have been transformed into biotite-rich quartz-diorite of a lighter shade. The matrix is purple granofels which may grade into more pegmatitic rocks. A remarkable fact is the parallelism of the internal structures within the irregularly distributed enclaves. These agmatites are thus typical replacement agmatites, figs. 32 and 33. East of Enehøj, more basic to ultrabasic enclaves are found and inclusions of calc-silicate and gabbro-anorthositic rocks (35858) have also been met with. Between Nordvestpassagen and Vinkelsø, the gneisses are partly hypersthene-bearing. Another feature which deserves mention is the occurrence of highly granitised rocks east of Trekantsøen, where they form a narrow zone where the granitisation has been controlled by an early fault.

Fig. 32. Replacement agmatite, north-east of Enehøj. The internal structures of the enclaves have preserved their parallel orientation (Photo: GGU, A.B.).

The area between Gammel Lejrskar and Pâkitsoq

This area, which towards the east is limited by the Pâkitsoq Pyribolite, forms the southern continuation of the unit described above. It consists mainly of hypersthene gneiss, which gradually becomes more and more uniform towards the south. Around Gammel Lejrskar, smallfolding is very pronounced and the gneiss may be rich in enclaves. The axial culmination observed within the antiform of the Pas Pyribolite is also found in the gneisses. Around the two parallel NE-striking faults, a more or less discontinuous thin layer of pyribolite may be traced in a partly gneissified state. The hypersthene gneiss east of this layer is rather homogeneous, but contains a few enclaves which represent untransformed relics, since their internal structures still conform to the southward plunging axis (fig. 34). North of Ilordlia, gabbro-anorthositic inclusions —often arranged in bands—are found in the gneisses enclosed between the Pas Pyribolite and the above-mentioned thin pyribolite. On the west coast of Ilordlia, dragfolded bands and streaks of finely banded calc-

Fig. 33. Agmatitic gneiss which has been moved after the formation of a replacement agmatite, near to the locality pictured in fig. 29 (Photo: GGU, A.B.).

silicate rocks are found, whereas gabbro-anorthosite has not been found here. This latter rock, however, forms abundant enclaves in the gneisses bordering on the two enclosing pyribolite bands on the north coast of the Pâkitsoq fjord. A less inclusion-rich hypersthene gneiss here occupies the central part of the gneiss layer in which an antiformal hinge zone may be discerned (see map Pl. 1).

On the west coast of the narrow strait which connects the Pâkitsoq fjord with Ilordlia, a narrow layer of pyribolite is found. It terminates before reaching the north coast. Whether this is due to folding or to shearing is not known. North of Ilordlia, and west of the Pâkitsoq Pyribolite, the hypersthene gneiss contains scarce small gabbro-anorthositic enclaves. Right on the coast west of the Pâkitsoq layer a narrow band of pyribolite is seen on the north shore of Ilordlia, but it has not been possible to trace this band further to the north. To the south, on the east coast of Pâkitsoq, the rocks dealt with in this paragraph form part of the Pâkitsoq antiform to be described later on.

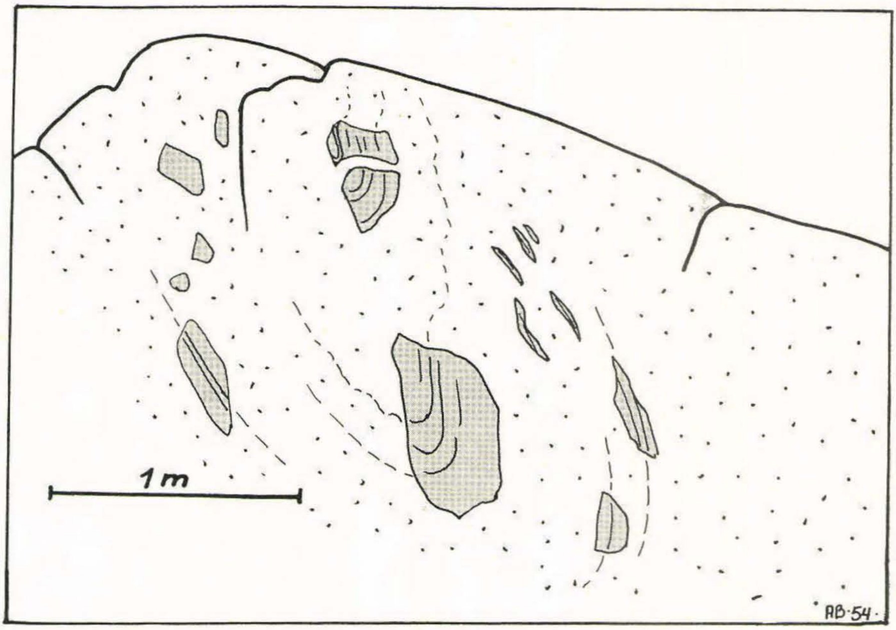

Fig. 34. Enclave-bearing hypersthene gneiss, northern side of the fault valley east of Vinkelsø. The structure lines added to the sketch show that the enclave structures reflect an older smallfold.

The Smalledal structure

North of Smalledal and Nordvestpassagen two structural units can be distinguished, the Smalledal and the Irdal structures. As was the case with the Tovqussaq dome, it is the marker horizons of pyribolite which enabled such a division to be made. In the Smalledal structure a central pyribolite layer can thus be traced in a complete loop from Kangeq to Trehøje, where it turns round and runs westwards again as far as Kangeq, here closing again in a smallfolded hinge zone. Another outer pyribolite layer takes part in a similar structure. Its western hinge zone is hidden under the sea at Kangeq, but the eastern closure may be traced from south of Krydssø, across Breddal to north of Bortehøj. On this last stretch the pyribolite is generally completely gneissified; only north of Breddal and north of Bortehøj are mappable relics of pyribolite preserved.

At the latter locality, pronounced double folding on a mesoscopic scale is seen in the surrounding gneisses. From this northern extension, to just east of Hjørnet and south of Lejrsø, the strike continuation of the pyribolite can be traced as a litho-structural boundary between the two different gneiss layers, which, between Lejrsø and Kangeq, are separated by the pyribolite.

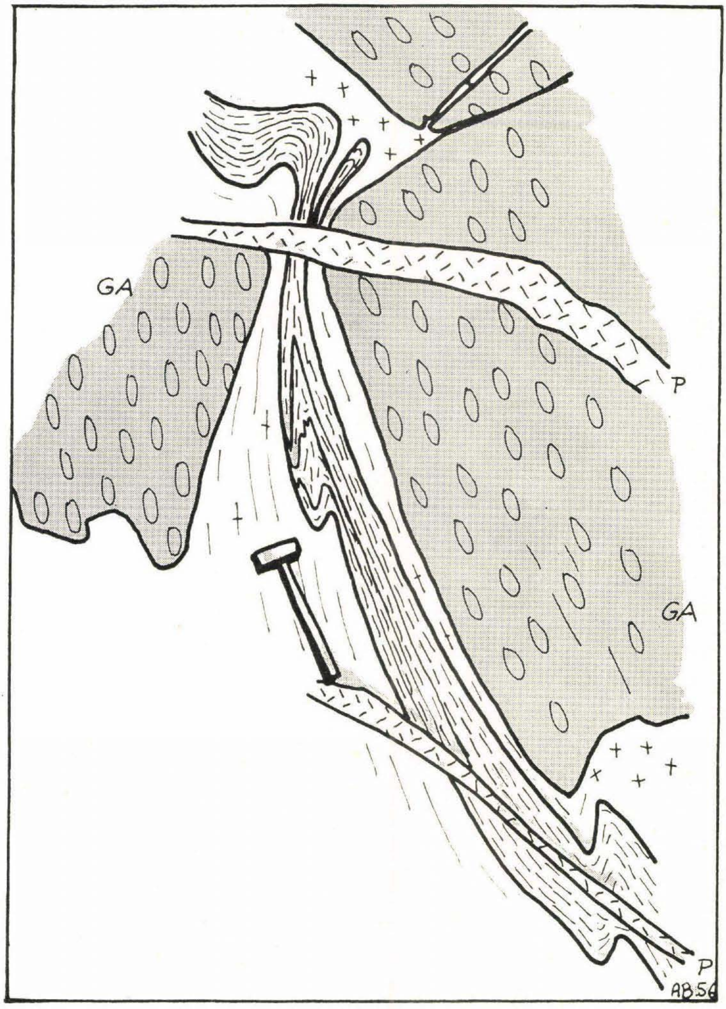

Fig. 35. Drag folds developed in mixed gneiss separating enclaves of gabbro-anorthosite, east of Nordvestpassagen.

From these results, it can be concluded that the rock succession south of Smalledal and Nordvestpassagen virtually circumscribes the Smalledal structure (as indicated by the two pyribolite layers). It also surrounds the exposed parts of the Irdal structure. This gneiss layer has been called the Frame Layer since on the map it surrounds the Irdal and Smalledal structures as well as the Tovqussaq dome. As will be explained later on, it may also be correlated with the Frame Layer of the central Tovqussaq dome described earlier. This correlation is based on lithological resemblance and structural reasoning and thus represents a more advanced stage of the analysis. However it was tested as a working hypothesis during the field work.

On the point north of Gule Hav the Frame Layer is represented by mixed, in part hypersthene-bearing gneisses, which are smallfolded around southward-plunging axes. They are limited by a thrust fault from the granitic gneisses to the south. The southernmost pyribolite layer of the Smalledal structure is separated here from the mixed gneiss by a conformable layer packed with enclaves of gabbro-anorthosite or composed of finely banded gabbro-anorthosite. The pyribolite, as well as the fringing gabbro-anorthosite, takes part in an antiformal structure, which explains the smallfolding in the mixed gneisses. Towards the east, the fringing gabbro-anorthosites gradually take the place of the pyribolite, and this layer, where crossing the Ankerbugtdal, is only indicated by a zone of gabbro-anorthosite enclosed in microcline-bearing gneiss.

In the antiformal hinge zone of Trehøje it may be seen how the individual blocks of gabbro-anorthosite have controlled the folding of the enclosing more plastic rocks. Fig. 35 shows an example where a band of amphibolitic gneiss in a purple pegmatic granofels has been squeezed into tight folds between the more massive enclaves. In spite of this relation, the fold axis measured in the amphibolitic gneiss conforms to the more regional antiformal structure. Drag folds with amplitudes from less than one to over ten metres are very common in the mixed gneisses of the Frame Layer south of Krydssø.

The Frame Layer continues east of Trehøje towards Breddal as a mixed mainly hypersthene-bearing gneiss and then becomes purple or more light coloured. In more or less transformed pyribolitic inclusions isoclinal smallfolds are seen (fig. 36). North of Bortehøj and against the thick pyribolite a narrow amphibolitic band may be followed for about one kilometre until westwards it passes into smallfolded agmatitic gneiss.

In Kipdal, on Hjørnet and Graahat, around Lejrsø and on Graaryg, the Frame Layer is first of all characterised by its great content of gabbro-anorthosite inclusions, which in places constitute more than 50 $^0/_0$ of the total rock. In a zone limiting the northwestern extension of the Smalledal structure (i. e. north of Hjørnet), mesoscopic double folding further complicates the structures of this layer. The gabbro-anorthositic enclaves, where abundant, give an unusual light grey colour to the mountains of this region (hence the names Graahat and Graaryg). A nearly total absence of lichens was also noticed. The enclosing gneisses are generally purplish but light coloured types (4094, A) are also met with, as for example west of Lejrsø and within the antiformal crest at Lejrsø. On the seaward slopes of Graaryg, slightly banded hypersthene-bearing gneiss is also found. Besides the gabbro-anorthositic inclusions, the gneisses of the Frame Layer contain bands, schlieren and fragments of pyribolite (4094, A) in all stages of amphibolitisation or granitisation.

Fig. 36. Agmatitic structures of replacive origin in isoclinally smallfolded gneiss, southern wall of Breddal (Photo: GGU, A.B.).

The gabbro-anorthositic rocks generally occur as boudins or enclaves half a metre to over one metre across. According to the structures within the enclaves, three types may be distinguished: 1) A porphyritic type, which is rare. 2) A type characterised by flaser structure which presumably has been developed from original porphyritic bands by shearing of these before they became boudiné (fig. 37). 3) Finely banded types. Whether the last mentioned type represents a more advanced stage of kinetically transformed originally porphyritic types, or whether the banding can be regarded as a primary structure is not known. Possibly both explanations are right. The porphyritic type has only been encountered in a few places. An example, where the outlines of rather fine grained plagioclase aggregates indicate the existence of former megacrysts from 10 to 20 cm large, was described under the basic division (35838) (Berthelsen, 1957, fig. 9). In some places, the rigid gabbro-anorthositic enclaves and boudins are still aligned, but in other places, as for example on the coast east of Kangeq, the individual fragments are seen to have been rotated in the enclosing more plastic purple gneiss (19235). Close to this locality, agma-

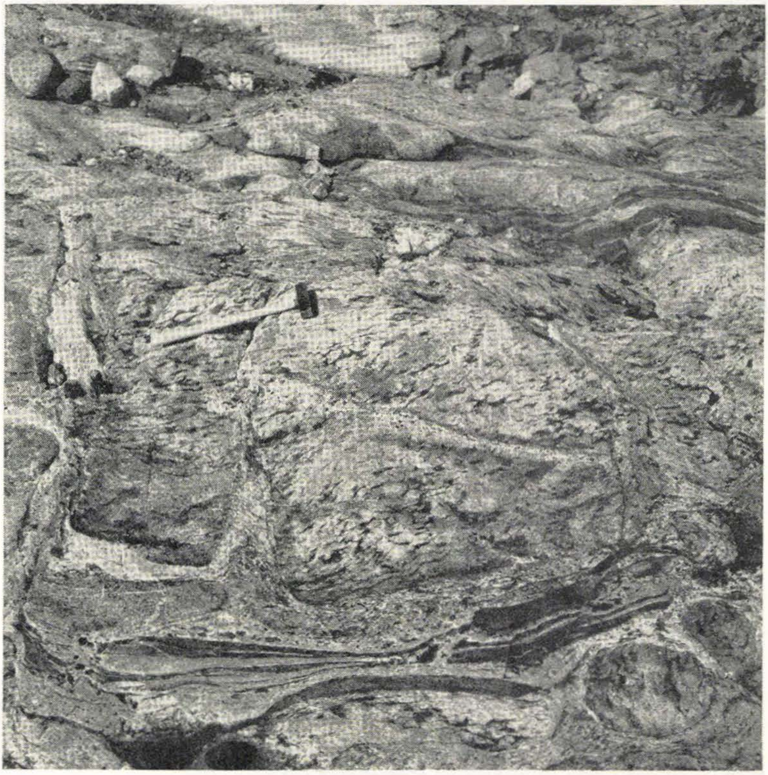

Fig. 37. Boudin of gabbro-anorthosite with internal flaser structure, near Graahat (Photo: GGU, A.B.).

tisation of former larger enclaves was also noticed. In some of these, the plagioclase content is unusually low, so that a nearly ultrabasic composition is approached.

The northern part of the outermost pyribolite layer of the Smalledal structure, which in this region forms the southern border of the Frame Layer, is, at Kangeq, exposed as a broad layer, with a thin intercalation of light coloured gneiss. Eastwards, it diminishes in thickness, but is still recognisable as a distinct layer until west of Lejrsø. South of this lake, however, it is less distinct and is partly gneissified. The gneisses to the south of it are generally rather light coloured or faint purple. They contain pyribolitic and amphibolitic streaks and schlieren and scattered inclusions of gabbro-anorthosite. This latter rock type is particularly well exposed on the north side of the fault-determined valley in which Lejrsø is situated.

On the southern side of the same fault, but east of Lejrsø, a remarkable agmatite is developed from the same gneisses. Pyribolitic and amphibolitic fragments, which obviously are derived from a banded succession, here lie in a scattered arrangement within a purple granofels. Although the banding of the inclusions for the most part conforms with a fold with steeply plunging axis, it is obvious that rotation and displacement of the individual fragments has taken place. This exposure is found just north of the northern hinge zone in the innermost pyribolite layer of the Smalledal structure. This particular structural situation seems to be responsible for the exceptional development of the agmatite.

Where the outermost gneiss layer of the Smalledal structure is exposed east and north-east of Trehøje, complex smallfolding is also seen. Basic bands and schlieren are folded here around N-plunging axes; a little to the south the axes are vertical but change then to SSE-plunging. The purple to light gneisses south of Krydssø grade westwards into slightly banded, more rust coloured, in part hypersthene-bearing rocks, which west of the Ankerbugtdal are replaced by light coloured banded gneisses with conformable rust zones richer in biotite. Whether these rust zones represent original more pelitic intercalations or have developed from transformation of amphibolitic bands is uncertain. Further west the light coloured gneiss layer passes into more mixed purplish gneiss, which finally on the west coast of the small point south of Kangeq changes into hypersthene gneiss.

Where it reappears on Kangeq, the layer is represented by banded mainly purple gneiss which contains a few gabbro-anorthositic inclusions. From the east coast of the small bay and eastwards, light purple gneisses are found. When the southern slopes of Graaryg are approached, the layer becomes more heterogeneous and passes into the rocks south of Lejrsø mentioned above.

Just south-east of Kangeq, the innermost pyribolite layer of the Smalledal structure forms a well exposed hinge zone. The axes of the prominent smallfolds plunge west, thus indicating the antiformal character of the closure. This isoclinal antiform participates in the S-plunging antiform structure indicated by the outermost pyribolite layer and the Frame Layer gneisses north of Gule Hav.

As mentioned earlier, the northern and southern flanks of the overturned antiform can be traced eastwards across the Ankerbugtdal-Lejrsø fault. In the southern flank, west of this fault, the diopside-rich pyribolite was seen to give way to finely banded gabbro-anorthosite. This transition takes place along the strike and within one metre. East of the fault, the pyribolite reappears in a more or less gneissified or dioritised state. It is divided here into two parallel layers by a narrow intermittent band of light gneiss, the development of which may be related to a shear zone. The light coloured gneiss encloses a band of magnetite-bearing skarn (19652). Just west of Krydssø some large barren quartz veins are found in the pyribolite.

In its complicated course around Trehøje, the innermost pyribolite layer is transformed into a mixed gneiss which carries only a little more basic material than the surrounding lithological units. However, mapping of small scale litho-structural features made it possible to trace the layer continuously into the northern flank of the Smalledal structure.

The central gneiss layer of the Smalledal structure is a purple to light gneiss, which contains small amounts of basic material and, around Trehøje, scattered inclusions of gabbro-anorthosite. In the crest zone of the western isocline it is very homogeneous.

The Irdal structure

Along its southern borders, the Irdal structure is separated from the surrounding gneisses of the Frame Layer by a narrow band of pyribolite, which in the coast section contains smaller bodies of ultrabasic rocks. This thin layer grades eastwards into mixed gneiss, which can hardly be distinguished from the underlying unit comprising mixed biotite and hypersthene gneisses. This unit is characterised by banding on a rather large scale (half a metre to several metres thick bands), fig. 38. The individual bands, which are rather discontinuous, include dark agmatised rocks, intermediate rock types, purple gneiss and hypersthene gneiss. The darker components of the gneiss are generally composed of pyribolite and its derivates or, more rarely, ultrabasic rocks. This rock succession is easily distinguishable from the gabbro-anorthosite-bearing Frame Layer due to its predominant rusty brown colour. The banding made it possible to trace the antiformal hinge zone indicated by the underlying pyribolite right from the cirque of Irdal to Bortehøj. Within this hinge zone, smallfolding is common.

The core of the Irdal structure is formed by a light coloured gneiss with scattered small ultrabasic inclusions. It is separated from the banded rusty brown gneisses by a more or less gneissified or granitised composite layer of pyribolite. The granitisation is especially prominent west of Irdal where the basic layer is cut at a small angle by a shear zone.

West of Kipdal, another antiformal hinge zone is shown by the occurrence of an almost gneissified pyribolite layer (around the 84 m hill). The core of this antiform is occupied by smallfolded to agmatitic gneiss.

Central and Eastern Tovqussap nunâ

The rocks which build up the central and eastern parts of Tovqussap nunâ show great analogies to those described from the western region. The only fundamental difference is that gabbro-anorthositic rocks are very scarce in the eastern area. They have only been found at three

Fig. 38. Banded gneisses near point 442, west of Hjørnet (Irdal structure), (Photo: GGU, A.B.).

localities and in close association with pyribolite; the localities are on the east coast of the narrow inlet to Ilordlia, at Buedal and east-north-east of Kronehøj.

In central and eastern Tovqussap nunâ, the following structural units may be distinguished: 1) The Pâkitsoq antiform, the western flank of which delimits the Tovqussaq dome, the Smalledal and the Irdal structures. 2) The Flankepas synform which borders the Pâkitsoq antiform to the east. 3) The Krebsesø antiform which stretches from Kangerdlugssuit in the north to south of Krebsesø. 4) The Riddersporen synform east of the Krebsesø antiform. 5) The Dioritnæs synform right on the south-east coast.

The Pâkitsoq Pyribolite and other rocks of the Pâkitsoq antiform.

The course of the Pâkitsoq Pyribolite delineates clearly the northern and central parts of the Pâkitsoq structure which is a composite disharmonious antiform. Within the bay of Ilordlia and between Vestsø and Østsø the Pâkitsoq layer forms two antiformal hinge zones which are separated by an intermediate synform, the Stjernesø synform, the closure of which is found just south of Stjernesø. East of the Pâkitsoqelv, the thick pyribolite contains a narrow layer of purple gneiss, which increases considerably in thickness within the Stjernesø synform where it is developed as a very homogeneous biotite gneiss of light colour. In the transition zone north of Ilordlia, the gneiss layer is squeezed out locally. The western antiformal closure within the Pâkitsoq Pyribolite lies on the 373 m mountain (south-south-west of Stjernesø) and on its southern slopes. The presence of this closure also explains the exceptional width of the Pâkitsoq Pyribolite in this region.

In the eastern antiform of the Pâkitsoq structure, the median gneiss layer is intricately folded together with the enclosing pyribolite which here splits up into several layers. The most persistent of the 'lowermost' of these fades out into mixed or agmatitic gneiss before reaching the north coast. Only the easternmost extends right to the coast.

The Pâkitsoq Pyribolite is usually developed as a somewhat banded rather fine grained rock (19273). South of Breddal, calc-silicate rocks were found as intercalations in the western flank of the pyribolite. As is usually the case along the Frame Layer, the pyribolite is also smallfolded at this locality. Typical shearfolding may also be seen, fig. 39. Just west of Stjernesø the smallfolding indicates a synformal structure within the Pâkitsoq layer. The branch forming the synformal closure at Glemte Høj consists of pyribolite which occasionally carries garnet. Ultrabasic rocks are common as inclusions of varying size. The structural relations of these ultrabasic bodies have not been studied in detail (as on Langø) except north-east of Stjernesø from where olivinitic and eclogitic rocks have been described (35841, 35842, and 35843).

The hypersthene gneiss which forms the core of the composite Pâkitsoq antiform is developed as a rather mixed, banded to streaky gneiss in the north (19274), but passes south into a very homogeneous hypersthene-bearing granofels (35872) which may contain scattered enclaves of calc-silicate rocks.

South of Ilordlia the antiformal structure may be traced over Ganghøj to Itivínguaq. Here the antiform is no longer composite since the discontinuous pyribolite layer south of Ilordlia is not affected by the Stjernesø synform. The gneisses overlying the Pâkitsoq Pyribolite belong to the same succession which has been described previously as the Frame Layer. In spite of careful search, no gabbro-anorthositic rocks have been found within the gneisses of the Pâkitsoq structure. The gabbro-anorthositic enclaves south of Ilordlia are associated with the short pyribolite layer. In the eastern flank of the big antiform, the gneisses become more and more homogeneous towards north-north-east. South-west of Ganghøj a rather thick pyribolite layer reaches the east coast of Pâkitsoq,

Fig. 39. Shear folds in the Pâkitsoq Pyribolite south of Breddal (Photo: GGU, A.B.)

but in the hinge zone of the antiform it passes into mixed gneisses and cannot be traced further on. At Itivínguaq, two pyribolite bands, the southern of which contains a prominent rust zone, are also developed within the hinge zone. An amphibolitic band which can be traced into garnet-biotite schist, has also been mapped across Itivínguaq in the eastern flank of the antiform. Further towards south-west the Pâkitsoq antiform dies before the Qâqârssúnguit mountain is reached.

The Krebsesø antiform and its flanking synforms

The complex nature of the Krebsesø antiform may be recognised by tracing the thick Krebsesø Pyribolite. Following this layer from the north coast (south-west of Kangerdlugssuit) towards Flankehøj, it can be noticed that the pyribolite layer splits up into two parts in the steep escarpment south of Buedal. The lower part closes antiformally on Midterhøj and continues via Buedal and Rullefjeld to the north coast. The upper part makes up Flankehøj proper and the southern slopes of this mountain. It reappears east of Krebsesø in some complex structures.

Fig. 40. Sheeting in light coloured gneiss at Skiverne (Photo: GGU, A.B.).

Between the upper and lower parts of the Krebsesø Pyribolite, a light coloured gneiss layer makes its appearance. On Midterhøj it overlies the partly gneissified and smallfolded lower pyribolite band. East of Krebsesø, it changes through purple gneiss into hypersthene gneiss, which latter occupies the core of a peculiar hinge zone around the 350 m mountain. North of Buedal, the same gneiss layer is locally developed as purple gneiss, particularly on the north coast, where it encloses ultrabasic boudins and a short band of garnet-biotite-sillimanite schist. Brecciated and pegmatite-veined larger ultrabasic enclaves also occur on the slopes north of Buedal.

The lower part of the Krebsesø Pyribolite is underlain by another important gneiss layer. On the western flank of the antiform, on Skiverne and around Buedal, it is developed as light coloured granitic gneiss. North of Skiverne, transitions into purple gneiss are locally found (19276, A and B, 19277, 19278 and 19279). Within the light coloured gneiss, sheeting may be strongly developed, as on Skiverne, see fig. 40. Here several large pyribolitic to amphibolitic enclaves are also found. As far as can be judged from the smallfolding noticed within some of them, they have retained their original position. In the eastern flank of the antiform, the gneiss layer is made up of purple to in part hypersthene-bearing mixed gneisses.

Fig. 41. Closure of the Krebsesø antiform at Buedal, viewed from north (Photo: GGU, A.B.).

Underlying these gneisses, a discontinuous pyribolite layer delineates the core of the Krebsesø antiform. On the western flank, the pyribolite overlies conformably a band of ultrabasic rocks (olivinite and even dunite). In the closure, just south of Buedal, gabbro-anorthositic rocks are associated with the pyribolite which here encloses several large ultrabasic bodies (boudins), fig. 41. In the eastern flank the pyribolite layer is partially sheared out. The actual core of the antiform is made up of purple gneisses, which within the hinge zone pass into light coloured granitic gneisses.

The occurrence of the conformable ultrabasic band in the lower part of the pyribolite is a remarkable feature, since elsewhere in Tovqussap nunâ the ultrabasic rocks always occur as disrupted bands, boudins or as accumulations within hinge zones (cf. Langø).

West of Flankehøj a synformal hinge zone is shown by a pyribolite-diorite layer. The synform structure is also evident from the highly smallfolded closure within an upper and thinner pyribolite band. This thin layer can be traced into the broad antiformal Krebsesø structure in the escarpment south of Krebsesø. Here it is more banded, smallfolded and partly gneissified. To the east, on the slopes above Sangmissoq, it can be followed into the Riddersporen synform.

The continuous trace of this thin layer is rather important, since it helps in the understanding of the complex relations around Krebsesø. On the 300 metre ridge east of this lake, two hinge zones are found within a thick pyribolite layer, which now can be correlated with more

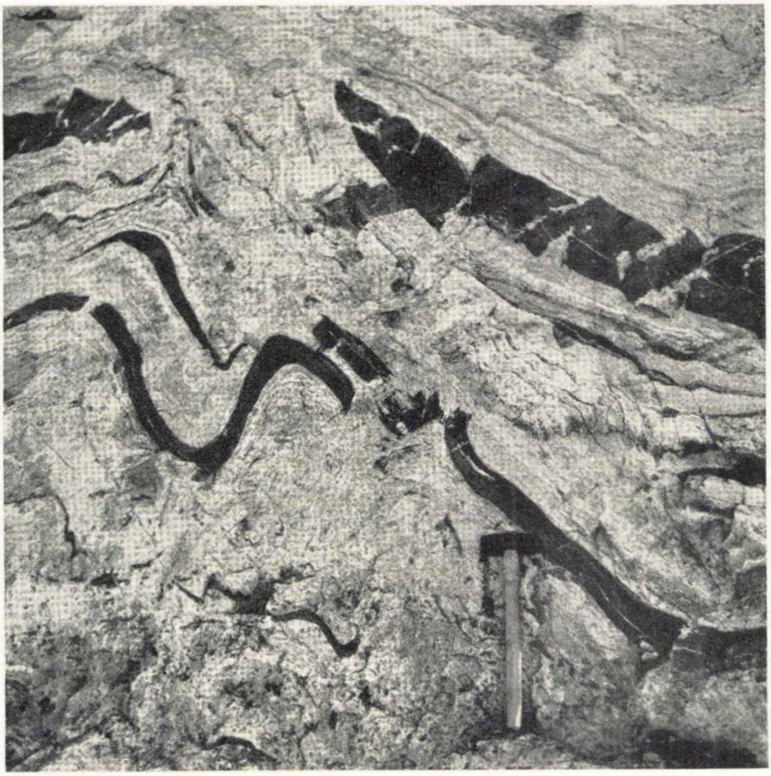

Fig. 42. Small scale structures in gneiss, escarpment south of Krebsesø (Photo: GGU, A.B.).

confidence with the Krebsesø Pyribolite (upper part). As will be shown later, the northern hinge zone belongs to a refolded recumbent synform, and we may expect that, due to refolding, it reappears under the west part of Krebsesø. This also means that the lower pyribolite-diorite layer in the Flankepas synform is equivalent to the Krebsesø Pyribolite (see the map, Pl. 2).

Above the thin pyribolite which delineates the Krebsesø antiform and its two flanking synforms, another pyribolite layer has been mapped continuously from the west coast of Sangmissoq to south of Pâkitsoq. Where forming part of the Krebsesø antiform, this layer is strongly banded and partly gneissified. Like other layers within the hinge zone, it is cut by metre-broad granitic pegmatites. In the Flankepas synform, the hinge zone is affected by the Gammel Lejrskar-Stjernesø-Krebsesø fault. Since the pyribolite closure reappears just north of the said fault, a downthrow of the northern fault block has taken place.

On the northern slopes of Kronehøj a light coloured gneiss separates the two above-mentioned pyribolite layers. Like the lower pyribolite,

Fig. 43. Small scale structures in gneiss, near the locality pictured in fig. 42 (Photo: GGU, A.B.).

this gneiss is strongly smallfolded and the thin basic bands have been disrupted, figs. 42 and 43. The style of this folding is somewhat reminiscent of the small scale structures described from the hinge zone on western Tugdlerúnarssuit (p. 91 and fig. 29), since at both localities the smallfolds are affected by more or less vertical zones of plastic movement. The Kronehøj type may, however, represent a more advanced stage with superimposed granitisation and deformation. Common to both types is the universal axial parallelism in the structures, a feature which could hardly be expected if the folds were developed by turbulent flow due to anatexis.

The broad pyribolite layer of northern Kronehøj (the Kronehøj Pyribolite) has also been mapped for a considerable distance. In the Riddersporen synform it is divided into an upper and lower part by a median gneiss layer. This latter wedges out on Riddersporen, where biotite-garnet-bearing schist is developed along the south-eastern border of the pyribolite. West of Kronehøj, the pyribolite layer is only slightly affected by the Flankepas synform. In the steep walls of the western spur of Kronehøj a tight closure, which takes the layer back across Kronehøj, was mapped. North-east of Kronehøj, the Riddersporen synform causes a renewed closure in the layer, which then continues more or less parallel to the east coast until Itivínguaq.

The course of the pyribolite on Kronehøj shows that the Krebsesø antiform dies out here. The original pyribolite layer is strongly gneissified and shows abundant folds, which all indicate a relative tectonic transport towards west. Smallfolding is also prominent in the eastern synformal closure where a few gabbro-anorthositic enclaves were found associated with the partially gneissified pyribolite. East and north-east of Sârdlup qâva, the pyribolite layer contains a thin median gneiss band.

South of Kronehøj and east and west of Sârdlup qâva, more or less mixed hypersthene gneisses are found, whereas the gneisses in the centre of the structure around Sârdlup qâva are very homogenous purple to hypersthene-bearing types. At the lake, east of the said mountain, the hypersthene gneiss contains small garnets.

East of Sârdlup qâva and east of the Kronehøj Pyribolite mixed in part hypersthene-bearing gneisses are found. They change into the more homogenous purple gneisses and granofelses of the east coast (14976, A and B).

At Dioritnæs several thin pyribolitic to amphibolitic layers occur. Their course describes a gentle synform, which may have formed part of a large complex structure now hidden under the sea. This possibility is suggested from the way the pyribolite layers split up and join, a feature possibly controlled by isoclinal folding.

V. THE OCCURRENCES OF DIORITIC ROCKS

Structurally, the dioritic rocks of Tovqussap nunâ may be divided into the following four groups:

1. Conformable layers and bands associated with pyribolite.
2. Semi-concordant to clearly transgressive bodies of lineated to almost structureless diorite which have been formed by replacement preferably of pyribolite occupying hinge zones.
3. More irregular bodies and dykes, which in spite of their cross-cutting nature, may have been formed by replacement processes.
4. Cross-cutting diorite aplites which may show dilational emplacement tectonics.

All the mapped occurrences of dioritic rocks are shown on plate 2 by red shading. The mode of "mise en place" of the diorite being dependent on regional as well as local physico-chemical conditions, the four structural types may represent a succession in time, but need not necessarily do so (cf. the Krebsesø diorite below). As appears from the descriptions to follow, the dioritic rocks were formed syn-, late-, or postkinematically in relation to a phase of folding occurring under granulite facies conditions. Thus they were developed as pyroxene diorites. The uralitisation and other retrograde alterations of the original pyroxene diorites were related to a later deformation under amphibolite facies conditions. In the following description of some of the typical occurrences, most attention is paid to the pre-uralitisation features.

1. Conformable layers

As an example of a conformable diorite layer, the diorite band forming part of the Great Pyribolite on the east side of Langø may be mentioned. A description of a sample from this band has been given above (19137, Intermediate Division). Where studied in detail on northern Langø, this layer exhibits a faint banding due to intercalations of leucopyribolite and calc-silicate rocks as 1—10 cm thick bands. Discontinuous rust zones have also been noticed. Close to the western contact with the enclosing pyribolite, abundant elongated basic and ultrabasic enclaves and lenses are found. Usually they vary in size from 10 centimetres to about half a metre. They all seem to be stretched (elongated) parallel to the fold axis. The eastern contact with the pyribolite is very sharp, but when traced in detail it may be seen to run in a zig-zag manner suggesting that the two units were folded together. The ultrabasic enclaves of this layer have previously been studied by H. SØRENSEN (1953).

The following observations point to the synkinematic origin of this diorite layer:

i. Its generally conformable relations and the foliated nature of the rock.
ii. The isoclinal smallfolding of the thin basic schlieren enclosed within the diorite.
iii. The apparent folding of the eastern contact.
iv. Pinch and swell and boudinage structures developed within the diorite in bands of slightly different composition.

The last feature, which can hardly be explained as a ghost structure, is particularly good evidence of the synkinematic nature of the diorite.

2. Semi-concordant to transgressive bodies

The second group comprises the diorites found at the following localities: Nordnor (Langø), Verdens Begyndelse, east of Qagssiarssuk, south of Ørenæs, Gule Hav and Krebsesø. Three of these localities have been selected for more detailed description.

The Nordnor diorite, Langø

As indicated by the geological map of Langø (fig. 12), diorite is exposed both north and south of the Nordnor. If the diorites in the two areas are assumed to be connected under the sea, the total occurrence is about 600 metres long and up to 100 metres broad. These dioritic rocks and the conformable layer on the east coast have previously been mentioned as hypersthene-bearing, almost pegmatitic gneiss and hypersthene-gneiss respectively by H. SØRENSEN (1953). The dioritic nature of these rocks was first detected by the present author in 1953. Samples from the Nordnor diorite have been described above along with the rocks of the Intermediate Division (19141 and 19145-three slides). They all consist of more or less uralitised pyroxene diorite. The mafics—including the pyroxenes—contribute to the lineation. In the field, the diorite is red-brown on the weathered surface. In fresh specimens it is greenish grey to more typically flesh coloured. The latter type owes its colour to the feldspar. Where it has been subject to retrograde alteration the diorite is light or nearly white. Apart from the lineation, the dioritic rocks are very homogeneously developed and display a distinctly "plutonic" appearance. They may contain scattered thin diopside-hypersthene-bearing veins. Enclosed by the diorite and alien to this rock, scattered inclusions of basic to ultrabasic rocks are found. Before the structural significance of these enclaves is discussed, it may be useful to describe the contact relations.

South of the Nordnor, the eastern contact is fairly well exposed. To the south-east, the diorite is in contact with purple gneiss. The contact is conformable, except where a wedge of gneiss locally extends into the diorite. Along the contact plane a few centimetres thick medium to almost coarse grained plagioclasitic rock with scattered hypersthene grains is found. Because of the dip (ca. 60° to ENE) of the contact this transition rock partially covers the steep eastern flank of the N—S ridge formed by the diorite. Abundant thin rectilinear quartz-feldspar pegmatites, which cut the contact plane, further complicate the relations.

Further to the north and again near the south coast of the Nordnor, the diorite borders on pyribolite. No plagioclasitic transition zone is seen here, but schlieren and small enclaves of the basic rock are found in the neighbouring diorite.

North of the Nordnor, the contact is exposed along its entire length. The diorite is bordered by a layer of calc-silicate-rich pyribolite (S of fig. 12) which to the east is succeeded by a narrow layer of hypersthene gneiss with smallfolded purple veins and basic schlieren. The diorite close to the contact is somewhat heterogeneous, due to more pegmatitic veins and patches which may also traverse the basic enclaves. To the north the diorite becomes transgressive in relation to the skarn-rich layer, which latter is seen to have been gradually "dissolved" into an agmatitic breccia with rather angular fragments separated by more coarse grained diorite (cf. Berthelsen, 1957, fig. 11).

The western contact is only exposed south of the Nordnor, where the diorite overlies a light coloured flaser gneiss with a steep easterly dip. Observations on the original contact relations are, however, difficult to make since a mylonite with a lower easterly dip runs more or less parallel to the contact. Where the mylonite makes a narrow angle with the contact, several sub-parallel zones of movement may be found (fig. 44) and here the original contact is preserved for a short stretch. However, it is still difficult to trace, since the mylonitisation has been accompanied by alterations resulting, amongst other things, in a bleaching of the diorite. This locality is also of interest because of the occurrence of a thin meta-dioritic aplite in the gneiss (fig. 44). This aplite has been shear-folded in response to the mylonitic movements and transformed in

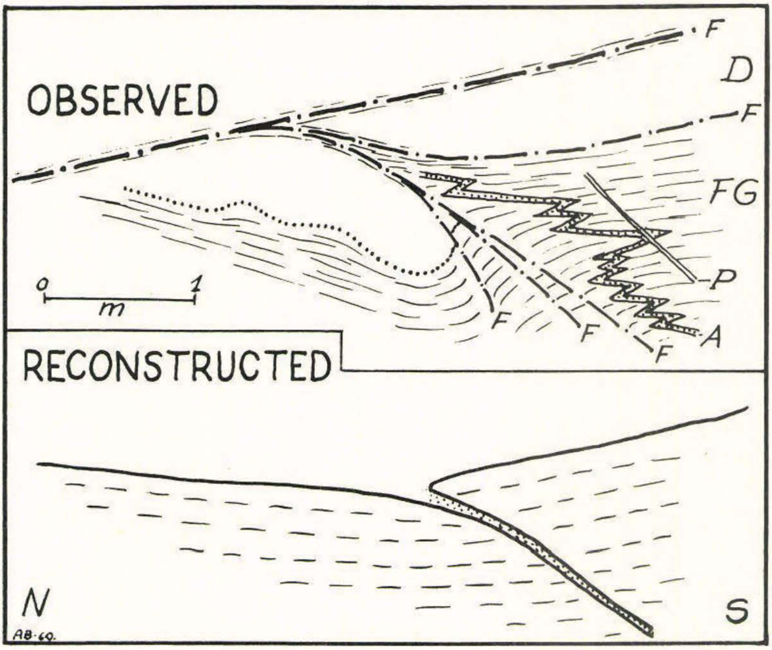

Fig. 44. Contact relations of the Nordnor diorite.

the same way as the surrounding rocks. It is cut by an undeformed microcline-bearing pegmatite. Fig. 44 shows an attempt to reconstruct the contact relations before the mylonitisation. This reconstruction suggests that the aplite represents an apophysis originally given off from the main diorite.

A little further to the south, an example of an older pegmatite is found. It carries quartz, plagioclase, microcline and biotite. The last mineral may form centimetre large flakes. This pegmatite is older than the mylonitisation, since it has been kinetically sliced together with the gneiss and the diorite. Presumably it was originally developed due to migration of quartzo-feldspathic material derived from the gneiss into a fracture within the diorite body. This mode of formation is suggested by the flexure in the gneiss banding around the root of the pegmatite. A similarly developed pegmatite in an ultrabasic rock from western Langø was described above (cf. H. Sørensen, 1953, fig. 8).

The diorite south of the Nordnor contains scattered enclaves of pyribolitic and ultrabasic rocks. Close to the western contact (just described) these are about 10 centimetres across, but in the central portions of the diorite they may attain a size of several metres. These enclaves, some of which may be quite angular, seem to "float" in diorite (fig. 45). The old internal fold structures of the enclaves are cut off by the diorite and the lineation in the latter may trend obliquely to the

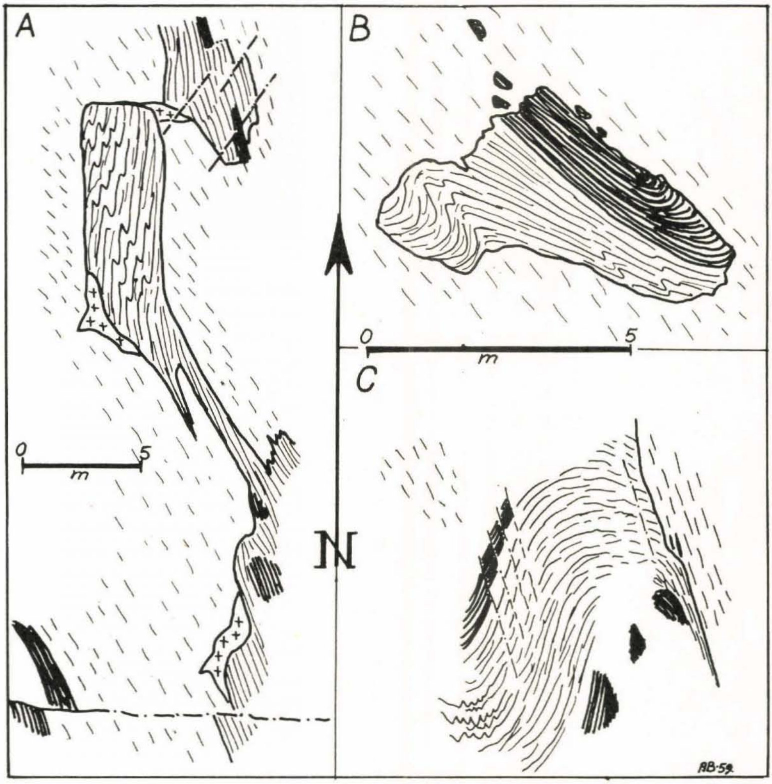

Fig. 45. Enclaves in the Nordnor diorite.

border of the enclaves. Thus it might be inferred that these enclaves are xenoliths or roof pendants. This, however, would not be correct.

As a first approach to the investigation of the spatial orientation of the enclave structures in relation to those of the rocks surrounding the diorite, the author measured the axes of the folds found in some of the inclusions. These axes turned out to be parallel to those measured in greater number within the enclaves of the agmatitic breccia north of the Nordnor. They are also almost parallel to the axes measured in the surrounding gneissic and pyribolitic rocks (fig. 13, Berthelsen, 1957), the axes obtained from the enclaves showing a slightly larger and somewhat more southeasterly plunge than those measured in the surrounding rocks.

In order to decide whether this variation was fortuitous or not, additional readings of the foliation within the enclaves were taken on a later visit. These readings were obtained from the enclaves shown in fig. 45, B and C, the only two enclaves which contained measurable folds with an appreciable amplitude. Although not statistically valid, this check indicated that the smallfolds measured agree with somewhat larger structures. Unfortunately the enclaves are too scattered and too few to allow a reconstruction of the major structure of which the small-folds undoubtedly once formed part. The structures in the surrounding rocks indicate, however, that the Nordnor diorite occupies an antiformal position (see page 72) and that the contacts of the diorite are therefore more or less concordant. Taking these relations into account it seems highly improbable that the scattered and often quite small inclusions,

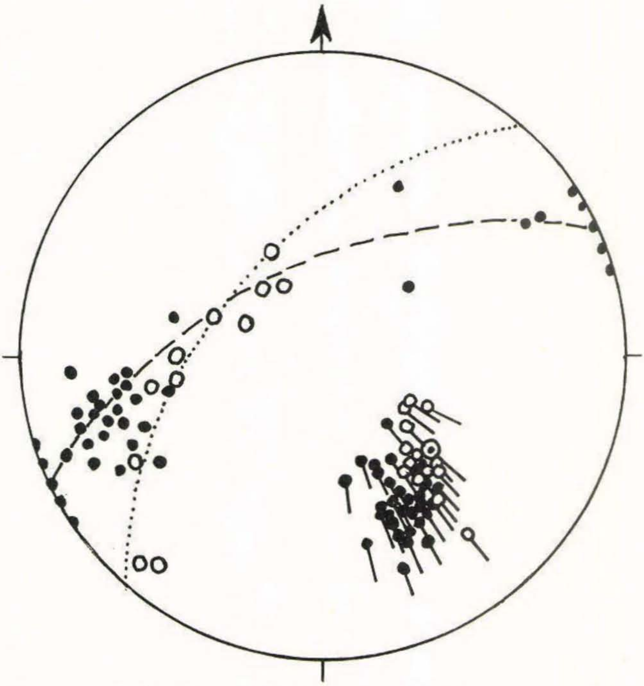

Fig. 46. Stereogram of structures within the enclaves and in the surrounding rocks (Wulff net, lower hemisphere, as in all succeeding stereograms).

which occur deep within the antiformally arranged diorite, represent roof pendants.

Compiling the available structural data into a stereogram (fig. 46) it becomes apparent that there really is a slight difference in the orientation of enclave axes and those measured in the rocks surrounding the diorite. In fig. 46 the open circles indicate poles of foliation within the enclaves, while the filled-out circles are foliation poles from the surroundings. The axes with open heads were measured within the enclaves — mainly north of the Nordnor. The axis with a dot in the head is constructed from the foliation poles defining the stippled great circle. The axes with the black heads were measured in the surrounding gneisses and pyribolites. There is a close agreement between the two axes maxima and the corresponding zones of foliation poles.

On Langø the fold axes vary slightly, the northern axes plunging steeply to SE, while the southern plunge at a lower angle to SSE. This variation may partly explain the difference in orientation of enclave axes and the axes of the surroundings, since the readings of the former come from northern Langø, while the latter were measured mainly in central and southern Langø. There might also, however, have been a slight original axial disharmony between the antiformal structure, now occupied by the diorite, and its surroundings.

The fact that all the enclave structures have preserved their original orientation parallel to each other remains unaffected by this problem just discussed. The scattered and quite independent enclaves having preserved their original position strongly suggests that the diorite was formed by isospacial replacement processes — preferably from pyribolite, to judge from the preserved relics.

Because of the smooth and rounded shape of the diorite ridge, reliable measurements of the lineation in the diorite are difficult to obtain. As far as could be ascertained, however, the lineation is oriented parallel to the fold axes of the enclaves. The fact that the pyroxenes, hypersthene and diopside, show preferred linear orientation suggests that this lineation was developed during the actual process of dioritisation. Thus the formation of the diorite seems to have taken place towards the decline of the movement which caused the folding now visible in the enclaves, i.e. the dioritisation may be referred to as latekinematic. The transformation of the pre-existing rocks into diorite must have been a process which did not cause any significant increase in plasticity. Otherwise, the angular enclaves would have been oriented according to their outer shape. The original predominance of water-deficient minerals in the diorite also suggests that the dioritisation was a rather dry process which took place under granulite facies conditions.

It should also be stressed that the composition of the enclaves is always more basic than that of the enclosing diorite. Leucocratic rocks may be in contact with diorite, but have never been found as enclaves within this rock. This applies also to all the other dioritic rocks of Tovqussap nunâ. In the author's opinion, it supports the idea of a metasomatic origin of the diorites, since leucocratic xenoliths are quite often found in basic igneous rocks derived from magmas which would be hotter than, and in many cases just as dry as, a dioritic magma.

The diorite at Verdens Begyndelse

The location of this occurrence may best be seen from fig. 11. Petrographically, this diorite resembles the Nordnor diorite, but may be slightly more coarse grained. Besides basic and ultrabasic inclusions, gabbro-anorthositic to pure anorthositic enclaves are commonly found at Verdens

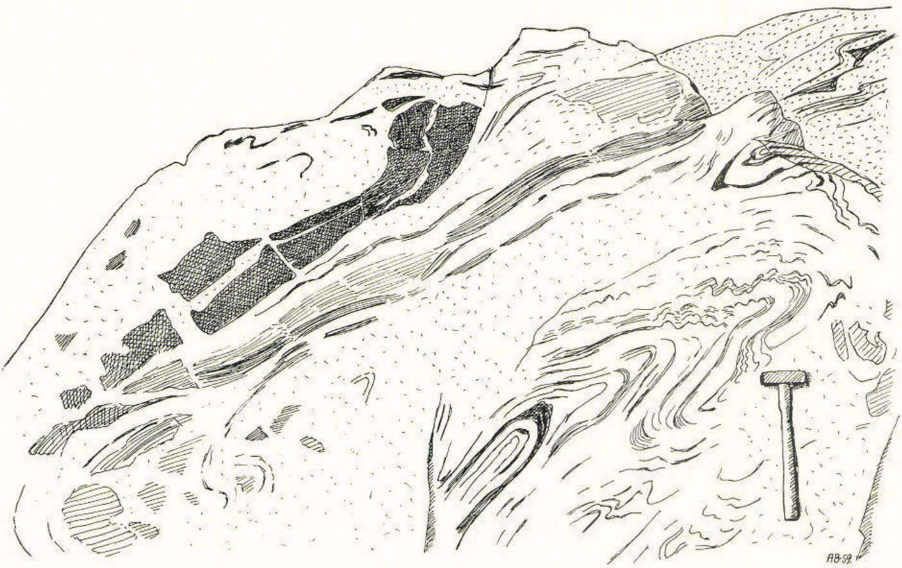

Fig. 47. Enclave structures in the diorite at Verdens Begyndelse.

Begyndelse. To avoid any confusion, it should be stressed that these anorthositic rocks, on account of their calcic plagioclase and fine, granular texture, are easily distinguishable from the plagioclasitic rocks which, for example, are associated at Dioritnæs with the dioritic rocks. The anorthositic rocks at Verdens Begyndelse correspond completely to the anorthositic to gabbro-anorthositic enclaves found in the surrounding gneisses. Very similar anorthositic enclaves have also been described from a diorite body south of Godthaab (Berthelsen, 1955).

At Verdens Begyndelse, only the eastern contact of the diorite can be studied, the western being covered by the sea, the southern fault-determined and the northern occurring on an inaccessible cliff. To the east, banded, streaky to smallfolded gneisses border with a conformable but sharp contact on the diorite. The gneiss dips steeply to the east, the smallfolding being nearly isoclinal.

The abundant enclaves within the diorite seem at first sight to be quite haphazardly arranged, but a closer study revealed that they are all arranged in an orderly way. Due to their relatively small size (10 cm to about half a metre), they seldom show internal fold structures, in contrast to the enclaves at the Nordnor. Only their overall arrangement betrays the existence of a relic fold pattern, which is characterised by nearly horizontal NW—SE trending axes and a very complicated style with development of recumbent smallfolds, fig. 47. Once understood, this pattern, forming a gentle antiformal arch, may be traced throughout the diorite. When the eastern gneiss, which occupies the flank of this structure, is met, the complicated hinge style is overprinted by a younger folding causing the steep easterly dip. This younger folding corresponds to the superimposed small scale structures described from the Pas Pyribolite south-west of Kolbesø (see page 79). East of the diorite border a small isolated outcrop with a very interesting rock association is found, fig. 48. Smallfolded gabbro-anorthositic to pure anorthositic bands are enclosed here in and more or less replaced by a dark brownish hypersthenite, which latter, when compared with the surrounding banded and enclave-bearing gneiss, has taken the place structurally of the gneissic

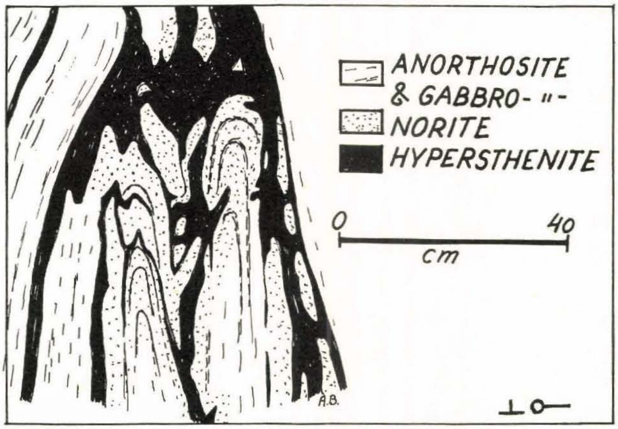

Fig. 48. Hypersthenite associated with the diorite at Verdens Begyndelse.

component. The hypersthenite not only destroys the fold pattern and is uninfluenced by deformation, but also may form small transgressive veins in other parts of the outcrop. Where the gabbro-anorthosite bands are more or less replaced by the hypersthenite, a transformation of the former into granular norite may be noticed. The formation of the hypersthenite, therefore, is undoubtedly a process which is younger than or at least outlasted the nearly isoclinal smallfolding of the gabbro-anorthositic bands. This conclusion may have some bearing on the dating of the neighbouring dioritisation. In the outcrop described, there is exposed a "negative", so to speak, of the process of dioritisation. Knowing that the dioritisation (at Dioritnæs, see below) was accompanied by a contemporaneous segregation, which lead to the formation of ultramafic schlieren and ultrabasic pods *within* the diorite, the author is inclined to consider the "ultrabasification" and the dioritisation at Verdens Begyndelse to be closely related. The formation of the hypersthenite may, however, belong to a late stage of the dioritisation. The preservation of the complicated hinge zone structures within the main diorite at Verdens Begyndelse suggests that the dioritisation here occurred more or less contemporaneously with the youngest folding which caused the refolding in the eastern banded gneisses. As indicated by the undeformed hypersthenite veins the metasomatic processes may have outlasted the refolding locally. In consequence, the dioritisation at Verdens Begyndelse may be dated as latekinematic.

The Krebsesø diorite

From a structural point of view, the diorite occurrence at Krebsesø may be divided into two parts.

i. The conformable diorite layer found within the Flankepas synform north-west of the lake.

ii. The locally transgressive diorite, which occurs within the hinge zone of the Krebsesø antiform, south of the lake.

The diorite of the western synform is developed as a thick layer which overlies pyribolite. Close to the border against the diorite, the pyribolite contains several ultrabasic lenses.

The diorite occupying the hinge zone of the Krebsesø antiform has a direct structural connection with the conformable layer, but exhibits quite different border relations. In the small lobe-shaped fold south-west of Krebsesø, the diorite is more melanocratic than usual. Possibly some older ultrabasic rocks were involved in the dioritisation at this locality. East of this smaller fold, a younger fault marks the border of the diorite for about 300 metres. Further east the diorite transgresses upwards and, in the shape of clean cut dykes, reaches the overlying thin pyribolite layer. In the few metres thick dykes the diorite is more fine grained than usual. Where cut by the dykes, the thin pyribolite layer is strongly folded and the competent ultrabasic inclusions are rotated. Tracing the dykes downwards on the escarpment, the author found that they passed gradually into the underlying diorite. On the larger point on the south coast of the lake, several amphibolitic and pyribolitic enclaves were found in the diorite. Judging from the orientation of the foliation within these enclaves, they all seem to be in their original position. On the northern tip of the point and small island east of it, the diorite carries some quartz. This could be explained by a retrograde influence from the younger fault nearby.

Structurally, the Krebsesø diorite takes the place of the Flankehøj Pyribolite. The greater part of the diorite has undoubtedly been formed by in situ replacement of this layer, but the diorite so formed has also been capable of transgressing through the overlying gneiss layer into the thin pyribolite (as seen in the mountain wall south of the lake).

During its transgressive phase the diorite migrated in some way or another into fractures in the overlying rocks of the hinge zone. This migration took place when the overlying rocks had lost their ability to yield by plastic deformation. Since the dioritisation is not only lithologically but also structurally controlled, the process may be dated as late-kinematic.

As mentioned in the beginning of this chapter, the different modes of occurrence of the dioritic rocks depend on the physico-chemical conditions prevailing during their formation. The different types, conformable layers, semi-concordant to transgressive bodies, irregular bodies, dykes and aplites may correspond to a succession in time (syn-, late-, to post-kinematic), but local variations around or within one and the same occurrence may also cause the simultaneous development of several of the types. The Krebsesø diorite serves as an example of an occurrence with variations in the local environment. The shear-folded aplite west of the Nordnor diorite (fig. 44) was most probably developed at the same time as the main body of diorite. This latter, however, is cut by younger dioritic dykes and aplites which reflect a changing regional environment (cf. fig. 55).

3. Irregular bodies and dykes

The group of more irregular bodies and dykes comprises the following occurrences: the Kosakfjeld, the Ankerbugtelv, the Lille Ekkodal occurrences, the Dome dyke (south of Tovqussaq Mt.) and the Dioritnæs occurrence.

The Kosakfjeld diorite

This occurrence is about 400 metres long and about 100 metres broad. The main rock is uralite diorite, (4042, 13416). The body is elongated in a northeasterly direction. The northern end abuts on one of the Ankerbugtdal faults with a considerable drag close to the fault which here dips at about 55° SE. Just west of the top of Kosakfjeld (143 m) a large ultrabasic body is enclosed in the diorite. On the south-west facing slopes of the mountain, the diorite contains several quite irregular pyribolitic enclaves, which may be more than ten metres large. These ultrabasic to basic enclaves are "exotic" blocks, since—unless it is postulated that the Little Pyribolite describes a complex fold in the area now covered by the Ankerbugt and Inderhavn—the presence of these rocks right within the Frame Layer cannot otherwise be accounted for. The drag against the Ankerbugtdal fault indicates that the eastern fault block has been shifted to the north and it seems unlikely therefore that the Kosakfjeld diorite has ever been connected with the dyke-shaped diorite occurrence west of the Ankerbugtelv. Instead of attempting risky reconstructions, the author prefers to leave the problem of the provenance of the enclaves at Kosakfjeld unsolved. It should be added that the field observations do not allow the enclaves to be interpreted as transported xenoliths. In the giant agmatite formed by the basic enclaves, the replacive nature of the diorite can easily be seen.

The Ankerbugtelv diorite

This occurrence is poorly exposed, but seems to be dyke-shaped. In the field the impression was obtained that, in the Great Pyribolite, the dioritisation had operated quite irregularly within the "dyke", thereby causing the formation of a large scale agmatitic structure with pyribolitic enclaves of all sizes surrounded by rusty brown diorite.

The Lille Ekkodal diorite

Occurring in the steep cliffs north of the valley, this is rather inaccessible. The lower part of the slope is covered by gravel and scree from the easily weathering dioritic rocks.

The Dome dyke

This dyke (north of Qaersup ilua) can be traced for more than two kilometres from the north-west of Qaersup ilua to south and east of Tovqussaq Mt. Thus it cuts across the northern part of the dome and traverses the Frame Layer before it ends at the Great Pyribolite. All along its course the dyke seems to be vertical or sub-vertical. The width varies from three to well over ten metres. Due to the crumbling nature of the diorite when exposed to weathering, and the distinct yellow-brown to rusty colours of the gravel, the dyke can be seen on aerial photographs (except within the core rocks of the dome) in spite of its relatively small thickness. Where the core rocks are granitised, the Dome dyke is transformed into quartz-bearing biotite diorite. In the Frame Layer, the dyke rock is pyroxene to uralite diorite. East-south-east of Tovqussaq Mt. the dyke dies out, but continues again in several short lens-shaped bits. The irregularities are too random to be called en echelon structures. Branching of the dyke has also been noticed. In the same region it was observed that the lineation and trend of the enclosing gneiss could also be found within the diorite, here indicated by the preferred orientation of the mafic minerals. Another interesting feature deserves to be mentioned. As shown on the map of Pl. 1 the Dome dyke cuts the Interior Pyribolite, but traverses neither the Little Pyribolite nor the Great Pyribolite although occurring between them.

All these features could be explained by assuming that the diorite dyke was formed by replacement of the wall-rock around an initially somewhat irregular joint or fissure. Such an origin would explain the leucocratic composition of the Dome dyke.

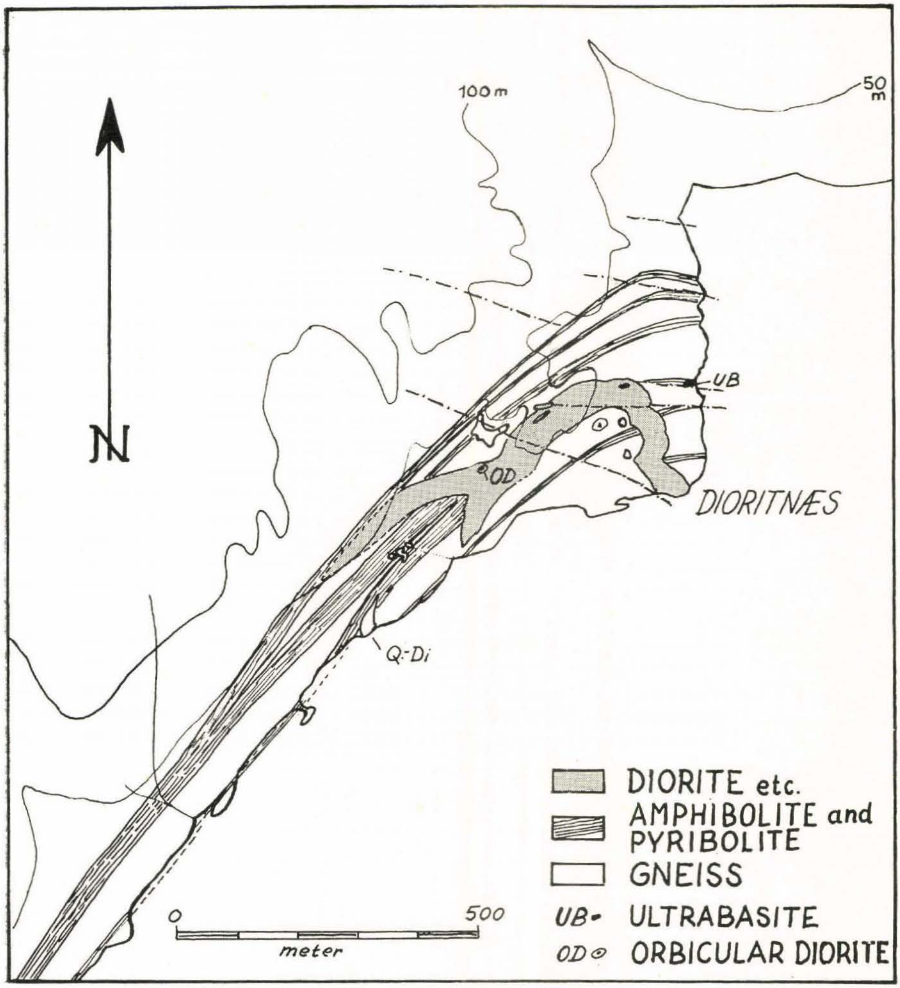

Fig. 49. Geological map of the Dioritnæs diorite.

The occurrence of diorite at Dioritnæs

Dioritnæs, a small promontory on the south-east coast of Tovqussap nunâ, has been named from an interesting occurrence of dioritic rocks. The promontory consists of low rounded hills separated by strike- and fault-valleys. Except for the coast section, the exposures are not very good when compared to the general high standard of the outcrops within the Tovqussaq region. The diorite has weathered into a coarse gravel which covers the hill slopes and fills the valleys.

During his visits to this locality in 1953 and 1954, the author formed the impression that the diorite and associated rocks occurred as a conformable body formed by replacement of pre-existing amphibolitic rocks where these occupied a hinge zone (BERTHELSEN, 1957, fig. 12). When, in 1957, the author revisited the locality to study more closely the orbicular rocks found here earlier, he soon realised that in places the diorite behaved as a transgressive body. The promontory, therefore, was re-mapped in 1 : 10.000. The resulting map is shown in fig. 49 on a reduced scale.

The external structure of the diorite

From fig. 49 it may be seen that the diorite forms an irregularly shaped dyke. The contact with the surrounding rocks, gneisses and pyroxene amphibolites, is neat and in several places clearly cross-cutting. To the east, in the coast section, the dyke is vertical, but further inland it dips to the south-west, as can be read from the map where the configuration of the dyke is controlled by the small hills. The dyke swings into an almost northeasterly strike to the north of an isolated hill (just above 50 m) on the promontory. As far as the exposures and the low topography enable one to judge, the dyke here has a medium to steep southeasterly dip. To the west, the dyke splits into a short southern branch which dies out before it reaches the coast and a somewhat more persistent northern part. The exact termination of the latter could not be mapped, as outcrops are scarce in this area.

A separate body consisting of quartz-diorite was moreover observed on the coast south-west of the main occurrence. Inland, the quartz-diorite wedges out, but seawards it may well continue as a dyke below the sea.

The internal structures of the diorite

A division of the internal structures can be made into: a) structures inherited from older rocks, and b) structures developed along with or later than the formation of the diorite. The first group comprises structures within inclusions of older rocks in the diorite. Two larger inclusions of pyroxene amphibolite are seen in the northern part of the main dyke, see fig. 49. The foliations measured within these inclusions conform well with the gentle synform formed by the surrounding rocks. The outcrop with the orbicular diorite, which will be described and discussed independently below, is found in the southwestern continuation of these inclusions.

In the well exposed coastal cliffs more detailed studies could be carried out on the main dyke. About ten metres from the eastern contact, several seemingly irregularly orientated inclusions of pyroxene amphibolite are found (fig. 50). A closer analysis of the orientation of the foliation within these inclusions revealed, however, that the foliation shows a well defined relic fold structure. This becomes very clear when the inclusions are viewed in the direction of the fold axis of this structure. Otherwise, the orderly arrangement is difficult to understand as the axis

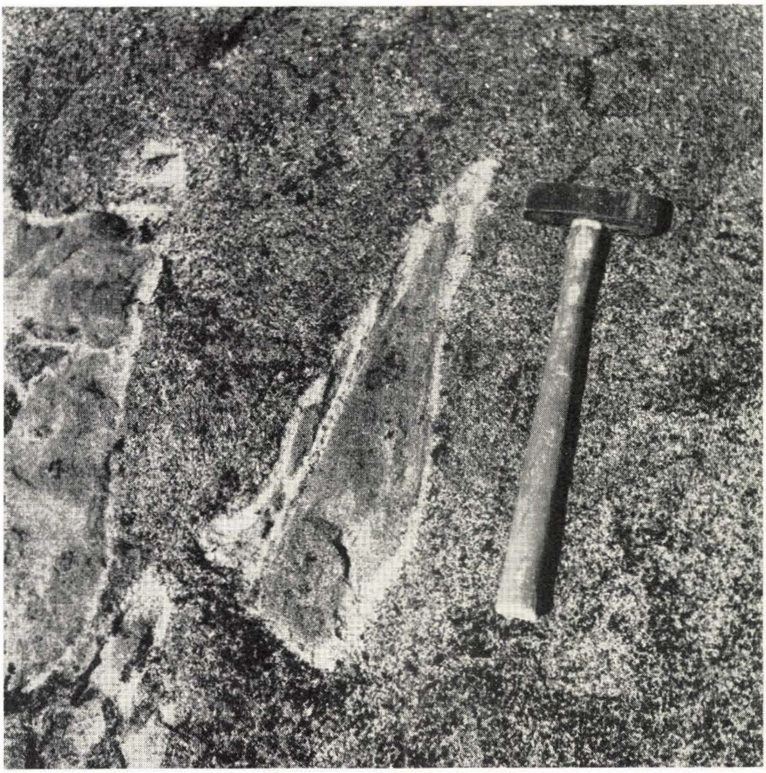

Fig. 50. Amphibolitic enclaves with light reaction rims, Dioritnæs (Photo: GGU, A.B.).

is more or less parallel to the slightly undulating surface of the exposure. A nearby vertical surface gives a better impression of the orientation and allows one to state that the leuco-amphibolitic inclusions really are inclusions, i.e. are separated by diorite, and not pseudo-inclusions (Berthelsen, 1957, fig. 14).

These observations are very important as they imply that the diorite separating the inclusions has been formed by replacement. In an attempt to decide if the fold structure preserved within the inclusions conforms structurally to the folds of the surrounding rocks, measurements were taken of the foliation seen in some of the inclusions. These readings are plotted in stereogram V of fig. 58 together with readings on foliations and axes from the surrounding gneisses and pyroxene amphibolites. Although the total number of readings is rather small, a pattern of two nearly parallel axes and a dispersed zone around the two correponding great circles becomes evident. The foliations measured within the pyroxene amphibolitic inclusions are all grouped within this zone.

This seems to indicate that the relic fold structure occurs in situ within the diorite dyke and in consequence that the diorite has been formed by isospacial replacement processes, even if the external shape of the dyke points towards a magmatic intrusive mode of emplacement.

For comparison the readings on the banding (described below) seen in the diorite have been plotted in addition, fig. 58, V. These structures, which can be dated as having been developed along with the formation of the diorite and which bear a close relationship to the dyke-shape of the diorite, cannot be matched with any pre-existing structures.

It may be argued that from a statistical point of view the total number of readings is far from being sufficient to permit the conclusion drawn above. However, when applied to structures defined otherwise (e.g. by marker horizons), the stereographic method may be used safely, when few but varied readings permit a zone (great circle) to be defined.

Apart from scattered occurrences of smaller amphibolitic inclusions, inclusions of another type were noticed in the central parts of the dyke. These consist of homogeneous ultrabasic rocks and therefore could not supply any evidence to the problem discussed above.

The structures which were formed along with and later than the formation of the diorite comprise internal variations in composition of the dyke, a peculiar type of banding and a foliation. In the coastal section, a zone about 1 m wide of anorthositic rocks occurs along both contacts. At the eastern contact, which can be studied best, the purple, homogeneous gneiss contains, close to the dyke, small (10—20 cm), elongated lenses of anorthosite. The lenses of anorthosite within the older "wall rock" pose a problem, if a magmatic origin is claimed for the diorite dyke.

The anorthositic border gradually passes into the main rock of the dyke — the diorite, which here contains mafic schlieren up to 1 m long elongated parallel to the contact. Further to the west, the diorite is very homogeneous but exhibits all the same a weak but measurable foliation due to the arrangement of the mafic minerals. This diorite encloses the group of pyroxene amphibolitic inclusions described above. Where the diorite borders these inclusions, a narrow reaction zone (1—2 cm), with a more pegmatitic leuco-dioritic rock, is seen.

The western parts of the dyke provide a spectacular display of very different rock types and structures. Rocks ranging in composition from pure anorthosite to ultrabasite are found, and banding may be developed. In one place a transition from diorite into ultrabasite is seen. The intermediate stage is represented by a speckled diorite with the feldspar forming light coloured round spots (2—3 cm in diameter). The ultrabasic rock contains an amphibolitic inclusion similar to those found in the surrounding diorite. The impression is thus obtained that the ultrabasic type has formed by a segregation process and can be compared to the mafic schlieren referred to above. It has, however, no genetic connection

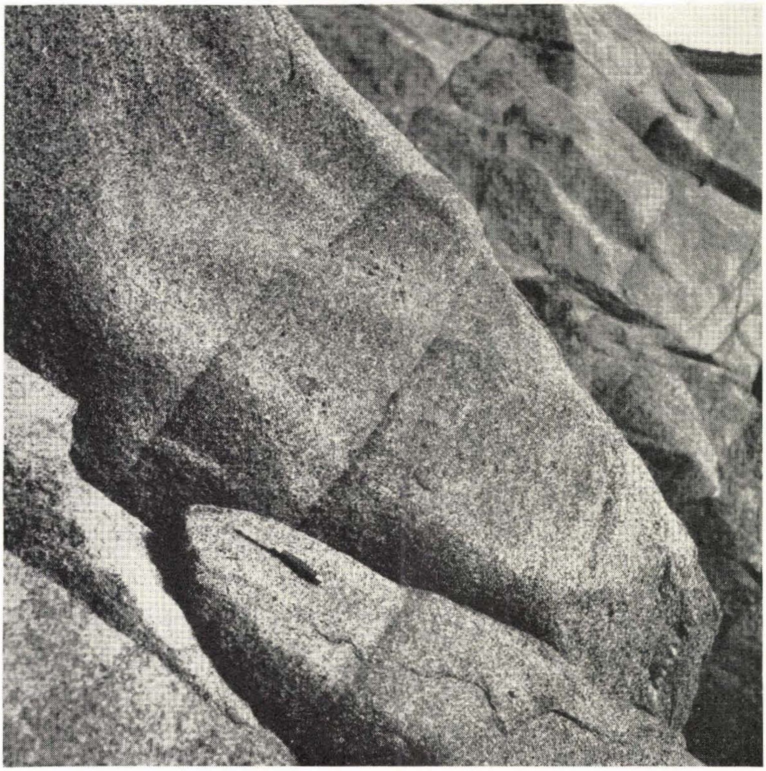

Fig. 51. Banding in diorite, Dioritnæs (Photo: GGU, A.B.).

with the scattered inclusions of ultrabasite which, like the amphibolitic inclusions, seem to be untransformed relics of pre-existing rocks.

Banding occurs in different types of development. One type is caused by increasing content of mafics across the band and resembles graded banding. Another type is brought out by a change in composition between different layers. The origin of the banding is rather obscure. Although locally it is well developed, as pictured in the photograph fig. 51, no continuous system could be seen in these structures. At one place, the graded banding is caused by an increase of light coloured minerals upwards while close by the dark coloured minerals increase upwards within a band. The only general feature noticed is that the banding seems to be parallel to the foliation which is occasionally displayed by the diorite.

Since all these features are caused by variations in mineral composition of the diorite or by preferred orientation of the minerals of the diorite, and since they are influenced structurally by the outer shape of the diorite, they must have been formed along with and possibly slightly later than the formation of the diorite.

Petrography of the Dioritnæs rocks

The investigated samples from Dioritnæs will be described in the following order:

First the "wall rock", a light purple gneiss sp. 14966. Then the anorthositic border rock sps. 14970 and 14969, and later on the pyroxene diorite (14971) which contains ultramafic schlieren (19649). 14967a represents the pegmatitic reaction zone between diorite and the leuco-amphibolitic inclusions (14967b). A melanocratic speckled diorite (14975) is next described. This passes into the ultramafic rock containing an amphibolite inclusion (35828).

The specimens 19648, 19656 and 35826 belong to the orbicular phase of the diorite.

19650 originates from the isolated quartz-diorite occurrence south-east of the main dyke.

The wall rock of the diorite dyke is a light purple gneiss of rather leucocratic composition. It is medium grained and shows a well developed open foliation. The minerals are feldspar, quartz, some biotite, a little hornblende and accessory apatite and ore. The foliation is indicated by seams of scattered stumpy flakes of yellowish to olive green biotite. A few relics of a spongy bluish green hornblende, partly replaced by biotite, in diablastic intergrowth with quartz may also be seen. The allover texture is lobate hemi-granoblastic. The feldspar comprises poorly twinned plagioclase (ca. 25 $^0/_0$ An) and microcline, which generally form a coarse diablastic intergrowth. Mesoperthite grading into diablastic intergrowth was also noted.

The gneissic wall rock may thus be classified along with the transitional type between the purple gneiss and the light coloured gneisses. In the present case, no deformation seems to have accompanied the recrystallisation.

The anorthositic border-rock. Of the two specimens, 14970 was taken closest to the contact. It is a violet, medium to coarse grained rock with a saccharoidal granoblastic texture. The minerals are plagioclase, with an An content of about 35 $^0/_0$, and very small amounts of hornblende, biotite, quartz in addition to apatite and ore. Minute grains of clinozoizite and sericite occur in the slightly altered plagioclase. The hornblende is opt. neg. and pleochroic from X: yellowish to Y: deep green and Z: deep greenish blue. The biotite is pleochroic from almost colourless to olive green. Most of the hornblende and the biotite seem to have been formed by replacement of primary pyroxene, possibly hypersthene, judging from the parallel extinction of some relics and the stout prismatic shape of the pseudomorphs. Quartz occurs as small grains or in vermicular intergrowths with the uralitic hornblende.

The other specimen (14969) consists of a rather dark, violet to bluish grey rock of medium to coarse grain and with a granoblastic texture. Faintly antiperthitic plagioclase, which may show a slight normal zoning in the larger grains (33—38 $^0/_0$ An), constitutes more than 95 $^0/_0$ of the rock. Hornblende, biotite, some relic hypersthene and diopside, quartz and accessory apatite, zircon and ore, together with secondary scapolite, calcite and saussurite-minerals make up the remaining 5 $^0/_0$. The relic diopside and hypersthene grains are rimmed by hornblende and in some cases by an extra rim of biotite. The uralitised hypersthene is packed with minute ore grains and once seems to have formed idioblasts, while the original outlines of the diopside seem to have been xenoblastic. The hornblende closely corresponds to that of specimen 14970. The biotite is olive green where associated with the uralite, but yellowish to deep grey-brown when surrounded by plagioclase.

The pyroxene diorite. (14971). This is a coarse grained rock with an uneven granoblastic texture. An acid andesinic plagioclase constitutes about 65 $^0/_0$ of the rock. The rest is made up of diopside, hornblende, hypersthene, biotite, quartz and ore. The diopside is greenish and pleochroic with schiller inclusions. The hypersthene is pinkish and faintly pleochroic. One grain of hypersthene is partly surrounded by a thin rim of diopside. Both pyroxenes may be replaced by a hornblende which is opt. neg. and pleochroic from X: yellowish green to Y: olive green and Z: (faint bluish) green. The greenish to brown biotite only occurs in a few grains.

This rock contains:

The ultramafic schlieren (19649). These are formed of a dark, coarse grained rock with a decusate texture. The minerals are hypersthene, which makes up about 2/3 of the rock, quartz, hornblende, andesine, biotite and ore. The hypersthene is pleochroic from pink to greenish and forms fairly hypidiomorphic grains. Quartz and andesine occur interstitially in relation to the hypersthene. The quartz shows undulate extinction. The hornblende is opt. neg., pleochroic (from X: yellow-green, to Y: green and Z: brownish green) and forms xenomorphic grains. The pleochroic biotite (yellowish to red-brown) exhibits almost poikiloblastic outlines.

Where diorite contains pyroxene amphibolite inclusions, a thin pegmatitic reaction zone surrounds the inclusions.

This reaction zone (14967a) shows a coarse grained rock of dark grey colour with an uneven granoblastic texture. More than 90 % of the rock consists of an antiperthitic plagioclase (ca. 35 % An). Diopside occurs in stout prismatic grains rimmed by hornblende or wholly replaced by hornblende and biotite, both of which seem identical to those described from specimen 14969.

Specimen 14967b, which represents the pyroxene amphibolite inclusions, consists of plagioclase, hornblende, diopside and hypersthene. The hornblende: pyroxene ratio just allows the application of the term pyribolite to this rock. It is medium to fine grained with a well developed lineation of the hornblende. Under the microscope, however, the preferred orientation of the hornblende is less striking, since it only affects some (about one third) of the grains. The general texture is granoblastic with an almost reticulate arrangement of lobate mafic grains around aggregates of saccharoidal plagioclase. Small more idioblastic grains of hornblende may be found within these aggregates. The pale greenish grey diopside often surrounds the pink hypersthene. The yellowish to brownish green hornblende seems to have corroded the pyroxenes. Furthermore, thin more bluish green uralitic rims are found on some pyroxenes. Ore forms a rather important accessory. Small flakes of red-brown biotite have in places grown on the ore grains.

The diorite containing the amphibolitic inclusions passes westwards into melanocratic speckled diorite (specimens 14975 and 14972). This variety is made up of 3—5 cm large plagioclase grains which contain scattered hypersthene crystals (1 cm approx.) and stumpy diopside grains (ca. 1/2 cm). The large feldspar grains are separated by saccharoidal plagioclase and concentrations of mafic minerals. Slender prismatic and idioblastic hornblende grains up to 2 cm long interweave the rock in a haphazard arrangement. The hornblende needles also intersect the larger and more stumpy pyroxene grains. The hornblende in its turn may be replaced by slender flakes of a transparent to yellow or orange brown biotite. The hornblende is pleochroic from X: greenish yellow to Y: brownish green and Z: olive green; $c \wedge Z$ varies from 25 to 35° and 2V from 58 to 40. The plagioclase shows deformed twin lamellae, but seems also to be zoned. In the large poikiloblastic grains it is more calcic around the idioblasts of hornblende, as far as can be judged from the extinction angles. The zoning seems to take place within the compositional range of an andesine.

With the gradual disappearance of the plagioclase speckles, the melanocratic diorite passes into an almost ultramafic variety which contains small inclusions of amphibolite.

This amphibolite (35828) is medium to fine grained and shows a weak lineation. Otherwise the texture is saccharoidal granoblastic. The

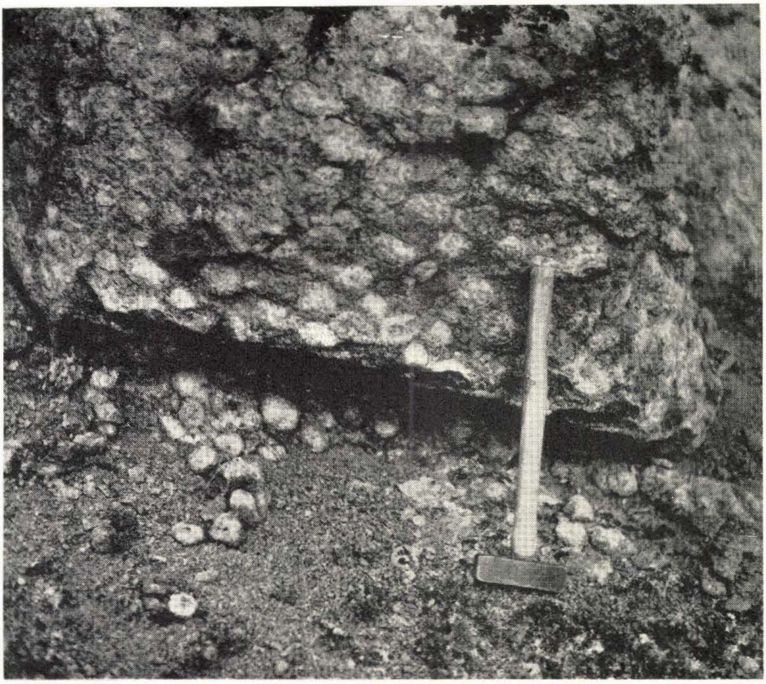

Fig. 52. Outcrop of orbicular diorite, Dioritnæs (Photo: GGU, A.B.).

minerals are plagioclase (ca. 50 $^0/_0$), hornblende (ca. 40 $^0/_0$), diopside and accessory biotite, apatite, zircon and ore. The plagioclase is a normally zoned andesine. The yellow-green to olive and almost bluish green hornblende contains some clear patches of relic diopside.

The orbicular diorite

The outcrop with orbicular diorite (fig. 52) is situated within the northern anorthositic border rock which at this locality shows a greater thickness than on the coastal section described above.

The orbicular rock contains elongated orbicles with an average size about 4 × 8 cm. The nodule rock is greenish to brownish grey and medium grained. The orbicles are contained in a slightly more coarse grained bluish grey rock, which seems to separate even the most closely packed orbicles.

All orbicles show a zonal arrangement with an outer "crust" and an inner "core". The thickness of the crust may only be a few mm in some orbicles and 1—2 cm in others but in both cases a radiating arrangement of the mafics within this peripheral zone may be noticed. The core consists of a granular-textured rock in all the sectioned nodules (19648, 19656 and 35826).

In the following, the inter-orbicular rock, the "crust" and the "core" will be described.

1. The inter-orbicular rock.

The inter-orbicular rock is a coarse to medium grained biotite diorite. The minerals are plagioclase, biotite, quartz, a little hornblende and accessory apatite, zircon and ore. In addition interstitial calcite has been observed.

On the sectioned surface of sp. 19648 a concentration of biotite in irregularly arranged aggregates is seen along a line running between the two orbicles. In a slide covering the same area, mortar seams were also noticed along this line. The surrounding biotite shows strain features. Locally, aggregates of smaller biotite flakes build up polygonal arches, but elsewhere no overall pattern in the orientation of the biotite can be discerned. The local development of the arch-shaped aggregates may have been produced by recrystallisation of large, formerly bent biotite flakes. The plagioclase is andesinic (34—35 $^0/_0$ An). The biotite is pleochroic from pale yellowish to dirty brown or almost black. Quartz occurs in a few large grains within the mortar seams and as vermicular to parallel intergrowths with biotite.

From the above description it may be concluded that mainly post-crystalline movement between the individual orbicles has taken place.

2. The crust of the orbicles.

The minerals found in this rock type are plagioclase (An ca. 40 $^0/_0$), hypersthene, biotite, a little diopside, hornblende, quartz and accessory ore. A little secondary interstitial calcite is also seen. The texture is dominated by the pyroxene which occurs in more or less parallel elongated grains, sometimes forming chains which stand out from a background of an extremely amoeboid mosaic of plagioclase.

The pinkish hypersthene is surrounded by a thin rim of transparent diopside, the cleavage of which is parallel to that of the hypersthene. The extinction of the hypersthene is generally oblique to the elongation of the grains, but neighbouring grains or groups of grains may show the same or nearly same optical orientation. The impression is obtained that the hypersthene is only present as a relic of former larger glomeroblasts. In this case, the highly interlocking amoeboid plagioclase grains must have replaced pyroxene. The peculiar texture of the plagioclase indicates that such transformations (if real) took place under very quiet conditions.

In addition, bluish green hornblende and nearly colourless to olive green biotite is seen to replace the hypersthene. In a few places the hornblende also forms a thin rim on the diopside fringing the orthorhombic pyroxene. Where biotite replaces hypersthene the small flakes are generally arranged parallel to the cleavage of the hypersthene, but they may also intersect the latter obliquely. In the latter case, the biotite contains vermicular to parallel intergrowths of quartz.

Quartz also occurs as amoeboid grains together with plagioclase but, in total, is found in smaller quantities than in the inter-orbicular rock.

3. The core of the orbicles.

The core is composed of plagioclase (35—40 % An), hornblende, biotite, a little quartz and some relics of hypersthene besides accessory apatite and ore. Furthermore, interstitial calcite is found. The texture is hetero-granoblastic. The hornblende is opt. neg. and pleochroic from X: pale yellowish to Y: emerald green and Z: greenish blue. It occurs as large xenoblastic grains formed at the expense of hypersthene, which is only found as a few relics. Some hornblendes show a complicated intergrowth with vermicular quartz. The pale yellow-green to green biotite replaces the hornblende and forms irregular flakes and aggregates together with calcite and quartz. The latter mineral also occurs interstitially within the granoblastic plagioclase. In one orbicle a 2 cm large porphyroblast of plagioclase was found in the very centre.

Summarizing this description, it can be said that the inter-orbicular rock shows a somewhat higher content in quartz than the crust but corresponds rather closely in this respect to the core of the orbicles. The biotites of the inter-orbicular and the orbicular rock types show differences in pleochroism which may reflect differences in composition. The plagioclase of the crust is slightly more basic than that within the matrix between the orbicles while the plagioclase of the core takes an intermediate position.

Various processes of replacement and recrystallisation can be guessed at, inferred or observed, the youngest replacements naturally being the most evident. These are uralitisation and biotitisation of the pyroxene, which are features common to both the orbicular and the normal diorite.

The origin of orbicular structures has been a subject of controversy for a long time. The author does not want to enter such discussions, but all the same, he would like to point out one possible mode of origin for the Dioritnæs orbicular structure. This idea was born during the field work, when the replacive nature of the diorite was detected. As mentioned earlier, there occur some inclusions of amphibolite east of the outcrop with the orbicular rock. Still further to the east, outside the dyke, a thin amphibolite layer can be traced to the coast. In the coast section, this layer is seen to contain a lenticular inclusion, about 5 m broad, of ultrabasic rock.

This small body of ultrabasite consists of a core of greenish brown, medium grained rock (19651—1) with about 50 % olivine, 30 % orthopyroxene, 15 % phlogopite, 5 % calcite and ore. This rock is cut by

2—3 cm thick veins consisting of a central part formed of diopside and borders of orthopyroxene. Surrounding the core, hypersthenite (19651—2) with a composition of about 90 % hypersthene, 7 % phlogopite, 3 % calcite, quartz, feldspar, biotite and ore is found. Within this rock a variety occurs which contains porphyroblasts attaining a size of 5 to almost 10 cm in length. They (19651—3) consist of an intergrowth of diopside and hypersthene of amoeboid character, with a preferred elongation of the hypersthene resembling very much the occurrence of this mineral in the crust of the orbicular rock described above. Small quantities of quartz, mesoperthite and red-brown biotite occur as irregular grains seemingly replacing the pyroxene intergrowth.

A thin zone rich in green hornblende and biotite separates the ultrabasic rock from the surrounding amphibolite.

If a similar porphyroblastic ultrabasic rock had occurred within the amphibolite now replaced by the diorite, the replacement of the large porphyroblasts may have given rise to an orbicular structure, each porphyroblast functioning as a small body of ultrabasic rock which became surrounded by reaction zones developed during the replacement process. This theory may be supported by the above-mentioned textural similarities between the hypersthene in the crust of the orbicles and the unreplaced porphyroblasts. The general elongated shape of the orbicles would thus be directly inherited from the parent porphyroblasts.

4. Cross-cutting aplites

Dioritic and meta-dioritic aplites have been observed at various localities in Tovqussap nunâ. They were first detected in connection with the detailed mapping of Langø, on which island they are quite common. Later on they were found north and south of Qaersup ilua and at Dioritnæs. Although their distribution may depend on the occurrence of larger dioritic bodies, it is very possible that a more thorough search throughout the whole map region would lead to their identification at various other localities.

Petrographically, the diorite aplites are clearly related to the dioritic rocks described above. When not affected by retrograde metamorphism or granitisation, they carry pyroxene and are quartz-free (see for example sp. 19144, Intermediate Division).

The aplites seldom exceed half a metre in width and dykes of 2—10 centimetres thickness were found to be the most common. Where not influenced by younger shearfolding, the aplites are rectilinear and exhibit several structural features which could be taken as indicating their intrusive origin. En echelon arrangement and combined en echelon and en bayonet structures have been found in some aplites on Langø, fig. 53. Clearly dilational emplacement tectonics have been observed in a 5 cm broad dyke north of Granulitsø. This dyke displaces an ultrabasic inclusion in the gneiss, see fig. 53, B. The outcrop conditions being favour-

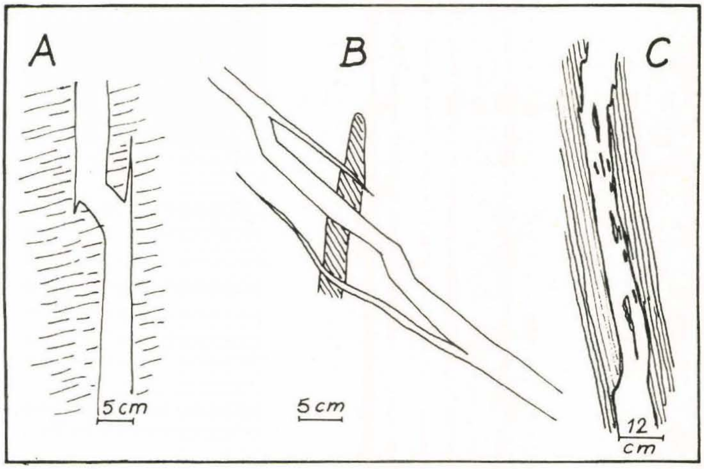

Fig. 53. Structures of diorite aplites.

able, the three dimensional dilation can be demonstrated beyond any doubt in this case. Another aplite, in the Great Pyribolite of the east coast of Langø, contains several small inclusions of the host rock, fig. 53, C.

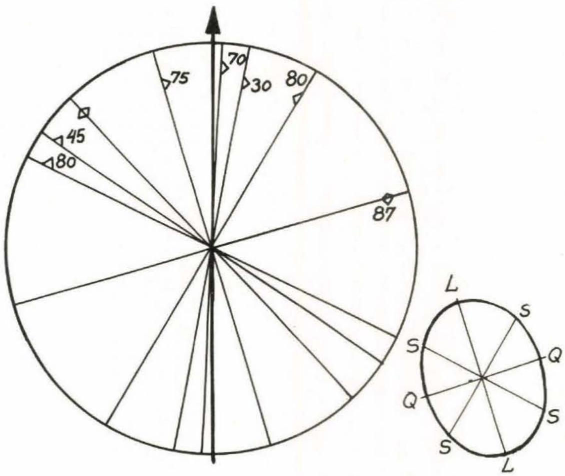

Fig. 54. Diagram showing the orientation of undeformed diorite aplites. Q, L and S on the attached strain elipsoid denote cross, longitudinal and diagonal joints respectively.

In this case, the preservation of the orientation of the inclusions makes a metasomatic origin feasible.

Since many of the observed aplites have been involved in younger shearfolding, the number of strike and dip readings of aplites is not as great as could be desired. However, a rose diagram constructed from the available readings from Langø (fig. 54) seems to bring out that the spatial orientation of the aplites is related to the fold tectonics, since Q, S and L are the preferred directions of the aplites.

As to the age of the aplites, it can be said that although aplites have been formed locally simultaneously with larger bodies of latekinematic diorite, the majority of the aplites are regarded as representing the latest

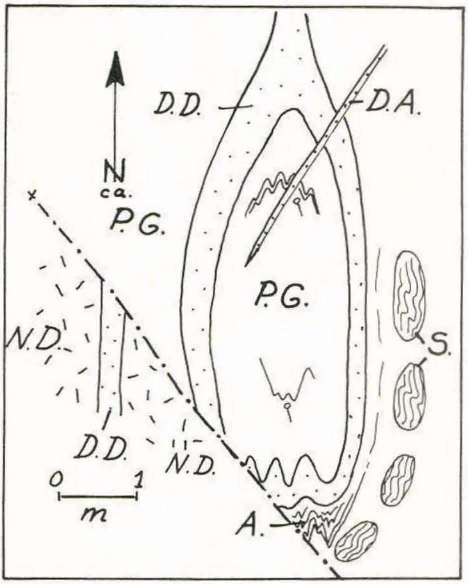

Fig. 55. Sketch map illustrating the relation between the Nordnor diorite and diorite dykes and aplites.

phase of the diorite formation. This point of view is supported by many observations of aplites cutting latekinematic diorite bodies. A very illustrative example of the age relations between a large body, a dyke and an aplite is found close to the Nordnor diorite, where the border of this is displaced by a NW-striking sinistral fault, see fig. 12 and fig. 55. As shown on the sketch fig. 55, which is a simplified map, the folded country rock to the Nordnor diorite (i. e. P. G.: purple gneiss, A: amphibolite, S: calc-silicate boudins) contains an irregular dyke of diorite (D. D.) which is slightly affected by posthumous shearfolding. A similar (or possibly the same) dyke cuts the Nordnor diorite south-west of the fault. The thin diorite aplite (D. A.) cuts the dyke and is not influenced by any movements. These relations suggest that the formation of the dyke which cuts the main diorite body took place closely after the 'mise en place' of the latter and therefore became affected by the very last spasm of movements, while the aplite is clearly postkinematic.

VI. THE CHRONOLOGICAL SIGNIFICANCE OF THE DIORITE ROCKS

The foregoing pages contain descriptions of the petrography, distribution and mode of occurrence of the Tovqussaq diorite rocks. It may now be appropriate to emphasize what the field and laboratory investigations have brought forward regarding the chronology of these rocks.

The petrographic study has shown that the transition from hypersthene gneiss via purple gneiss into light coloured gneiss most probably represents a sequence in time. Retrograde alterations with or without the introduction of new material have in many places changed originally hypersthene-bearing types into biotite-bearing gneisses or granofelses. This process, where most advanced, was accompanied by deformation of old S-planes, granulation and recrystallisation. There is thus evidence of two separate periods of metamorphism, an older corresponding to granulite facies and a younger characterised by amphibolite facies.

The granulite facies metamorphism culminated with the formation of pyroxene diorite. The dioritisation seems to have outlasted the deformation otherwise accompanying the granulite facies metamorphism. From a chronological point of view, the dioritic rocks, which were formed when the movements were waning or had stopped, i.e. the late- and postkinematic types, are particularly important. The present state of the contact planes of the dykes and aplites supply us with valuable information about the presence and intensity of post-dioritic deformation. The degree of alteration (uralitisation, etc.) of the dioritic rocks indicates the relative importance of later retrograde metamorphism.

On the assumption that the syn-, late- and postkinematic nature of the diorites indicates that they were formed within a definite span of time during which the deformation waned and finally ceased, they may be used as a time mark in the relative dating of other events. In this way the space and time relations of the two periods of metamorphism can be studied.

In the field, granulite and amphibolite facies rocks are separated in space, but convincing evidence that this separation in space also cor-

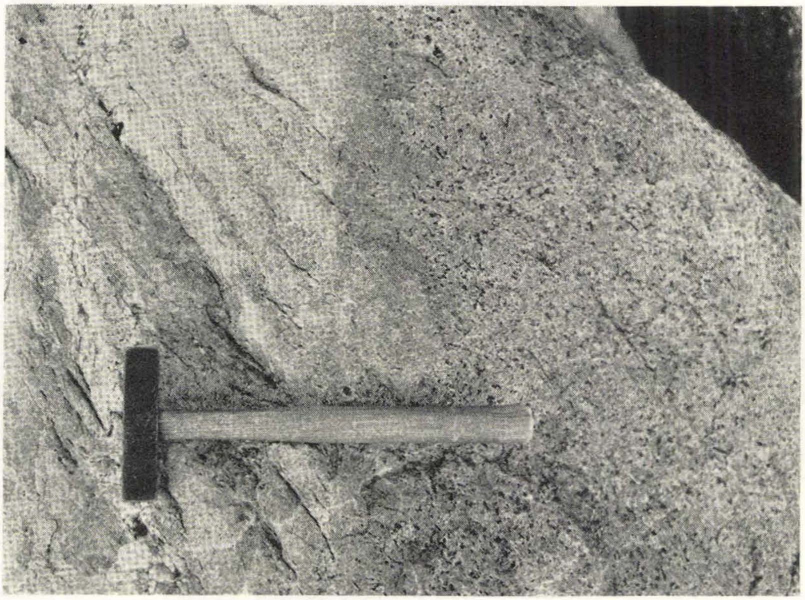

Fig. 56. Retrograde altered diorite dyke in granulite, southern Langø (Photo: GGU, A.B.).

responds to an appreciable separation in time can only be brought forward through a study of the dioritic dykes and aplites. This study has revealed that, where the enclosing gneiss shows mineral associations developed by retrograde alteration from granulite facies, the dykes and aplites have been affected by the same alterations. Observations on the deformation from which the originally more or less straight dykes and aplites have suffered, have also demonstrated a connection between younger shear movements and alterations. Although alteration need not always be accompanied by shear movements, there are a number of cases in which these two processes are obviously associated. A few examples of this relation may be given.

In the granulitic rocks on Langø several meta-dioritic dykes and aplites have been found. On Sydtangen a metre-broad NE-trending dyke cuts a type of granulite which seems developed from purple gneiss. The dyke, fig. 56, shows more fine grained borders, but contains garnets and biotite similar to the enclosing rock. The borders are bleached and whitish, while the colour of the central part is caused by the flesh coloured feldspar. Quartz may be observed in the border zone, but is absent in the central part of the dyke. It may also be noted that the foliation of the granulite in places seems to continue into the dyke. This is most probably due to a slight shearing of the dyke during its alteration. The dyke itself is cut by thin meta-dioritic aplites.

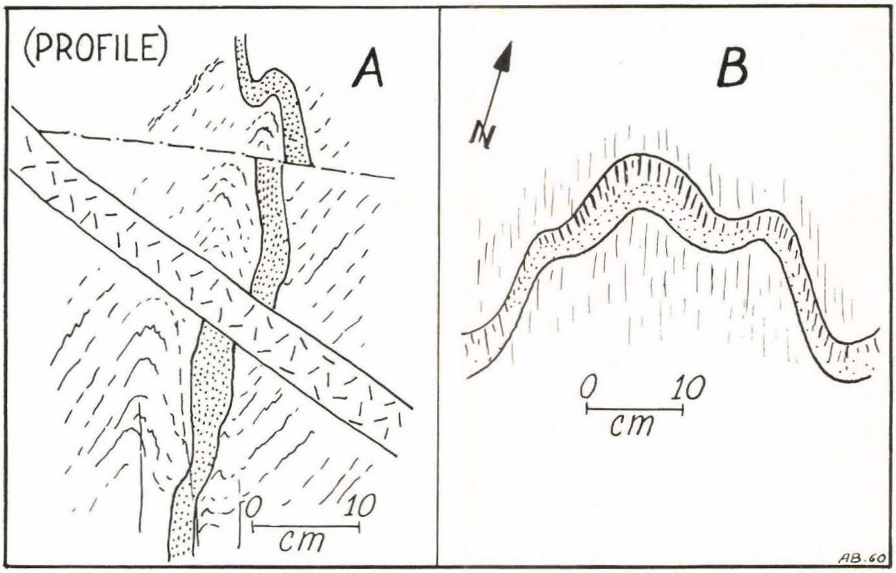

Fig. 57. Deformed diorite aplites.

South and north of the Nordnor, where the granulite exhibits a distinct Plättung texture, the cross-cutting meta-diorite aplites are deformed. They carry biotite and garnet and generally show grey colours. Some, however, contain relic flesh coloured feldspar in their central parts. An example north of the Nordnor, fig. 57 B, shows that not only slip movements (shearing) but also compression has taken place. In this thin aplite, which is nearly vertical, the northern part is recrystallised and almost pegmatitic in grain, and the centimetre-large biotite flakes exhibit a fan-like arrangement. The southern part is less altered and carries flesh coloured feldspar, garnet and biotite. It may be noticed that the pegmatitic zone represents the lower part of the aplite, when seen in relation to the SSE-plunging fold axis. This fold may be referred to as a 'Fächerfalte' or a 'divergentschiefrig Falte' (Sander, 1950, pp. 292—293). Sander stresses the importance of this type of fold, since "sie eine zeitliche Vergleich mit der Regelung relativ plötzlich einsetzende weitere Biegung (Verengung der Biegung) der Falte kennzeichnen und damit die regionale tektonische Einengung".

At the same locality another aplite, which is parallel to the position of the axial plane of the shear folds, is found. This aplite is not deformed but has been altered along its contact.

All observations suggest that the development of the minerals and the textures typical of granulite took place later than the formation of the diorite aplites. The 'Fächerfalte' indicates that not only penetrative movements but also flattening accompanied the 'granulitisation'.

Fig. 57 A. shows a meta-dioritic aplite found in the core rocks of the Tovqussaq dome south of Qaersup ilua. The slightly deformed aplite is cut by a small fault, which is older than a microcline-bearing pegmatite. The shear movements which affected the aplite have plicated the foliation in the gneiss.

The Dome dyke, which cuts across the northern part of the Tovqussaq structure, where found in the purple core rocks is only slightly altered along the contacts, while the central parts consist of uralite diorite. In the granitic shell of the dome, the dyke is granitised. In the hypersthene gneisses of the Frame Layer, it carries pyroxene. The more or less vertical position of the Dome dyke shows that no important deformation of the dome structure has taken place after the formation of the dyke, i. e. the dome was already formed when, during the second period of metamorphism, the granitic core rocks were developed. The deformation attached to the second metamorphism (amphibolite facies) was therefore restricted to small scale shearing of posthumous character.

Other diorite occurrences have in a similar manner preserved their original late- or postkinematic structures nearly unaffected by later deformation, although on a small scale local deformation may be seen. These relations are very important, since they suggest that the development of the major structures of Tovqussap nunâ date back to the period of granulite facies metamorphism, or even earlier. Their geometry and kinematics can be analysed independently of the superimposed amphibolite facies metamorphism.

VII. GEOMETRICAL ANALYSIS OF THE MAJOR STRUCTURES

As mentioned in the description of the geological map, the litho-structural mapping made a division of Tovqussap nunâ into major structural units possible. In the west, the Tovqussaq dome, the Smalledal and the Irdal structures were distinguished. In the east, the Krebsesø antiform with its flanking synforms and the Dioritnæs synform were defined. Of these structural units the Krebsesø antiform will be discussed first. There are several reasons for doing this. First of all the marker horizons have been mapped almost continuously through the Krebsesø structure, a fact which greatly facilitates the structural interpretation. Furthermore in this structure the conditions for carrying out an analysis of older refolded structures are at their best.

A. The Krebsesø antiform and its flanking synforms

North of Midterhøj, the Krebsesø structure appears to be a simple antiform, but in the southern part of the structure the relations are more complex. The complexity is shown by three abnormal features:

1. The 'splitting up' of the Krebsesø Pyribolite east of Midterhøj due to the appearance of a median gneiss layer which only occurs in the central and eastern portions of the antiform.
2. The peculiar hinge zone in the upper part of the Krebsesø Pyribolite on the 350 m mountain east of the Krebsesø lake.
3. The aberrant orientation of the smallfolds within certain parts of the structure.

All these features are inconsistent with the idea of the Krebsesø structure as a simple antiform. It is obvious that within it some more complex structures are concealed. In order to analyse the structure a combination of different methods has to be used, beginning with a consideration of minor structures.

The small scale structures

These include linear and planar structures. Unfortunately the number of observations and measurements on such structures is not as great

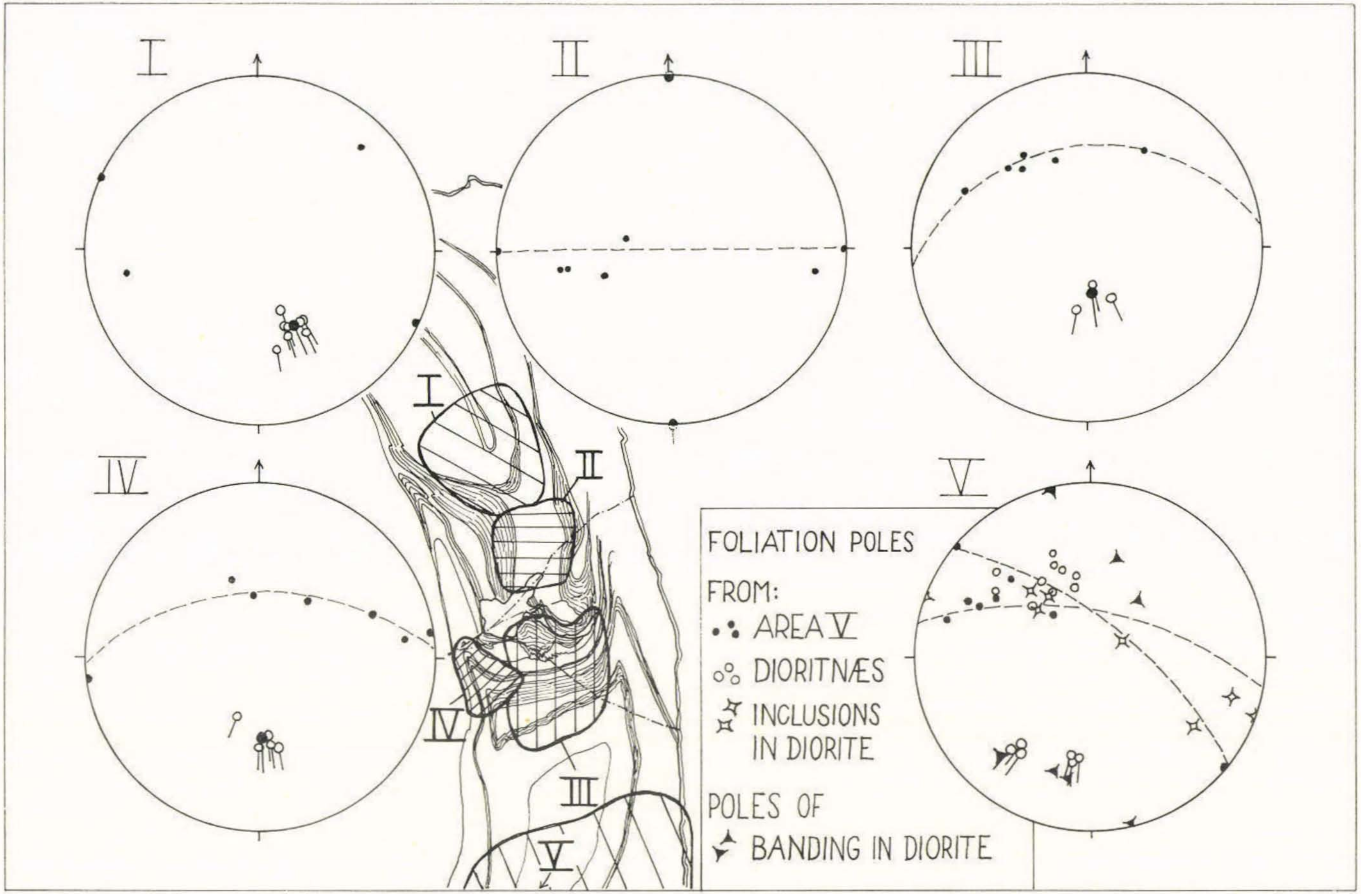

Fig. 58. Stereograms from eastern Tovqussap nunâ.

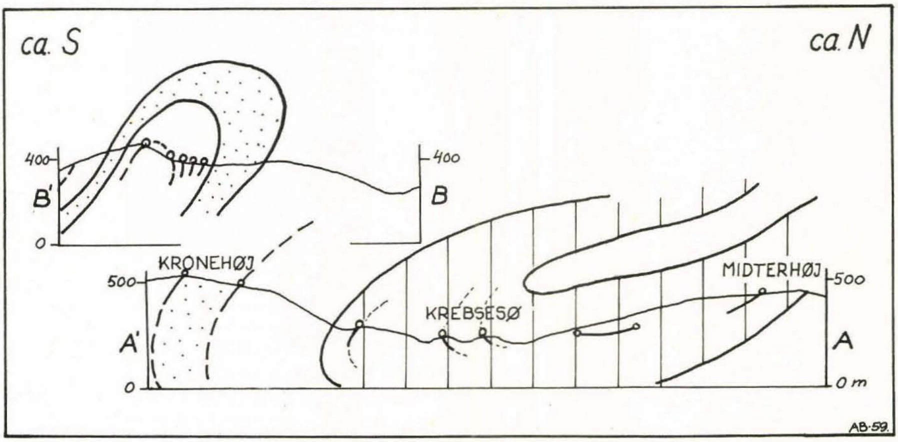

Fig. 59. Longitudinal profiles through the Krebsesø antiform. The thick lines illustrate the observed variation in the plunge of the smallfolds. For the location of the profiles see fig. 60.

as it might have been, had more time been available for field work. Nevertheless the good exposure ensures that the readings taken are representative, and these serve as an important starting point for the structural analysis.

Around Buedal (fig. 58, stereogram I) the smallfolds plunge SSE. In the field these folds could be recognised as drag folds developed in connection with the formation of the main antiform. Stereogram II, fig. 58, shows plots of foliation poles from the area between Midterhøj and Krebsesø. The axis of the antiform is seen here to be horizontal. To the south of this area (II) and north of area III (see key map in fig. 58) the axial relations are very complex. On the small point on the east shore of Krebsesø, the pyribolite shows smallfolds with a steep northerly plunge. Further to the south, on the small island, variations from a steep northerly plunge to a vertical orientation were noticed in a vertical N-S profile. The plunge decreases down the plunge line in the smallfolds. South of Krebsesø and on Kronehøj the plunge is again southerly, but is still rather steep (stereogram III, fig. 58). The nearby Flankepas synform is characterised by less steeply plunging axes (stereogram IV, fig. 58). West of Kronehøj, in the corner between areas III and IV, complex axial relations are found. Going from north to south, the southerly plunge of the smallfolds is seen to increase until it passes 90° (vertical) and becomes steep northerly. Then, due to a tight culmination, it changes rapidly to steep southerly.

All these features demand an explanation. In fig. 59 the variation in the plunge of the smallfolds is shown in two longitudinal profiles, one through the Krebsesø antiform (A—A′) and one through the southern part of the Flankepas synform (B—B′). Pictured in this way, the variations are seen to form a pattern which suggests the existence of older recumbent folds within the rock succession affected by the Krebsesø antiform. The recumbent folds evidently have had an axial trend oblique to that of the younger Krebsesø antiform and its flanking synforms. The smallfolds with an anomalous orientation are seen in fig. 59 to occur close to the hinge zones of the recumbent folds, while more normally orientated axes are found in the flanks of these structures.

The smallfolds which most probably were developed in connection with the recumbent folding were seemingly destroyed during the refolding since practically no traces of such older anomalous small scale structures have been met with. (As explained below, the above-mentioned smallfolds are not older structures). The only example so far found was seen at Flankehøj, where the pyribolite shows a tight smallfold which is cut by a transverse foliation. The new axes (and foliation) imposed on the rocks naturally conform to the main antiform and its flanking synforms — except within the refolded hinge zones of the recumbent structures. Here the form imposed on the layer by the recumbent folding has controlled the later-formed B-axes and caused them to bend around the old hinges, i.e. the new axes reflect in their orientation the position of the older hinge zones.

The alternative view, that the variation in the axial pattern could be due to refolding of older small scale structures, need not be considered in the present case. The results obtained at a later stage of the analysis are inconsistent with such a view.

The significance of the outcrop pattern

As shown in fig. 59 the axial pattern suggests that the hinge zone of a recumbent synform occurs at Krebsesø, (profile A—A′) while a hinge zone belonging to a recumbent antiform should be expected west of Kronehøj (profile B—B′).

Following this line of thought the outcrop pattern can be viewed from a new angle. The existence of older recumbent structures within the Krebsesø antiform would naturally influence considerably the course of the marker horizons. The outcrop patterns of double fold structures of this type are well known from D. REYNOLDS and A. HOLMES (1954) experiments with plasticene models, and it is a simple operation to reconstruct the pattern produced by a plunging antiform which encloses a recumbent fold. Such a fold would appear with a hinge zone within both antiformal flanks. A glance at the map shows, however, that the relations at Krebsesø are somewhat more complex. In order to extend the analysis, a special trick has to be used. If the map is redrawn without any attention being paid to the gneiss layer at Midterhøj, a bilaterally symmetrical outcrop pattern can be arrived at, fig. 60 — that is if the Krebsesø Pyribolite forms a closure below the western part of the Krebsesø. This closure would represent the hinge zone of the recumbent synform within the western flank of the younger antiform, while the synformal

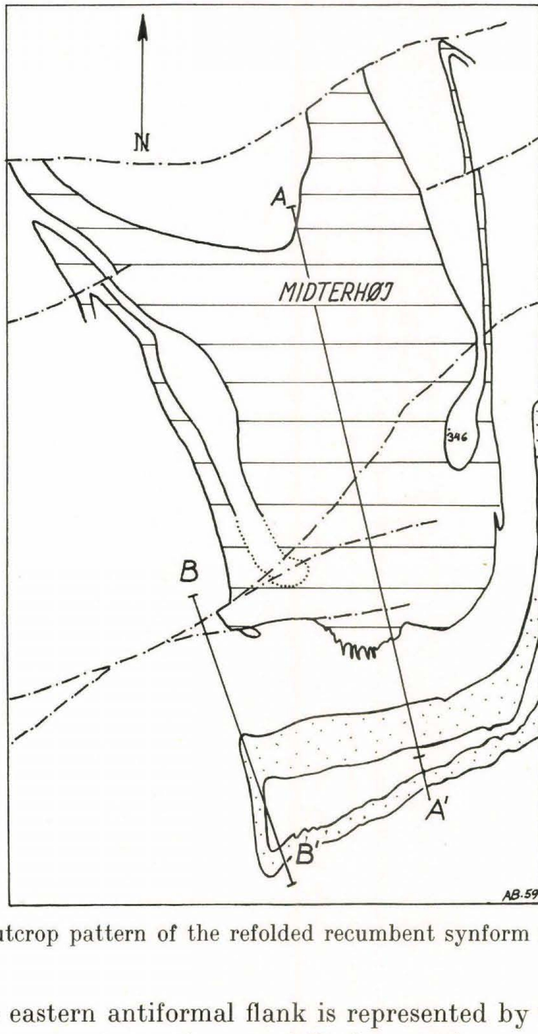

Fig. 60. Outcrop pattern of the refolded recumbent synform at Krebsesø.

closure in the eastern antiformal flank is represented by the hinge zone on the 346 m mountain north-east of Krebsesø. The trend of the hinge axis in the recumbent structure would then be ENE, as indicated by a line joining the two mentioned closures. In fig. 60 the symmetry of the antiform is also brought out from the occurrence of two subsidiary folds, one on each flank of the structure. These folds may, however, be drag folds formed in connection with the major antiform.

It should be stressed that the interpretation just outlined is based on the assumption that the dioritic rocks at Krebsesø belong from a structural point of view to the Krebsesø Pyribolite. This assumption concurs with the idea expressed earlier that the Krebsesø diorite has been formed by metasomatic transformation of pyribolite.

The axial variations west of Kronehøj suggest the presence of a nearby recumbent antiformal hinge zone. Such is actually shown by the course of the Kronehøj Pyribolite, which closes here. This closure is only slightly affected by younger refolding (Flankepas synform). Tracing the duplicated Kronehøj Pyribolite eastwards, it may also be seen that south of Krebsesø the effect of the Krebsesø antiform has diminished considerably. The Riddersporen synform, however, still influences the southernmost recumbent structure. In spite of this influence, the old antiformal hinge zone does not reappear on the map to the east. This means that the hinge axis of the recumbent antiform here lies above the present erosion level. Although around Kronehøj, the outcrop pattern is less indicative of double folding than around Krebsesø, there is west of Kronehøj strong evidence in support of the idea that the variations in the orientation of the smallfolds are controlled by hinge zones belonging to refolded recumbent structures.

Still older isoclinal structures

Two 'abnormal' features still remain to be explained, 1) the 'splitting up' of the Krebsesø Pyribolite at Midterhøj and, 2) the peculiar hinge zone on the 350 m mountain east of Krebsesø. These features were deliberately disregarded in order to reconstruct the outcrop pattern of the antiformally refolded recumbent synform, fig. 60. Returning to the problem posed by the gneiss layer at Midterhøj, the idea that it occupies the core of an isoclinal structure suggests itself. The 'splitting up' of the Krebsesø Pyribolite could then be explained as a narrow closure. This would mean that the Krebsesø antiform is in reality a triple folded structure.

If this interpretation is correct, the oldest fold structure, which here will be called the Midternæs isocline, should be influenced by refolding by the recumbent synform as well as by the Krebsesø antiform. The tight closure west of Midterhøj indicates the present synformal character of the isocline where this latter occurs within the western flank of the Krebsesø antiform and within the lower flank of the recumbent synform. Refolding would then cause the isocline to appear antiformally in the eastern flank of the Krebsesø antiform and in the overturned flank of the recumbent synform. If this behaviour of the isocline can be demonstrated, the interpretation can be checked.

In order to carry out such a check, a special method, the contour map method, has to be used. This method and its application to refolded structures has recently been discussed by the author (Berthelsen, 1960a). Since, in this publication, the Krebsesø structure was used as an example of the application of the method, no further comments on the procedure in construction of contour maps will be given here.

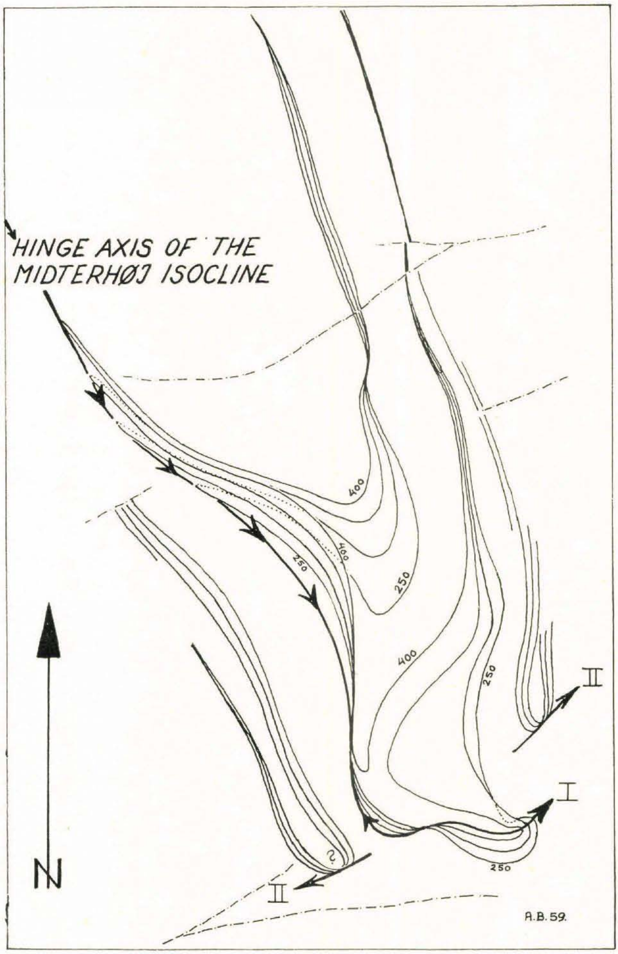

Fig. 61. Structural contour map of the Midterhøj isocline. I: hinge axis of oldest isocline, II: hinge axis of recumbent synform.

The contour map in fig. 61 shows the shape of the Midterhøj gneiss layer which occupies the core of the isocline. This map demonstrates that the isocline due to refolding and re-refolding changes from a synform into an antiform. The antiformal closure is represented by the 'peculiar' hinge zone on the 350 m mountain east of Krebsesø. The outcome of the check thus favours the interpretation leading to the idea of the Midterhøj isocline.

Since the Midterhøj isocline was refolded two times and since its hinge axis only appears twice within the area discussed, the original trend of its hinge axis, before any refolding, is difficult to estimate with any accuracy. All that can be said is that the trend may have varied around a NW—SE direction. This estimate is arrived at by unrolling the younger structures, as shown schematically in fig. 62 A and B.

On Riddersporen the Kronehøj Pyribolite 'splits up' just like the Krebsesø Pyribolite does east of Midterhøj. Does this mean that the branching of the pyribolite on Riddersporen also represents a tight fold closure? If so, the median gneiss layer occupies the core of an isocline, which, due to its being refolded twice, would make the median gneiss reappear on the map within the same layer and in the upper flank of the recumbent antiform. Although not marked on the map as a separate layer, there actually occurs a thin median gneiss band within the Kronehøj Pyribolite east and north-east of Sardlup qâva (p. 112). This median gneiss band takes exactly the structural position to be expected by double refolding of the median gneiss at Riddersporen. It is therefore very probable that both gneiss layers represent the core filling of one and the same isoclinal fold (Kronehøj isocline). This interpretation is shown in fig. 62 B, where the structures are viewed from north-east. By unrolling the recumbent antiform, the hinge axis of the oldest isocline is seen to become more or less parallel to the hinge axis in the Midterhøj isocline. As may be seen from fig. 63, which shows a longitudinal profile through the Riddersporen synform, the oldest isocline and the younger recumbent structures have exerted a recognisable influence on the very youngest structure, the Riddersporen synform. The plunge of the axis within this latter structure reflects in its variation the older transverse structures.

Although the conception of the geometrical relations arrived at above is largely interpretative, it represents in the author's opinion the closest possible approach to an understanding of the complex outcrop pattern of the map. Up to now, it has explained all the observed features and has been confirmed when checks could be carried out. It seems also to be the only interpretation possible if most of the salient features are to be explained.

The author therefore accepts this interpretation as a base for further analysis — knowing well that the conclusions arrived at from now on belong only to the second order of probability.

Returning to fig. 62 B, where the recumbent structures have been unrolled, it may well be worth considering the implications of the isoclinal structures within the Krebsesø and the Kronehøj Pyribolites. Assuming that these oldest traceable structures are not just chance features, but conform to still larger fold structures, their influence on the other members of the rock succession should also be studied. Disregarding the gneiss layers, none of which possess any lithological characteristics, and restricting this study to the marker horizons, it may be noticed that

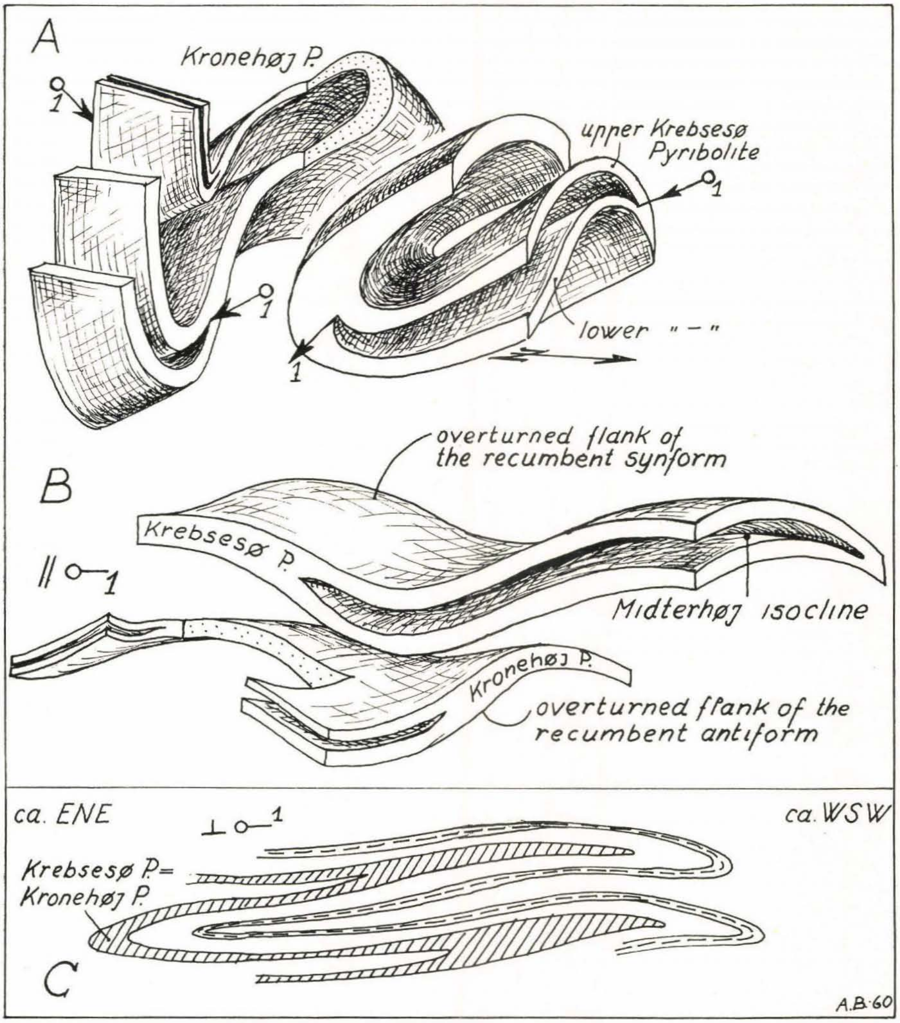

Fig. 62. For explanation see text.

the lithology of these is highly suggestive of large scale isoclinal folding. As shown in C of fig. 62, two thin banded pyribolite layers separate the upper and the lower isocline shown respectively by the Krebsesø and the Kronehøj Pyribolite. Similar thin pyribolite bands overly and underly the isoclines. These thin layers are marked by broken lines in fig. 62 C.

The similarity between the thin pyribolite layers and the presence of the two isoclines suggest that this rock succession in reality hides two large isoclinal structures and that, in consequence, the Krebsesø and the Kronehøj Pyribolites are one and the same layer while all the thin pyribolites belong to another layer. The total stratigraphical sequence

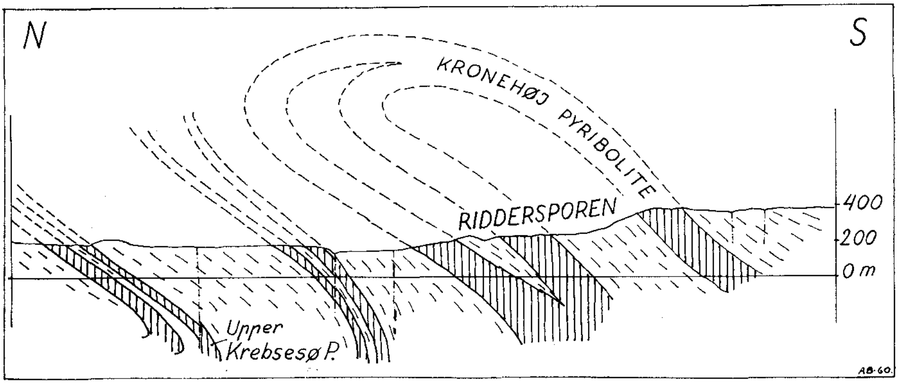

Fig. 63. Longitudinal profile through the Riddersporen synform.

to be traced in this region would then be restricted to five layers — the gneiss layers being included.

This conclusion necessitates a renewed study of the outcrop pattern, in order to see whether or not it is consistent with the ideas arrived at above. First of all, the external hinges (frontal noses) of the isoclines shown by the Krebsesø and the Kronehøj Pyribolites may be looked for in the western part of the structures discussed. Going to the Flankepas synform, one finds that the Krebsesø Pyribolite, which here is composed of diorite as well as pyribolite, terminates in the western flank of the synform. This termination may be interpreted as the frontal nose of the isoclinally folded Krebsesø Pyribolite. This frontal nose occurs within the reversed flank of the recumbent synform, which in its turn is refolded by the Flankepas synform. Due to the moderate SSE plunge of the youngest synform, the corresponding frontal nose within the normal flank of the recumbent synform falls under the sea in the Eqaluk fjord further to the north.

The frontal nose of the Kronehøj Pyribolite isocline is to be studied next. It may be found in the termination of the Kronehøj Pyribolite just east of Itivínguaq, where it appears in the upper flank of the recumbent antiform. This latter is here strongly refolded by the southern extension of the Riddersporen synform. The axis of the frontal nose of the isocline thus ascends within the inverted eastern flank of the youngest synform, and causes the Kronehøj layer to close within itself where it terminates east of Itivínguaq.

Since, west of Kronehøj, the Kronehøj Pyribolite defines the closure of the recumbent antiform, the frontal nose of the isocline contained within the pyribolite does not appear here, but lies below the present erosion level. Further to the north, the frontal nose of the Kronehøj Pyribolite isocline is situated under the normal flank of the recumbent synform so it cannot appear any more within the Tovqussaq peninsula.

An attempt to visualize these complex spatial relations has been made in the structural stereogram fig. 64. By means of contour maps, the outcrop pattern of the Krebsesø and the Kronehøj layers within a 250 m level has been constructed. Applying all available information,

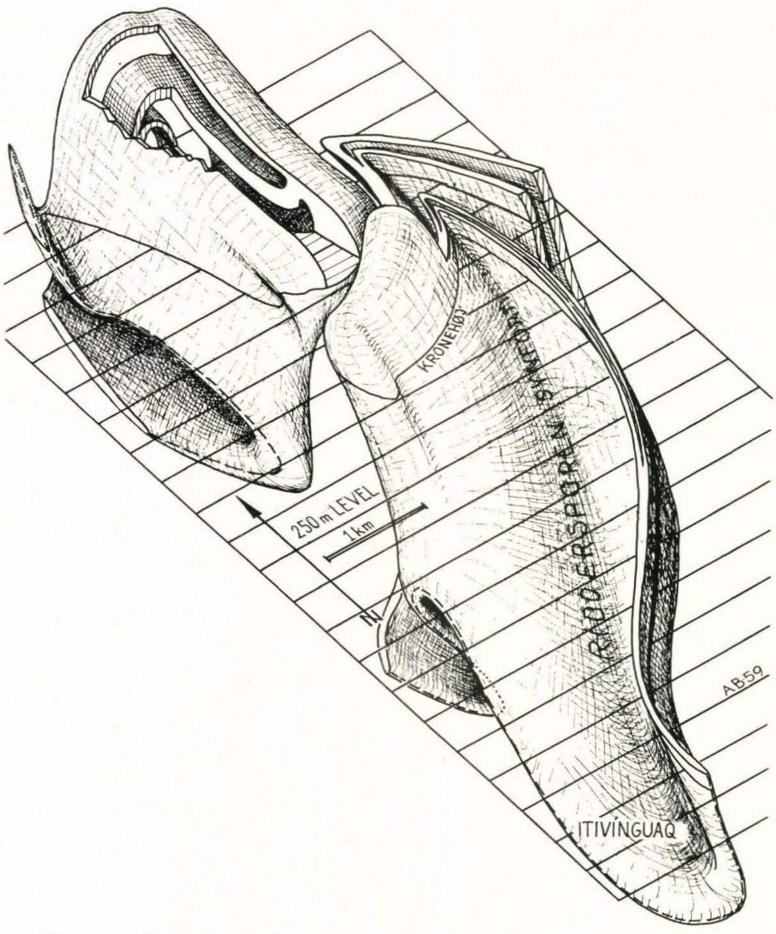

Fig. 64. Structural stereogram of eastern Tovqussap nunâ.

observed facts and interpretative results, a three-dimensional reconstruction of the refolded behaviour of the oldest isoclines has then been attempted.

Based on this reconstruction, a stratigraphical correlation of the individual members of the entire rock succession can also be made. The results of such an analysis are incorporated in the tentative stratigraphic map of Pl. 4. Applying the stratigraphic sequence arrived at in fig. 62 C, the thin pyribolitic marker horizon can also be traced throughout the complex structures. It is, however, not completely continuous.

Around Sardlup qâva it is missing, but its position can still be inferred from the lithological differences of the surrounding gneisses. North of the small crooked bay, Nagtoralínguit, it is partly missing in the smallfolded closure. North of Riddersporen, where it is duplicated between the two isoclines, its two parts are nearly in contact. This feature may indicate the proximity of a subsurface isoclinal hinge zone, where the two parts join up. East of the Flankepas, the same layer cannot be traced around the postulated frontal nose of the Krebsesø Pyribolite. It is also missing entirely within the light coloured gneisses of the western closure of the recumbent synform in the west flank of the Krebsesø antiform. But its absence here could hardly invalidate the interpretation given when it is born in mind that the same layer was also missing for a long stretch around Sardlup qâva.

The thin branching pyribolite layers at Dioritnæs most probably represent the south-easternmost occurrences of the thin pyribolitic marker horizon on the peninsula. As suggested in the description of the geological map, these layers may form part of a larger complex structure now hidden under the sea.

Summarizing, it can be said that the geometrical interpretation of the structures in eastern Tovqussap nunâ allows the following three stages of folding to be discerned:

I. Formation of isoclines with axes trending about NW.
II. Recumbent folding around ENE to NE trending axes.
III. More gentle to closed folding with SSE to S plunging axes.

The kinematic evolution of these structures will be discussed in a separate chapter together with the kinematics of western Tovqussap nunâ.

B. The Irdal and the Smalledal structures

The geometrical analysis of eastern Tovqussap nunâ revealed that the youngest antiform and synforms here refold older recumbent structures, which latter in their turn refolded still older isoclinal structures. Similar complex relations may therefore be expected in the western parts of the peninsula. Studying the structural map of Pl. 2 with the analytical experience just acquired in mind, attention is attracted by the Smalledal and the Irdal structures, since the outcrop patterns of these two structures are highly suggestive of double folding.

The Irdal and the Smalledal structures have already been described with regard to their extent, lithology and the mode of occurrence of their small scale structures (pp. 98—104). Before passing to the geometrical analysis, it may therefore suffice to emphasize that the terms Irdal and Smalledal structures refer to the mainly E-W-trending structures occurring around Irdal and north of Smalledal respectively. The terms Western and Eastern Antiforms refer to the SSE-plunging antiforms at Kangeq and around Trehøje (see Pl. 2).

The geometrical analysis of the Irdal and the Smalledal structures will begin with a study of the orientation of the small scale structures. For the Irdal structure, two stereograms have been prepared (fig. 65 I and II). The stereogram from the western part of the structure shows that this is here an overturned ESE-plunging antiform. Northwards directed dips are only preserved within the closure of the central composite pyribolite layer. Towards east (stereogram II) the plunge becomes due east, due to refolding around the SSE-plunging Eastern Antiform. On the crest of this latter structure, the hinge axis of the Irdal antiform plunges below the present erosion level so its trend within the eastern —here overturned—flank of the younger structure can only be inferred.

In the central parts of the Smalledal structure, the axes plunge to south-west (stereogram IV), and the structure is here overturned to the north, as indicated by the concentration of the foliation poles in the northern part of the stereogram. At Kangeq, the overturned Smalledal antiform is influenced by refolding by the Western Antiform. This refolding causes the Smalledal axes to show due west to WNW plunges (stereogram III). Around Trehøje, the Smalledal structure is influenced by the Eastern Antiform. The effect of this refolding is clearly seen in stereogram V, where the scattered foliation poles only indicate the orientation of the younger structure. The strong scattering of the older smallfolds in this stereogram is most probably connected to the local gneissified nature of the otherwise competent marker horizons of the Smalledal structure. The approach in lithology and competence between the pyribolite and the gneiss has made an intensive smallfolding of all rock layers possible during the refolding. At Kangeq, the readings obtained within the closure of the inner thick pyribolite of the Smalledal structure helped to define the axes belonging to this latter structure in spite of a similar refolding (cf. stereogram III and V).

Where the north-eastern and outermost closure of the Smalledal structure is affected by the Eastern Antiform, small scale double folds have been developed, e.g. north of Bortehøj. These double fold structures help to solve the question as to the relative age of the Smalledal structure and the Eastern Antiform.

The seemingly 'orderless' agmatitic structure previously described from the gneisses south of Breddal (p. 102) is situated within the corresponding closure of the gneisses south of the outermost marker horizon. Assuming that the agmatitic structures were developed within originally double folded rocks, their apparent lack of order may easily be explained.

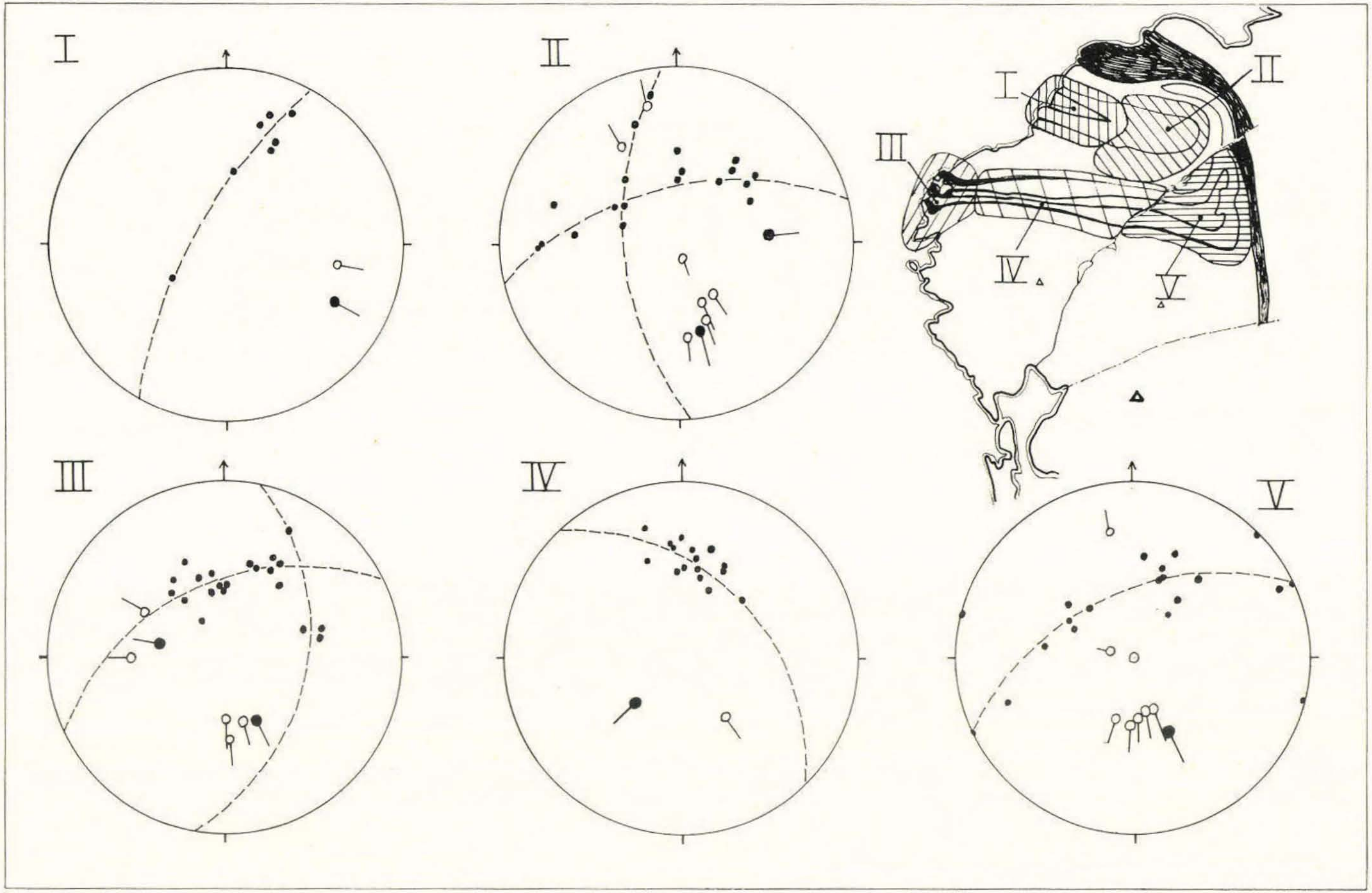

Fig. 65. Stereograms from north-western Tovqussap nunâ.

Due to the rather varied topography of north-western Tovqussap nunâ, the contour map method may be applied successfully in the analysis of the Irdal and Smalledal structures. Using the upper surfaces of the outermost marker horizons as reference horizons, contour maps have been prepared for both structures. From these maps the structural stereogram of fig. 66 was constructed. The position of the two northernmost noses of the Irdal and Smalledal antiforms is, however, inferred and therefore hypothetical. Nevertheless, the diagram brings out clearly

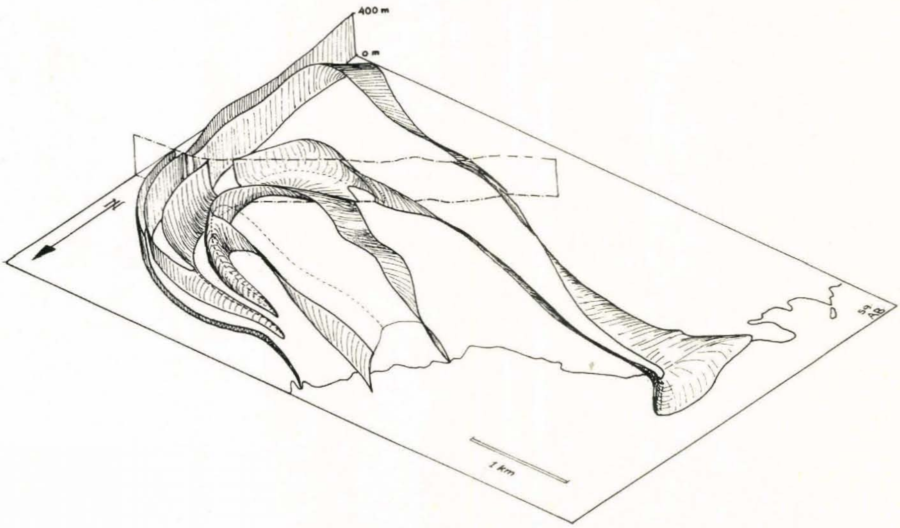

Fig. 66. Structural stereogram of the Irdal and Smalledal structures.

that the reference horizons chosen within each structure represent one and the same surface. Due to the younger antiformal refolding around the Eastern Antiform, the synform connecting the two older structures culminates at Lejrsø above sea level but still below the erosion level. This synform is occupied by the gabbro-anorthosite-bearing gneisses of the Frame Layer. The hinge axes of the two overturned antiforms, the Irdal and the Smalledal antiforms, trend more or less ENE. The younger Western and Eastern Antiforms plunge to SSE where developed within the overturned flanks of the older structures. Where imposed on the hinge zones of the older structures, the plunge of the younger axes has, however, been controlled by the old hinge dips, as for example at Bortehøj (see Pl. 2).

These relations are very similar to those described above from eastern Tovqussap nunâ, where the recumbent folds and their hinges influenced the orientation of the younger axes. Thus one may reasonably compare the overturned Irdal and Smalledal antiforms to the recumbent structures at Krebsesø and Kronehøj. In the north-western region the plunge of the younger structures is somewhat steeper because here the older structures were not recumbent but only overturned. (Naturally, the present overturned nature of the older structures could also be explained by a later tilt of originally recumbent structures in connection with the refolding). In this region, obvious traces of still older isoclinal folds corresponding to the Midterhøj and the Kronehøj isoclines are also absent. No definite answer can, however, be given as to the possible occurrence of isoclinal repetitions within the rock succession of the Irdal and Smalledal structures before the remaining part of the western area has been analysed.

The central part of this remaining area is occupied by the Tovqussaq dome. This structure is flanked on both sides by the southern continuation of the Western and Eastern Antiforms. South of the dome, the Blindtarmen antiform appears. These western structures are separated from the eastern ones by a median structure, the Pâkitsoq antiform, see Pl. 2.

C. The Tovqussaq dome and its surroundings

Turning now to the remaining part of the western area it may be appropriate to start the analysis within the Tovqussaq dome. The rock succession involved in this structure was described in detail on pp. 58—92, where mention of the small scale structures shown by these rocks was also made.

The central part of the dome, which is delineated on the map by the course of the Interior Pyribolite, is in two halves divided by the deep bay Qaersup ilua. The application of the term 'dome' to this structure (BERTHELSEN, 1950) may be disputable since the length to width ratio just exceeds 2 : 1 (which divides domes from brachy-anticlines), but has been retained for the sake of clarity in the nomenclature.

The southern part of the dome appears to be a nearly cylindrical structure, see fig. 67, (stereograms I, II, III and V). Only on the southern part of Akuliaruserssuaq may a slight inconsistency between the constructed and the measured axes be noticed. By superimposition of the stereograms, however, all the foliation poles group around a well defined great circle and the measured axes form a maximum around the corresponding axis. The slight scattering of all plots may be explained easily by the culminative effect of the dome.

The most surprising result is that the Ørenæs structure (stereogram III) conforms so closely to the surroundings. In this connection it may be recalled, however, that both closures within the Little Pyribolite, from which aberrant readings could have been expected, are submerged under the sea. Therefore the author regards the apparent uniaxial nature of this structure as fortuitous. In his opinion, the Ørenæs structure represents an original ENE-trending overturned antiform

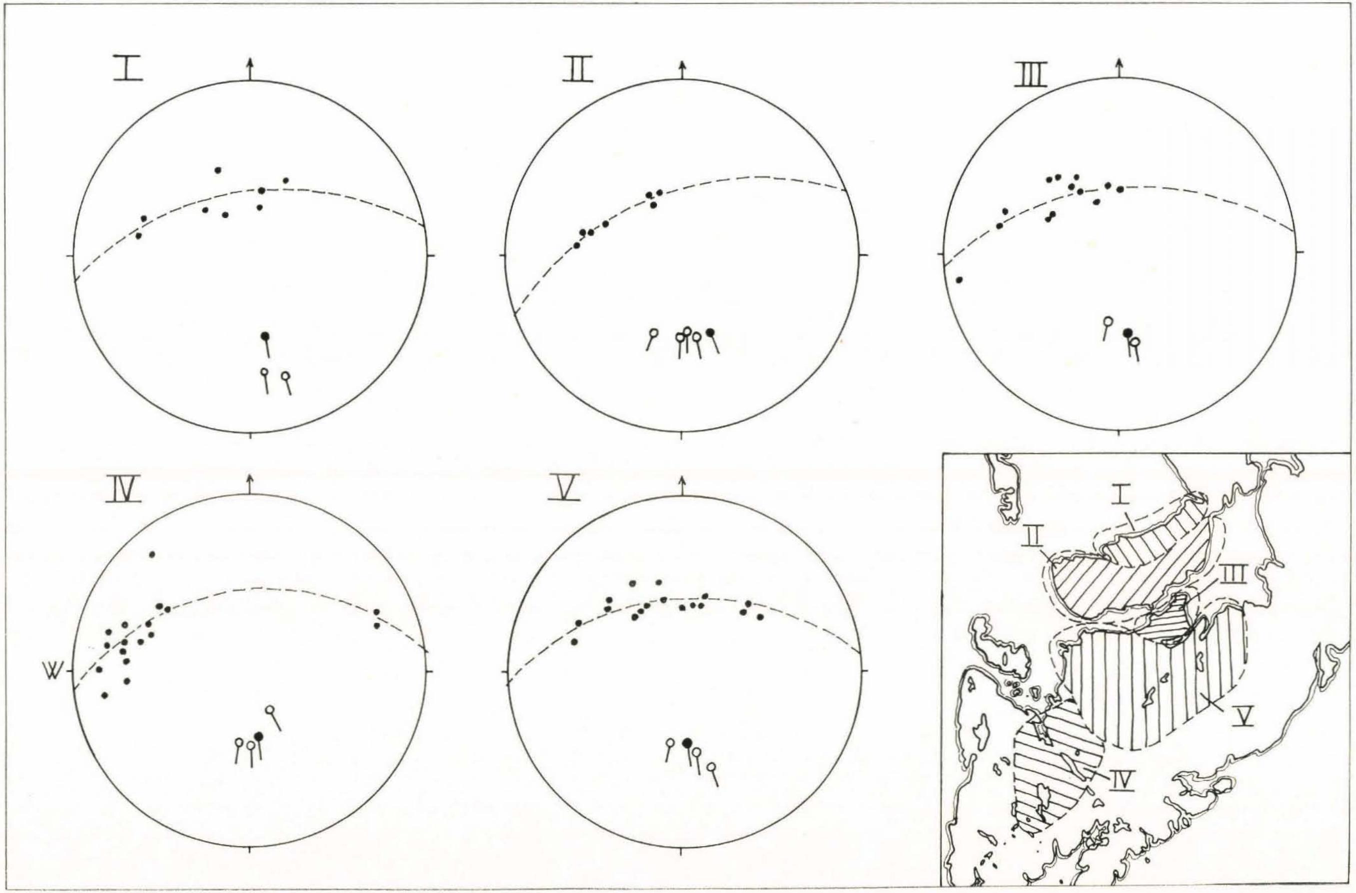

Fig. 67. Stereograms from the southern Tovqussaq dome and the Blindtarmen antiform.

which during the formation of the dome became dragged out parallel to B along the crest of the dome and thereby attained its apparent uniaxial character.

In the Blindtarmen antiform the relations are somewhat more complex, (see stereogram IV, fig. 67). The spreading out of the foliation poles close to W indicate a deviation from cylindrical towards conical shape

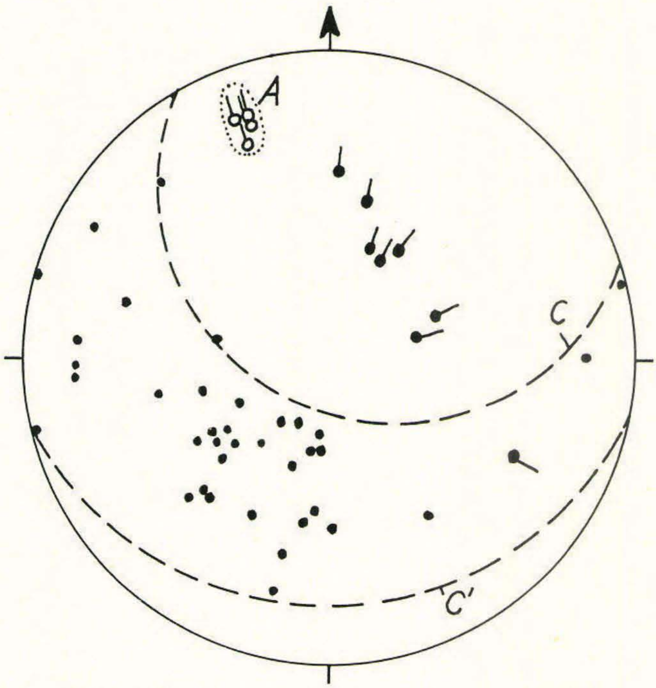

Fig. 68. Stereogram from the northern dome core.

in this structure, i.e. the rock members of this structure become more tightly squeezed together in the southern closure.

Returning to the northern half of the central dome, still more complex relations are met with. If all readings taken within or immediately around the Interior Pyribolite are plotted, the stereogram shown in fig. 68 is obtained. Here, the foliation poles are too scattered to be grouped around a single great circle, and, apart from a local maximum of axes with moderate NNW plunge, a great variation in the orientation of the smallfolds is manifest. The broad zone indicated by the foliation poles can, however, be limited by two small circles (C and C^1 of fig. 68). This suggests that a conical shape with varying angular radius is approached. Knowing from the field that north of Navlen the dip decreases with increasing altitude it may be inferred that the axis corresponding to the partially developed cone surface is convex towards NE.

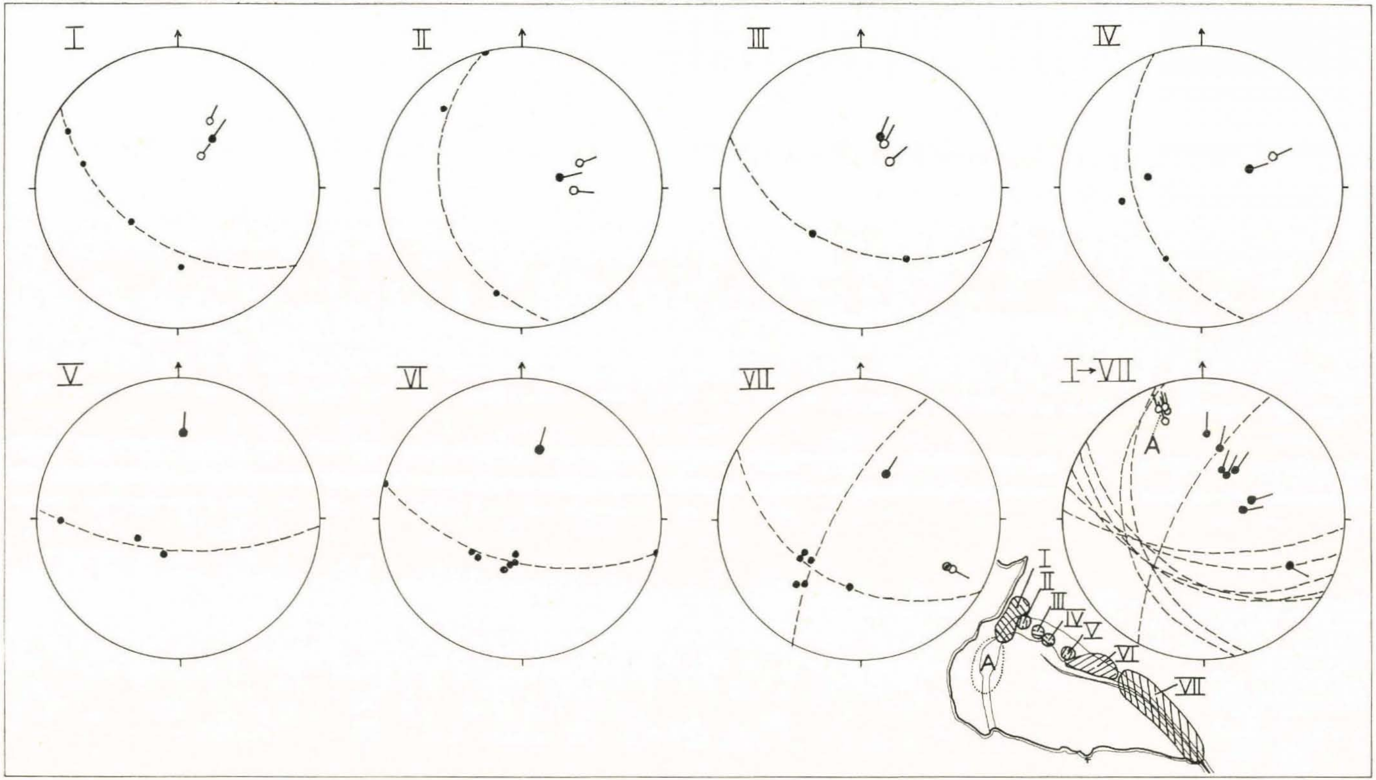

Fig. 69. Stereograms from the northern part of the Tovqussaq dome.

Along the western part of the dome the Interior Pyribolite is strongly folded, and it is the axes of these drag folds which cause the maximum (A) of fig. 68. Along its northern and eastern course, the pyribolite shows also dragfolding but of a different style. The axes of these drag folds plunge in any direction between north-east and south-east.

The stereograms I to V in fig. 69 show some of these smallfolds which have been measured directly or indirectly at various localities within the Interior Pyribolite. As appears from the compiled diagram I—VII, they show a great variation in plunge, although they all indicate a relative movement of the overlying rock masses from left to right when viewed down the plunge.

A relative dating of the NNW-plunging axes and the axes with NE to SE plunge has not been possible in the field since their development seems to be restricted to two separate sectors. An indirect dating according to their style (gentle open asymmetric folds versus folds with sharp hinges to isoclinal folds) would also be very risky. The only means of arriving at a relative dating is by considering their age relations to the major structure, analysing this latter independently.

The major structure can be studied with the aid of the contour method. Fig. 70, A and B, shows contour maps constructed from the upper and lower surfaces respectively of the Interior Pyribolite north of Qaersup ilua. These maps bring out that the northwestern part of the dome is overturned and not in a simple way as if the dome itself had become overturned. The concave bend of the overturned part, which is best seen in fig. 70 C where the constructed outcrop pattern of the marker horizon at sea level is shown, suggests that a north-east trending hinge zone belonging to an overturned or recumbent structure has been refolded antiformally around a NNW-plunging axis. The NNW-plunging smallfolds in the western flank of the structure can thus be considered as having been developed in connection with this refolding. The smallfolds in the northern and eastern part of the structure evidently have a different origin. Their variation in trend is too great to be accounted for solely by spreading due to refolding. The author assumes therefore that they were imposed on the recumbent structure during an intermittent phase of deformation, which caused the development of an initial copula-shaped structure. Levelling out of the effect of the later antiformal refolding would then re-establish the original orientation of these drag folds. Viewed in this manner they indicate concentrically upward directed movements along the flanks of the initial dome.

This interpretation implies that the following three local factors have played a role in the development of the present Tovqussaq dome: 1, The existence of a NE-trending hinge zone belonging to an overturned or recumbent fold in the Interior Pyribolite. 2, Formation of a copula-shaped bulge on this structure due to movements directed towards the

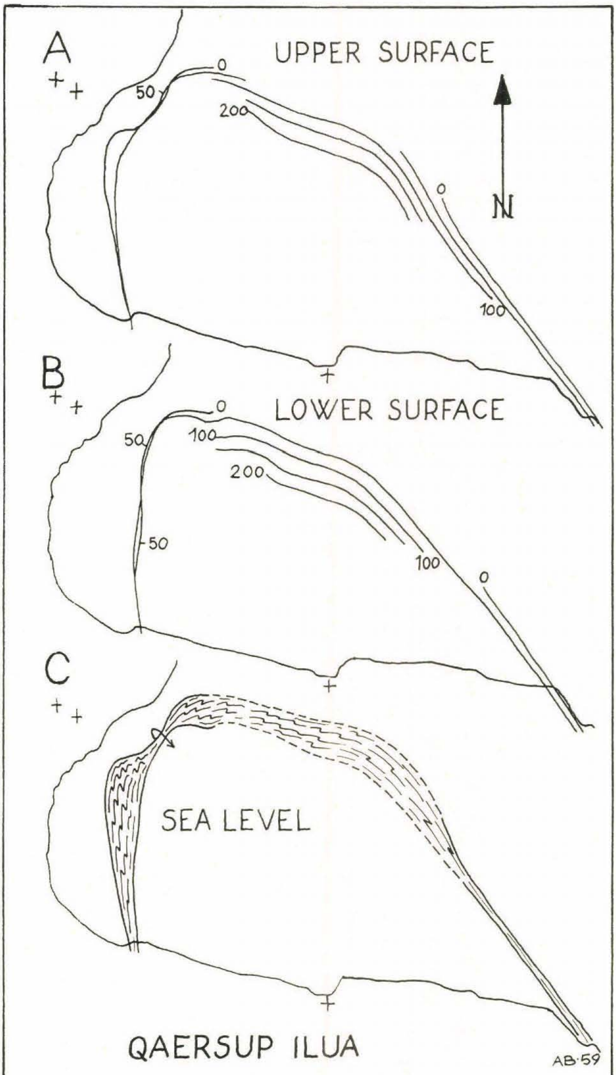

Fig. 70. Structural contour maps of the Interior Pyribolite, north of Qaersup ilua. A for the upper surface, B for the lower. C shows the constructed outcrop pattern of the pyribolite within the 0 m level.

top of the bulge. 3, Antiformal refolding controlled by the earlier developed structures. This composite origin would also explain the more or less conical surface developed in the northeastern sector of the dome.

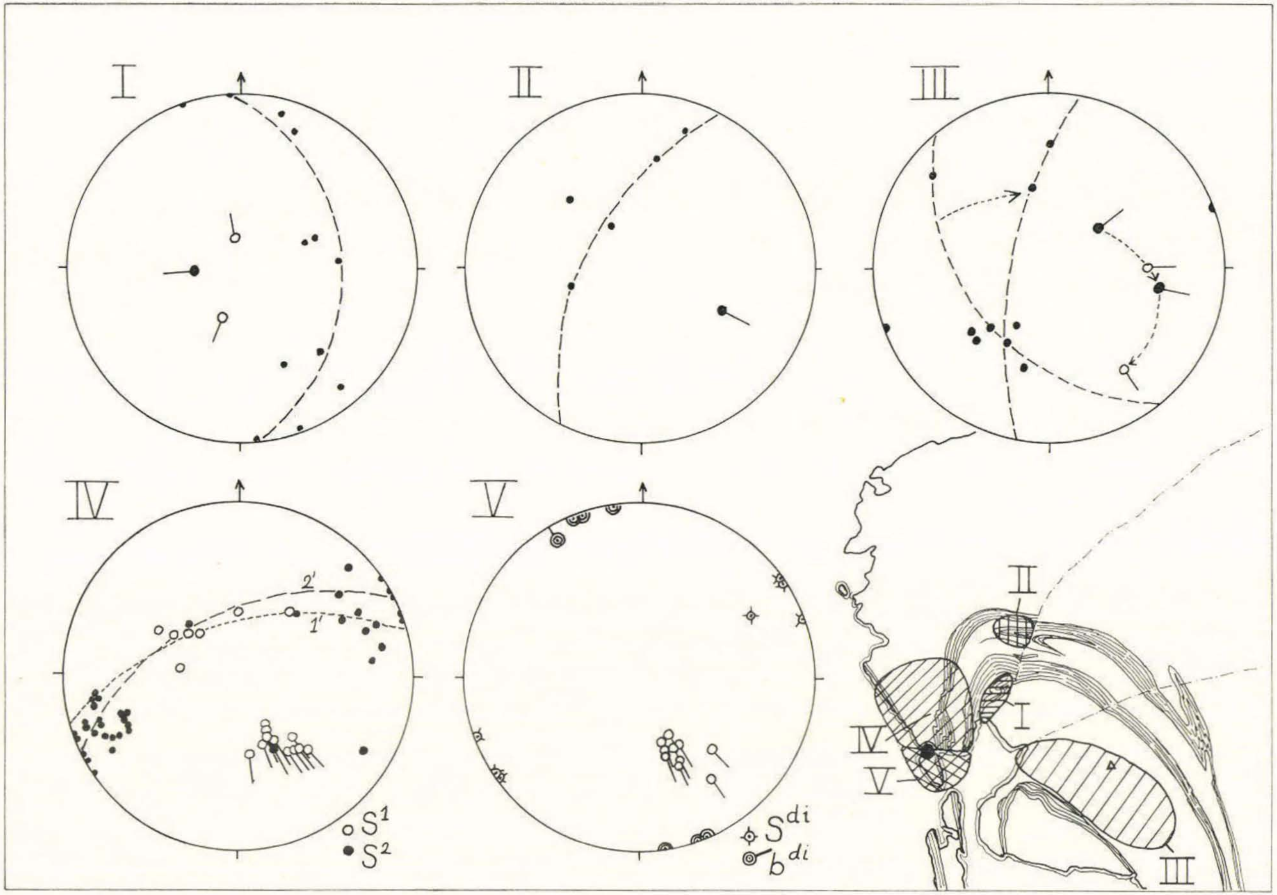

Fig. 71. Stereograms showing 'aberrant' axes within the northern dome and its surroundings.

The remainder of the northern part of the Tovqussaq dome and the area between this structure and the Smalledal structure will be considered next. As mentioned in the description to the geological map, minor structures are difficult to measure within the rocks of the Frame Layer in the region around the Tovqussaq Mt. The stereogram from this region, therefore, contains only a few plots which are not really representative (see fig. 71, III). Nevertheless, the diagram brings out that the axial plunge varies from NE to ESE and even to SSE in the easternmost part of the region. This variation is very similar to that described from the Interior Pyribolite.

Going to area II of the same figure, i.e. the synform between the Great and the Pas Pyribolites west of Ankerbugtdal, and ESE-plunging axis can again be recognised, see stereogram II. West of this region, on Sorthat, nearly east-west trending horizontal smallfold axes have been observed.

In the lower part of the Ankerbugtdal several larger folds have been mapped within the Great Pyribolite and the Little Pyribolite. Stereogram I, fig. 71, shows that the plunge of these folds is steep westerly. Turning now to stereogram IV of fig. 71 it might be tempting to draw a great circle corresponding to a similar steeply W-plunging axis in this diagram. This, however, cannot be done, because the spreading out of the foliation poles in the eastern sector of this stereogram is merely due to a conical widening (virgation) of the northwestern flank of the antiform structure of this region. The stereogram thus serves as an example of the need for caution when stereographic projection is used for geometrical constructions and not in a statistical way. If no geological 'ground control' is applied—in the present case by aid of the marker horizon west of Strøget—serious errors may result.

Axial disharmony

Stereogram IV of fig. 71 serves, however, to illustrate quite another feature, i.e. the complex nature of the Western Antiform. Where this antiform affects the Pas Pyribolite south of Kolbesø, a younger shearfolding of somewhat older gentle structures is seen, (see also page 79). Within the hinge zones of the gentle folds, banding (S^1) enabled measurements to be taken. These readings are shown as open circles in stereogram IV, fig. 71, while measurements of the foliation (S^2) developed in connection with the shearfolding are indicated by the symbol generally used for foliation poles (filled-out circles). The latter group themselves close to the WSW periphery of the diagram. The two great circles 1' and 2' which are based on the S^1 and S^2 plots respectively, suggest the existence of a slight axial disharmony between the two sets of structures. This disharmony becomes more marked further to the north-east, where near Sorthat a later imposed mineral lineation is at an angle to the large scale lithological banding and layering, p. 79.

South of area IV, dioritic rocks are found at Verdens Begyndelse. As mentioned in the description of this diorite occurrence, an older smallfolding is preserved within the diorite while younger smallfolds have been superimposed on the eastern gneisses. In stereogram V, fig. 71, plots of poles of foliation measured from the smallfolded enclaves within the diorite are shown by the symbol S^{di}, whereas axes measured in the enclave rocks are marked by the symbol b^{di}. Smallfold axes from the eastern gneisses and the remaining part of Hestenæs are shown as axes with open head (normal symbol). In this diagram a difference in orientation between the older smallfolds (enclosed as relics within the diorite) and those superimposed later on may be noticed. The first-mentioned are horizontal and trend NNW, while the last-mentioned plunge to SE. It seems natural to compare this disharmony to that just described from area IV and around Sorthat.

The slight difference in the orientation of the enclave axes and the axes from the rocks surrounding the Nordnor diorite (p. 118 and fig. 46) now becomes more comprehensible. Around the Nordnor, the axes with the most southeasterly plunge, however, represent the oldest structures.

Areas IV and V of fig. 71 both belong to the Western Antiform which can be followed from Kangeq along the western flank of the Tovqussaq dome. South-west of this latter structure, the Western Antiform affects the rocks of western Tugdlerúnarssuit (see fig. 29). Realising the two step evolution of the regions just discussed, the complex hinge structures on Tugdlerúnarssuit (fig. 29) can better be understood. The recrystallised shear zones now appearing as zones of flow traversing the smallfolds of the hinge zone were, most probably, formed in connection with the imposition of the younger structures (e.g. S^2 of stereogram IV, fig. 71) in rocks occupying a more deep-seated position within the antiform.

Between Kalotbugt and Gule Hav, the Western Antiform has caused the development of complex structures. Although this part of the antiform appears fairly simple from stereogram I of fig. 72, field observations on small scale double fold structures (e.g. on the little island in Gule Hav) clearly show its composite nature. Unfortunately, the greater part of the structure is sea-covered, so its original outcrop pattern can only be guessed. The only marker horizon in this region is the diorite, which forms a concordant layer along the west coast of Gule Hav (hence the name 'Yellow Sea'). To the south this layer contains enclaves of gabbro-anorthosite. Because of the occurrence of gneiss on the east coast of the small island and because the diorite layer cannot be traced into Smalledal or south of Kalotbugt, the author has interpreted the outcrop pattern of this layer in the manner shown on Pl. 2, where it is seen to

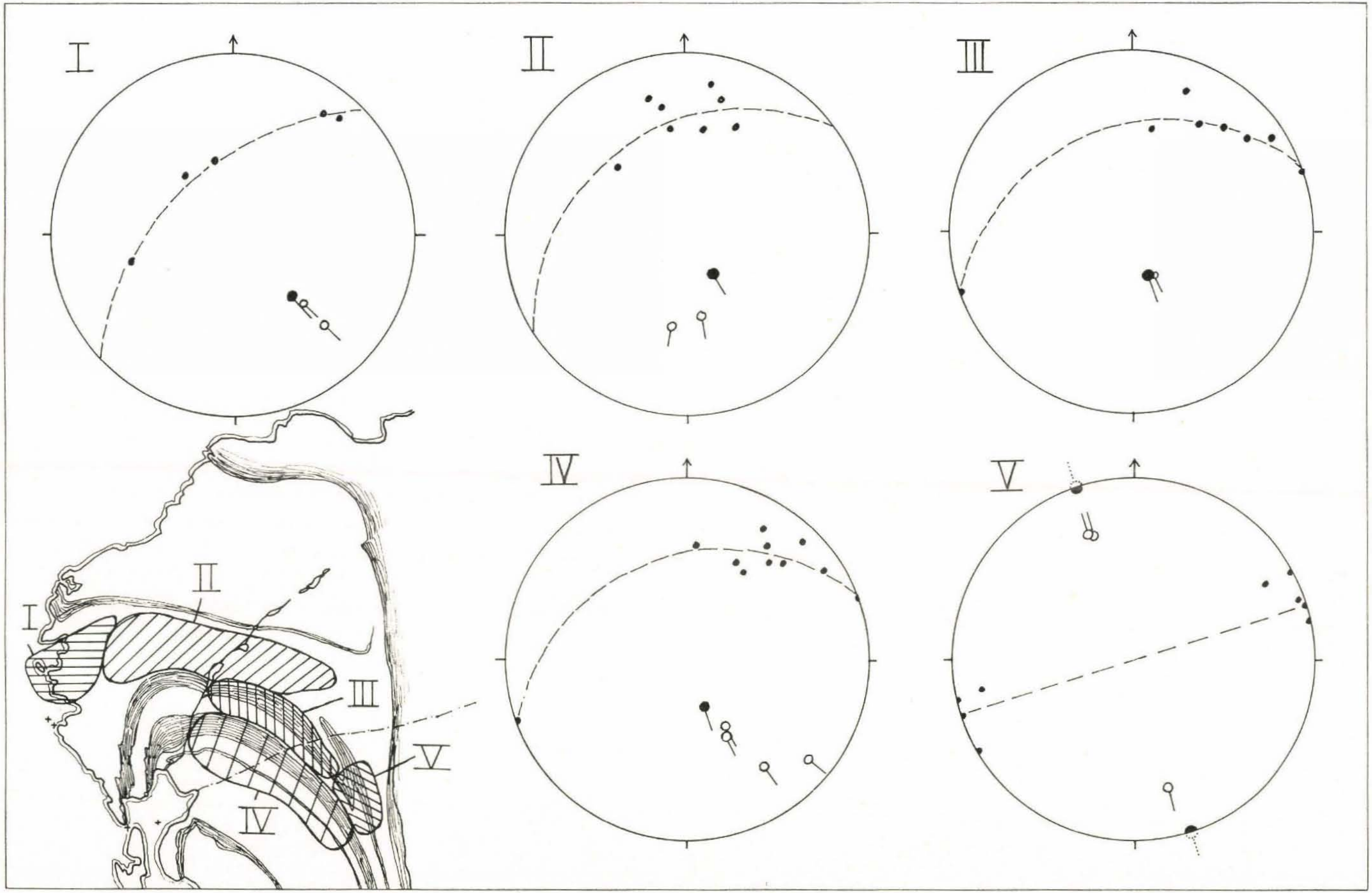

Fig. 72. Stereograms from the regions around Gule Hav, Gammel Lejrskar, and the Pas Pyribolite.

form an arrowhead-shaped structure. This structure is explained as having been formed by antiformal refolding around a SE-plunging axis of an original recumbent structure with a nearly E—W trending hinge axis. As will be shown on a later page, this interpretation fits well into the general geometrical pattern.

After this digression into the Western Antiform we will return to the dome structure proper and the region which to the north separates it from the Smalledal structure. Knowing now that the Smalledal structure is an overturned antiform, it is possible to predict the existence of a synformal structure between this and the dome. Since the northern part of the dome is overturned, the intervening synform will not be apparent from the measured dips. The amphibolitic—in part agmatitic—gneisses north and west of Skraaland show, however, a closure south of Lille Ekkodal. The mere presence of this closure within the eastern flank of the Western Antiform shows that, although strongly refolded by the younger SE-plunging axes, it represents the hinge zone of the synform separating the dome and the Smalledal antiform. This relation being realised, the synformal position of the granitic gneisses of Skraaland may also be inferred. The scepticism expressed on p. 93 as to the litho-structural validity of the petrographic base for the mapping of this region can now be disregarded. The complexity of the structures does not appear from the stereogram (II, fig. 72), which only shows that the region is dominated by a rather steep SSE-plunging axis. Similar axes may also be constructed from the foliation poles from areas III and IV of fig. 72, which cover sectors of the Pas and the Great Pyribolites respectively. In diagram IV, the distribution of the measured smallfolds and lineations is, however, suggestive of the presence of older refolded linear structures.

The folds within the Pas Pyribolite south-west of Vinkelsø show nearly horizontal axes trending NNW—SSE, see stereogram V, fig. 72. The position of these structures in relation to the dome can be compared to the horizontal smallfolds occurring within the diorite at Verdens Begyndelse.

The Eastern and the Western Antiforms

The Eastern Antiform, which refolded the Smalledal structure at Trehøje, may be traced across Nordvestpassagen into the region immediately east of Vinkelsø. From here it extends isoclinally south on to Pâkitsoq, south of which it causes a tight closure within the Pas Pyribolite. This hinge zone appears to be homoaxial with the southern part of the dome. The folds within the Pas Pyribolite west of Vinkelsø occur within the western flank of the Eastern Antiform and were presumably developed in connection with the formation of this structure. They are, however, axially controlled by the dome structure, since their axes describe a complex culmination.

Returning to the Western Antiform, very similar relations are found here. The folding of the Pas Pyribolite between Kolbesø and Verdens Begyndelse and on northern Langø represents an analogous influence of the Western Antiform on the west flank of the dome. Further to the south, the Western Antiform also refolds the Pas Pyribolite, wrapping this layer around its hinge zone.

These relations suggest that both the Western and the Eastern Antiforms were formed later than the initial dome structure, but were developed more or less contemporaneously with the refolding of the dome; this was also accompanied by the formation of the Blindtarmen antiform.

Correlation of the litho-structural units of the western area

Having now studied the geometrical aspects of the western area, it may be useful to consider how the litho-structural units established within this region fit into the structural pattern.

Unfortunately, parts of the western structures are sea-covered, so no continuous tracing of the marker horizons has been possible within this area. The litho-structural units of the western area show, however, certain individual features, particularly in the gneisses, which have made a fairly safe reconstruction of the outcrop pattern possible. Arguments for the reconstruction shown on Pl. 2 were given in connection with the description of the geological map (Pl. 1).

The northern part of western Tovqussap nunâ, where the outcrop pattern is more complete, will be considered first. Here the gabbro-anorthosite-bearing gneisses of the Frame Layer circumscribe the Irdal and Smalledal structures. They occupy the synform which separates these two structures and can also be traced directly into the northern surroundings of the Tovqussaq dome. Along both flanks of the dome, they form the cores of the Western and Eastern Antiforms. In the Western Antiform, the Frame Layer gneisses extend south on to western Tugdlerúnarssuit, while in the Eastern Antiform they reach the south shore of Pâkitsoq. Within this distribution a decrease in the abundance of gabbro-anorthositic enclaves may be noticed going from north to south — or south-east. It should also be mentioned that gabbro-anorthositic enclaves occur within other litho-structural gneiss units in the northern part, where they are particularly abundant in the Frame Layer gneisses.

The Frame Layer gneisses surrounding the central Tovqussaq dome correspond in their lithology completely with those just mentioned from the surroundings of the dome. As is suggested by the author's application of the term Frame Layer to both these units (see Pl. 2), they may be considered as representing one and the same rock succession. This being the case, a repetition should occur within the rocks forming the outer part of the dome structure.

Is there any evidence of such a duplication within the rock layers of the northern dome? This question can only be answered in the affirmative.

Turning to the geological map (Pl. 1), a symmetry around a median line through the 2nd Intermediate Layer may immediately be observed. On both sides of this layer thick pyribolitic members occur, the Great and the Pas Pyribolites respectively. Then follows a gneiss layer which in turn is succeeded by a thin pyribolite layer (Little Pyribolite) or a thin layer of gabbro-anorthosite. Knowing from the Smalledal structure that pyribolite and gabbro-anorthosite from a structural point of view can be considered as 'isomorphous', the symmetry shown by the rock layers in the northern part of the Tovqussaq dome can be said to fulfil expectations perfectly.

Since the outcrop pattern favours the idea of a repetition, the next step in the analysis will be to look for more direct evidence of the existence of a recumbent structure which could cause the duplication. Such evidence is close at hand. Around the Ankerbugtdal there occur within the light coloured gneisses of the 2nd Intermediate Layer two (at present synformal) hinge zones, which indicate the existence of a more deep-seated hinge zone which causes the Great and the Pas Pyribolites to join up into one layer, fig. 73. The axes of the visible parts of this complex hinge zone plunge moderately to ESE (fig. 71, stereogram II). A little further to the west, E—W trending horizontal axes have been observed. Allowance being made for the deflection of these linear structures by the doming and refolding, an original more or less E—W trend of the hinge axis in this large recumbent structure may be inferred.

Since the symmetry described in the outcrop pattern around the 2nd Intermediate Layer is also valid for the rest of the dome (with the exception of the disappearance of the outermost thin pyribolite layer in the southern portions of the structure) it may be concluded that the dome structure has been formed entirely within the flanks of the large recumbent structure.

An amplitude of more than 10 km for the recumbent fold can therefore be estimated, i.e. the structure may be termed a nappe.

Before extending the interpretation which lead to the conception of this large nappe, it may well be worth considering the relations around the northern part of the dome in more detail. Could any further evidence for the existence of a buried nappe front be found here? As shown in fig. 73, which presents a sketch profile through this northern part, the frontal hinge zone of the nappe should occur at some depth below Skraa-

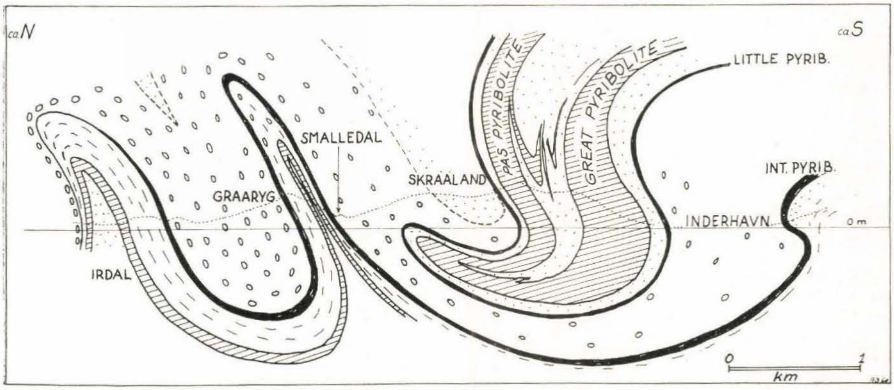

Fig. 73. Longitudinal profile through northwestern Tovqussap nunâ.

land, the northernmost parts of the nappe being influenced by the synform separating the Smalledal structure and the dome.

Knowing that refolding around the Eastern Antiform at Lejrsø caused the synform separating the Irdal and the Smalledal antiforms to rise above sea level (see structural stereogram of fig. 66), it is natural to look for the effect of refolding of the complex Skraaland synform (fig. 73) by the Western and Eastern Antiforms. Doing this, the arrowhead-shaped structure shown by the diorite at Gule Hav immediately attracts attention. In the southern hinge zones of pyribolite within the Western and Eastern Antiforms this rock may be represented by diorite where closing antiformally around these structures, and thus there is a likelihood that the dioritic layer at Gule Hav represents an original thick pyribolite layer occupying a somewhat similar structural position.

The arrowhead-shaped structure at Gule Hav may therefore be regarded as the frontal hinge zone of the large nappe which, due to antiformal refolding around the Western Antiform, has been lifted above the present erosion level. This interpretation concurs with the actual observed small scale double folds within the Gule Hav structure — and with the more or less E—W trend assumed for the hinge axis of the nappe.

The gabbro-anorthositic band outside the diorite north of Kalotbugt and the agmatised pyribolite and amphibolite in the granitic gneisses east of the Gule Hav would, on this interpretation, represent the outermost thin basic marker horizon forming part of the nappe structure.

In Pl. 2 the outcrop pattern connecting the west coast and the islands (Kalotten and Ivigssuartôq) has been constructed tentatively in accordance with the interpretation of the Gule Hav structure just outlined. The nappe shown by the thick pyribolite layer is here assumed to rise once again to the surface in the Kalot structure. The hinge zone of the nappe would appear here as a closure joining the pyribolite of the two islands and the pyribolite of the west coast. This closure would fall north-west of the Kalot island — just outside the map area.

This interpretation, which is naturally open to discussion, means that between Kalotnæs and Sorthat, the Western Antiform runs within a larger synform and that the symmetry of the outcrop pattern here is due to still more complex structures than hitherto believed.

Leaving the western portions of the dome 'foreland', we will now investigate the possible effect of the Eastern Antiform on the frontal parts of the large nappe structure. Apart from some abnormal WNW-plunging smallfolds at Nordvestpassagen, no traces of the buried nappe structure have been found in this region. If not contrary to the nappe hypothesis, this must mean that the nappe front, although refolded, does not rise to the present erosion level.

The Eastern Antiform would therefore cause the frontal part of the nappe to plunge deep below Breddal, where the nappe is enclosed in the eastern flank of this antiform. To the north-east, the Påkitsoq antiform would make the thick pyribolite layers of the nappe structure reappear within this composite structure. But before entering into any further discussion on the connection between the western structures and the median structure, the implications of the great nappe regarding the correlation of the different litho-structural units of the southern dome region should be studied.

As shown schematically in fig. 74, the great nappe has its root below the southern part of the dome. Assuming that this nappe was originally formed in connection with the Irdal and Smalledal folds and applying the simplest possible solution connecting it with these structures (as done in fig. 74), the following deductions can be made. Where occurring within the dome, the Frame Layer gneisses occupy a recumbent synform which closes deep below the southern part of the dome. This assumption agrees well with the observed symmetrical nature of the Frame Layer south of Qaersup ilua. The reconstructed longitudinal profile of the western region shows furthermore that there are virtually only five stratigraphic units represented in this region.

1: An upper gabbro-anorthosite-bearing gneiss, which southwards grades into mixed hypersthene gneiss. Overlying it in the northern part, a granodioritic to granitic layer can possibly be distinguished.
2: A thin basic layer which is developed either as pyribolite or gabbro-anorthosite.
3: A gneiss layer which passes from mixed granodioritic rocks in the north into more granitic gneisses in the south.
4: A thick pyribolitic layer with abundant ultrabasic intercalations.
5: A lower mainly granitic gneiss layer, which includes the granulites s. s. of Langø.

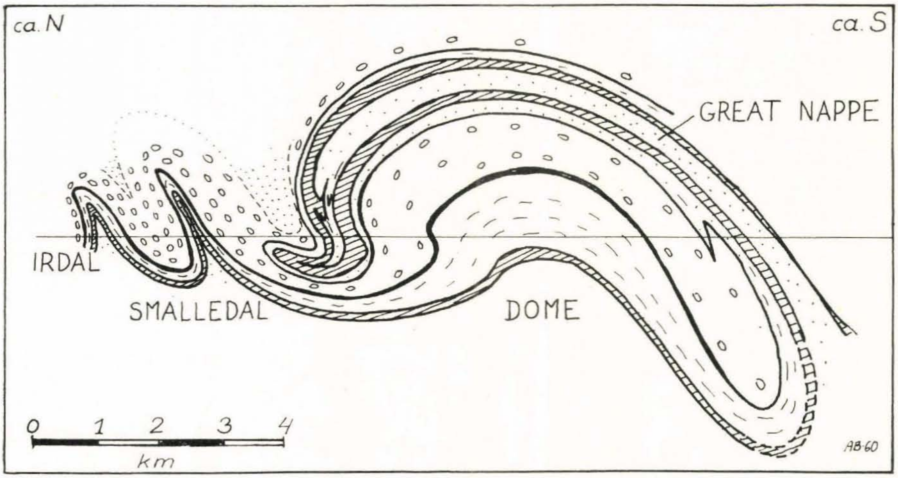

Fig. 74. Longitudinal profile through western Tovqussap nunâ.

This stratigraphic succession is, however, only valid if the western region was not affected by the very oldest phase of isoclinal folding which has left its traces in eastern Tovqussap nunâ. As will be shown later on, when the structural connection between the western, the median and the eastern region has been discussed, it is doubtful if any larger NW-trending isoclinal structures have ever affected the rock succession of western Tovqussap nunâ. The lack of older isoclines in this region is also suggested by the great similarity between the stratigraphic column just described and that established from the triple folded eastern structures.

D. The Pâkitsoq antiform

An analysis of the minor structures of this composite antiform is shown in the stereograms I to V of fig. 75. In the northernmost area (I) the axes are seen to plunge moderately to SSE. The stereogram from area II shows a significant scattering of the foliation poles near to the periphery of the diagram. The measured smallfolds, however, show a corresponding variation which may be attributed to a slight axial disharmony due to the general southerly widening of the structure. It becomes distended. In area III, a similar effect may be explained more easily since the measurements taken within the closure of the intermediate Stjernesø synform (subarea IIIb) correspond to a SSW-plunging axis (b'), while readings from the remainder of area III (IIIa) define a S-plunging axis (a'). Around Ganghøj (stereogram IV) and south of this region (stereogram V), where the Stjernesø synform is not developed, a clearer picture is arrived at.

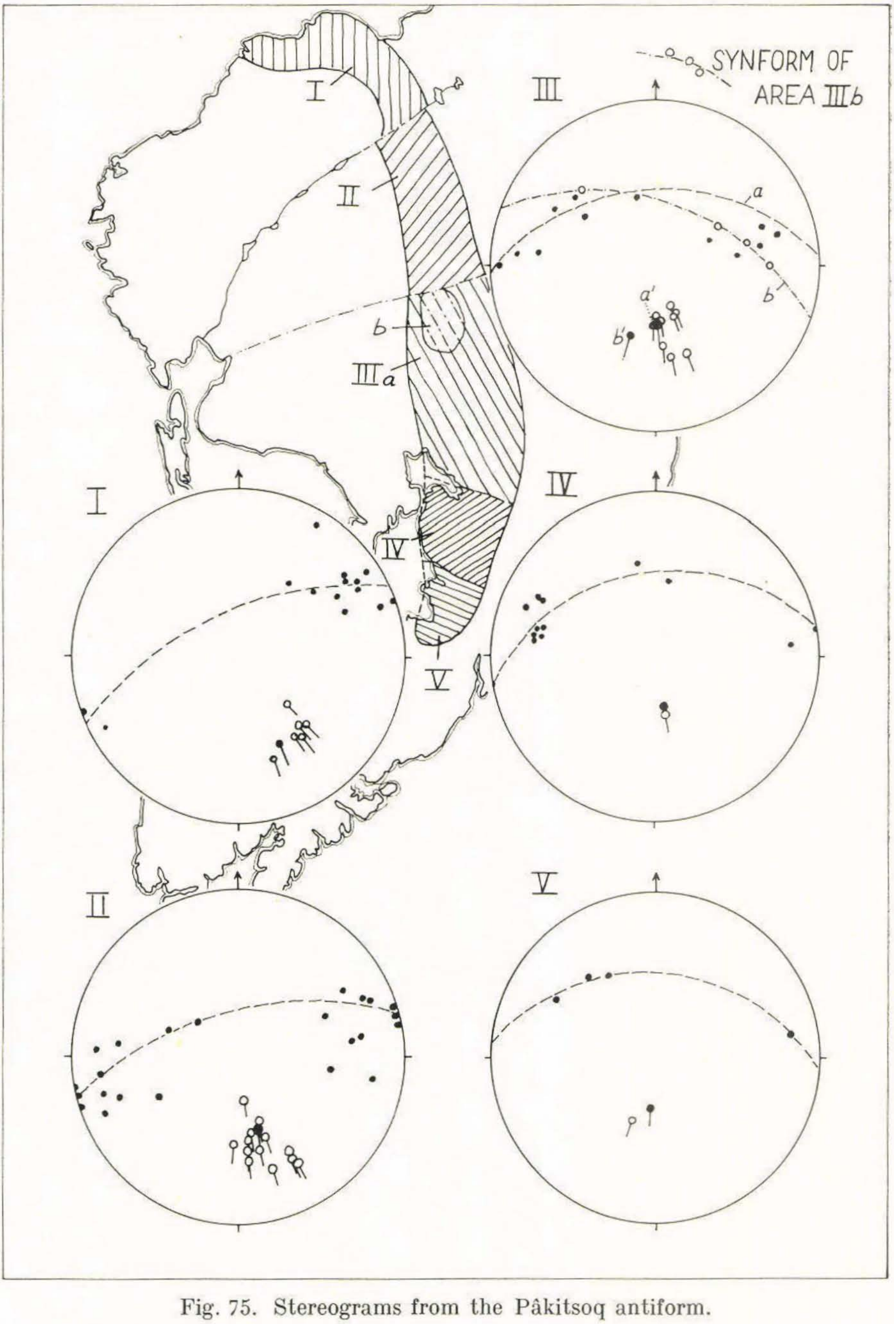

Fig. 75. Stereograms from the Pâkitsoq antiform.

These stereograms indicate that although composite the Pâkitsoq antiform is a fairly uniform structure with SSE to S-plunging axes, it

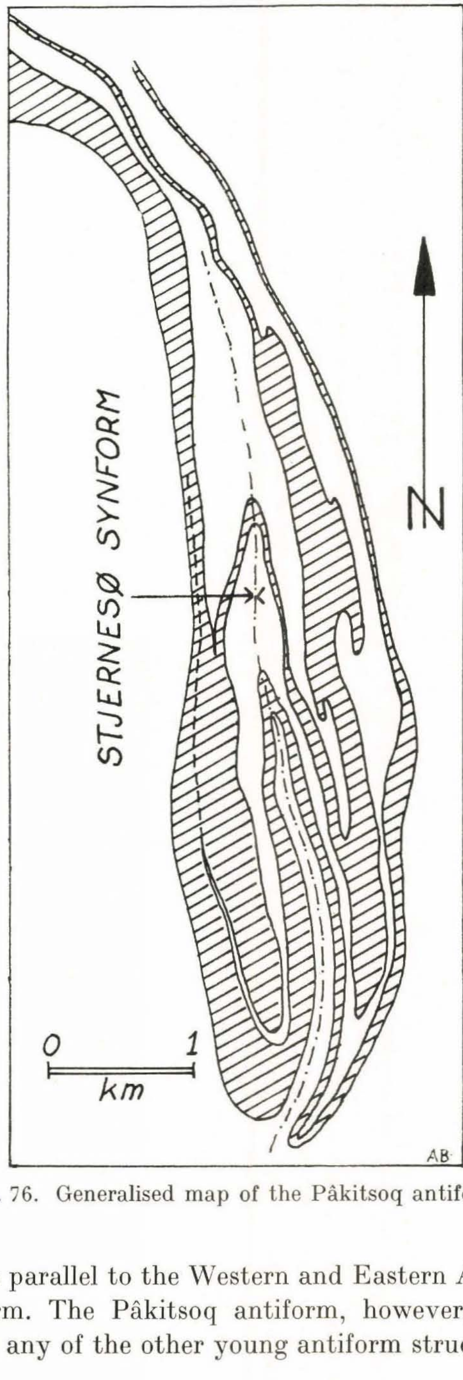

Fig. 76. Generalised map of the Pâkitsoq antiform.

runs more or less parallel to the Western and Eastern Antiforms and the Krebsesø antiform. The Pâkitsoq antiform, however, is more tightly compressed than any of the other young antiform structures.

The complex nature of the Pâkitsoq antiform can only be realised from a study of the outcrop pattern. Fig. 76 shows a slightly generalised map of the structure as outlined by one marker horizon, the Pâkitsoq Pyribolite. Within the Stjernesø synform, this marker horizon is represented by a lower and an upper layer, both of which participate in the eastern antiformal fold, the lower in an extremely complex manner. The gneiss layer separating the upper and the lower parts of the Pâkitsoq Pyribolite also forms part of the western antiformal fold, but terminates within the western flank of this fold. If the termination of the gneiss layer is due to a tight closure, a duplication of the pyribolite layer may be expected within its northwestern course which is also unusually broad here. Still further to the north-west, the basic marker horizon should close within itself, as suggested on Pl. 2, where this northwestern closure has been placed arbitrarily in the Eqaluk fjord. Possibly detailed mapping of the islands north of the fjord would help to fix the position of the closure more accurately, but in this case allowance should be made for the displacement along the Eqaluk fault running along the southern edge of the islands.

On Pl. 2 there is shown in addition another closure which makes the upper and lower parts of the Pâkitsoq Pyribolite join up in the eastern flanks of the main antiform. The existence and position of this closure is naturally also open to discussion. The reader may think that the author deliberately places all critical hinge zones under the sea or in lakes, i.e. where their existence cannot be checked at all. In defence of his reconstructions the author may, however, plead that hinge zones with aberrant axes would favour deep-cutting erosion.

In the present case, the two closures have been postulated in order that the Pâkitsoq structure should conform with the western region. The occurrence of the intermediate gneiss layer in the southern and eastern parts of the composite antiform suggests that this latter refolds a large recumbent structure. The frontal parts of this recumbent structure—as shown by the Pâkitsoq Pyribolite—form the two postulated closures at Eqaluk, one within each of the flanks of the younger antiform. The roots of the large recumbent structure should be looked for deep below the southern closures of the young (here composite) antiform, i.e. south of Ilordlia or Pâkitsoq.

Since the younger Pâkitsoq antiform plunges to SSE or S, the refolded recumbent structure can be described as an antiform with parallel limbs and an amplitude of more than 10 km or simply as a nappe. This interpretation has been chosen in preference to several other possible explanations, since it concurs completely with the ideas developed in the analysis of the western region. Here, good evidence was found in support of the existence of a great nappe structure, and when discussing the effect of refolding imposed on this nappe by the Eastern Antiform it was concluded that a reappearance of the nappe should be expected within the Pâkitsoq structure.

The extreme refolding which the nappe was subjected to within the isoclinal to squeezed and composite Pâkitsoq structure renders a more detailed correlation of the rock succession of the nappe from the western to the median area difficult. The thin pyribolite (or gabbro-anorthosite) layer, which separates the nappe from the Frame Layer gneisses in the west (see fig. 74), is imperfectly developed in the median structure. It may occur in the lower part of the Stjernesø synform closing just south of Breddal and in the antiformal hinge zone just south of Ilordlia, but it is entirely lacking along both flanks of the Pâkitsoq structure. The gneiss layer, which separates the two parts of the Pâkitsoq Pyribolite, would correspond to the 2nd Intermediate Layer of the western region, and the Pâkitsoq layer naturally corresponds to the Great and the Pas Pyribolites.

The Frame Layer gneisses of the western region actually continue into the gneisses around Ganghøj, to which they should correspond according to the nappe theory. They should also be represented by the mixed gneisses forming the core of the median antiform south of the Eqaluk coast.

The three discontinuous pyribolite layers occurring within the hinge zone south of Ganghøj cannot be matched with any other known litho-structural units and are considered by the author to belong to the heterogeneous Frame Layer succession.

On the existence of still older isoclinal structures in the western and central region

A study of the geological maps of Pl. 1 and Pl. 2 shows that the structural interpretation given above is completely consistent with the outcrop pattern, although in the western and median regions the interpretation only involved two main phases of folding: an older causing overturned to recumbent folds trending about ENE and a younger causing open to closed or even squeezed refolding around SE to S-plunging axes. In the central part of the western region an intermediate phase with doming could be distinguished as well.

Features which suggest triple folding have not, however, been noticed, except in the eastern region. Taking the establishment of the ENE-trending recumbent structures for granted, a check can be carried out to see whether or not they enclose older isoclinal structures. For this purpose the structural connection between the western, middle and eastern regions must be clarified first. In the east, the recumbent structures comprise a lower recumbent synform and an upper recumbent antiform. The recumbent synform is underlain by the great nappe of the middle and western region. In the west the great nappe structure overlies the Skraaland synform, which is succeeded by the Smalledal antiform, the Graaryg synform and the Irdal antiform.

In the east, the Midterhøj and Kronehøj isoclines were refolded by the recumbent structures and naturally exerted their influence on the resulting outcrop pattern. Had similar isoclines occurred in the west,

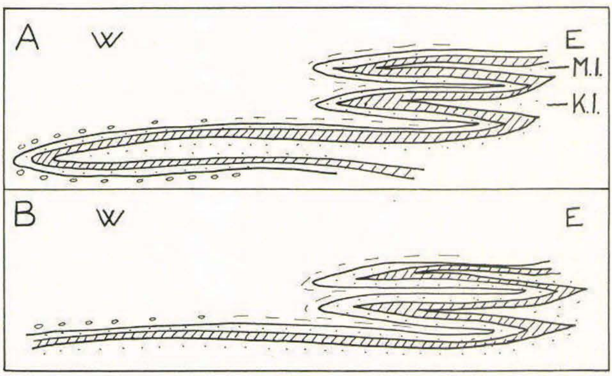

Fig. 77. Schematic profiles showing two alternative interpretations of the structures developed during the first isoclinal folding.

they would also influence the outcrop pattern produced by the subsequent recumbent folding.

In fig. 77 two alternative E—W profiles through Tovqussap nunâ before the onset of the recumbent folding are presented, one in which large scale isoclinal folding occurs throughout the region (A), and one in which the isoclinal structures are restricted to eastern Tovqussap nunâ (B). Starting from the situation pictured in profile A, and folding up the great nappe of the western and middle region, we arrive at a quadruple repetition of the rock units within the nappe. This would mean that all four pyribolitic marker horizons of the nappe belong to a single horizon and that the 2nd Intermediate Layer represents the Frame Layer gneisses. This result does not agree at all with the observed outcrop pattern around the dome.

Nappe-folding of the rock succession shown in profile B would, on the other hand, explain all the mapped features.

The Smalledal structures could be fitted better into the pattern resulting from recumbent refolding of the sequence shown in profile A, since the central gneiss layer of the Smalledal structure actually carries enclaves of gabbro-anorthosite—just like the Frame Layer gneisses—but they are far from being as abundant as within the latter. Their appearance within the core gneisses is more likely to be due to the overall increase of this rock type towards north (or NE) within the area studied.

The outcome of this check, therefore, does not favour the idea of the existence of any larger isoclinal structures in the western part of the Tovqussaq peninsula.

The rock succession established for the western region (p. 172) can thus be directly compared to that previously (p. 151) deduced for the eastern region. Realising that within an area of 25 × 40 km at least (the isoclinal and recumbent folds being unrolled), this succession is built up of conformable rock members, it seems most probable that it actually represents the original stratigraphic sequence. The total thickness of the series occurring within the Tovqussaq peninsula is of the order of 1000 metres.

This completes the geometrical analysis of the major structures. It need only be mentioned that the author's conception of the structures of the western and middle Tovqussap nunâ is represented in the structural stereogram of Pl. 3. This isometric diagram is based on construction (plunge-projection, contour maps, etc.), but is to some extent interpretative.

Pl. 4 shows the stratigraphical division of Tovqussap nunâ. The correctness of this division depends entirely on the results of the geometrical analysis.

VIII. THE KINEMATIC EVOLUTION OF THE TOVQUSSAQ STRUCTURES

A review of the kinematic evolution of the Tovqussaq structures, based on the results of the structural analysis, will now be given. This review also incorporates the post-dioritic phase of the structural evolution, which, because of its relative simplicity, has not been considered worth an analytical description.

The ultimate result of the geometrical analysis described in the foregoing chapter was the establishment of an originally more or less flat-lying conformable series comprising five stratigraphic members, the total thickness of which amounts to about 1000 metres. Taking the usually rather steep plunge of the Tovqussaq structures into account, this figure may at first sight appear surprisingly low, but repeated duplications due to refolding account for the moderate thickness actually arrived at.

The Midterhøj phase

The first tracable structures developed within this series are the Midterhøj and Kronehøj isoclines of eastern Tovqussap nunâ, fig. 78, (1). The axes of these structures seem to have had an original NW—SE or NNW—SSE trend. The amplitude of the isoclines must exceed 5 km since the synformal closure separating them never appears within the region discussed. The series affected by the isoclinal folding in the east is believed to have remained in its original stratigraphic order in the west, where it was not influenced by the folding. Since this phase of folding was first recognised from the Midterhøj isocline, it will here be called the Midterhøj phase.

The Smalledal phase

The Midterhøj phase was succeeded by another phase of folding during which several recumbent (to overturned) structures with approximately WSW—ENE trending axes were formed. This younger phase will here be referred to as the Smalledal phase. As shown in fig. 78 which

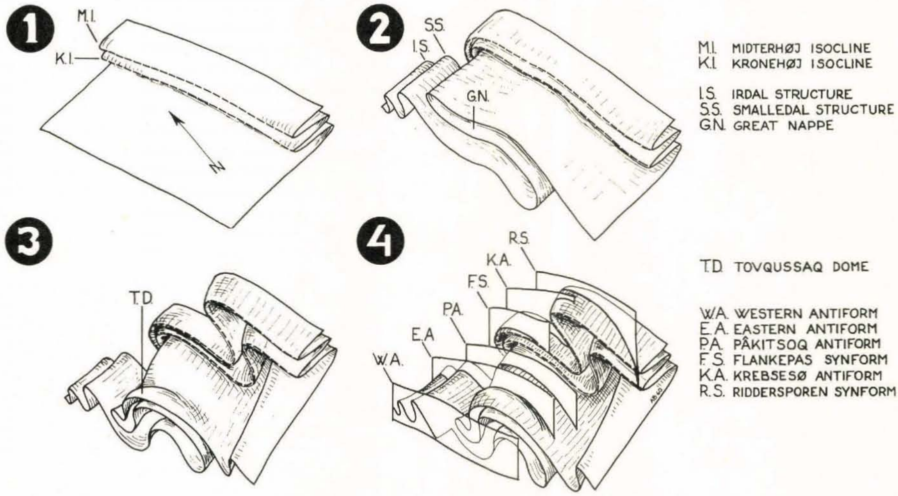

Fig. 78. Schematic diagrams illustrating the kinematic evolution of the Tovqussaq structures

contains four schematic structural stereograms illustrating the successive development of the Tovqussaq structures, the folds developed during the Midterhøj phase (1) were refolded by the later recumbent structures (2 and 3). In these diagrams, it is assumed that the northernmost recumbent structures were the first-formed during the Smalledal phase (2). They comprise the Irdal, the Smalledal and the great nappe structures. The refolding of the western portions of the great nappe is believed to have taken place in connection with the somewhat later development of the overlying recumbent antiform structure (3). Overriding of the great nappe by the upper structures would then explain the refolding of the nappe and the involution formed by its sole. The effect of this refolding can also be traced in the hinge zone shown by the Interior Pyribolite in the northern part of the central dome, and in the originally ENE-trending Ørenæs fold, see fig. 74. For some reason which is not fully understood, this refolding only affected the western area, the median zone being left unaffected by it. Possibly this relation is due to the effect exerted by the structural 'front' build-up east of the median region by the refolded frontal noses of the Midterhøj and Kronehøj isoclines (see fig. 78 (3)).

The assumed difference in time between the formation of the northernmost and the southern recumbent structures is supported by the fact that the recumbent antiform above Kronehøj has a more southwesterly and a steeper plunge than the underlying folds.

The initial doming

The initial doming of the refolded part of the great nappe most probably took place at the same time as the refolding, outlasting it somewhat. The doming was accompanied by a transport of rock material towards the top of the structure. In response to this transport, drag folds were developed within the central parts of the dome. These drag folds were more or less controlled by the grain inherited by the rocks from the Smalledal phase, their asymmetry, however, being dependent on the kinematics of the dome phase. Possibly the folds shown by the Little Pyribolite in the Ankerbugtdal were also developed during the initial doming.

The Pâkitsoq phase

The evolution outlined above brings us up to a stage just before the onset of the refolding around SE to S-plunging axes. The phase of folding responsible for this refolding will here be called the Pâkitsoq phase, since the Pâkitsoq structure owes its formation to this phase in particular.

The refolding of the eastern units caused the development of the Krebsesø antiform and its two flanking synforms, of which the Riddersporen synform is the most important. In fig. 78 (4) the axial planes of these younger structures are shown schematically on the diagram of the earlier-formed structures. For a study of the extremely complex structures resulting from the refolding, the reader may consult the structural stereograms of fig. 64 and Pl. 3.

As pointed out earlier, the pre-existing structures have largely controlled the shape of the structures formed during the Pâkitsoq phase. Perhaps the influence of the older structures has been greatest on the Pâkitsoq antiform proper. This structure, which can be traced throughout nearly the whole N—S extension of the Tovqussaq peninsula, shows a remarkable disharmony when compared to the western and eastern structures. It has also been more tightly compressed than any of the other structures formed during the same phase. These relations are clearly dependent on the setting of the pre-existing structures. To the east the Pâkitsoq antiform is bordered by the 'front' build-up of the refolded old isoclines, and to the west the Pâkitsoq antiform is bordered by the initial Tovqussaq dome. Where strongly squeezed (around Breddal), the Pâkitsoq antiform is controlled by contemporaneous refolding (Eastern Antiform) of the Irdal and Smalledal structures. This means that the median part of the great nappe has been forced into the composite Pâkitsoq antiform just as if it had been compressed between the jaws of a vice.

The disappearance of the intermediate Stjernesø synform in the upper part of the Pâkitsoq structure indicates that the latter was formed mainly by flexure slip folding.

The control of the older hinge zone structures on the smallfolds and lineations formed during the Pâkitsoq phase has already been discussed in detail, so this point needs no further comment here.

In the western region, the folding of the Pâkitsoq phase caused the development of the Western and Eastern Antiforms and it affected the earlier formed dome structure, in particular its southern portions. The initial doming was accompanied by the formation of a depression within the great nappe surrounding the dome — probably due to influx of material towards the dome. Compression of these depressions during the Pâkitsoq phase caused antiforms to rise from them. One of them, the Western Antiform, is limited by the Kalot structure, in which the great nappe reappears. The antiformal folding of the central part of the depression makes the frontal part of the great nappe reappear within the Gule Hav structure.

To the south, where the Pâkitsoq phase refolding has left more traces than within the northern part of the dome, the two flanking antiforms have wrapped the outer shell of the dome around their hinge zones. In this way they join up in the conical S-plunging Blindtarmen antiform. Refolding of the southern dome also caused a drag out of the Ørenæs structure parallel to the imposed S-plunging axes.

The dome structure, as it appears now, has a complex origin. The analysis, however, brings out one important feature, i.e. the initial doming with diapiric movements falls more or less in a period when a shift from folding with ENE axes into folding around SSE-plunging axes took place. During such a shift development of diapir structures becomes feasible.

The Pâkitsoq phase of folding was accompanied by the formation of granulite facies minerals, i.e. during this phase of folding the Tovqussaq rocks were exposed to the highest grade of metamorphism reached within the area. This also implies that the preceding phases of folding, the Midterhøj and the Smalledal phases, took place under conditions of progressive metamorphism. These relations will be discussed in more detail in the following chapter.

The Langø sub-phase

The central and southeastern parts of the Western Antiform have been studied in greater detail (see the maps of figs. 11 and 12). Throughout this region evidence was found which indicated that the formation of this antiform took place in two steps. This is brought out by the presence of two sets of folds which show a varying degree of axial disharmony. The later imposed folds were often formed as shear folds, and their formation may have been accompanied by the imprint of a foliation parallel to the slip planes or by the development of recrystallised broader shear zones resembling zones of flow. Several examples of this late sub-phase of the Pâkitsoq phase were mentioned on p. 166. Here another example from western Langø ought also to be cited. Very detailed mapping of the ultrabasic rocks contained within the more or less gneissified western part of the Pas Pyribolite has shown that these rocks were first thrown into plastic folds about 5 to 10 metres broad. During this folding the ultrabasic rocks migrated towards the hinge zones, preferably the antiformal ones. At a later stage, small scale shear folds were imposed on the larger disharmonic folds, see fig. 24 A and B. Discussing the behaviour of the ultrabasic rocks during the Pâkitsoq phase, it may also be recalled that the structures within the ultrabasic rocks pictured in fig. 22 D, were taken as an evidence of two phases of folding (p. 83).

The formation of boudins from the smallfolded calc-silicate layers of northern Langø (fig. 14) could also be pointed to as a result of the youngest sub-phase of the Pâkitsoq folding. Much of the evidence leading to the establishment of this sub-phase being found on Langø, it may appropriately be termed the Langø sub-phase.

The formation of the latekinematic diorites (e.g. the Nordnor diorite and the diorite at Verdens Begyndelse) was contemporaneous with or outlasted the Langø sub-phase.

The intermediate period

As mentioned in the chapter dealing with the chronological significance of the dioritic rocks, the occurrence of dioritic aplites allows one to distinguish between the Pâkitsoq phase and a post-aplitic phase of deformation, the posthumous phase (see below). Between these two phases falls the intermediate period, during which the aplites were formed. The dioritic dykes, which are older than the aplites, may be influenced in places by the last spasms of the Langø sub-phase (see fig. 55). Although the observed material is scarce, the orientation of the aplites seems to be dependent on the axial pattern of the pre-existing structures (see fig. 54). The structures of the aplites (fig. 53) suggest that they were formed in a tensional environment, the dioritisation proceeding along fractures formed subsequent to and controlled by older fold structures.

It is a well known fact from experimental data that the strength of a given rock is much less in a tensional field than in a compressional one. Applying this principle to the Tovqussaq rocks, one realises that changing tectonic environment could alone account for the drastic change in the mode of 'mise en place' of the dioritic rocks developed during various structural stages. In the author's opinion, there is therefore no reason to assume that the intermediate period corresponds to any prolonged time interval accompanied by deep-cutting erosion at the surface.

As to the regional effect of the intermediate period no definite statement can be made at present, but scattered observations suggest that this hiatus in the structural evolution may be traced throughout a large region. North of Tovqussap nunâ, at Alángua, dioritic rocks enable one to recognise the hiatus. These rocks will be described in the forthcoming part III of this series of publications. To the south, i.e. south of Godthaab, the intermediate period was accompanied by the emplacement of basaltic dykelets and sills, which on Qilángârssuit cut a late-kinematic diorite (Berthelsen, 1955). The author also regards the thin grey aplitic dykes with pegmatitic borders described by H. Ramberg (1956, fig. 1 of pl. 9) from the Godthaab region as having been formed by granitisation of original dioritic aplites. In Tovqussap nunâ, the contacts of granitised aplites are often seen to have been altered in a similar way.

The posthumous phase

After the formation of the dioritic aplites, Tovqussap nunâ underwent renewed deformation. Since the movements belonging to this subsequent phase were to a great extent determined by the structures formed during the Pâkitsoq phase of folding, the younger episode will be called the posthumous phase. Some structures developed during this phase have already been described in connection with the discussion of the chronological significance of the dioritic rocks. Here it may be added that the posthumous phase was responsible for the development of the shear-folding previously described from the core rocks of the dome and the penetrative movements leading to the grinding down of the intermediate gneiss layers. Some slip along the contacts between the litho-structural units of the dome may also have occurred during this phase — in response to the strain imposed on the light coloured gneisses due to the accumulative effect of contemporaneous penetrative movements. But otherwise the effect of the posthumous phase on the major structures formed during earlier episodes of folding seems to be negligible.

The posthumous deformation took place under amphibolite facies conditions. Where the earlier formed granulite facies rocks (including the diorites) became influenced by the posthumous deformation, they were retrograde metamorphosed into amphibolite facies rocks. Because of the restricted influence of the posthumous movements, true granulite facies rocks have been preserved in many places and all types of transitions between these and true amphibolite facies rocks are found. The most prominent of these transitional types is the purple gneiss. The retrograde metamorphism of hypersthene-bearing gneiss into purple gneiss was often accompanied by a change from saccharoidal textures into more typical interlobate or even amoeboid textures. This change was not, however, followed by reduction of the grain size. Quartz may show undulate extinction and plagioclase bent twin lamellae, but otherwise penetrative shear movements have exerted only a slight influence in this process of retrograde metamorphism. The purple gneiss metamorphism corresponds therefore more or less to a 'static' recrystallisation, essentially characterised by introduction of water. The term hornfelsing is deliberately avoided since it would imply quite another origin for this textural change.

The close field association and the complete transition between the hypersthene-bearing gneisses and the light coloured gneisses, with the purple gneiss as an intermediate stage, justify the author's disinclination to speak of hornfels textures. The transition from purple gneisses into light coloured gneisses was accompanied by shearing and grinding down of originally larger grains. This is clearly seen in thin section, although the process was generally outlasted by recrystallisation. The provenance of the granitising agents, which caused the present more granitic composition of the light coloured gneisses, will be discussed in chapter IX in connection with a general discussion of the metamorphic evolution.

Towards the decline of the posthumous phase, numerous biotite-bearing quartz-microcline pegmatites were formed. A number of these became influenced by later movement of more or less mylonitic character (see below under the Tovqussaq mylonites). Some lens-shaped pegmatites were not affected by these late movements. As examples of such lenses, the pegmatites east of Gule Hav and those in the escarpment north of Kronehøj can be mentioned. In some of these pegmatites there occur microcline grains which are over 10 cms large, in addition to well developed biotite crystals.

The Tovqussaq mylonites

This name is given here to a group of sheared and mylonitised pegmatites and granitised fault and shear zones which were all formed towards the end of the posthumous phase. They have previously been described by the author as plastomylonites (Berthelsen, 1950). Very similar shear zones from the Godthaab region have been photographed by Mrs. M. L. Ramberg (H. Ramberg, 1952, fig. 67).

According to the author (loc. cit. p. 561), the formation of the Tovqussaq mylonites "is a process later than granitisation, but must have followed immediately upon this latter, being the last chance of a fairly plastic deformation". Later studies have shown that this statement is by and large correct, although the shearing leading to mylonitisation and faulting took place intermittently in relation to the waning granitisa-

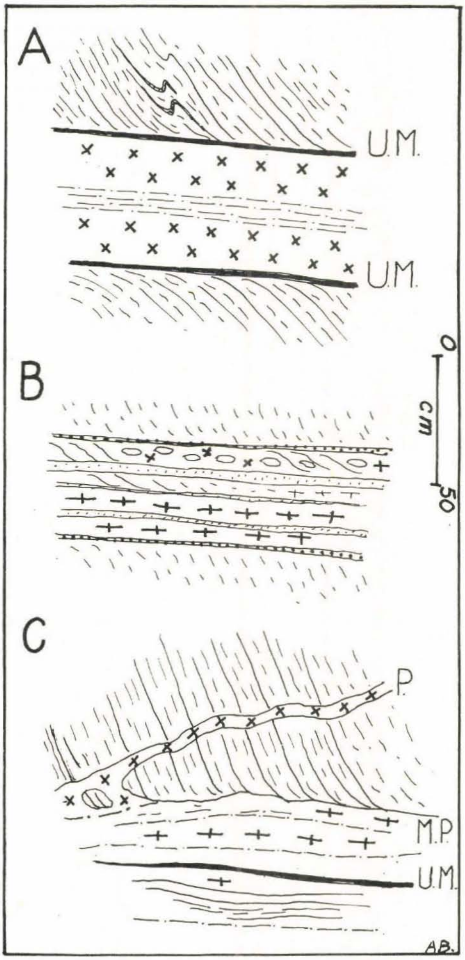

Fig. 79. Tovqussaq mylonites.

tion, so that some movement was paracrystalline while most was of postcrystalline character.

As stated in the first paragraph of this section, the shearing leading to the development of the Tovqussaq mylonites may have given rise to the development of shear zones and mylonites as well as faults. Common to all types (excluding the clean-cut faults) is the plastic drag seen in the structures of the surrounding rocks close to the planes or zones of movement. Retrograde alterations, most commonly marked by a pronounced bleaching due to discolouration of the feldspar, are also prominent close to the Tovqussaq mylonites. The width of the mylonites varies from a few centimetres to several metres. Where not developed from original pegmatites, they are often accompanied by granitisation. This may even be the case along the clean-cut faults. Where a southward-dipping fault cuts the Pâkitsoq Pyribolite east of Snirkelsø, the pyribolite of the hanging wall has been strongly granitised with the resultant formation of a replacive agmatitic breccia. Where paracrystalline movements have taken place within the mylonites fine grained dyke-like blastomylonites were formed. They are very similar to the recrystallised mylonites described by N. EDELMAN (1949, p. 33) from migmatitic gneisses in SW Finland. Like the Finnish occurrences, the Tovqussaq blastomylonites resemble acid dykes.

Fig. 79 A, B, and C, shows some typical examples of mylonite-pegmatites. In all three sketches a drag may be seen close to the mylonite. Within the zone of rupture, flaser structures, drawn-out augen, slickensides, foliated pegmatite (MP), fine grained crush zones (dotted), and ultramylonitic to pseudotachylitic seams (UM), are developed. Fig. 79 C shows an example where the postcrystalline nature of the mylonitisation can be ascertained, because a branch of the original pegmatite has been left unaffected by the shearing.

In addition to the larger mylonites and faults shown on the map of Pl. 2, about fifty Tovqussaq mylonites were studied in the field. No attempt to carry out systematic measurements of the striae on the slickensides was made since the composite origin of these fractures was realised at an early stage of the field work. The observations were restricted to strike and dip measurements and determination of the amount and direction of the displacement along the mylonites with the aid of the drag.

The equal area projection of fig. 80 shows a stereogram (lower hemisphere) where the normals to the measured mylonite-pegmatites have been plotted as black dots. The small arrows attached to some of the dots indicate the direction of the strike-slip movements of the hanging block (in relation to the foot block). This stereogram shows that the mylonites can be grouped into three sets according to their orientation. The 'average' mylonite planes of each of these sets are also indicated on the diagram. The NE-striking set of mylonites (NE-M) dips to SE, while the set striking more or less WNW (WNW-M) shows NNE dips. The ENE-striking set (ENE-M) dips steeply to NNW.

Along the NE-striking mylonites the direction of strike-slip movements (of the hanging block) seems always to be towards SW, while the strike-slip movements observed along the WNW- and ENE-striking planes are conflicting. Using ANDERSON's classification (ANDERSON, 1951), the NE- and WNW-striking mylonites may be said to form a conjugate set of shear planes, the acute angle between which is divided by the direction of the principal stress, i.e. the mylonites indicate in their varying orientation a transition from normal faulting towards wrench faulting. This interpretation of the fracture system would mean that the hanging blocks of the WNW-striking mylonites should show a WNW direction of strike-slip movement. Movements with this direction of

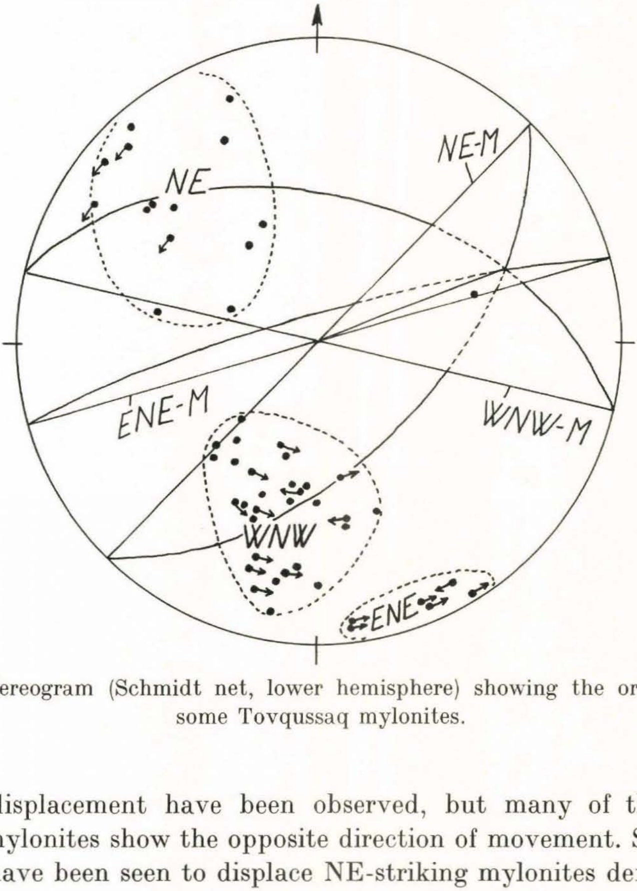

Fig. 80. Stereogram (Schmidt net, lower hemisphere) showing the orientation of some Tovqussaq mylonites.

relative displacement have been observed, but many of the WNW-striking mylonites show the opposite direction of movement. Since some of these have been seen to displace NE-striking mylonites dextrally the author assumes that the conflicting observations on the relative strike-slip movements along the WNW mylonites are caused by a preferential reworking of some of these latter in connection with younger wrench faulting.

This reworking most probably took place in connection with the formation of a regional system of wrench faults, the Fiskefjord faults, which displaced earlier intruded post-orogenic basic dykes dextrally. These cratogenic (i.e. post-orogenic) events have been described in a recent paper by Berthelsen and Bridgwater (1960) and therefore need not be discussed here.

Although reluctant to speculate on the orientation of the stresses responsible for the structures analysed, the author here offers some comment on the possible stresses involved in the formation of the Tovqussaq mylonites.

The direction of principal stress which can be postulated from the orientation and kinematics of the conjugate shear formed by the Tovqussaq mylonites does not seem to be consistent with the orientation and kinematics of the shear folds formed during an earlier stage of the posthumous phase. The compressed shear folds ('divergentschiefrige Falte') on Langø point towards a more or less horizontal directed compression, while the conjugate system of mylonite planes suggests a different orientation of the direction of principal stress. This possible change in the stress field towards the end of the posthumous phase may be due to the fact that the Tovqussaq rocks became more and more brittle.

The Tovqussaq mylonites can be viewed in consequence as being the last observable traces of an orogenic evolution — more or less representing the transitional stage before purely cratogenic conditions obtained.

Summary

This concludes the review of the kinematic evolution of Tovqussap nunâ. At this point it should be emphasized that the kinematic picture arrived at represents the most advanced stage of the structural analysis and that as such it may be affected by possible misinterpretations during an early stage of this analysis.

The content of this chapter is summarized in the chronological scheme of table I. This table also includes the chronological data for the cratogenic evolution (Berthelsen and Bridgwater, 1960).

This entire evolution almost certainly belongs to the pre-Cambrian, since the youngest post-orogenic dykes as far as is known are involved in the Nagssugtôqidian orogeny north of Søndre Strømfjord (H. Ramberg, 1948; Berthelsen, 1957; Berthelsen and Bridgwater, 1960).

It is very probable that the orogenic evolution of the Tovqussaq area took place during the Ketilidian orogeny which formed the ancient basement rocks of SW Greenland (Wegmann, 1938). This point of view has previously been put forward by Ramberg (1948) and Noe-Nygaard (1952). Later studies in the intervening regions and absolute age determination (Moorbath, Webster and Morgan, 1960) seem to support this correlation (Berthelsen, 1960b).

Concluding this chapter, the author wants to point out that Tovqussap nunâ can stand as a type area for a region characterised by 'tectonique superposée' in the sense of Wegmann (1953) and Oulianoff (1953).

Table I.
Chronology of Tovqussap nunâ.

Division	Structural characteristics of period or phase
Younger group of basic dykes	Dyking along E-W, NE and NW directions
Fiskefjord faulting	Dextral movements along NE to ENE-trending wrench faults
Older group of basic dykes	Dyking along N-S and ENE-trending lines
Tovqussaq mylonites	Shearing, mylonitisation and faulting (para- to postcrystalline)
Posthumous phase	Shearfolding and largely paracrystalline penetrative movements
Intermediate phase	Dioritisation along tension-conditioned fractures
Langø sub-phase	Repeated deformation, latekinematic dioritisation
Pâkitsoq phase	Open to closed or squeezed folds generally with SE to S-plunging axes. Synkinematic dioritisation
Doming	Diapiric movements in the 'dome'
Smalledal phase	Recumbent folding (ENE axes) with partial refolding of first-formed structures as in the great nappe
Midterhøj phase	Development of isoclines with NW or N-S trending axes
Supracrustal period	Sedimentation and extrusion (? intrusion)

IX. PETROLOGICAL AND PETROGENETIC CONSIDERATIONS

The progressive metamorphism and its relation to the kinematic evolution

Although the structural analysis has revealed that the Tovqussaq rocks passed through a prolonged and complex structural evolution corresponding to various stages of progressive metamorphism, no relic mineral parageneses which, with any safety, can be attributed to these stages, have been found during the petrographic study. It seems as if the recrystallisation accompanying the Pâkitsoq phase of folding has obliterated all such traces with the overall production of granulite facies minerals.

As to the nature of the metamorphism corresponding to the Midterhøj and the Smalledal phases, one can only guess, although the style of the structures developed during the two phases may give a hint to the metamorphic conditions during these phases. Structures similar to those developed during the Midterhøj phase are known from many low- to nearly non-metamorphic terrains, and thus the Tovqussaq isoclines may have been formed under low-metamorphic conditions.

The recumbent structures belonging to the Smalledal phase show quite a different style. During this phase the rocks yielded to deformation in a much more plastic manner, otherwise structures of nappe dimensions with overturned flanks intact could hardly have been formed. The present pyribolite layers became somewhat thickened around the hinge zones (see fig. 59) and were interfolded with the surrounding gneisses (frontal part of the great nappe, see fig. 73). These relations point to a higher temperature and pressure during the Smalledal phase. Analogous structures are described from many amphibolite facies regions, for example the Sokumvatn area, N. Norway (Rutland, 1960). Recumbent folds or nappe structures have also been described by Demay (1942), Michot (1951, 1957b) and Haller (1955) from migmatitic catazonal areas. A valuable review of the fold movements in the various structural zones of the basement has recently been published by Kranck (1957).

Relying on the analogies mentioned, the author considers the Smalledal phase as having taken place during a stage of progressive metamorphism corresponding to amphibolite facies, the mobile stage of the structural evolution — in Tovqussaq accompanied by migmatisation. This concept would also explain the onset of the doming with diapiric movements which outlasted this phase of folding.

The metamorphic evolution outlined here naturally culminates with the granulite facies metamorphism of the Pâkitsoq phase. This means that the Midterhøj, the Smalledal and the Pâkitsoq phases most probably belong to one and the same orogenic cycle. The structures which were developed during the different phases, and which were superimposed on each other, represent the successive imprints of the different structural levels (Stockwerke) through which the Tovqussaq rocks passed on their way 'down' into the granulite facies régime. (The same effect would naturally result from the ascent of the PT curves). As mentioned in the foregoing chapter, the posthumous phase, including its final stage with the formation of the Tovqussaq mylonites, may reflect the structural events during the start of the 'rise' of the region (or the retreat of the PT curves).

Mineral facies of the Tovqussaq rocks

The discussion of the mineral facies relations of the Tovqussaq rocks can be restricted to a treatment of the granulite facies parageneses and the changes imposed on these due to the retrograde metamorphism attached to the posthumous phase.

The quartz-bearing rocks will be considered first. They comprise the gneisses, the granulite s. s. and various schists.

The granulite facies gneisses

Most of the granulite facies gneisses are bi-pyroxenic, i.e. they carry rhombic as well as monoclinic pyroxene. 'Intergrowths' of hypersthene and diopside resembling exsolution textures have been noticed and diopside often forms a more or less complete rim around hypersthene. The hypersthene-bearing gneisses contain only small amounts of potash feldspar, usually in the form of antiperthite. Sphene and muscovite are totally absent. These gneisses are usually quartz-dioritic although some of them may approach enderbite in composition. Garnet forms a rare constituent of some of the hypersthene-bearing gneisses of the Frame Layer rocks.

In places, the granulite facies gneisses can be traced into purple gneisses along the strike direction. In other places, hypersthene-bearing and purple gneisses are interbanded. In both cases, the purple gneisses seem to have been formed by retrograde metamorphism from originally pyroxene-bearing rocks. Chronologically, this retrograde alteration dates from the posthumous phase. The term, the purple gneiss metamorphism, is used here to describe this retrograde process.

The purple gneiss metamorphism

The effect of the purple gneiss metamorphism can be illustrated by the following pair of reactions (RAMBERG, 1952).

$$\begin{array}{lll} \text{high temp.} & \text{low temp.} & \\ \text{hypersthene + anorthite + water} \leftrightarrows & \text{hornblende + quartz} & (1) \\ \text{hypersthene + potash feldspar + water} \leftrightarrows & \text{biotite} & (2) \end{array}$$

Reaction (1) has taken place preferably where there was an insufficient amount of potash feldspar to allow the formation of biotite. It was accompanied by a decrease in the An content of the plagioclase. In some samples, reaction (2) seems to have followed immediately on the uralitisation of the pyroxene, leading to replacement of the newly formed hornblende by biotite. The biotite formed in this way may show parallel intergrowth with the quartz originating from the preceding uralitisation. Where reaction (2) succeeded reaction (1), an increase in the supply of potash feldspar (potassium) can be inferred.

If reaction (2) is taken as indicative of the limit between granulite and amphibolite facies, the purple gneisses take an intermediate position between these facies. If not, which is more probable, they represent rocks incompletely adjusted to amphibolite facies, their relics of hypersthene only having been preserved because of the dipsenic nature of the retrograde metamorphism. The term 'dipsenic' is used here in the same sense as by I. ROSENQVIST (1952), who introduced it to describe water-deficient metamorphism (dipsenic means thirsty).

The dipsenic nature assumed for the purple gneiss metamorphism would match the rather 'static' character of this. The light coloured gneiss metamorphism could then be taken as representing the end point of the retrograde metamorphism where strong penetrative movements allowed the access of sufficient water for reaction (2) to be completed.

The application of ROSENQVIST's principle of dipsenic metamorphism to the purple gneisses has the advantage that using it the restricted field occurrence of these rocks may be explained. Except for the purple gneisses known from the districts surrounding Tovqussap nunâ and for some very similar rocks on the SE coast of Greenland (personal communications from A. NOE-NYGAARD and H. SØRENSEN, who have studied some rock samples collected by R. BØGVAD in this region), rocks comparable to the purple gneisses described in this paper seem scarce. The author has searched the available literature on granulite-amphibolite facies rocks and on charnockitic rocks without finding descriptions of similar rocks. Since the colour and mineralogy of these rocks are so distinctive, they could hardly escape notice.

Two features of the purple gneiss metamorphism still have to be mentioned. The first concerns the potash feldspar. Where hypersthene-bearing gneisses have been changed into purple gneisses, the plagioclase usually becomes more antiperthitic and interstitial potash feldspar appears. Most of the potash feldspar grains show irregular extinction or microcline grid twinning. From his study of some granulite and amphibolite facies gneisses from SW Norway, K. Heier (1957) suggests that the inversion from monoclinic to triclinic potash feldspar takes place at about 500°, i.e. at slightly lower temperatures than the transition from granulite into amphibolite facies. Hence monoclinic "orthoclase" may be found in high grade amphibolite facies rocks. In Tovqussaq untwinned potash feldspar is rare. The purple gneiss metamorphism seems to have been accompanied by a 'triclinisation' of the potash feldspar — in just the same way as the 'light coloured gneiss metamorphism' (see below), which also produced grid twinned microcline. If the inversion from monoclinic to triclinic feldspar can be relied on as a thermometer, the presence of triclinic potash feldspar in the purple rocks would confirm the assumed isophysical nature of the dipsenic purple gneiss metamorphism and the 'wet' light coloured gneiss metamorphism.

The second feature to be mentioned may or may not be directly connected with the purple gneiss metamorphism. It is the occurrence of diopsidic rims on hypersthene grains. Similar textures have been described by P. Quensel (1951, p. 253, fig. 12) from some intermediate charnockites from Varberg, SW Sweden. K. Parras (1958, p. 100) has described diopsidic exsolution lamellae in hypersthene from charnockites in SW Finland. He assumes that an original calcic hypersthene became unmixed with falling temperature (opt. cit. p. 114). The lamellar to more irregular 'intergrowths' between diopside and hypersthene in some of the Tovqussaq hypersthenes could be interpreted in the same way, and the diopsidic rims could then represent a slightly more advanced stage in this reconstitution due to the onset of the retrograde metamorphism. The external hornblende fringes sometimes found on the diopside rims may support this assumption.

The light coloured gneiss metamorphism

This expression is used here to designate the more advanced stages of the retrograde metamorphism which caused the formation of the present mineral assemblage of the light coloured gneisses. Pre- to paracrystalline penetrative movements were attached to this phase.

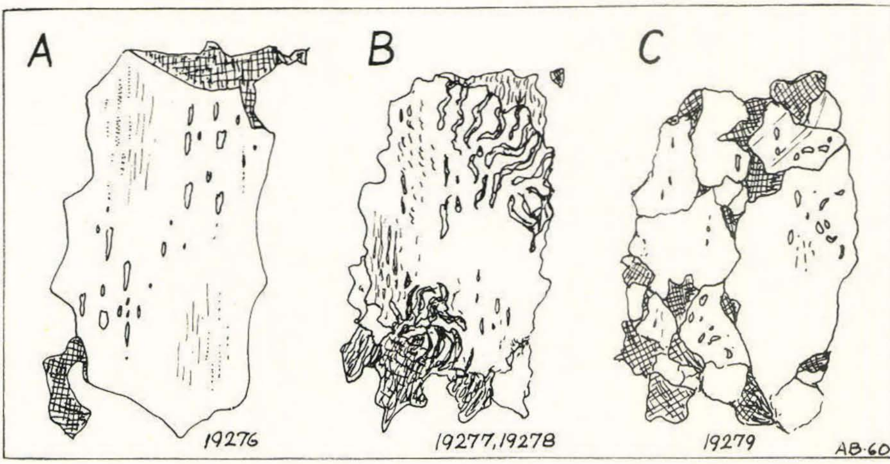

Fig. 81. Schematical sketches of antiperthitic feldspar changing into mesoperthitic to diablastic feldspar before complete rearrangement into two separate phases. Microcline is dark shaded.

The light coloured gneisses are generally more fine grained than the pyroxene-bearing and the purple gneisses. They are granodioritic to granitic in composition and carry feldspar, quartz, greenish biotite and accessory apatite and ore. Where the metamorphic reactions are most advanced, the feldspar forms two separated phases (oligoclase and microcline), but in the transitional stages which connect the light coloured gneisses to the purple rocks, diablastic intergrowths and mesoperthitic textures are common (fig. 81, for a fuller description see page 30). Where the recrystallisation is advanced, the quartz is often packed with tiny grains of rutile, zircon and ?tourmaline in addition to liquid (or gas) inclusions. Sphene has only been found in two specimens. When present, epidote is associated with the mafic minerals and its formation cannot be regarded as being due to saussuritisation of the plagioclase. The light coloured gneisses are thus simply amphibolite facies rocks.

Although no relics of former pyroxenes have been found in the light coloured gneisses, it seems reasonable to assume that, before the onset of the retrograde metamorphism, they carried granulite facies minerals. This may also be suggested by the field relations and by their rather high content of Ba, which exceeds that normally found in granitic rocks (Goldschmidt, 1954, p. 253), see table II. The granulite facies equivalent of a light coloured gneiss would be a charnockite s. s.

The occurrence of small zircon individuals in the quartz of the light coloured gneisses poses a separate problem. Zircon is usually regarded as one of the most resistant survivors to regional metamorphism, but in the present case it is quite clear that the metamorphism has lead to the formation of new zircon individuals.

The inclusion-rich quartz of the light coloured gneisses reminds one of the quartz found in many granites believed to be of magmatic origin (see for example JOHANNSEN, 1958, Vol. II, pp. 131—137). The occurrence, in several of the light coloured gneisses, of quartz cataclasts of an older generation side by side with recrystallised inclusion-rich individuals however, rules out the possibility of an anatectic origin of the latter. The typical blasto-cataclastic textures shown by much of the inclusion-rich quartz also suggest the total absence of any anatectic liquid phase.

The conclusion must therefore be drawn that, as stated above, the amphibolite facies metamorphism was capable of causing the development of a new zircon generation. In this process, the introduction of water, fluorine and boron (considered necessary for the formation of tourmaline) may have played an important role.

The petrographic study of the pyroxene-bearing and the purple quartz-dioritic rocks showed that these are very poor in zircon, their constant accessories being apatite and ore. The results of some spectrographic analyses stand in striking contrast to these observations. The quartz-dioritic rocks show a Zr content of 200 to 300 ppm, while all analysed light coloured rock types only contain 80 ppm Zr, i.e. rocks which contain microscopically visible zircon contain less Zirconium than rocks seemingly devoid of zircon.

These contrasting results of the petrographic and the spectrographic studies possibly indicate that during very high grade metamorphism Zirconium does not enter into the usual Zr-mineral (zircon), but becomes 'absorbed' in one form or another in one of the common rock-forming minerals. For this idea the author is indebted to mag. scient. H. MICHEELSEN, with whom he has discussed the problem. Since Mr. MICHEELSEN has kindly agreed to carry out a thorough investigation of this problem, it is left aside here.

In this connection it may also be mentioned that HOWIE (1955, p. 743) reports "considerable" amounts of Zirconium in quartz from charnockite and garnitiferous leptynite and 400 ppm Zr in plagioclase from charnockite from Madras, India.

The author would like, however, to draw attention to the distribution of Zr within the different types of rocks. From table II it can be seen that the Zr content decreases the more granitic the rocks become in composition. The lowest content (10 ppm Zr) is found in the granulitic rocks of Langø. This distribution seems incompatible with a magmatic origin for these rocks, since, if magmatic, the granitic end members would be expected to be enriched in Zr. If, on the other hand, a sedimentary origin is assumed, the distribution of this particular trace element could be explained better. The quartz-dioritic rocks which may have been derived from greywackes (resistates in a geochemical sense) show

Table II.

Trace elements of some acid rocks from Tovqussap nunâ.
(Analyst Mr. IB SØRENSEN, cand. polyt. et lic. techn.)

(ppm)

Rocktype	Specimen	Ba	Sr	Ti	Zr	Co	Ni	Cr	Mo	V	Cu
hyp.-gneiss	4029 C	100	80	2000	200	tr	tr	5	÷	50	10
hyp.-gneiss	19246	80	200	3000	300	20	10	5	÷	30	1000
purple gneiss	19276	200	100	1000	200	÷	tr	tr	÷	10	20
light purple gn.*)...	4094	500	500	30	80	÷	tr	÷	÷	÷	30
light col. gneiss.....	4041	500	500	800	80	÷	50	tr	÷	tr	20
light col. gneiss.....	19302	800	800	500	80	÷	tr	÷	÷	tr	5
granulite s.s........	4037	300	100	100	10	÷	tr	tr	÷	÷	5
granulite s.s........	19264	800	200	10	10	÷	tr	÷	÷	÷	5

*) This sample does not contain any mafic minerals.

NB! The results shown in this table and tables III and IV are based on semi-quantitative analyses (visual comparison with a standard plate prepared from synthetic standards). The accuracy is $\begin{matrix}+100\\-50\end{matrix}$ %.

The sensitivity is as follows:

Ba: 10, Sr: 5, Ti: 10, Zr: 10, Co: 10, Ni: 5, Cr: 1, Mo: 5, V: 10, and Cu: 1 (all in ppm).

a relatively high Zr content (200 to 300 ppm), while the most Al-rich of the granitic rocks, which probably represent pelitic sediments (hydrolyzates), only contain very little Zr. Similar distribution trends due to varying conditions of sedimentation have been described by SAHAMA (1945).

It should, however, be added that the biotite schists which have been studied under the microscope all contain zircon as a common accessory. If these schists are genetically related to the granulites (which actually is suggested by their mutual relations in the field), the present distribution of Zr cannot be used as an evidence of a sedimentary origin of the gneissic and granulitic rocks. From table II it may also be seen that the Ti content decreases simultaneously with Zr. Normally, a much higher Ti content is found in hydrolyzate sediments (SAHAMA, 1945).

The position of the Tovqussaq granulites in the facies classification

Before proceeding with this topic, it should be stated clearly that the author regards the granulites at Tovqussaq as amphibolite facies rocks formed during the posthumous phase of deformation. In arriving at this conclusion the author uses the reaction (for the quartz-bearing rocks)

$$\text{hypersthene} + \text{potash feldspar} + \text{water} \leftrightarrows \text{biotite}$$

to draw the boundary between granulite and amphibolite facies. This means that the purple gneisses are dipsenic amphibolite facies rocks and the light coloured gneisses true amphibolite facies rocks. The granulites are amphibolite facies rocks because they recrystallised under amphibolite facies conditions but nevertheless carry some minerals thought to be typical of granulite facies. The occurrence of triclinic (grid twinned) potash feldspar in all the three mentioned rock types would, according to HEIER's results (opt. cit.) classify them as amphibolite facies rocks. Why then call the garnet bearing gneisses granulites?

First of all it should be admitted that the author has been unable to investigate the composition of the garnets found in the rocks termed granulites by him. It might have been more prudent to describe them as garnet-bearing granitic gneisses. But the occurrence of Plättung textures within these rocks, a texture which is so typical of the granulite sensu stricto, made him risk applying the term granulite — irrespective of the difficulties arising with the system of facies classification.

From the occurrences of retrograde altered diorite aplites within the granulite at Langø, it may be inferred that the granulites fall within the same isograde as the light coloured gneisses, to which they also show a great resemblance in their feldspars. In the field the Plättung texture characteristic of granulite can also be seen to be related to the compressional shearing of the diorite aplites. But how then could the paragenesis garnet + potash feldspar survive in the presence of water without having been completely changed into biotite + sillimanite?

RAMBERG (1952, p. 158) assumes that the following two reactions take place abruptly around the border line between granulite and amphibolite facies, but he mentions also that increasing Mg/Fe ratio lowers the equilibrium temperature of reaction (3) relative to (4).

low temp. high temp.

biotite + sillimanite $\rightleftarrows$ garnet + potash feldspar + water (3)

biotite $\rightleftarrows$ hypersthene + potash feldspar + water (4)

Of these two reactions, the first (3) can be said to illustrate the amphibolite and granulite facies parageneses of the present garnet-bearing rocks, while the second exemplifies the retrograde alteration of an original charnockitic rock into a light coloured gneiss.

In Tovqussap nunâ, the field relations suggest that during the posthumous phase, the 'granulite facies side' of reaction (3) was in stable coexistence with the 'amphibolite facies side' of reaction (4). These conditions might well have been fulfilled if the granulite had a higher Mg/Fe ratio than the light coloured gneisses. Such a compositional difference cannot be proved directly due to the lack of chemical analyses, but it may be inferred indirectly. On Langø, the granulite encloses boudins of a schistose rock which, due to its content of cordierite and cummingtonite, can be considered as a biotite-bearing pyriclasite which has been exposed to a magnesium metasomatism. Whether the introduction of Mg into these boudins was due to a primary excess of this element in the surrounding rocks, or whether it should be seen as the consequence of a larger scale Mg metasomatism is hard to tell. Possibly the inferred enrichment in Mg within the 2nd Intermediate Layer on Langø and Tugdlerúnarssuit could be explained as a result of the strong dioritisation to which the pyribolite bands within and around this unit were subjected during the Pâkitsoq phase (see below under the diorites).

Finally it should be mentioned that lowering of the equilibrium temperature of reaction (3) just stabilised the right hand side of the reaction. Biotite rims on the garnets and the occasional occurrence of sillimanite and muscovite within the most sheared rocks show that small changes in physical or chemical environment could drive the reaction to the left.

This discussion of granulite versus amphibolite facies may appear academic to some readers. It has nevertheless been dealt with at some length because, in the present case, a combination of the structural and petrographic results have made it possible to apply the mineral facies principle to rocks which at first sight appeared to be in conflict with this system of classification. Imperfect adjustment of a rock to a new facies may have many causes — lack of time, lack of water or a particular chemical situation which stabilises earlier formed minerals. All these factors must be considered if a reliable concept of the metamorphic evolution is to be arrived at.

The effect of the posthumous deformation and the retrograde metamorphism on the schists

The cummingtonite-cordierite-bearing schists referred to above seem to represent the sheared and retrograde metamorphosed equivalents of the biotite-pyriclasite schists found in the purple and in the hypersthene-bearing gneisses. Due to their original high content of diopside these schists may be regarded as highly metamorphosed marly sediments.

The biotite schists have clearly recorded the posthumous deformation. In the quartz-free bands, the mafics are biotite and garnet. The biotite shows bent and contorted lepidoblasts, while the more competent garnet grains are much less influenced by the deformation. In places, the impression may even be obtained that the garnet has grown syn- to posttectonically. The fact that biotite has been welded in between the pieces of broken garnet grains and has grown as small idioblasts within these shows, however, that the garnet in reality is pretectonic in its development.

In the quartzo-feldspathic layers of the schists, hypersthene is formed instead of garnet. Here the effect of the posthumous deformation may also be traced. The hypersthene porphyroblasts have been broken and partly replaced by biotite. In these rocks zircon is a common accessory both in the quartz-free and the quartz-bearing bands. It may be contained in biotite as well as in hypersthene or quartz, as fairly large and rounded grains.

The formation of the diorites in relation to the mineral facies classification

Among the various other rock types the diorites take a specific position, since they were formed syn- to postkinematically in relation to the Pâkitsoq phase. Even if formed at the expense of older rocks, they are new rocks which only participated in the last part of the structural evolution. The latekinematic diorites will be discussed first, because the process of their formation is understood best. They were all formed by metasomatic replacement of pyribolite layers where the latter occur in antiformal hinge zones. In the Western Antiform, pyribolite has thus been transformed into diorite in the arrowhead structure at Gule Hav, at Verdens Begyndelse, around Nordnor and at Qagssiarssuk. The formation of these diorites can be attached to the Langø sub-phase.

The conversion of pyribolite into diorite must depend on the introduction of salic material, and eventually a removal of femic material in order to maintain the isospacial nature of the replacement process. The material introduced would correspond to the composition of the assumed dispersed phase in the surrounding gneisses (Si, Al, Na and some K). The expression 'dispersed phase' is used here to designate the material capable of migration irrespective of whether it migrates by diffusion or by liquid flow as an anatectic melt (ichor). In the author's opinion, there is no evidence which favours the second alternative (see also p. 197).

The transformation of pyribolite into diorite can be illustrated in a simplified manner by the following reaction, where introduced quartz reacts with the hornblende of the pyribolite. This reaction is well known for granulite facies, (Parras, 1959, p. 108).

$$\text{Hornblende} + \text{quartz} \rightleftarrows \text{plagioclase} + \begin{cases} \text{diopside} \\ \text{hypersthene} \\ \text{biotite} \end{cases} \quad (5)$$

(in pyribolite) (introduced) → (in diorite)

This reaction would explain the higher content of light coloured minerals in the diorite and the occasional coexistence of hypersthene and biotite in these rocks. (The greater part of the biotite of the diorites is still being considered as having been formed during retrograde metamorphism).

The onset of the dioritisation during the Langø sub-phase in certain structurally controlled foci, i.e. the antiformal hinge zones, seems to have depended on the particular style of folding during this sub-phase. Development of foliation planes and zones of intense movement oblique to the lithological layering caused not only conditions more suitable for long range migration, but enabled the dispersed phase driven out from the gneissic cores of the antiforms to migrate into the overlying pyribolite and react with this.

The structural position of the synkinematic diorites suggests that an early formation of diorite took place due to metasomatic reactions along the locally sheared flanks of the major structures. (Pyribolitic and ultrabasic inclusions in the synkinematic diorites do not favour the idea of an original dioritic composition of these layers).

Since the dioritic rocks thus formed are definitely more leucocratic than the original pyribolite, some mafic material may have been removed during the dioritisation. This means that there should have been an opportunity for the contemporaneous formation of more basic or even ultrabasic rocks. The hypersthenitic rocks at Verdens Begyndelse (p. 121) could be viewed as the result of a basic front accompanying the dioritisation nearby.

The origin of the postkinematic dioritic dykes and aplites presents a separate problem. These rocks were formed in a tensional environment by replacement along the walls of fissures or by migration of material into the fissures.

Assuming that the dispersed phase present in the Tovqussaq rocks could migrate towards low pressure regions, i.e. the opening cracks and fissures, the material introduced into these or gathering around them would probably have a composition which not only was controlled by the diffusion coefficient of the elements in question, but also was influenced by the regional bulk composition of the surrounding rocks. Assuming that the low pressure conditions in the fissures were maintained over a sufficiently long period (due to constant opening) even the slowest migrating elements may have become members of the dispersed phase.

The relative abundance of basic rocks in Tovqussap nunâ might in this way have influenced the composition of the dispersed phase and may have caused its dioritic composition. If the system had been closed, more granodioritic or quartz-dioritic fissure fillings would be expected in the gneisses.

The field relations at Dioritnæs suggest that the postkinematic dioritisation proceeded very similarly to the latekinematic one where it involved pyribolitic or amphibolitic rocks. Ultrabasic segregation rocks were formed from the excess mafics and small scale basic fronts may even have been responsible for the development of the irregular banding found within the dyke.

The author (Berthelsen, 1957) has previously pointed out that the exothermal nature of the dioritisation may be the reason why it outlasted the deformation (i. e. the Langø sub-phase). Although still unable to offer any physico-chemical evidence in support of this view, he feels convinced that it contains the clue to this particular problem.

Mineral facies relations of the basic rocks

The pyribolites

The pyribolites all seem to be characterised by a rather high Mg/Fe ratio. This may be seen from the only analysis published so far (H. Sørensen, 1953, p. 45), and may also be inferred from the diopsidic composition of the clinopyroxene, the hypersthenitic composition of the orthopyroxene as well as from the common occurrence of Mg-hastingsitic hornblende in these rocks. At first sight the retrograde metamorphism does not seem to have left any visible traces in the pyribolitic rocks. Their paragenesis, plagioclase, ortho- and clinopyroxene, hornblende and accessory apatite and ore, points towards recrystallisation under granulite facies conditions. The field relations show also that the thick pyribolite layers remained nearly uninfluenced by the deformation of the posthumous phase. Due to the absence of penetrative movements, which could make possible the introduction of water into these rocks, the critical granulite facies mineral, hypersthene, has survived.

That the original granulite facies pyribolite has nevertheless suffered from a dipsenic metamorphism of retrograde character, is suggested by the peculiar variations now found in the optic constants of the hornblendes of this rock. The hornblende lattice, which is capable of accomodating many different elements, has, so to speak, absorbed all instabilities. One could therefore speak of a 'cryptic' retrograde metamorphism. Occasional thin hornblende rims to the pyroxenes are further evidence of a retrograde trend in the metamorphism.

Due to the weak influence of the retrograde metamorphism it has been possible to trace some textures developed during granulite facies. In some samples, the hypersthene can be seen to be in a state of growth. This feature was noticed by H. Sørensen (1953).

The only garnet-bearing amphibolite studied (13418) does not carry diopside and its plagioclase is strongly zoned. H. Sørensen, who also studied this sample, considers the garnet and hornblende to be secondary to hypersthene. To this the author can only add that the rock has been affected by the posthumous deformation and that the biotite (formed at the expense of hypersthene) was influenced by this deformation. A partial recrystallisation of the hornblende may also have taken place.

The gabbro-anorthosites

From a mineral facies point of view, the gabbro-anorthosites take a rather dubious position since the association calcic plagioclase and diopside can be considered as stable in granulite as well as amphibolite facies. Only textural observations on the relative age of the different minerals allow one to distinguish between a primary granulite facies paragenesis and later modifications imposed on this during the post-humous phase. Calcic plagioclase, diopside, Mg-hastingsitic amphibole and apatite are regarded as belonging to the original granulite facies association. Saccharoidal grains of scapolite possibly also formed part of this paragenesis.

The following features are attributed to the retrograde metamorphism: 1. The possible unmixing of the plagioclase (see p. 46), 2. Uralitisation of diopside, 3. Formation of biotite, and 4. Scapolitisation of the plagioclase. The last three features necessitate a slight metasomatic introduction of water and potassium. Such a metasomatism is known to have taken place in the surrounding gneisses during the retrograde metamorphism. Where gabbro-anorthosites have not been exposed to metasomatic alteration, their facies relations are difficult to understand.

In a few samples an incipient saussuritisation of the plagioclase may be noticed. This feature possibly is related to retrograde alteration caused by late faulting.

Disregarding the slight retrograde or metasomatic effects, it is remarkable how the original granulite facies paragenesis is maintained within the gabbro-anorthositic rocks in spite of their varying mode of occurrence. Hypersthene has never been found in these rocks, nor has calcite. This constancy in mineral composition (if not in the relative proportions of feldspar and mafics) indicates that the chemical composition of the gabbro-anorthosites varies within rather fixed limits. They are rich in Ca and show a rather high Mg/Fe ratio. Compared to the pyribolites, they are definitely more rich in Al and poorer in total iron.

The calc-silicate rocks

Regarding their parageneses, the calc-silicate rocks show a much greater variation than the gabbro-anorthosites. They carry a very similar calcic plagioclase, but their diopside is more hedenbergitic and generally seems to be Cr-bearing (bluish green colour and high Cr content in the rocks). Quartz, calcite and an orange brown coloured garnet are commonly found in the calc-silicates while they are scarce or absent in the gabbro-anorthosites. The occasional occurrence of sphene in the calc-silicates may be explained by the Ca-rich nature of these rocks, which feature may make sphene a stable mineral in granulite facies (RAMBERG, 1952). The peculiar inclusion-rich type of scapolite so common in the calc-silicate rocks has never been observed in the gabbro-anorthosites. From field observations and from what is seen under the microscope, two types of calc-silicate rocks are present in Tovqussap nunâ. The one occurs as banded layers in the granulite of Langø, the other as enclaves in different gneisses. In the latter type quartz and garnet are more abundant and diablastic reaction textures are strongly developed. In spite of these differences, the two types have such specific minerals as the bluish green diopside and the inclusion-rich scapolite in common.

In the author's opinion the occurrence of two types of calc-silicate reflects original lithological differences and/or variations due to differences in the metamorphic development within rocks in different structural positions. Both types are, however, considered to have a common origin, i.e. they have been derived from more or less impure calcareous sediments through varying metamorphic and metasomatic processes. Lacking analyses of these rocks and of their minerals, the author refrains from any further discussion of their metamorphic or metasomatic evolution. The descriptions given on an earlier page should suffice to show that they actually are high grade calc-silicate or skarn rocks.

On the distribution of some trace elements in the basic rocks and the calc-silicate rocks

Regarding their content of trace elements, the gabbro-anorthosites, the pyribolites and the calc-silicate rocks show great similarities. Table III contains the results of several spectrographic analyses of these three rock types as well as those obtained from some ultrabasic rocks and a gabbro-anorthosite of magmatic origin (xenoliths in basic dykes of Gardar age from Kobberminebugt SW Greenland). In table IV, the results for the ultrabasics, the calc-silicates, the pyribolites and the Tovqussaq gabbro-anorthosites are summarized by calculating the average values. (These averages are reduced to the nearest round figure).

From these two tables it may be seen that the content of Co, Ni, Cr and V is rather constant within each group and that it increases when going from the gabbro-anorthosites through the pyribolites to the calc-silicate rocks. As might be expected, the ultrabasic rocks are rather different in their relative content of these elements. They are enriched in Ni relative to Cr.

Comparing the four analyses of table III, it becomes evident that in the ultrabasic rocks, the Ni content is related to that of olivine. When olivine gives way to pyroxene, Cr increases relative to Ni.

The increase in Co, Ni, Cr and V accompanying the increasing ferromagnesian content from gabbro-anorthosites to pyribolites seems quite natural. It is, however, remarkable that the calc-silicate rocks, to which a sedimentary origin has confidently been ascribed, yield still higher values for these elements. This has made the author very doubtful of the validity of trace element determinations as evidence for or against a particular origin of a rock. Caution is needed when the most 'sedimentary looking' type shows the most 'magmatic' trace element content.

Table III.

Trace elements of some ultrabasic to intermediate rocks from Tovqussap nunâ.

(Analyst Mr. IB SØRENSEN, cand. polyt. et lic. techn).

(ppm)

Rock type	Specimen	Ba	Sr	Ti	Zr	Co	Ni	Cr	Mo	V	Cu
Ultrabasic rocks											
Amph.-olivinite	35841	÷	÷	500	tr	100	1500	600	tr	30	20
Hbl.-pyr.-olivinite ..	19281	10	÷	500	tr	50	1000	1000	÷	30	30
Saxonite	19263	÷	÷	200	÷	30	800	1000	÷	10	tr
Hbl.-hypersthenite..	4035	20	÷	1000	30	30	300	600	tr	200	tr
Calc-silicate rocks											
Banded layer	18221 A	<10	100	1000	<10	30	200	1000	÷	100	5
Banded layer	18221 B	10	150	1000	<10	50	300	1000	÷	150	1000
Banded layer	18222	<10	30	800	tr	30	300	1000	÷	100	10
Skarn enclave......	35858	10	100	1000	10	10	30	200	÷	150	10
Pyribolites											
Pyribolite	19273	<10	30	3000	10	30	100	200	÷	100	100
Pyribolite	19299	30	50	2000	10	30	200	200	÷	80	20
Pyribolite	19242	10	30	500	10	30	100	600	÷	100	20
Pyribolite	4048	30	200	400	10	30	200	800	÷	100	30
Gabbro-anorthosites											
Gabbro-anorthosite	19262	200	80	100	tr	30	100	200	÷	20	20
Gabbro-anorthosite	19234	30	100	500	tr	10	30	300	÷	80	30
Gabbro-anorthosite	4097	10	50	100	÷	10	60	100	÷	20	10
Gabbro-anorthosite	19245	20	100	200	tr	20	100	300	÷	50	10
Diorites											
Latekinematic	19145	100	300	3000	<10	10	100	100	÷	200	20
Postkinematic	14971	100	150	300	<10	30	300	3000	÷	150	tr
Biotite-bearing											
pyriclasite schist	35845	50	<10	2000	10	10	100	1000	?tr	30	tr
Gabbro-anorthosite...	19632	500	500	3000	100	10	<10	tr	÷	50	1000
of magmatic origin ..	19636	500	500	800	30	tr	÷	tr	÷	tr	10

Table IV.

Average content of some trace elements in basic and ultrabasic rocks from Tovqussap nunâ.

(Analyst Mr. Ib Sørensen, cand. polyt. et lic. techn.).

(ppm)

Rock types/Elements	Ba	Sr	Ti	Zr	Co	Ni	Cr	Mo	V	Cu
Ultrabasics	<10	÷	550	tr	50	900	800	?tr	70	<10
Calc-silicates	≤10	95	950	<10	30	200	800	÷	125	(5–1000)
Pyribolites	20	80	1425	10	30	150	450	÷	95	40
Gabbro-anorthosites ...	65	80	250	tr	20	70	225	÷	60	20

Recalling what was said on a preceding page about the distribution of Zr in the acid rocks, the author is inclined to regard the present trace element content of the various Tovqussaq rocks as being largely dependent on the metamorphic and metasomatic processes to which the rocks have been subjected.

The trace elements, as well as the 'major' elements, formed part of the dispersed phase and could be re-distributed according to varying physical conditions which caused the development of new minerals. When a 'major' element entered into a new mineral its affinity to a given trace element may have been radically changed. If, for example, olivine forms from dolomite (+ quartz), its Mg might become more liable to attract Ni. It should also be remembered that no particularly drastic transport or migration is needed to cause considerable changes in the distribution of the trace elements. If an originally lime-rich layer is altered into calc-silicate through skarn-metasomatism, the siderophile elements would be expected to migrate towards the limestone along with the 'major' elements. In order to cause an enrichment of Cr of 1000 ppm in an originally Cr-free limestone, the surrounding rocks (now gneiss or granulite) need to be impoverished by only 10 ppm over a distance 50 times the width of the calcareous layer. When once started, the process of migration may carry on in a similar way to a process leading to the formation of concretions (Ramberg, 1952).

The ideas just outlined deviate considerably from what generally has been said about the behaviour of trace elements during high grade regional metamorphism (Sahama, 1945; Howie, 1955; Parras, 1959). But the author prefers to attribute to the trace elements the same ability to migrate as the 'major' elements, rather than to explain the skarn enclaves as having been formed from ophicalcites or other carbonate-bearing rocks of magmatic origin. Addition of volcanic ash to a calcareous sediment could not explain the present content of Ni and Cr in the calc-silicate rocks because this would mean that the Ni and Cr content of the volcanic rocks proper (which in the present case would be the pyribolites) should be higher than in the calc-silicates. Shrinkage in volume due to transformation of an original tufaceous limestone into calc-silicate rocks might have caused a slight relative increase in the content of the siderofile trace elements, but it could never explain the present richness of Cr in the calc-silicates.

It could be argued that the Cr and Ni content of the Tovqussaq gneisses is much too low if these rocks have supplied the Ni and Cr now found in the calc-silicates. To this the author can only answer that maybe the gneisses are deficient in these elements just because they have been extracted and are now to be found in the ubiquitous basic rocks. From other granulite facies areas, Cr contents of 200 to 500 ppm have been recorded from quartz-dioritic gneisses (PARRAS, 1959) and granulites (SAHAMA, 1945). Thus there are instances in which these rocks contain sufficient Cr to allow a considerable secondary enrichment in other layers.

Petrogenetic considerations

The completion of the structural analysis brought out that the Tovqussaq rocks form an approximately 1000 metres thick succession, the individual members of which show an overall conformable relationship (excepting the 'young' dioritic rocks). Although metasomatic processes may have taken place during the progressive metamorphism the author is more inclined to consider the present lithological characteristics of the different rock types as inherited from their original composition. This applies particularly to the basic rocks. In the acid and intermediate types, metasomatism has played a greater role.

The pelitic rocks (lutogenites) are obviously old argillaceous sediments and there is not, in the author's opinion, any doubt about the sedimentary origin of the calc-silicate rocks.

Regarding the gneisses, the author is most inclined to regard them as meta-sediments. This is partly because there seems to exist a complete gradual transition from the metasomatised meta-sedimentary granulites into the quartz-dioritic types. The possibility that dacitic and rhyolitic flows and corresponding tuffs were interbedded with the series cannot, however, be excluded, but when the author most favours a meta-sedimentary origin, he draws from experience obtained in other parts of the southern Sukkertoppen district. (The results from these studies will be published as part III of this series).

One problem related to the origin of the acid Tovqussaq rocks is the provenance of the material which caused the granitisation of some of these rocks during the posthumous phase. Although in many cases the 'granitisation' can be viewed as merely the result of recrystallisation with only slight introduction of new material (water and eventually some potassium), there are also instances where appreciable amounts of new material were introduced. Considering the water, an introduction from outside sources could be postulated because of the assumed dry nature of the original granulite facies rocks. Two features, however, have to be born in mind: 1) the gneisses are interstratified with conformable pyribolite layers containing hypersthene, and 2) the retrograde effect on the pyribolite is so slight that the possibility of migration of water through these layers can be more or less disregarded. Only where gneissification of the pyribolite has taken place can water have migrated from one gneissic member into another. The most probable answer to this problem seems therefore to be that the original granulite facies rocks were not all as dry as otherwise assumed. Repeated re-distribution of the ever present water due to changing tectonic conditions might easily result in an association which resembles that caused by long range migration and introduction of juvenile water. Similar ideas may also be applied to explain the behaviour of potassium.

In Tovqussap nunâ, the present distribution of the granitic and granitised rocks suggests that the greatest influence on the distribution of the granitising agents was exerted during the Smalledal and posthumous phase of folding.

The origin of the pyribolites will be discussed next. H. Sørensen (1953) regards these rocks as metamorphosed marly sediments. According to Sørensen they are related to the diopside-bearing amphibolites of the Alángua complex further to the north. A volcanic origin of the Tovqussaq pyribolites, however, might also be inferred. The interstices between individual pillows of sub-marine lava flows are quite often filled with lime-rich material and tectonisation and metamorphism of such rocks may well give rise to amphibolitic or pyribolitic rocks with interstratified bands of calc-silicates. The author therefore wants the possibility of a volcanic origin of these rocks to be left open. Possibly intercalations of marls, ash beds and pillow lavas (and eventually early intruded sills) contributed to build up the present pyribolitic rocks. If the sedimentogenous origin suggested by H. Sørensen were the correct one, a migration of Ni and Cr over long distances would have to be assumed. Taking the quantitative importance of the pyribolitic rocks into account one realises that it would be difficult to derive the Ni and Cr needed to build up their present content of these elements by diffusion from the surrounding gneisses alone. Since no evidence of long range diffusion of the corresponding 'major' elements has been found in Tovqussap nunâ, the author is more inclined to regard the Ni and Cr content of the present pyribolites as in the main original.

In an earlier paper the author attributed a sedimentary-metasomatic origin to the gabbro-anorthosites (BERTHELSEN, 1957). This idea was based on the similarity between some of these rocks and the calc-silicates. Subsequent laboratory work has made him somewhat more doubtful about this similarity (see above). Having heard that gabbro-anorthositic rocks ('blotchy gabbros') have been found as sills within the low-metamorphic rock assemblage of the Labrador trough, much thought was given to the possibility that the Tovqussaq gabbro-anorthosites are of magmatic origin.

After searching the literature, the author also found that, although described under many different names, rocks similar to the Tovqussaq gabbro-anorthosites (and calc-silicate rocks) are rather widespread. Examples from the pre-Cambrian of Finland (A. A. T. METZGER, 1945, p. 46; M. HÄRME, 1954, p. 37 and K. PARRAS, 1958, p. 58) and of Peninsular India (S. SEN and S. RAYCHAUDHURI, 1952) as well as some from the old Paleozoic mountains of Australia (A. J. R. WHITE, 1959) and from the Hercynian rocks of the Pyrenees (J. ALLAART, 1958, and H. J. ZWART, 1959) may be cited here. The views presented by these different authors as to the origin of the rocks in question vary considerably; tectonic mixing of basic igneous rocks with limestone, intrusion of primary gabbro-anorthositic material, addition of volcanic ash to calcareous sediments and iso-chemical metamorphism or metasomatism of shaly limestones or marls have all been suggested.

Discussing the facies relations of the Tovqussaq gabbro-anorthosites, the author stressed their uniform mineral composition. This combined with the occasional occurrence of relic large plagioclase grains (? phenocrysts), might be taken as evidence for the magmatic origin of these rocks. If of magmatic origin, the Tovqussaq gabbro-anorthosites most probably represent former flows and ash beds and not sills. This parentage is indicated by the small thickness of the original beds and layers. The seemingly constant competent behaviour of the gabbro-anorthositic rocks throughout the structural evolution concurs well with this view. On the other hand, the calc-silicates, which are believed to be of sedimentary origin, underwent a change in competence — apparently just before the onset of the Langø sub-phase.

Other features, such as the occurrence of calcite pockets surrounded by successive reaction rims of plagioclase, diopside and hornblende (fig. 82) show, however, that the mineral assemblage found in the gabbro-anorthosites could actually have been produced by metasomatism of original calcareous material.

As for the ultrabasic rocks, an origin through 'ultrabasification' of pyribolite (hypersthene-amphibolite) has been suggested by H. SØRENSEN (1953). This ingenious theory is not, however, supported by the more

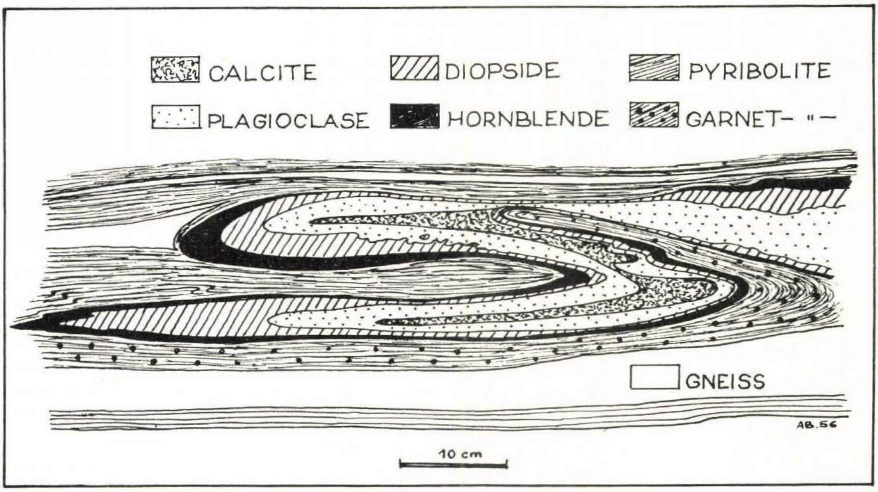

Fig. 82. Folded pocket of calc-silicates in pyribolite, Langø.

detailed field observations made by the present author. The ultrabasic rocks apparently formed independent layers in the original supracrustal rock series and during the Pâkitsoq phase of folding they behaved as plastic rocks; late, during the Langø sub-phase, they gained in competence. Bowen and Tuttle's experiments with the MgO-SiO_2-H_2O system have made it evident that olivine and enstatite can form as stable minerals at relatively low temperatures (Bowen and Tuttle, 1949). Supracrustal rocks with an ultrabasic chemistry may therefore recrystallise into olivinitic or peridotitic rocks during high grade regional metamorphism. The ultrabasic chemistry of the supracrustal rock could represent either siliceous dolomites or ultrabasic effusives (ophiolites) — or a mélange of both. Kinetometamorphic differentiation operating during an early stage of folding may also have contributed to the formation of rocks of ultrabasic composition (Tuominen and Mikkola, 1950; Mikkola, 1955). With his actualistic point of view and remembering the common occurrence of effusive ultrabasic and basic ophiolites in young mountain belts (A. Gansser, 1960), the author is most inclined to consider the Tovqussaq ultrabasites as merely recrystallised 'ultrabasic' layers of an ophiolitic rock assemblage.

Concluding Remarks

"The safest way in geology is the old way of inductive science, by studying nature's methods in her own great workshops, where they are available to our direct observation. Only by such field studies, conjoined with microscopical research, and of course giving due consideration to the

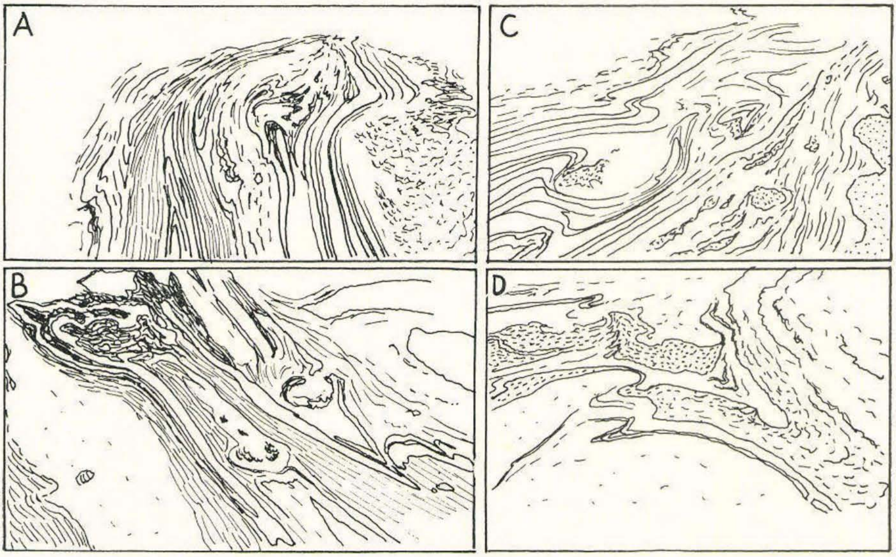

Fig. 83. Double fold structures in rock salt, anhydrite, dolomite and 'Haselgebirge', Ischler Salzberg, Austria.

important results of modern physicochemical science, we reach a better knowledge of the petrogenesis of the depth" (J. J. Sederholm, 1923, p. 151).

The recognition of the overall conformable relations between the individual rock members of Tovqussap nunâ depends on the structural interpretation. For example, had not the eastern isoclines ascribed to the Midterhøj phase been recognised as true fold structures, quite different conclusions could have been arrived at. The 'splitting up' of the pyribolite layers could have been taken as evidence for an intrusive origin of either the basic or the acid rocks forming the 'wedge' structures.

Thus double folding may give rise to outcrop patterns which simulate a discordant relationship between, for example, granitic rocks and layers of obvious metasedimentary origin. If the regional outcrop patterns suggest the presence of double folding, apparent discordances cannot be used as arguments for an intrusive emplacement of the granitic bodies. Possibly such seemingly discordant granitic bodies actually form the cores of unrecognised strongly refolded older isoclines.

The conformable relations between the Tovqussaq rock members suggest that the structures shown by these rocks have been developed without the introduction of any larger synkinematic intrusions. This conclusion is of some importance since, in many respects, the Tovqussaq structures are highly reminiscent of some Finnish structures which are considered to represent a particular type of 'intrusion tectonics' (see, for example, M. Härme, 1954). Thus there may exist convergent struc-

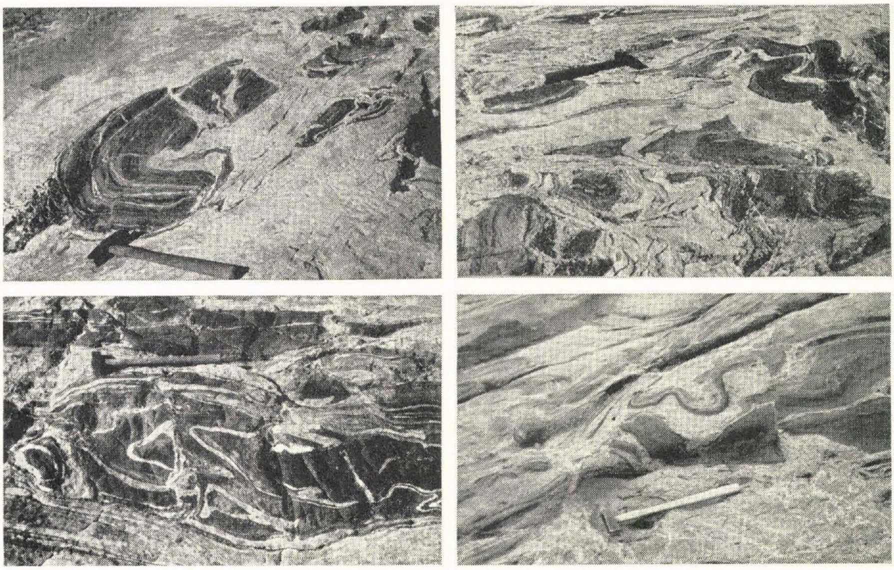

Fig. 84. Double fold structures in high-metamorphic gneisses from West and South-West Greenland.

tural styles as well as convergent rock types. Naturally divergent views on identical features may also exist.

Since E. Wegmann's classic paper on diapirism appeared in 1930, the analogies in the structural behaviour of salt domes and granite diapirs have been widely recognised by petrologists. In recent years increasing attention has also been paid to the structural influence of anatectic (palingenetic) processes (see, for example, A. A. T. Metzger, 1947). While not wishing to actually deny the significance of such processes the author wishes to point out that their influence may easily be over-estimated. Seemingly orderless small scale structures in gneisses and migmatites (the so-called wildfolds) may in many cases be shown by detailed structural studies to belong to a regular, if complex pattern. If such rocks are studied primarily from a petrological point of view, they may erroneously be termed anatexites.

In order to assure oneself that structural disorder is caused by anatexis, structural analysis is necessary. Paracrystalline deformation leading to plastic movements within an agmatite of originally replacive origin may, for example, cause the development of orderless structures completely without the intervention of anatexis.

As an illustration of this problem, the author presents in fig. 83 four sketches redrawn from H. Mayrhofer (1953, pl. 31, Bild 4, 8, 3 and 1). These sketches show the pattern of double folded rock salt, anhydrite, dolomite and argillaceous and arenaceous 'Haselgebirge' from the Ischler salt stock in Austria. These structures were formed by folding of solid rocks with accompanying recrystallisation according to Riecke's principle.

For comparison with these drawings, four photographs taken from double folded gneisses in the pre-Cambrian of W and SW Greenland are shown, fig. 84. Some geologists might think that these gneisses are anatexites. However, considering the great analogies between the structures in the Austrian salt and 'Haselgebirge' on the one hand and the Greenlandic granulite and amphibolite facies gneisses on the other, the author can see no reason to invoke any anatectic mechanism to explain the Greenlandic structures.

When the author started his work in Tovqussap nunâ in 1949, it was mainly with the intention of gaining an understanding of the formation of the Tovqussaq dome, which structure was first recognised by Dr. HANS RAMBERG. Having published his first preliminary report (BERTHELSEN, 1950) he received a post card from professor BALK who asked why normal and overturned dips were not indicated on the map. These few critical words inspired the author to do his best to clear up the stratigraphical relations not only in the dome but also in the region surrounding it, see Pl. 4. In consequence, the main object of the work changed gradually and discussion of the dome forms only a small part of the present paper.

In the intervening years much has been written about pre-Cambrian domes. ESKOLA's paper on the problem of mantled gneiss domes (P. ESKOLA, 1949) has been followed by numerous contributions (e.g. P. MICHOT, 1957).

In 1954, summarizing his experience from Western Greenland where dome structures are common in several regions, the present author wrote as follows (translating from the Danish):

"The formation of dome structures in gneissic terrains presumably depends on one or several of the following factors. —

1. Marked lithological differences between neighbouring stratigraphic members, or similar differences caused by migmatisation of particular stratigraphic members, or the presence of primary discordances. All these variations can be grouped under the common heading: physico-chemical disharmony.

2. Superimposition in time and interference of structural styles belonging to different structural levels (Stockwerke), i.e. tectonic disharmony.

3. True double folding."

At that time the author assumed that the physico-chemical disharmony presented by the different members of what was then called the Tovqussaq series was responsible for the development of the dome structure. The results arrived at in this paper suggest that all three factors played a part. However, had the influence of the posthumous phase been so strong that renewed diapiric movements had resulted, the composite origin of the structure might have been less evident.

As mentioned on the first page of this paper, it is the author's belief that an increasing number of carefully performed analyses of well exposed regions may help to establish a collection of structural styles. He hopes, therefore, that this paper may have contributed to the building up of a general picture of comparative tectonics.

X. РЕЗЮМЕ́

В настоящей работе рассматривается структурное развитие одного небольшого участка Товкуссапа Нунâ / Tovqussap nunâ / в южной части Суккертопп района / Sukkertoppen / в Зап. Гренландии / рис 1 /. Этот необыкновенно хорошо обнаженный участок состоит из до-камбрийских гранулитных и амфиболитных фаций горных пород. В введении излагается применённый способ картографирования и обсуждается петрографическая номенклатура. Далее описывается петрография 100 выбранных проб горных пород. Затем следует детальное описание геологической карты / табл. 1 и 2 /. Специальная группа горных пород — диориты — описывается особенно подробно, так как она имеет большое хронологическое значение.

Структурный анализ начинается описанием геометрических условий в восточной части Товкуссапа Нунâ. При помощи конструкций Вульфовой сетки, структуральных контурных карт и профилей проводится анализ сложных структур этого участка. Анализ показывает, что эти структуры образовались благодаря трём последовательным фазам складчатости. Полученные результаты применяются позднее при анализе западной и центральной части Товкуссапа Нунâ. Благодаря этому анализу, кажется вероятным, что горные породы Товкуссапа Нунâ, которые в настоящее время сильно метаморфны, представляют собою первоначальную геосинклинальную серию, мощностью, примерно, в 1000 м. и включающую пять взаимно согласующихся стратиграфических главных частей / табл. 1у / Эта серия образовала сначала лежачие изоклинальные складки, с осями идущими на С.С.З. / Мидтерхёй — фаза, Midterhøj phase / см. табл 1 и рис — /. По аналогиям предполагается, что эта складчатость произошла во время раннего периода метаморфного развития / шиферная фация. / эти структуры образовали позднее крупные лежачие складки, с осями идущими примерно в направлении В.С.В. и с амплитурой, которая может превысить 10 км. Эта фаза складчатости — Смалледалевская фаза / Smalledal phase / совпадала, должно быть, с метаморфозой амфиболитной фации.

Вторичная складчатость обширной западной площади, во время последней части Смалледалской фазы, создало, очевидно, условия для последующего диопиризма, при котором произошло первое формирование Товкуссапского купола. Этот диопиризм выражался в переносе пластических материалов к верховью купола между слоями некоторых горных пород / гнейссов /. Этот диопиризм произошел, по всей вероятности, во время изменений картины тектонических движений и повёл, повидимому, к последующей новой складчатости, на этот раз с осями, направленными на Ю.В. — Ю. Складчатость эта началась в условиях гранулитной фации. Структуры, которые образовались во время этой повторной складчатости, / Пакитсокская фаза Pâkitsoq phase /, показывают их отчётливую зависимость от более древних структурных форм. В конце Пакитсокской фазы можно выделить суб — фазу / суб-фаза Ланге /, при которой крутостоящие фольгационные плоскости вдавливались в ранее образовавшиеся складки. Можно считать, что образование поздне-кинематических диоритов имело место там, где слои пирроболитов, благодаря их движению вдоль крутостоящих фольгационных плоскостей в антиформных зонах сгиба, были более легко подвергнуты проникновению силинных материалов из нижележащих гнейсов центральной зоны складчатости. После-кинематические диоритные жилы и аплиты, которые, предполагается, образовались метазоматически, представляют собою переходный период времени — когда Товкуссапский участок был подвергнут тенциональному давлению. Первоначальные планарные аплиты дают возможность выделить еще одну фазу движения, которая местами сильно заметна в детальной картине структуры, но которая не причинила больших изменений в более крупных структурах. Эта »post-hume« -фаза образовывалась в условиях амфиболитных фаций и сопровождалась более или менее полной ретроградной метаморфозой более древних горных пород гранулитных фаций. Там, где гиперстенные гнейсы были подвергнуты лишь слабым внутренним движениям, они переобразовались в роговую обманку — биотит-содержащие »пурпуровые« гнейсы; а там, где внутренние движения были более сильными, они перекристаллизовались в светлые биотитные гнейсы. Гранулитные горные породы на Лангё / Langø / и в Тугдлерунарской свите (Tugdlerúnarssuit) образовались также во время »post-hume« -фазы, когда местное повышение отношения между Mg/Fe допустило сохранение / или новообразование / теперешнего состава минералов гранулитовой фации этих горных пород, несмотря на господствующее PT / давление — температура /, которое иначе бы повело к образованию типичных амфиболитных фаций минералов / как и в светлых биотичных гнейсах /.

Эта ретроградная метаморфоза сопровождалась местами также и гранитизацией, вследствие перегруппировки подвижного гранитного материала при господствующих тектонических условиях того времени. Под конец »пост-гумной« фазы образовались пара-до-пост-кристаллические милониты и сбросы. Как кажется, они представляют собою последние следы орогенных явлений в этом участке. Впоследствии произошло внедрение двух серий основных дайков, разделенных по времени образованием региональной системы разрывных сбросов, идущих в С.В. направлении. Этот последний кратогенный период эволюции участка был описан ранее в отдельной работе / Бертельсен и Бриджвотер 1960. / Berthelsen and Bridgwater, 1960 /.

Рассматриваемое здесь орогенное структурное развитие данного участка толкуется как отдельный геологический цикл. Предполагается, что различные фазы складчатости представляют последовательные структурные воздействия, которым были подвергнуты горные породы, проходя через различные тектонические стадии горной цепи. Кинематический анализ позволяет, таким образом, проследить здесь как вступительный период опускания, так и последующий период подъёма. Орогенное оформление участка Товкуссап произошло, вероятно, в связи с Кетилидианским периодом складчатости горной цепи, которое было установлено Е. Вегманном (E. Wegmann) в Ю.З. Гренландии.

В последней главе сперва обсуждается прогрессивная метаморфоза. Затем рассматривается минерально-фациональное положение горных пород Товкуссапа. Далее обсуждаются результаты некоторых анализов микроэлементов (trace element). По мнению автора возможно доказать, что настоящее содержание микроэлементов в горных породах только отчасти отражает их первоначальное распределение в исходном материале. Даже Zr, который обыкновенно считается очень стойким элементом при метаморфных переобразованиях, вероятно, тоже подвергся частичной перегруппировке. Высокое содержание Ni и Cr в некоторых скарновых горных породах можно считать свидетельством того, что эти элементы выделились из окружающих гнейсов во время скарновой метазомотозы. Что же касается происхождения ультроосновных горных пород, то автор отвергает теорию Н. Сёренсена, что они образовались из гипестена — амфиболитов. / H. Sørensen / Предполагается, что ультро-основной состав этих горных пород произошел во время геосинклинальной стадии, во время которой они могли быть представлены или кремнистыми доломитами, или, что более вероятно, ультро-основными офилитами изверженного происхождения. Основные горные породы / пириболиты и габброанортозиты / могли произойти или из изверженных геосинклинальных пород / или из ранее вытесненных силлов /, или из мергельных или известковых отложений. Сравнительно высокое содержание в них Ni и Cr указывает всё же, что первоуказанное происхождение более вероятно. Так как количество основных пород очень значительно, то трудно предполжить, что вокруглежащие гнейсы смогли бы доставить им достаточное количество Ni и Cr, чтобы эти элементы достигли той концентрации в пириболитах и габбро-анортозитах, каковая существует в настоящее время, если не предположить миграцию этих элементов через чрезвычайно большие пространства. Так как гнейсы и гранулиты в Товкуссапе и в ближайших областях показывают генстическую связь с явными седиментогенными породами, то предполагается, что кварцодиоритовые — гранитные породы являются метаморфозными и метазоматически измененными серыми песчанниками и слонцами.

В связи с этой петрогентической дискуссией, автор приводит несколько общих суждений, между прочим о том, поскольку анатектические процессы играли роль во время гранулитной и амфиболитной фазы метаморфных процессов. Исходя из аналогии между структурами метаморфных отложений и высоко метаморфных гренландских гнейсов, выводится заключение, что предположение о влиянии анатектонических процессов при образовании сложных гнейсовых структур является излишним.

В конце обсуждаются причины образования купольной структуры в участке гнейсов, на основании результатов кинематического анализа Товкуссапского купола.

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PLATE 1.

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GRÖNLANDS GEOLOGISKE UNDERSÖGELSE
THE GEOLOGICAL SURVEY OF GREENLAND

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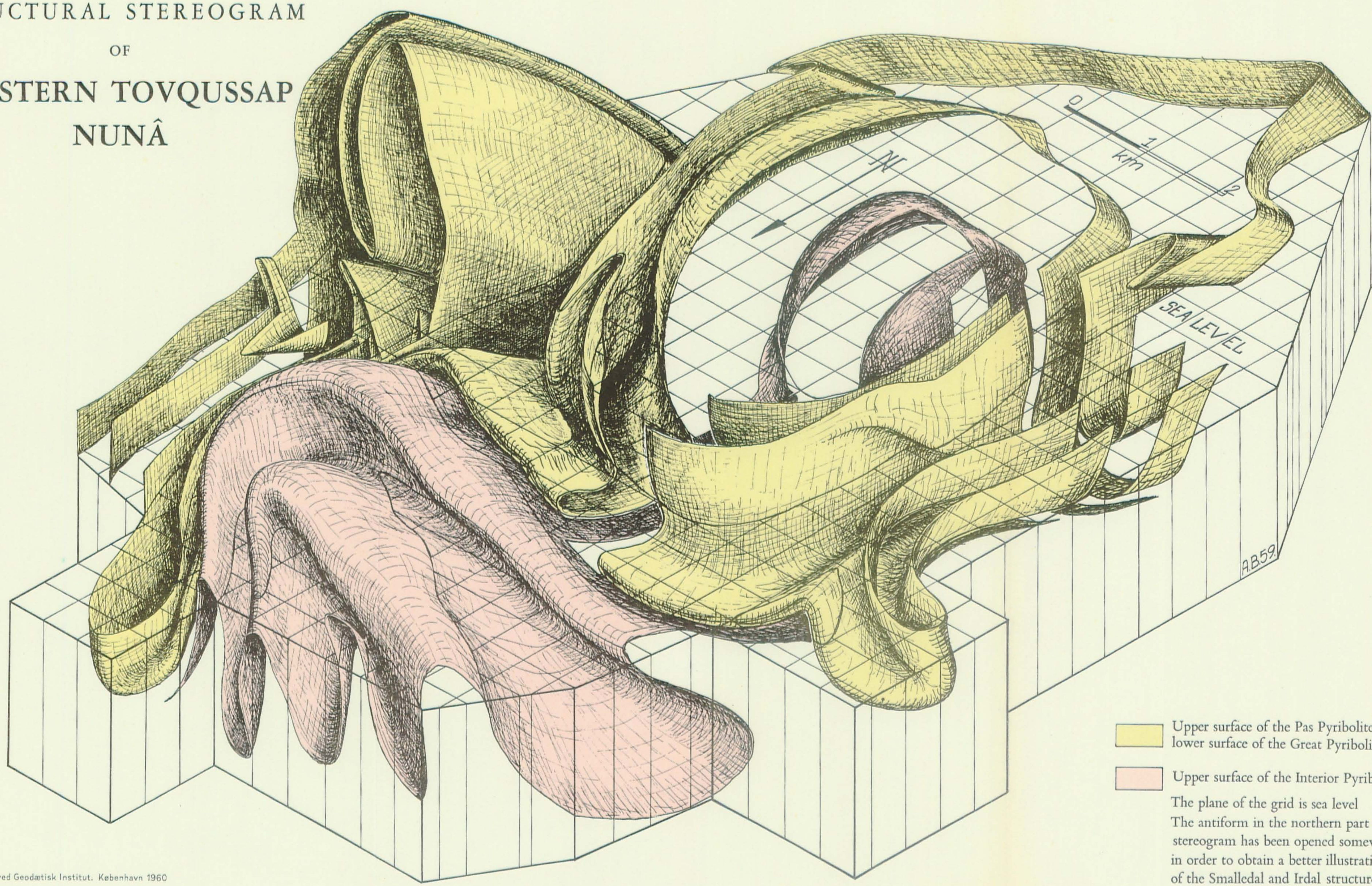
GRØNLANDS GEOLOGISKE UNDERSØGELSE
THE GEOLOGICAL SURVEY OF GREENLAND

PL. 3

STRUCTURAL STEREOGRAM OF WESTERN TOVQUSSAP NUNÂ



Upper surface of the Pas Pyribolite
lower surface of the Great Pyribolite

Upper surface of the Interior Pyribolite

The plane of the grid is sea level
The antiform in the northern part of the stereogram has been opened somewhat in order to obtain a better illustration of the Smalledal and Irdal structures.

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GRØNLANDS GEOLOGISKE UNDERSØGELSE
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PLATE 4

STRATIGRAPHIC MAP OF TOVQUSSAP NUNÂ