GRØNLANDS GEOLOGISKE UNDERSØGELSE
BULLETIN NO. 77

ANORTHOSITE XENOLITHS AND PLAGIOCLASE MEGACRYSTS IN PRECAMBRIAN INTRUSIONS OF SOUTH GREENLAND

PART I

BY D. BRIDGWATER AND W. T. HARRY

PART II

BY D. BRIDGWATER

WITH 79 FIGURES IN THE TEXT, 8 TABLES
AND 6 PLATES

С РУССКИМ РЕЗЮМЕ

DANISH GEOLOGICAL CONTRIBUTION
TO THE INTERNATIONAL UPPER MANTLE PROJECT

Reprinted from
Meddelelser om Grønland, Bd. 185, Nr. 2

KØBENHAVN
BIANCO LUNOS BOGTRYKKERI A/S
1968

GRØNLANDS GEOLOGISKE UNDERSØGELSE
BULLETIN No. 77

ANORTHOSITE XENOLITHS AND PLAGIOCLASE MEGACRYSTS IN PRECAMBRIAN INTRUSIONS OF SOUTH GREENLAND

PART I

BY D. BRIDGWATER AND W. T. HARRY

PART II

BY D. BRIDGWATER

WITH 79 FIGURES IN THE TEXT, 8 TABLES AND 6 PLATES

С РУССКИМ РЕЗЮМЕ

DANISH GEOLOGICAL CONTRIBUTION
TO THE INTERNATIONAL UPPER MANTLE PROJECT

Reprinted from
Meddelelser om Grønland, Bd. 185, Nr. 2

KØBENHAVN
BIANCO LUNOS BOGTRYKKERI A/S
1968

Abstract

Felspathic inclusions in Gardar igneous rocks are found over an area 70–150 km broad and at least 200 km long. This area is crossed by E–W trending transcurrent faults and contains the major Gardar alkali intrusive centres. The inclusions are of two main types: anorthosite xenoliths and plagioclase megacrysts. Some minor intrusions contain over 80 % of included felspathic material.

The xenoliths reach a maximum size of several thousand square metres in plan. The majority are granular anorthosites, composed of labradorite (An_{61}), olivine (Fa_{27-40}), Fe-Ti oxides, and rare ortho- and clinopyroxenes. Xenoliths of laminated anorthosite, composed of labradorite (An_{56}), olivine (Fa_{30}) and interstitial ilmenite and clinopyroxene, are found in gabbros from a restricted area close to younger syenite intrusions. Secondary anorthosite xenoliths, composed of derived fragments of granular anorthosite, rare laminated anorthosite or plagioclase megacrysts, set in a younger gabbroic host, are common. The interstitial host in the secondary anorthosites may be reduced to a thin film of altered mafic material surrounding closely packed felspathic fragments.

The plagioclase megacrysts, which reach 2 m in length, range in general from labradorite to calcic oligoclase. Single examples of bytownite have been noted in one Gardar dyke where they are found in small anorthositic aggregates with orthopyroxene (Fs_{22}) and olivine (Fa_{27}). The megacrysts are commonly clear and glassy; some are black and show a smoky discolouration in thin section. The colouration is thought to be due to the state of the iron in the felspar which is in turn related to the amount of water present. No major chemical differences have been found between black and clear felspar. Both varieties of megacrysts and many of the felspars from the anorthosite xenoliths show abnormal structural properties when examined using either the universal stage or X-ray techniques.

The type of inclusion is clearly related to the composition of the host rocks; olivine gabbro hosts contain large anorthosite xenoliths and scattered calcic labradorite megacrysts; less basic gabbroic hosts contain large sodic labradorite megacrysts and secondary anorthosites; while trachydoleritic hosts contain smaller andesine megacrysts.

The xenoliths and megacrysts are regarded as products of early crystallisation of the same Gardar magmas as later gave rise to their hosts. The granular anorthosites are thought to be the brecciated remnants of a felspathic roof formed by the flotation of plagioclase in a little-fractionated Gardar basalt magma at depth. It is suggested that this magma may have become stratified while still in the liquid phase so that there was an increase in the amounts of Na and K in the upper part of the chamber. As the magma differentiated the density difference between magma and felspar became less and the plagioclase crystals remained suspended. This resulted in the formation of large sodic labradorite crystals which were never compacted together to form solid anorthosite.

The early removal of a large amount of plagioclase-forming material and the suggested upward concentration of alkalies are believed to be two of the major processes controlling the formation of the Gardar alkali rocks.

CONTENTS

PREFACE

The premature death of W. T. HARRY in 1964 was a great shock to all those connected with geological research in Greenland. All who worked with HARRY will know the loss that geology in general, and in particular that of the Greenland Precambrian suffered. He was one of the increasingly rare geologists who are equally at home on both igneous and metamorphic rocks, an extremely valuable accomplishment when mapping the areas studied by HARRY in South Greenland where, as his papers record, some of the most spectacular igneous rocks in the world were emplaced into a well-exposed and varied metamorphic basement.

When HARRY originally put forward the idea that the felspathic dykes should be treated as a whole and not lost in a multitude of regional descriptions it was decided that he should be responsible for collecting all the available field information and working them into a composite picture while D. BRIDGWATER should carry out the laboratory work and mineralogical descriptions. Both authors were to be responsible for the comparisons and conclusions while HARRY should act as general editor. A preliminary draft was prepared in the spring of 1964 and only lacked the completion of the laboratory work and the compilation of field observations from the Kobberminebugt area. The new field observations however changed the interpretation of the problem and as the laboratory investigations were much more detailed than originally planned the whole paper had to be recast and the conclusions completely revised. It was decided to publish the paper in two parts in order to give HARRY credit for his work without ascribing conclusions to him with which he might have disagreed. The first part remains the joint responsibility of both authors; the second is the sole responsibility of BRIDGWATER.

HARRY's original script remains in a few passages which can be recognised by those who are familiar with his writing by their distinctive style, clarity of expression and economy of words.

K. ELLITSGAARD-RASMUSSEN
Director

The Geological Survey of Greenland
1966

PART ONE

By D. BRIDGWATER and W. T. HARRY

I. INTRODUCTION

a) Regional setting

The major events recognised in the history of the Precambrian of South Greenland can be summarised as follows:

1. Pre-Ketilidian sedimentation, intrusion of a large layered igneous complex with some resemblances to the Bushveld complex, folding, metamorphism, granitisation and erosion. The pre-Ketilidian may contain the relics of more than one cycle of sedimentation, metamorphism and subsequent denudation; however it is impossible to subdivide these in South Greenland.

2. Ketilidian geosynclinal sedimentation and basic lava extrusion, preceeded and accompanied by the emplacement of a regional swarm of basic dykes some of which carry large aggregates of plagioclase. Ketilidian folding, metamorphism, and the formation of regional granitic gneisses and granites **(the Julianehåb Granite)**. Remobilisation of considerable areas of pre-Ketilidian rock. Intermittent intrusion of basic and ultrabasic rocks appears to have continued throughout the plutonic activity of the Ketilidian.

3. Intrusion of a series of **basic and intermediate dykes**. These dykes mark a change in the tectonic conditions and may be used at least locally to separate the main Ketilidian plutonism from later events. Some of the dykes carry scattered plagioclase megacrysts.

4. The Julianehåb Granite was subjected to a rise in thermal conditions with a consequent reactivation and metamorphism. A series of dominantly allochthonous granites associated with noritic and monzonitic intrusions form a characteristic late plutonic suite resembling the rapakivis of Scandinavia. These intrusive bodies have given isotopic ages between 1600 m.y. and 1650 m.y. The high thermal conditions appear to have lasted until approximately 1500 m.y. in the granites close to Julianehåb.

5. Gardar igneous activity and sedimentation. The end of plutonic conditions in South Greenland was marked by the sedimentation of

coarse arkosic sandstones, the extrusion of a thick series of basaltic lavas, and the emplacement of several major alkalic intrusions together with numerous dyke swarms. Reliable isotopic age determinations on Garder rocks range from 1255 m.y. to 1020 m.y. (Bridgwater 1965).

The distribution of the alkali magmatism, the felspathic dykes, and the original distribution of the Gardar continental series and lava extrusion was controlled by a series of WNW–ESE trending faults. These faults occur in a belt approximately 70–100 km wide (measured at right angles to their trend) on the west coast of South Greenland. Isolated reports of syenite dykes and felspathic inclusions in dolerites from the east coast suggest that the belt may continue across Greenland under the Inland Ice. Movement began along these faults before the last plutonic activity in the area and continued intermittently throughout the Gardar period (Henriksen, 1960). The dominant movement direction was apparently a sinistral transcurrent dislocation, however, it is probable that there was locally a vertical component of at least a kilometre on some of the faults shown by the variation in the base of the Gardar continental series in the Igaliko area. The fault system, alkali magmatism, and type of sedimentation have broad resemblances to the tectonic features and rock types found in modern rift valleys. Some of the larger Gardar alkali intrusions straddle the transcurrent faults and thus lie at least partly outside the main fault controlled block. The faults themselves although generally showing considerable horizontal displacement are probably not pure transcurrent faults, movement is not constant along their trend and they are considerably more complex than the diagrammatic representation shown on Plate 4.

The sequence of events within the Gardar period is complex. In the area in which the felspathic dykes are most abundant the early Gardar is characterised by WNW–ESE olivine dolerites, which run parallel to the transcurrent faults. Several major alkaline intrusions including the nepheline syenite-carbonatite complex of Grønnedal-Íka (Emeleus, 1964) and possibly the earlier members of the Igaliko complex were intruded during this time. In the Kobberminebugt and Tugtutôq areas the WNW–ESE dolerites were followed by renewed movement of the transcurrent faults and the emplacement of large olivine gabbro dykes and plugs, some of which contain syenitic pods in their centre. These dykes are among the earliest to be intruded in the ENE–WSW direction which became dominant in the mid-Gardar. They are sheared and displaced by movement along a series of faults trending in the same general direction. Although the total movement was probably small the faults appear to have controlled the topographic features of South Greenland out of all proportion to their size. Many of the fjords in the Kobberminebugt to Narssarssuaq region trend in the same ENE–

WSW direction although there may only be small shear zones present in the rocks at the fjord heads.

By the mid-Gardar dyke activity was pronounced and extensive swarms were intruded between Ivigtut and Julianehåb. Mid-Gardar magmatism resulted in a considerable variety of dyke rocks ranging from olivine gabbros to comendites. Many areas show several generations of basic dykes which may be separated by intermediate or syenitic dykes and no strict progression from less differentiated to more differentiated magmas can be seen. However, there is an increase in the proportion of alkali granite and syenite emplaced late in the mid-Gardar and many areas show dyke swarms of closely related rock types which show trends towards more differentiated members late in the succession. Some differences are noted from west to east. In the Nunarssuit-Kobberminebugt area the alkali gabbros and syenites are all oversaturated and there is a preponderance of potash micro-syenites; in the areas east of Narssaq the majority of dykes are undersaturated and there is an increase in the proportion of soda microsyenites.

Small syenitic and alkali granite intrusions are found in the mid-Gardar, notably at Bangs Havn and Narssaq. Anorthosite xenoliths and plagioclase megacrysts are particularly abundant in gabbroic dykes associated with these mid-Gardar alkali intrusions.

Movement along the sinistral transcurrent faults diminished during the mid-Gardar and there was a change in the tectonic control towards the end of the period so that the late Gardar was marked by the emplacement of major alkali intrusions which are approximately equidimensional in plan. Anorthosite xenoliths are generally absent from the late Gardar intrusions except for fragments obviously derived from overlying earlier Gardar rocks though plagioclase megacrysts and felspar aggregates occur in the marginal gabbros of the Igaliko and Klokken intrusions. A few altered megacrysts are found in the late Gardar camptonite dykes (Upton, 1965).

6. The final magmatic activity in the area gave rise to a series of **olivine dolerite dykes** intruded approximately parallel to the present coastline. These dykes post-date the majority of faults in South Greenland and are thought to be post-Cambrian in age. No felspathic inclusions have been found in them.

b) Scope of paper

Various rather imprecise descriptions of porphyritic dolerites can be found in the early geological accounts of South Greenland (see Wegmann, 1938, for bibliography). Modern observation began in 1957 when a Gardar dyke composed of anorthosite xenoliths and large plagioclase crystals in a subordinate dolerite matrix was found by K. Ellitsgaard-Rasmussen near Nyboes Kanal in east Alángorssuaq. The locality was visited a few weeks later by one author (W. T. H.) and further work by the Geological Survey of Greenland (G.G.U.) in South Greenland has revealed widespread similar occurrences in Gardar dykes of various generations. Initially attention was particularly attracted by the large size of the plagioclase megacrysts (up to 2 m long). For this reason the dykes became known during the field work as **"Big Felspar Dykes"** (BFD) and have been referred to as such in recent literature, by Ayrton (1963) for example. It is now apparent that the abundance rather than the size of the plagioclase megacrysts is the feature of greatest significance and the dykes might be more aptly termed **"Felspathic dykes"**. This name is used throughout the present paper as a general term covering dykes with a noticeable proportion of anorthosite xenoliths, plagioclase megacrysts or any combination of the two.

Although these intrusions represent only a fraction of all the Gardar dyke activity in the area they are of considerable interest not merely intrinsically but also for their wider implications—their bearing on the regional geology of South Greenland and, perhaps, the origin of felspathic dolerite phases in other countries. As they therefore merit more than a piecemeal description dispersed in numerous accounts of separate areas they are collectively described and discussed in the present paper together with related phenomena in the same region. The main purpose in the laboratory studies was to provide a basis for a comparison of the anorthosites with rocks of more certain origin, especially the Gardar gabbros. In some respects the results only show our ignorance of the properties of normal Gardar rocks rather than giving any firm answer to the question of the origin of xenoliths.

The following have worked in the areas indicated on Plate 4 and Figs. 19 and 35. In the interest of consistency of descriptions one of us (D.B.) has visited many of the more accessible occurrences noted by other workers.

J. H. Allaart	Igaliko peninsula and the area north of Bredefjord.
Steen Andersen	Jespersen Dal district to the east of Igaliko.
S. A. Ayrton	Qagssimiut and the area to the west of the Tigssaluk Granite.
J. P. Berrangé	"Vatnahverfi", to the south of Igaliko.
E. Bondesen	The islands of Tôrnârssuk and Sermerssût and the area close to the Inland Ice north of Arsuk Fjord.

D. Bridgwater	Inner Kobberminebugt and the area along the margin of the Inland Ice north of Kûtsiaq.
K. Coe	Kûtsiaq.
K. Ellitsgaard-Rasmussen	Klokken intrusion.
C. H. Emeleus	Tigssaluk Granite and Grønnedal-Íka.
C. H. Emeleus and W. T. Harry	Igaliko complex.
J. Ferguson	Ilímaussaq intrusion.
W. T. Harry	Alángorssuaq.
S. Bak Jensen	Outer Kobberminebugt and Kînâlik.
T. C. R. Pulvertaft	Nunarssuit and the ground to the east.
H. G. Scharbert	Area north and east of Narssaq peninsula.
J. W. Stewart	Narssaq peninsula.
B. G. J. Upton	Tugtutôq and Narssaq.
B. J. Walton	Area north of Igaliko complex.
W. S. Watt	Qaersuarssuk.
J. S. Watterson	Ilordleq (Kobberminebugt).
M. Weidmann	Area surrounding the Tigssaluk Granite.

In addition, information about inclusions in pre-Gardar intrusions to the north of Ivigtut has been supplied by E. Bondesen, N. Henriksen, S. Bak Jensen and W. S. Watt.

c) Acknowledgements

The geologists listed above have contributed data and material from Gardar intrusions in the areas which they studied or have provided us with detailed instructions allowing us to make our own observations and collections and we are indebted for their generous cooperation.

Permission to publish this account is gratefully acknowledged from Mag. scient. K. Ellitsgaard-Rasmussen, director of the Geological Survey of Greenland (G.G.U.). Special acknowledgements are due to B. G. J. Upton, T. C. R. Pulvertaft, C. H. Emeleus, W. S. Watt and B. F. Windley for helpful discussion of the origin of anorthosites in Greenland and to J. P. Berrangé for discussion on the anorthosites of the Canadian Shield. One author (D. B.) wishes to thank members of the Geological Institute, Copenhagen for the first class facilities provided and for their help in laboratory determinations; in particular the help given by M. Danø and E. Krogh Andersen (X-ray determinations), H. Bollinberg (spectrographic determinations), Aa. Jensen and A. C. R. Ketelaar (determinations of Fe-Ti oxides) and H. Micheelsen (optical studies), is acknowledged. Chemical analyses were carried out by B. I. Borgen and I. Sørensen. A. E. Escher kindly redrew Plate 3. The second author (W.T.H.) wished to record his thanks to the Court of St. Andrews for leave to pursue his studies in Greenland; to his senior colleague Professor C. F. Davidson, for useful criticism and to J.E. Richey and F. Walker for constructive conversation.

II. FIELD OCCURRENCE OF FELSPATHIC MATERIAL IN GARDAR INTRUSIONS

a) Distribution

Anorthosite inclusions and plagioclase megacrysts occur at least locally in the majority of rock types intruded during the Gardar igneous activity. A few occur in the early Gardar dolerites and large masses are found in the roof rocks of the late Gardar agpaites of the Ilímaussaq intrusion; however, the greatest concentration is found in the mid-Gardar ENE–WSW dyke swarms from which most of the material described in this paper is taken.

The occurrences lie mainly in the 80 km wide tract of country between the Inland Ice and the sea from Kobberminebugt to the Igaliko district (Plate 4). They are more sporadic in the Ivigtut area, where felspathic dykes have been reported from the areas surrounding the Tigssaluk Granite, close to the Grønnedal-Íka alkali intrusion, and from Kînâlik on the north side of Kobberminebugt. Isolated occurrences have been reported from Sermiligârssuk, Midternæs and the islands of Sermerssût and Tôrnârssuk. The exact limits of the area in which inclusions derived from anorthosite are found in Gardar dykes are not easy to define due to the fact that many presumed Gardar dykes both south and north of the main area contain abundant plagioclase megacrysts which might represent xenolithic material. However if the boundary is drawn at the first dyke to carry abundant anorthosite xenoliths then it is found that the inclusions are restricted to the same belt of country as the Gardar alkali rocks, that is within the area cut by WNW–ESE transcurrent faults. There is a significant increase in the abundance of felspathic dykes close to the Gardar alkali intrusions.

The field diary of R. Bøgvad, made available to us through the kindness of Professor A. Noe-Nygaard, records observations made in 1932 of placioclase megacrysts in basic dykes on the east coast of Greenland. The most interesting observation by Bøgvad is a description of a megacryst-bearing dyke cut by an augite syenite dyke. The locality is due east of the main Gardar activity seen on the west coast and suggests that the Gardar rocks preserved in the area described in this paper may only represent a fraction of the total. Even disregarding Bøgvad's observations the area over which the felspathic dykes are found is quite impressive, forming a strip 200 km long and a maximum width of 150 km. It seems unlikely that the present coast line which limits the area to the west represents an original boundary.

Within the general area described above, there are two major concentrations of felspathic dykes each forming ENE–WSW trending belts between 10–15 km wide. The first of these occurs on the south side of Kobberminebugt while the second trends through the island of Tugtutôq to the Inland Ice north of Narssarssuaq.

b) Age relationships between the felspathic dykes and other Gardar intrusions

Relatively few felspathic inclusions are found in the early Gardar intrusive rocks. The swarm of WNW–ESE olivine dolerites which extend throughout the area locally carry megacrysts of plagioclase up to 40 cm in length which may be aggregated in clusters surrounded either by the normal host rock or by large mafic minerals identical to those found in the host. These are so scattered that little significance would be attached to them were it not for the much greater development of felspathic dykes later in the Gardar period. Occasional blocks of laminated or massive anorthosite have been found in the early Gardar basic dykes. The inclusions bear some resemblance to the larger masses brought up in mid-Gardar dykes but the material is too sparse for detailed comparison. Isolated felspathic inclusions have been noted in the gabbroic margins of the early Gardar composite syeno-gabbro dyke of Tugtutôq (Upton, 1962) and in the earliest olivine gabbro dykes trending ENE–WSW in the Isortoq area. Plagioclase megacrysts are also reported from NW–SE trending potash microsyenites north of Narssarssuaq which Walton (1965) describes as early members of the Gardar dykes in the area.

Intersections between the mid-Gardar dykes and the major intrusions provide some of the only evidence for dividing the period. The mid-Gardar dykes are cut by the Nunarssuit complex (Harry and Pulvertaft, 1963), the Tugtutôq central complex (Upton, 1962) and the Ilímaussaq intrusion (Ferguson, 1964). Some of the mid-Gardar dykes cut the early members of the Igaliko nepheline syenites but are themselves cut by the youngest rocks found in the Igdlerfigssalik centre of that complex. The Narssaq syenite and alkali granite intrusion cuts many mid-Gardar dykes but is itself cut by a few dykes including some felspathic rocks. In the Ivigtut area dykes probably belonging to the early and mid-Gardar cut the Grønnedal-Îka nepheline syenite-carbonatite intrusion (Emeleus, 1964), while they are cut by the Ivigtut cryolite body

(BERTHELSEN, 1962) and the late Gardar syenites of Kûngnât (UPTON, 1960).

Subdivision of the mid-Gardar dykes into different generations traceable over a large area is more difficult due to the repetition of dyke sequences within one area and the non-uniform distribution of individual dyke types. However, the felspathic dykes themselves, which show the same pattern of development in many parts of the area may ultimately help to provide a basis for the division of the mid-Gardar.

Detailed local sequences

The mid-Gardar dykes of the Kobberminebugt area belong to several different generations, the interrelations of which have not been fully worked out. The majority of the early dykes are medium-grained, rather nondescript olivine dolerites, some of which carry scattered felspathic material. These dykes trend in a general NE–SW direction. They are followed by the main mid-Gardar gabbro swarm of the area which contain less olivine and often show distinct alkali tendencies. In some places the gabbros form anastomosing complexes in which individual dykes were emplaced with little or no time interval between them. The general trend is ENE–WSW though many dykes have a sinuous course. Many are felspathic and some contain very large concentrations of xenolithic material. Several contain syenitic or alkali granite centres. The gabbros are associated with thin felspathic potash microsyenite-trachydolerite dykes with which they appear to be closely connected genetically. Intrusion of the microsyenite-trachydolerite dykes continued after the emplacement of the major gabbroic swarm; they are, however, cut and displaced by a few dolerite dykes and the Isortoq giant syeno-gabbro dykes which mark the end of the mid-Gardar igneous activity in the area. The Isortoq syeno-gabbros do not carry felspathic inclusions, though occasional anorthosite fragments are found in the intrusion breccias emplaced in a line southwest of the dykes.

The Gardar dykes of the Kobberminebugt area were affected by faults in four main directions. The ENE–WSW sinistral transcurrent faults continued activity after the end of igneous activity in the area and it is impossible to match individual dykes across some of the fault zones. At Augiâta tasia the movement on a group of faults, thought to be associated with the main transcurrent faults, can be seen to have been much greater in the early Gardar olivine dolerites and the first members of the mid-Gardar swarm. The felspathic dykes are sheared but rarely show marked movement. One distinctive felspathic dyke is displaced 200–300 m dextrally by a NW–SE transcurrent fault, in the opposite direction to the displacement seen on early Gardar dykes, suggesting that there was a change in movement direction at some time in the mid-Gardar.

Some movement occurred along zones parallel to the general ENE–WSW direction of the mid-Gardar dyke swarm. This movement can be used, at least locally, to distinguish between the early Gardar troctolitic ENE–WSW dykes and the mid-Gardar olivine dolerites; however, the movements appear to have taken place in a series of small dislocations rather than as one main break which could be used as a time marker. In some cases different generations of dykes have been displaced in opposite directions. Most of the felspathic dykes post-date appreciable movement along these faults although they may be crushed where they cross the fault plane. Local, approximately E–W, transcurrent faulting in the Isortoq area displaced the early Gardar dolerites and probably the first dykes of the mid-Gardar swarm. A thin felspathic unsheared dyke is found in part of the fault zone suggesting that it was intruded after the main movement. However, a prominent microsyenite-trachydolerite dyke (dyke 12, Plate 4), which is found on either side of the fault, is apparently displaced sinistrally at least 1 km. The apparent displacement may be due to en échelon emplacement.

With the exception of late movements on the WNW–ESE transcurrent faults, the youngest faulting in the Kobberminebugt area is a series of approximately N–S transcurrent faults. The majority of these show dextral displacement which affects nearly all members of the mid-Gardar swarm but which apparently did not displace the late mid-Gardar giant syeno-gabbros although shearing them.

The mid-Gardar swarm on Qaersuarssuk (WATT, 1968) contains at least six felspathic dykes with a general NE–SW trend. The south-west extension of the swarm is lost under the sea and only one of the more northern members of the swarm can be traced as far as the Qagssimiut archipelago. The number of dykes in the swarm decreases towards the north-west. The felspathic dykes all belong to the microsyenite-trachydolerite group and some of them may represent the same phase of dyke injection as dykes of similar type in the Kobberminebugt area. WATT distinguishes two, or possibly three generations of felspathic dykes which post-date both the early Gardar dolerites in the area and several generations of microsyenite dykes. The first generations of felspathic dykes occupy the same position in the local Gardar chronology as NE–SW trending dolerites. They can be separated from the younger felspathic dykes by sinistral movement on NNW–SSE faults, the emplacement of an olivine dolerite and sinistral movement on WNW–ESE faults. The northernmost felspathic dyke, which represents the youngest Gardar intrusive rocks of the area, is displaced 1500 m by a WNW sinistral fault which is thought to be the same as the main transcurrent fault in the Kobberminebugt area. The microsyenitic margins of the felspathic dykes in the Qaersuarssuk swarm are generally considerably more sodic than those seen in the Kobberminebugt area.

Further east, in **the ground north of Bredefjord** mapped by H. G. SCHARBERT and J. H. ALLAART, mid-Gardar felspathic dykes are rather sporadically distributed. They cut an earlier generation of ENE-trending dolerites but are cut by later microsyenites and dolerites sometimes intruded along the same fissure. The majority of felspathic dykes in this area are microsyenite-trachydolerites.

The second large concentration of felspathic dykes occurs on the island of **Tugtutôq** and extends in an ENE swarm along the **Narssaq peninsula** to the Inland Ice north of Nárssarssuaq. The chronology of this swarm is well established due to the work of UPTON (1962) and WALTON (1965). On **Tugtutôq** the felspathic dykes can be divided clearly into two groups, the olivine gabbros which are used as a datum for the beginning of the mid-Gardar, and a microsyenite-trachydolerite suite with many characters in common with the younger felspathic dykes in the Kobberminebugt area. The olivine gabbros, which form one of the major features of the island, cut early Gardar olivine dolerite dykes and the Hviddal syeno-gabbro (Fig. 35). On the western end of the Narssaq peninsula, where the gabbro forms a sheet-like body, it is cut by the syenites and granites of the Narssaq intrusion. The gabbros are not found with certainty further east, but large masses of felspathic gabbro found as inclusions in the lujavrites of Ilímaussaq may represent basic sills which were originally connected to the main gabbroic body at Narssaq. The olivine gabbros are cut by a variety of later mid-Gardar dykes including several generations of microsyenite-trachydolerites. Both potash- and soda-rich varieties have been noted; the time relationship between them is unknown. Many are felspathic. The felspathic dykes were succeeded by porphyritic alkali rhyolite dykes and a microsyenite to microgranite suite. UPTON (1962) suggests that some of the alkali dykes may represent preliminary tappings of the magma source which was subsequently to produce the rocks of the central ring complex. Some late porphyritic olivine dolerites trending parallel to the main mid-Gardar swarm cut microsyenites in the Skovfjord area.

On the **Narssaq peninsula** most of the mid-Gardar dyke swarm is cut by the Narssaq alkali intrusions. A few felspathic dykes, however, cut both the syenites and alkali granites of this complex. J. W. STEWART (personal communication, 1960) distinguished two types of felspathic dykes in the Narssaq area, the first carrying a high proportion of xenoliths and cleavage fragments of felspar while the second contains a much higher proportion of megacrysts and subhedral felspar fragments. It seems probable that these represent the gabbroic and microsyenite-trachydolerite hosts noted elsewhere. Felspathic sills, thought to be the lateral equivalents of some of the mid-Gardar dyke swarm, are found in the early Gardar lavas surrounding the Ilímaussaq intrusion. The mid-Gardar dykes and apparently the Narssaq intrusion are cut and displaced by the WNW–ESE transcurrent faults. These can be followed throught the late Gardar Ilímaussaq intrusion as a shear belt but do not seem to have caused a major displacement of the intrusion.

According to WALTON (1965) felspar megacrysts or aggregates of felspar occur in practically all generations of Gardar dolerites and microsyenites in **the area north of Narssarssuaq**. However, the most characteristic hosts are soda microsyenites. These post-date mid-Gardar NE–SW dolerites and potash microsyenites. They are approximately contemporaneous with a second generation of NE–SW dolerite dykes. They are followed by dykes bearing plagioclase inclusions surrounded by overgrowths of alkali felspar. E–W mylonites, presumably representing the transcurrent faults seen elsewhere in the Gardar period, cut the mid-Gardar swarm.

The Gardar dyke chronology worked out by J. H. ALLAART on the **Igaliko peninsula** is highly complex. A single laminated anorthosite xenolith has been found in an early Gardar dolerite. The early Gardar dolerites are followed by two generations of microsyenite and a persistant swarm of ENE mid-Gardar dolerites. The microsyenites are just saturated augite-hornblende bearing rocks the first generation of which trends E–W, the second ENE. The younger generation contains scattered megacrysts of sericitised plagioclase. The main concentration of felspathic inclusions occurs in the dolerites, both anorthosite xenoliths and plagioclase megacrysts are locally abundant. The microsyenites and the dolerites have both been displaced by movement along the E–W transcurrent faults which belong to the same fault system as the one which displaces the dykes to the east of Kobberminebugt. On the Igaliko peninsula the displacement appears to be up to 4 km although measurement is complicated by the fact that many of the dykes swing into the fault fissures but are dislocated by late movement.

Following the faulting and the emplacement of a series of hedrumite dykes there were two generations of porphyritic syenites trending NE and ENE. Both generations carry felspathic material which is especially abundant in the younger ENE dykes. The foreign material is mainly found as small altered plagioclase megacrysts averaging 0.5 cm in the older generation, 1 cm or above in the younger. Occasional blocks of granular anorthosite are seen. Several of the dykes show multiple injection often with late felspathic centres intruded into fissures occupied by an early non-felspathic phase. Late dolerite is locally intruded into the same dyke fissure.

The youngest igneous activity on the Igaliko peninsula includes several generations of alkali dykes some of which are probably contemporaneous with the major intrusions of Ilímaussaq and Igaliko. Several of the dykes contain large masses of felspathic material which is seen under the microscope to be perthitic. This may have been derived from anorthosite blocks which were replaced by interaction with the dyke material.

The majority of the mid-Gardar dyke swarm on Igaliko peninsula are truncated by the Igaliko syenite (Ussing, 1912, p. 267), and the metamorphic effect can be seen in the dykes for some distance along their strike.

The Igaliko syenite complex, mapped by one of us (W.T.H.) in cooperation with C. H. Emeleus, comprises four main intrusion centres, each of which is made up of several generations of syenite emplacement. These intrusive centres in order of probable age are: the Motzfeldt centre, the North Qôroq centre, the South Qôroq centre, and the Igdlerfigssalik centre. These are shown on Plate 4. There are also several satellite bodies associated with the main centres, the most notable of which is the Klokken intrusion (mapped by K. Ellitsgaard-Rasmussen). Felspathic dykes cut the first three centres although they are not numerous enough to be certain that they cut all the component syenites. The fourth (Igdlerfigssalik) centre which occurs between Igaliko village and the mouth of Qôroq was emplaced in two phases, the older phase, preserved on the south coast of Qôroq, is cut by felspathic dykes, while the younger phase which makes up the major part of the centre is younger than nearly all the dykes in this area, including those bearing felspathic inclusions.

In the ground between Narssarssuaq and Qôroq there are approximately five felspathic dykes cutting the syenites of the North and South Qôroq centres and the western extension of the Motzfeldt centre. In the south of the area, where the dykes cut the South Qôroq centre, the felspathic dykes have a rather irregular trend which varies between N–S and NNE–SSW. North of the major E–W transcurrent fault which bisects the South Qôroq centre the dykes swing towards ENE–WSW. There is a general decrease in the number of dykes north of this fault.

The felspathic dykes are generally under 20 m wide in this area, one of them locally reaching 25 m where it cuts the North Qôroq syenites. They are remarkably persistent along their strike although it is sometimes difficult to trace individual dykes due to a series of dislocations along E–W faults. One of the felspathic dykes is sinistrally displaced 2 km by the E–W transcurrent fault.

The felspathic dykes are among the oldest of the minor intrusives cutting the Qôroq syenites; they are followed by several generations of trachytic dykes. One felspathic dyke is seen chilled against an earlier trachytic dyke. The trachytic dykes show the same change in trend towards the north and appear to have been displaced approximately the same distance by the E–W transcurrent fault.

The younger rocks of the Igdlerfigssalik centre themselves locally contain abundant felspathic inclusions especially on the eastern margin of the intrusion at the well-known rare mineral locality of Narssârssuk (locality 22, Plate 4).

In the "Vatnahverfi" area south of Igaliko (Berrangé, 1966 a) felspathic insets are found as isolated inclusions in a variety of Gardar rocks. The oldest of these, designated as "Old dolerites" by Berrangé, is a group of irregular basic bodies and vertical dykes up to 40 m wide trending between NE and ENE. These may correspond to both the early Gardar and early mid-Gardar dykes seen in other areas further north. They are severely faulted and epidotised by later activity in the Gardar.

In the north-east of the "Vatnahverfi" area two of the NE-trending dykes belonging to the mid-Gardar swarm contain felspathic inclusions. One of the dykes is a microsyenite-trachydolerite resembling many of the hosts found in the areas to the north and west, the second is a nepheline syenite.

The relationship between the dyke chronology of the main belt of Gardar activity in the Nunarssuit-Narssarssuaq area and the dyke chronology north of Kobberminebugt is not yet clear. The early Gardar WNW–ESE olivine dolerites can be traced from one area to the other. They show a change in trend further north, where they become E–W or ENE–WSW. The early Gardar olivine dolerites which contain sporadic felspar inclusions cut the Grønnedal-Íka complex and several generations of lamprophyric dykes. They are cut by at least two younger generations of dolerite and a variety of trachytic dykes. Many are severely faulted. The felspathic dykes are generally described as early in the sequence, cutting early generations of lamprophyre and a few of the early olivine dolerites, but older than nearly all other dykes in the area. In the Tigssaluk Granite area a felspathic dyke apparently cuts and displaces a phonolitic dyke but elsewhere they are earlier than the alkali dykes and the NE–SW olivine dolerites. Petrologically the majority of felspathic dykes north of Kobberminebugt appear similar to the microsyenite-trachydolerite swarms further south. If they belong to the same period of emplacement then it appears that there is a considerable difference in the subsequent magmatic history of the areas north and south of Kobberminebugt. A few anorthositic inclusions are found in the NE–SW olivine dolerites which are later than the typical felspathic dykes in the Ivigtut area. One particular occurrence, from a dyke on the eastern end of Tôrnârssuk, is described on p. 33.

c) Types of inclusion and their distribution in Gardar hosts

There are two main types of felspathic inclusion found in Gardar igneous rocks; anorthositic xenoliths and plagioclase megacrysts. The majority of host dykes carry both; although the proportions may vary so that in one locality the inclusions are nearly all xenoliths while either along the strike or at a different distance from the contact the inclusions may be mainly megacrysts. It is taken as a fundamental assumption in this paper that the intimate association of the two types of inclusion throughout the Gardar is due to a close genetic connection between them.

i) Anorthosite xenoliths

The xenoliths show considerable textural variation ranging from massive granular anorthosite to layered olivine-gabbro inclusions. These are described in detail below as it seems probable that the textures give a more complete record of the geological history of the inclusions than any other feature. The textures may be sub-divided into those which appear to be primary, that is formed before inclusion in Gardar hosts, and secondary, formed by the interaction of felspathic material and either the host itself or an earlier Gardar magma.

BUDDINGTON (1939, p. 19) defines anorthosite, gabbroic anorthosite and anorthositic gabbro as rocks with 0–10, 10–22.5 and 22.5–30 units per cent of mafic minerals respectively. Following this classification the primary xenoliths are mainly anorthosites although some may range into gabbro anorthosite. Secondary xenoliths are generally more mafic

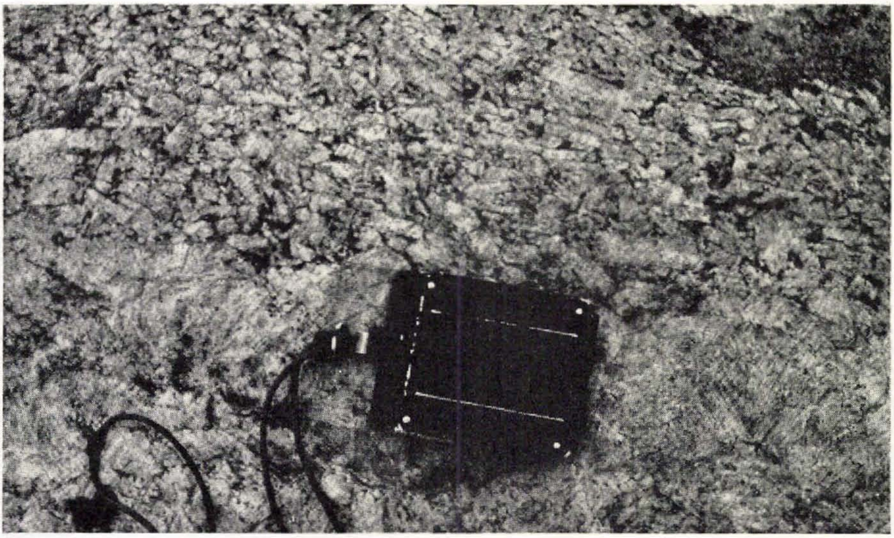


Fig. 1. Layering in anorthosite xenolith from a gabbroic host dyke at locality 2 at the eastern end of Kobberminebugt. Note the laminated subhedral plagioclase tabulae in the more mafic layer at the top of the picture. The compass is 6.4 cm broad.

Fig. 2. Layering in anorthosite block from alkali-gabbro host (dyke 6) at Eqalugssuit taserssuat. Note lack of preferred orientation of plagioclase crystals. The pencil is 0.75 cm wide.

and in places become anorthositic gabbro as the felspar content is diluted with other material.

Primary xenoliths belong to two groups. The first group, the most typical examples of which are anorthosites with a granular texture,

Fig. 3. Texture of hypersthene-bearing anorthosite block from an olivine gabbro host at locality 7, 5 km along the ice margin to the east of Kobberminebugt. The hypersthene is concentrated in pockets between the felspar crystals.

are found throughout the whole area in which the felspathic hosts occur; the second, which are practically confined to the area around Narssaq and the eastern end of Tugtutôq, are laminated anorthositic olivine gabbros.

Granular anorthosites. Massive, almost pure anorthosite, comprising randomly orientated, approximately equant anhedral plagioclase crystals occasionally more than 20 cm, but generally 3–5 cm in diameter, form the bulk of the large xenoliths found in Gardar dykes. They are particularly abundant in gabbroic hosts from the eastern end of Kobberminebugt. Some lithological layering may be present due to changes in the local content of mafic minerals (Figs. 1 and 2). The dark minerals may be restricted to films surrounding plagioclase anhedra or they may be concentrated in pockets or distinct layers within the anorthosite. The felspars in the relatively mafic layers show a more euhedral, commonly slightly tabular, form. The segregation of dark minerals into pockets is similar to that described by KRANCK (1961, p. 304) from the Canadian Shield except that the commonest mineral in the Greenland granular anorthosite xenoliths is olivine. The texture seen in the block from which one of the few samples of hypersthene has been collected is shown in Fig. 3. Many of the granular anorthosite xenoliths show considerable variation in grain size and compound block structures resembling those described by BALK (1931) are quite common. They arise

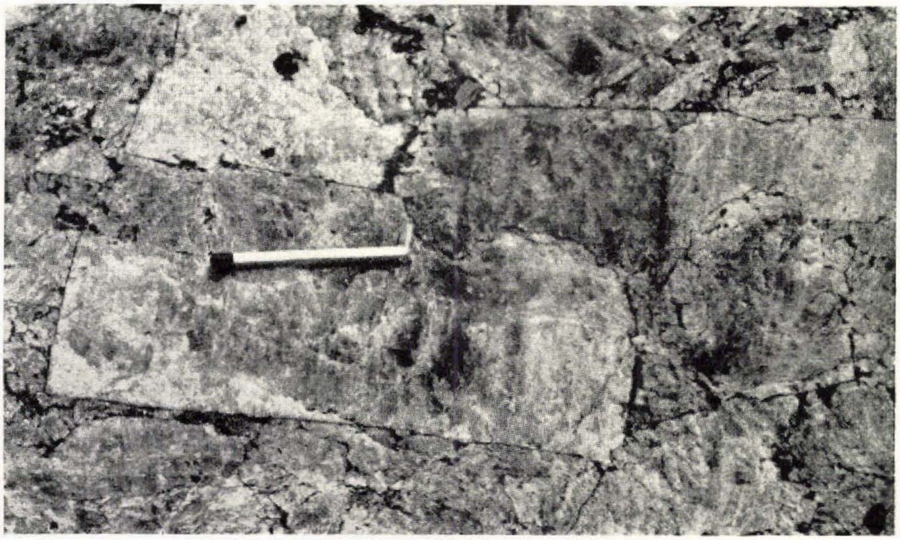


Fig. 4. Euhedral megacryst set in groundmass of anhedral plagioclase and a little mafic material. Anorthosite xenolith from alkali-gabbro host (dyke 6) at Eqalugssuit taserssuat.

where one textural or compositional variant of anorthositic rock encloses another. Often the earliest rock is coarser grained and less mafic than the surrounding material. Porphyritic varieties are moderately common in which large euhedral or subhedral plagioclase megacrysts are enclosed in a groundmass of smaller equant plagioclase grains (Plate 1a and Fig. 4). There is a tendency for the more basic xenoliths to be finer grained than the pure anorthosites.

Many of the fragments have a slightly deformed appearance which appears to have been formed earlier than the inclusion of the anorthosites in their present hosts. This deformation only exceeds the amount which could be explained by compaction of a large igneous body in a few cases. A block from the Bangs Havn complex (p. 39) approximately 1 m^3 in size has been sheared to such an extent that it resembles a fine-grained saccharoidal limestone with a slightly yellow colour and a pronounced impressed foliation. The surrounding xenoliths were completely unaffected. Other isolated fragments of deformed rock, especially of the more mafic varieties of anorthosite such as the olivine plagioclase rock illustrated in Fig. 5, have been found suggesting that some of the anorthosites were subjected to mechanical stress before their inclusion. However, the majority of fragments show no convincing evidence to suggest that they passed through a prolonged regional period of deformation between their formation as anorthosites and their inclusion as xenoliths.

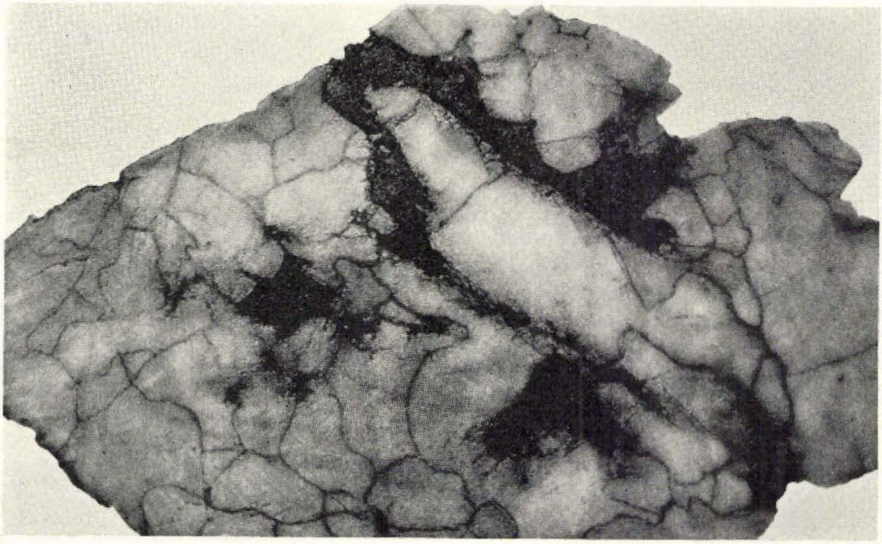


Fig. 5. Olivine-plagioclase xenolith with texture suggesting slight crushing. Xenolith from doleritic host, SE side of Eqalugssuit taserssuat. Specimen figured measures 11 cm in width.

Laminated anorthosites. The largest and best described inclusion of laminated anorthosite is well-exposed on the **Assorutit peninsula** (Fig. 35) at the eastern end of Tugtutôq (Upton, 1961, 1962 and 1964). The block consists of alternating units of anorthosite with a well-defined lamination of the individual plagioclase tabulae separated by thinner units of gabbro anorthosite in which olivine, the principal mafic mineral, encloses plagioclase tabulae with a much less marked lamination. The laminated anorthosites are not deformed. They occur within a restricted area, they enclose blocks of granular anorthosite and may be completely fresh while granular anorthosite blocks and plagioclase megacrysts in the same host dyke are sericitised. They appear to have passed through a much less complex history than the granular rocks and may have little bearing on the origin of the majority of anorthosite inclusions in the Gardar hosts except that they are basic rocks found in the same province.

Secondary xenoliths. The distinction between primary features developed before the inclusion of the anorthosite xenoliths in Gardar magmas, and features developed by subsequent interaction between xenolithic material and the surrounding host is often difficult to define and in fact may be artificial if the anorthosites themselves are an early-formed phase of Gardar basic rock. However, since much of the discussion of the origin of the felspathic inclusions depends on comparison with massive anorthosites in other parts of the world it is necessary to

Fig. 6. Secondary anorthosite xenoliths composed of granular anorthosite, plagioclase megacrysts, and rare gabbro blocks set in gabbroic host itself included in a younger gabbroic phase of the Bangs Havn complex. The gabbroic host to the felspathic material was only partly consolidated when broken up by the second gabbroic intrusion resulting in the sporadic release of plagioclase insets seen in the top right of the photograph (photograph, T. C. R. Pulvertaft).

try to separate those properties that are original from those resulting from later processes.

A large number of the larger xenoliths are made up of two components; older inclusions of felspathic material set in a younger, often gabbroic matrix. In many cases the secondary origin is clear; the gabbroic matrix to the compound xenolith, although itself an inclusion in a younger dyke, may be so close in texture and mineralogy to its host that there is little doubt that it is an early phase of the same intrusion. However, there is every gradation between this type of secondary inclusion and the block structure described as a primary feature on p. 22 where the matrix of a compound inclusion is much closer in character to the blocks it surrounds than to the host dyke. Although the boundary between primary and secondary features is arbitrary, there are several

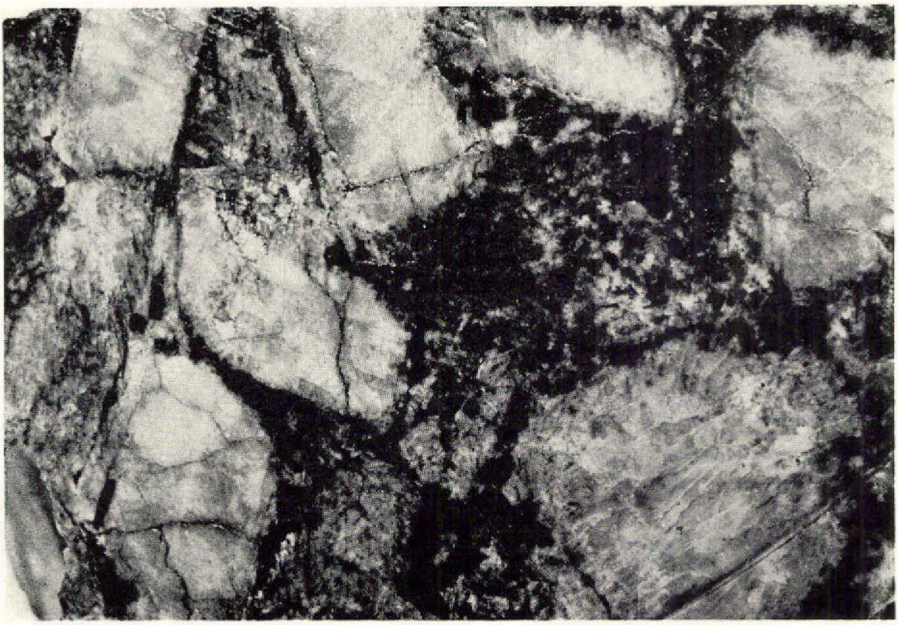


Fig. 7. Laboratory photograph of a secondary anorthosite xenolith from Bangs Havn, 1.5 × natural size. Note the rounded and sometimes broken plagioclase megacrysts set in a younger gabbroic matrix.

types of compound xenoliths the characters of which are clearly secondary and which have considerable bearing on the history of the xenoliths.

Types of secondary xenolith

Gabbro anorthosite aggregates. Large xenoliths composed of inclusions of granular anorthosite, plagioclase megacrysts and, in the Narssaq-Tugtutôq area, occasional laminated anorthosite blocks, all set in a gabbroic matrix are moderately common (Figs. 6 & 7). In some examples large single crystals of augite, olivine, and Fe-Ti oxides from the gabbroic component of the compound xenolith enclose older felspathic material ophitically.

Felspathic aggregates. Compound xenoliths of crushed granular anorthosite and plagioclase megacrysts set in a fine matrix of comminuted plagioclase are one of the commonest types of anorthosite inclusions found in Gardar dykes. Mafic material is often reduced to a film of indeterminate dark green coating surrounding some of the larger felspars. These aggregates are found throughout the felspathic dykes in the area between Kobberminebugt and Narssarssuaq. Typical examples from the Kobberminebugt area are illustrated in Figs. 8 and 9 and may be compared to the anorthosite xenolith shown by WALTON (1965, fig. 34).

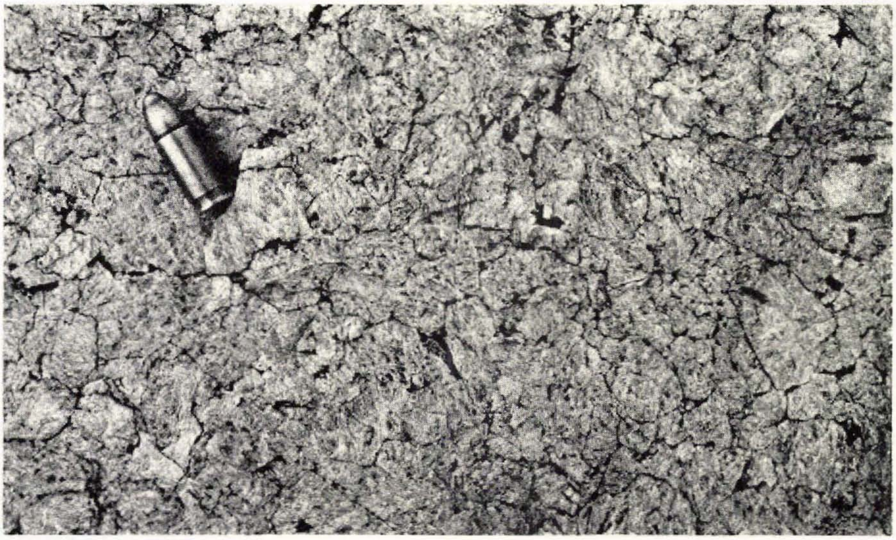


Fig. 8. Packed felspar crystals in anorthosite xenolith from dyke 8. Note the anhedral form of the felspars and the scarcity of mafic material. The revolver bullet is 9 mm in diameter.

Fig. 9. Detail of polished surface of packed felspar megacrysts from xenolith in dyke 8. Note the gabbroic host material which is found both interstitially and along crystallographic planes within individual megacrysts. 0.7 × natural size.

Fig. 10. Overgrowth of euhedral plagioclase surrounding an aggregate of closely packed plagioclase fragments. The megacryst is 15 cm long. Dyke 12, west shore of Isortoq fjord.

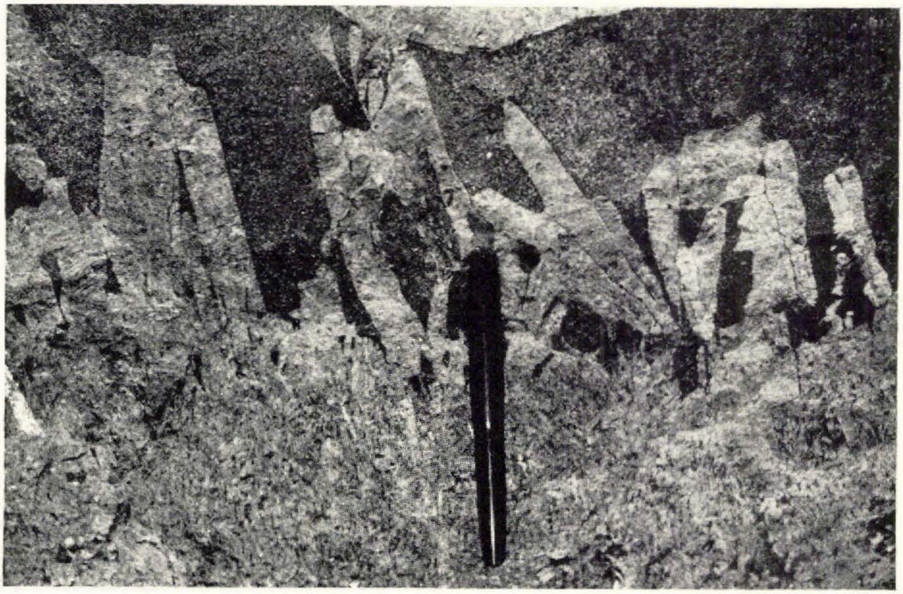


Fig. 11. Overgrowth of plagioclase megacrysts on a block of granular anorthosite. The granular anorthosite has a slight lamination parallel to the pen. Dolerite from the host fills the interstices between the plagioclase megacrysts to form a pseudo-ophitic texture. Locality 13, ice margin 18 km east of Kobberminebugt.

Rimmed aggregates. Several xenoliths have been noted consisting of a central core of either granular anorthosite or comminuted fragments, similar to the aggregates described above, surrounded by an overgrowth

Fig. 12. A 38 cm long plagioclase megacryst in a trachydolerite from Qaersuarssuk (photo. W. S. WATT). Note the nearly euhedral outline, the persistent 010 cleavage plane and the inclusions within the megacryst of anorthositic material (bottom left-hand corner).

of plagioclase. This overgrowth may occur as a single subhedral crystal (Fig. 10) or it may consist of several large crystals growing out from the early centre (Fig. 11).

ii) Plagioclase megacrysts

The plagioclase megacrysts probably form about 75 % of the total felspathic material found in the Gardar rocks. Some of them are certainly derived from the mechanical and chemical breakdown of anorthosite xenoliths. However, the proportion of material which can be shown

Fig. 13. A large anorthosite block in a dolerite dyke at head of Kobberminebugt (locality 2). The inclusion-free margin of the dyke is seen at top left-hand corner of photograph. The large anorthosite block is composed almost entirely of plagioclase with scattered pockets of olivine. Throughout much of the mass the plagioclase shows a persistent orientation parallel with the straight margin of the xenolith.

to have been formed from early solid anorthosite is remarkably small. The majority of megacrysts found as single crystals in Gardar hosts show several properties not seen in the constituent felspars of either the granular or the laminated anorthosites and their formation is treated as a separate problem allied to the formation of the anorthosites. The most spectacular feature shown by the megacrysts is their size. Clear, glassy, euhedral or subhedral plagioclase crystals and cleavage fragments up to 1 m long have been reported from several localities (Fig. 12), while large anorthositic blocks having the general outline of plagioclase crystals and a consistent crystallographic orientation throughout the mass have been noted up to 5 m long (Fig. 13). Apart from size the megacrysts show several other distinctive features. They are quite commonly black, or at least contain relic black material now surrounded by a clear rim (Plate 1b). They frequently contain inclusions of either gabbroic material or mafic minerals similar to those found in the surrounding host rock. These inclusions may occur throughout the megacrysts or may be concentrated within a marginal zone surrounding an inclusion-free centre. The megacrysts occasionally enclose small areas of less well organised felspathic material with textures and mineralogy similar to the granular anorthosite.

Fig. 14. Pseudo-ophitic aggregate of felspar megacrysts in a trachydolerite host, Qaersuarssuk (photo. W. S. Watt). Note the normal host rock between the crystals.

The megacrysts may be packed together to form a variety of secondary anorthosites. Compound xenoliths of granular anorthosite and broken megacrysts are quite common and in some dykes the megacrysts may form a loose texture in which rafts of randomly orientated plagioclase crystals are set in the host rock or in mixtures of host and large single crystals of augite, olivine or Fe-Ti oxide. This texture is termed pseudo-ophitic (Fig. 14).

iii) Distribution of xenoliths and megacrysts in Gardar hosts

Although the majority of Gardar hosts contain a mixture of xenoliths and megacrysts there is a general correlation between the petrology of the host and the relative abundance of xenoliths and xenocrysts.

Fig. 15. Packed felspathic fragments (mainly megacrysts) in trachydolerite dyke, Narssarssuaq. The lens cap measures 4.5 cm in diameter. This dyke has inclusion-free microsyenite margins. A poor lamination is seen in the tabular felspars parallel to the dyke margins.

The **felspathic early mid-Gardar olivine dolerites** are characterised by a high proportion of large xenoliths. These are often highly irregularly distributed and are commonly found either along the margins of the dyke or in small early apophyses which may be completely choked by inclusions. Megacrysts may also be present, however, the majority are fairly small and would generally be regarded as phenocrysts were it not for the presence of larger masses of felspathic material. Local accumulations of granular anorthosite, plagioclase megacrysts, laminated anorthosites and secondary anorthosites occur and may reach spectacular proportions (for example at Narssaq) but these are the exceptions rather than the general rule. Where this does occur it seems probable that the inclusions are concentrated near the roof of their host.

The **alkali gabbros** which form the commonest host rocks in the Kobberminebugt area contain an almost complete range of primary anorthosites, giant megacrysts, black megacrysts and many different types of secondary anorthosite. The more alkali members of the swarm show a marked increase in the numbers of large, clear, glassy megacrysts and secondary anorthosites formed from megacryst fragments.

The **microsyenite-trachydolerite host dykes** are characterised by a high proportion of moderate sized megacrysts and small xenoliths.

Pseudo-ophitic intergrowths between megacrysts and host rock are commoner in this group of dykes than any other. The felspathic inclusions show a typical distribution pattern in this group of dykes, with relatively few inclusions in the more alkali margins and increase in both the size and proportion of the insets, towards the basic centres. The felspars are frequently subhedral and may show a rough flow orientation and sorting according to size. A high proportion of the plagioclase insets are altered and show white against the dark background of the host. A typical felspar-rich member of the group is shown in Fig. 15.

d) Description of selected localities

The complete description of all the field occurrences of felspathic dykes visited by G.G.U. geologists during the last 10 years is beyond the scope of this paper. Many of the features are repetitive and the descriptions given in the following selection apply in all important respects to other areas. Most of the localities described are situated south of Kobberminebugt, in the region around Narssaq or within the Igaliko intrusion; areas of which we have first hand knowledge.

i) Localities north of Kobberminebugt

The most interesting anorthositic material found in Gardar intrusions north of Kobberminebugt was collected by E. Bondesen from an olivine dolerite at the east end of the island of **Tôrnârssuk**. The host belongs to one of the younger Gardar dyke swarms in the area, probably corresponding to the mid-Gardar south of Kobberminebugt. Most of the dyke, which trends NE–SW and reaches a maximum width of 50 m, is free from inclusions, but where the dyke branches and sends out thin apophyses into the surrounding gneisses there is an accumulation of small anorthosite xenoliths (averaging 5 cm in diameter) and megacrysts of olivine and hypersthene which may reach 3 cm in length. The mineralogy of the xenoliths and the megacrysts differs markedly from that of the host. The anorthosite xenoliths are considerably more mafic than the majority seen in the dykes further south; they contain between 25–35 $^0/_0$ of olivine and a little orthopyroxene. The texture is sub-ophitic with the majority of plagioclase grains surrounded by olivine and orthopyroxene. Apart from their high content of mafic minerals the inclusions contain a much more calcic plagioclase than is normally found in the anorthosite xenoliths (see Table 2, p. 75). A few of the other Gardar dykes on Tôrnârssuk and the neighbouring island of Sermerssût contain felspathic material which may be concentrated near the margins or where the dykes die out in a "horsetail" of small apophyses.

ii) Localities between Nunarssuit and Sioraq

The belt of country south of Kobberminebugt contains one of the main concentrations of felspathic-bearing dykes in the Gardar. The principal localities described are numbered on Plate 4. If the locality refers to a major dyke the same number applies to the dyke along its complete course. There are between eight and ten major felspathic dykes in the area (depending on correlation of individual dykes across faults) and a large number of smaller occurrences.

Locality 1. The two dykes which trend approximately E–W along the length of **Kobberminebugt** may be described together. Both show considerable variation along their length. The majority of exposures at the western end of the dykes comprise medium-grained dolerite with rounded anorthosite xenoliths, usually 0.5–2 m but sometimes up to 6 m in diameter, distributed across the entire width of the dyke and composing from 5–80 % of the area of individual large outcrops. Plagioclase megacrysts are scanty and occasionally are clearly seen to be derived from adjacent anorthosite xenoliths. The fine-grained anorthosite xenoliths disintegrate much more readily than the coarse-grained ones. Rare secondary xenoliths are found in the westernmost exposures of the southern dyke. Black felspar megacrysts with a vitreous lustre are found at the western end of the dykes. Sample 758 AT/59 (supplied by courtesy of the KRYOLIT SELSKAB ØRESUND) illustrated in Plate 1b, is taken from the southern dyke. Some of the plagioclases in the host rock are dusty and resemble the megacrysts.

For much of their length, however, the dykes show a well-marked division into marginal alkali dolerite zones almost devoid of foreign felspathic inclusions and central zones with abundant inclusions. The different zones lie in sharp planar contact but no chilling of one zone against another has been observed. At any one place the marginal zones are equal in width and together compose about one fifth of the total width of the dyke. Occasionally a plagioclase megacryst lies across the boundary between them and the central zone.

The marginal zones are fine- to medium-grained dolerite. Plagioclase megacrysts are very sparse, the largest seen is 20 cm long. Only one anorthosite xenolith was observed, a rounded block 20 cm wide.

The central zones are of alkali dolerite with plagioclase megacrysts from a few centimetres to 50 cm long. The megacrysts compose more than 50 % of the volume of the zones and, although often small, are distinct in size from the felspar crystals of their dolerite host. Often they are elongate and aligned parallel to the length of the dyke, especially near the margins of the zone. Some anorthosite xenoliths occur. These are usually rounded and less than 0.5 m in diameter. Other xenoliths are rare.

Locality 2. The felspathic dykes found at the eastern end of **Kobberminebugt** may be the continuation of the two dykes described from locality 1. However, glacial drift, the presence of a large transcurrent fault, and the en échelon emplacement of the dykes themselves makes this uncertain. The northern dyke at locality 2 is well-exposed on the cliff face of the nunatak forming the northern shore of inner Kobberminebugt. The dyke is approximately 100 m wide, and the host rock consists of a rather altered gabbro. In some places the contact between the dyke rock and the local country rock (a hornblende-rich diorite) is very difficult to define exactly. There is commonly a marginal zone to the dyke which is comparatively free of felspathic inclusions.

The host rock is unusually coarse-grained, resembling the matrix of typical secondary anorthosites. This coarse texture is especially noticeable along the borders of some of the xenoliths where the host rock has the characters of a gabbro pegmatite. Some of the insets are surrounded by a mafic rim. The dyke shows multiple injection; blocks of the coarse gabbro host together with felspathic inclusions are surrounded by a fresh dolerite comparatively free of felspathic material. The inclusions show a crude sorting averaging 10–15 cm at the centre and 2–5 cm at the margin. Cleavage fragments from large single crystals of plagioclase and xenoliths of granular anorthosite are both abundant. Many of the xenoliths have a slightly rounded outline or show fissures filled with host material suggesting that they were subjected to con-

Fig. 16. Packed anorthosite blocks in dolerite (locality 2, east end of Kobberminebugt). Note the small proportion of host to inclusion.

siderable brecciation before their inclusion. Both inclusions and host rock have been subjected to considerable alteration which appears to have been the result of accumulation of volatiles during the crystallisation of the host.

The southern dyke at locality 2 consists of three en échelon members which outcrop for small distances on the greenschist islands at the end of Kobberminebugt. The two eastern members contain one of the most spectacular accumulations of anorthositic material seen in South Greenland. The dykes themselves are not particularly prominent, reaching a maximum width of approximately 15 m. The centres of the dykes are choked with large anorthosite masses sometimes in blocks over 15 m long. Foreign material forms between 75–80 % of the total area of dyke exposed (Fig. 16). In some places there is a marginal zone with little or no included material; in other places the fragments have been thrust against the wall rock and commonly show signs of crushing and mechanical abrasion. Some of the fragments appear to have been broken by physical impact against each other within the dyke releasing clouds of felspar cleavage fragments in the process. Generally the blocks show no preferential alignment but some of the largest xenoliths have a tabular form, the long axis of which lies parallel to the trend of the dyke. Late veins of dolerite comparatively free from felspathic inclusions cut some of the xenoliths and their surrounding host rock. Nearly all the foreign felspathic material in these dykes is a coarse-grained massive granular anorthosite or felspar cleavage fragments directly derived from the larger blocks.

Locality 3. The most important felspathic dyke found in **Alángorssuaq** varies between 2–10 m in width. The host, an ophitic dolerite, is chilled slightly against the Julianehåb Granite country rocks but not against the felspathic inclusions. The inclusions occur throughout the width of the dyke, and are not confined to a central zone. Laterally their distribution varies; they average about half the total visible area of the dyke, but may locally reach 80 % of the rock exposed. The inclusions (Fig. 17) vary from blocks of coarse-grained anorthosite from a few centi-

Fig. 17. Fragments of granular anorthosite in 10 m dolerite from eastern Alángorssuaq (photo. W. T. Harry).

metres up to 4 m long, plagioclase megacrysts up to 1 m long and plagioclase cleavage fragments which show a range in size from tens of centimetres down to material barely distinguishable from the plagioclases of the host. Xenoliths and megacrysts are attacked by the dolerite, which sends small apophyses into them, but their instability in the dolerite seems physical rather than chemical; reaction rims around them are lacking.

Both host and inclusions have been affected by a late alteration which has saussuritised the felspar and uralitised the mafic minerals in the host. This alteration is common in the gabbroic rocks east of the Nunarssuit intrusion. Most of the included material weathers to a chalky white and is pale gray or white with a dull lustre on fresh fracture. Some large megacrysts, however, (*e.g.* those nearly 1 km south of Rinks Havn) contain irregular, vaguely defined patches of dark gray or black plagioclase with a highly vitreous lustre. This dark plagioclase is translucent, displays pronounced undeformed polysynthetic twin lamellae and is in crystallographic continuity with the rest of the plagioclase crystal in which it occurs.

Dyke 4 is one of the most complicated of felspathic dykes in the area as the rock types forming the host change fairly rapidly along the strike. The dyke can be traced 35 km from the sea inlet just south of **Nyboes Kanal** to the ice margin at **Eqalugssuit taserssuat.**

For 6 km along its western extent dyke 4 has dolerite or syenite marginal zones 20–40 m wide on either side of an amphibole granite centre 50–100 m wide. The edge of the granite is rarely seen but where it is visible a hybrid rock separates the granite from the marginal basic zone. The granite sends veins into the marginal zones and the Julianehåb Granite country rock. The marginal zones sometimes contain felspathic inclusions, which at some localities comprise abundant plagioclase megacrysts and at other localities comprise anorthosite xenoliths. At one place numerous Julianehåb Granite xenoliths occur in the dolerite marginal zone of the dyke. These xenoliths show all stages of digestion by their host. Comparable phenomena are described below from dyke 6. At another locality the composite dyke has altered the Julianehåb Granite for many metres by partial melting of the older rock and subsequent contamination of the dolerite. An older, metamorphosed amphibolite dyke close to the composite dyke is penetrated by veins of the remobilised basement granite.

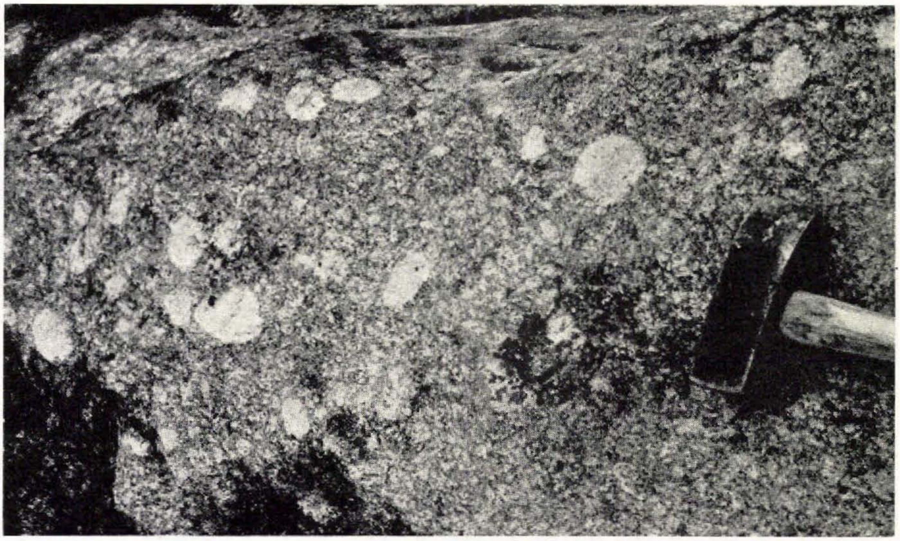


Fig. 18. Rounded and altered megacrysts of plagioclase in larvikite host of dyke 4 at Eqalugssuit taserssuat. The megacrysts are concentrated within 50 cm of the contact between the syenite and the gabbroic margin of the dyke.

The granite centre of the composite dyke thins out eastwards and the dyke then consists wholly of syenite without changing notably in thickness. This syenite continues eastwards for about 1.5 km and contains a central lens of granite about 200 m long. For the next 6 km of its course the dyke is contaminated by assimilation of Julianehåb Granite xenoliths, which are abundant; a few anorthosite xenoliths also occur. On the western side of Augiâta tasia the dyke is an alkali gabbro relatively uncontaminated by granitic material, but on the east side of the lake there are numerous inclusions of Julianehåb Granite in a heavily contaminated groundmass. The presence of large amounts of granitic material within the dyke appears to preclude the inclusion of significant amounts of anorthosite. Those parts of the dyke which contain large numbers of granite inclusions are often difficult to trace in the drift covered areas to the east.

3 km east of **Augiâta tasia** dyke 4 is again seen as a normal felspathic dyke with an alkali gabbro matrix. The dyke cuts several generations of NE-trending dolerites but is itself affected by the intrusion of a composite syeno-gabbro. Local pods of a larvikitic syenite are found in the centre of the dyke; this syenite is characterised by the development of felspar rhombs up to 2 cm long which often weather out of the more basic matrix. The dyke has not been traced with certainty across the lake running north-west from the head of Isortoq because of lack of exposure, however, in the high ground approaching the ice margin, 3 km north-east of Isortoq, a broad composite dyke with 2–3 m gabbroic margins and a rhombic larvikitic syenite centre is found directly along the strike. The gabbroic sheath contains scattered megacrysts and a few anorthosite fragments. In the syenitic centre the inclusions are rare and those found have been severely attacked by the alkali host. On the ice cleaned margin of the north shore of **Eqalugssuit taserssuat** the larvikite dyke broadens out to approximately 600 m wide. The gabbroic margins to the larvikite dyke are often severely altered by the emplacement of the alkali rocks. In places there appears to have been some mixing and locally the syenite contains many

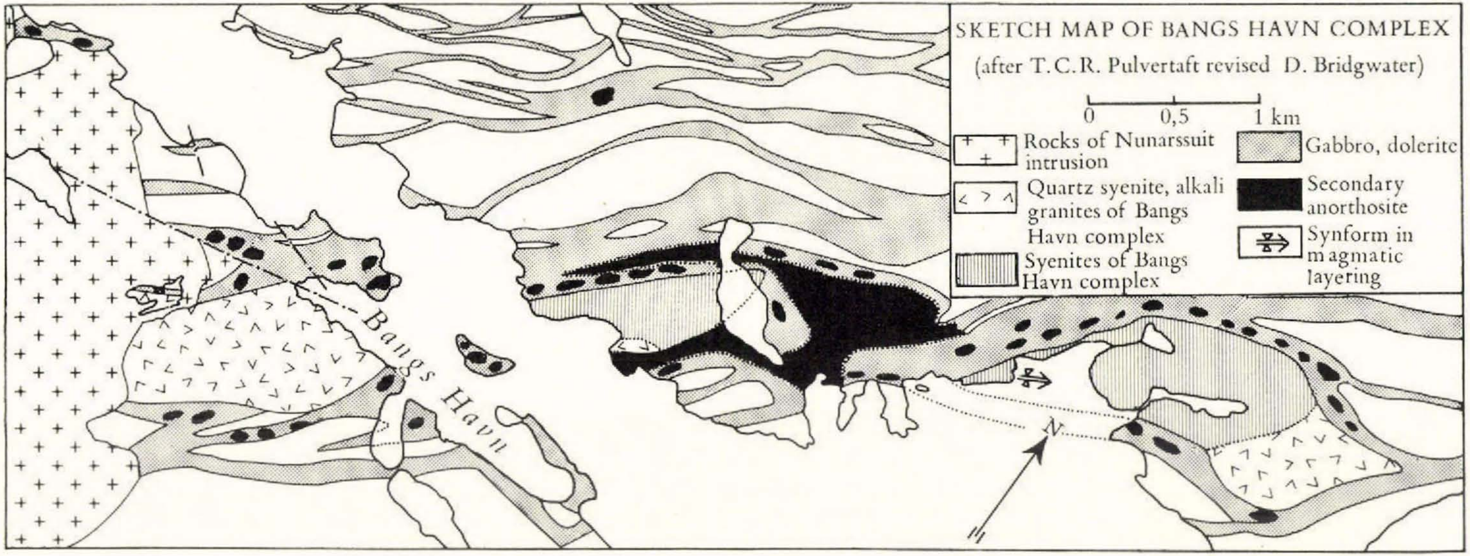


Fig. 19. Sketch map of the Bangs Havn complex.

rounded and heavily sericitised plagioclase megacrysts within a metre or so from the contact (Fig. 18). In other places the syenite breaks through the gabbroic margin. It appears that the two rock types were emplaced with little time interval between them; the best evidence for this is the way in which the gabbro follows the irregularities of the dyke form, suggesting that it was still plastic when the larger mass of syenite was intruded into the fissure. The last phase of intrusion of the dyke was the formation of a series of thin alkali granite veins and a single small dyke of plagioclase, chlorite, Fe-Ti oxide rock with a bulk composition close to an alkali gabbro. Three chemical analyses of the main rock types are given in Table 4, p. 124, Nos 8, 9 and 10.

The dyke cuts several earlier generations of Gardar rocks in the area, including one of the other major felspathic dykes (dyke 6) see Fig. 22, p. 42. It is cut by the east-west transcurrent fault which crosses Eqalugssuit taserssuat. The continuation of the larvikite dyke to the east has not been found although a rather brecciated mass of dykes containing some anorthosite fragments occurs on the south side of Eqalugssuit taserssuat close to the Inland Ice.

A dyke which, although devoid of xenoliths, might be the westward continuation of dyke 4 forms the south coast of the island Qeqertarssuánguaq near SE Alángorssuaq. It trends ENE, is about 70 m wide and contains a centre of quartz porphyry that in places passes rapidly into syenite. The margins of the dyke are uralitised ophitic dolerite locally felspathised by the syenite. The porphyry centre enclosed blocks of the dolerite and lies in irregular contact with the latter, without evidence of chilling.

Locality 5. **The Bangs Havn complex**, seen at Bangs Havn in Nunarssuit and on the mainland to the north-east, provides interesting data. This complex, about 1 km wide, is a lenticular congeries of anastomosing gabbroic dykes enclosing areas of syenite and amphibole granite (Fig. 19). On the mainland felspathic inclusions are most abundant in the gabbro surrounding the western area of syenite, sporadic and less common in the gabbro round the eastern syenite. The syenite grades into granite over several metres but has a relatively sharp contact with the gabbro, which develops marginal pegmatite and is highly uralitised whilst the other rocks are unaltered. Spectacular trough banding is developed in the mafic eastern syenite.

The felspathic inclusions in the gabbroic rocks are gray or white plagioclase insets and, perhaps to a lesser extent, anorthosite xenoliths. A large proportion of the felspars locally show a purple tinge thought to be the result of the breakdown of an original black pigmentation. The insets measure up to 35 cm long and present a broken or corroded outline. Many are cleavage fragments. The corners of some have been rounded so that they present a roughly oval shape. A sorting according to size is often evident, the insets within one outcrop then being roughly of the same dimensions. Inclusions of mafic or doleritic material can occur in the insets, frequently in trains parallel to the (010) cleavage.

The rocks richest in plagioclase insets form fairly well defined areas from less than 1 m up to more than 50 m wide surrounded by gabbro either poorer in, or devoid of, foreign felspar. These areas may be sharply bounded or, more commonly, pass gradually over distances of up to a few metres into the surrounding gabbro. The latter occasionally becomes finer grained close to the felspathic areas (Fig. 20).

In general the felspathic rocks comprise about 75 % by volume of broken plagioclase crystals set in an uralitised mafic ground, but felspathic material can be even more abundant and rocks of all compositions in the range felspathic gabbro to anorthosite are present. Some difficulty can arise in distinguishing anorthositic

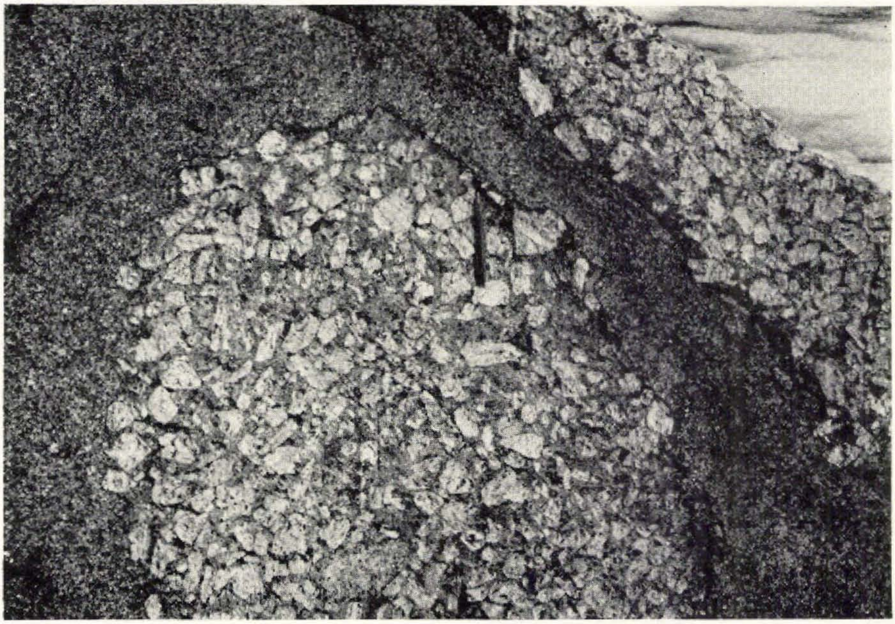


Fig. 20. Secondary anorthosite xenolith from the Bangs Havn complex. The xenolith consists of a moderately well-sorted aggregate of felspar fragments set in a gabbroic groundmass. This in turn is included in a younger gabbro. In places, for example along the upper right-hand margin of the xenolith, the second gabbroic phase is chilled against the older rock.

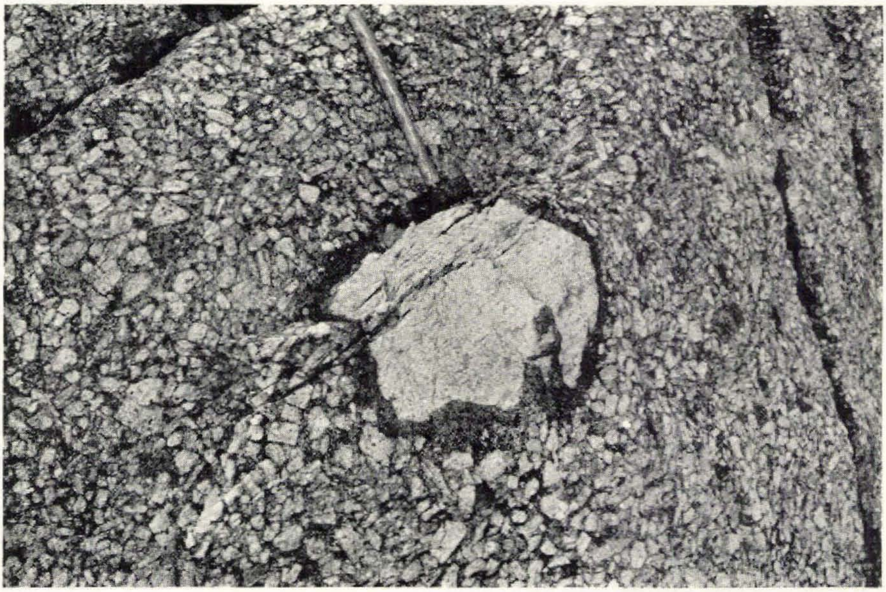


Fig. 21. Inclusion of granular anorthosite with chilled gabbroic rim set in a felspathic secondary anorthosite.

members of this series (which is clearly a secondary series formed by incorporation of fragments of anorthosite in basic magma) from xenoliths of the original anorthosite. Such secondary anorthositic rocks, however, can contain some gabbroic xenoliths and thin mafic rims may occur round their component fragments (Fig. 21). A thin mafic zone can occasionally also occur at the contact between normal gabbro and the secondary anorthosite xenoliths themselves.

The gabbro immediately surrounding the larger fragments within the secondary anorthositic areas is commonly finer grained than that around the other felspathic fragments in the xenolith-rich rock.

Dyke 6 can be traced from the north-east shore of the sea inlet **Tarajornitsoq**, 2 km to the east of Bangs Havn, to the ice margin at **Eqalugssuit taserssuat**, 30 km to the north-east. It varies between 10–100 m wide, averaging 50–70 m. It probably contains a greater volume of felspathic material than any other dyke mapped so far in South Greenland, and a high proportion of the detailed petrological and mineralogical studies described later have been carried out on samples collected from this dyke. No definite connection can be established between dyke 6 and any particular member of the plexus of Gardar dykes forming the easternmost extension of the Bangs Havn complex as the distribution of insets in these is sporadic and no other petrological peculiarities can be used. The general strike, and the fact that the Bangs Havn complex contains many inclusions of a similar nature to those in dyke 6, suggests however that a connection exists at depth.

The dyke is fairly constant in rock type with none of the lithological variation which characterises dyke 4 and which is described below from dyke 8, 200 m to the south. The host is a rather alkaline gabbro occasionally with narrow anorthosite-free margins. In some places thin dykes of fine-grained microsyenite with rhombic felspar phenocrysts occur near the margins of the main dyke. These alkali rocks show a variable relationship to the main dyke; in places they cut across the gabbroic host and the anorthosite inclusions but generally they appear to be a slightly earlier phase of intrusion in the same dyke fissure. Similar features are described in more detail from dyke 8 from the nunatak, 10 km to the east of Kobberminebugt. In some localities the dyke is chilled against the country granite, in others considerable rheomorphism has taken place with veins of remobilised granite back-veining the dyke.

The dyke has several en échelon breaks and locally shows sudden changes in thickness. It is cut by the gabbro margin and syenite centre of dyke 4.

At several points along its length the dyke runs parallel to and abutting dolerites with little or no included felspathic material. It is commonly difficult to decide whether these represent non-felspathic components of the dyke intruded essentially at the same time as the felspathic dyke or whether there are two or more dyke generations in the same fissure. Probably both occur. The relationship is further complicated in areas, such as at the head of Isortoq, where the dyke splits into two, one arm being comparatively free of inclusions at the junction but regaining its felspathic nature along the strike. In this respect the dyke resembles the anastomosing dolerites of Bangs Havn where there is a sporadic distribution of anorthosite not restricted to any one member of the complex. The best exposed example of this phenomenon occurs in the ice-polished slabs directly to the south of the intersection between dykes 4 and 6 (Fig. 22). At this point dyke 6 consists of three components each of which may become independent of the others and chill along both margins against the local gneiss. The largest of these three units is an inclusion-rich southern component, approximately 50 m wide with a chilled but highly irregular joint controlled contact against the local dioritic gneiss. The dyke host rock is a hornblende gabbro which does not appear particularly fresh in the field, probably due to autometa-

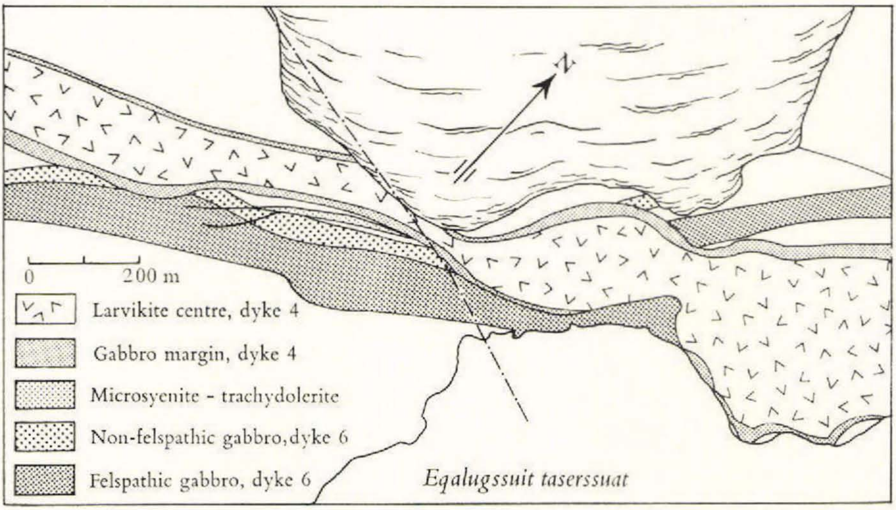


Fig. 22. Sketch map of dykes 4 and 6 at Eqalugssuit taserssuat.

morphism caused by the late accumulation of volatiles in the dyke fissure. Following the emplacement of the felspathic gabbro there was the intrusion of a 20 m wide gabbroic dyke. This may be traced westwards for several kilometres as an independent dyke but where it meets the felspathic dyke there is considerable difficulty in separating the two units. The contact between the two rock types is highly complex and it appears that the older inclusion-rich component of the dyke was still plastic when the second pulse of magma was introduced. The inclusion-free unit intrudes into the older rock without chilling against it; the contacts are generally non-planar and are very difficult to follow even on the ice-polished surfaces.

Closely associated with the two main components there was the intrusion of a microsyenite-trachydolerite dyke which reaches a maximum width of 5 m. This

Fig. 23. Inclusion-rich area in dyke 4 at Eqalugssuit taserssuat. Note the good size-sorting and the sharp though not chilled boundaries to the inclusion-rich area.

Fig. 24. Detail from the accumulation of felspathic material seen in Fig. 23. Most of the insets are either fragments of large felspar crystals or aggregates of felspar fragments. The compass is 6.4 cm in width.

chills against the country rock. It shows a distinct zoned structure with margins of fine-grained microsyenite containing rhombs of alkali felspar; an intermediate zone of alkali gabbro containing a few, rather altered, plagioclase megacrysts and a few alkali felspar rhombs; and a central zone with many black plagioclase megacrysts but no alkali felspar rhombs. The marginal syenitic zones are not constant along the length of the dyke. The age relationships between the microsyenite-trachydolerite and the main components of dyke 6 are not clear. Thin late veins of a dense almost glassy green-black material apparently connected with the microsyenite-trachydolerite cut the main felspathic component but have irregular contacts with the slightly later non-felspathic dyke. It appears that all three components were emplaced within a short time interval.

The felspathic component of dyke 6 contains a very large variety of fragments. Both xenoliths and megacrysts occur. There is a high proportion of large megacrysts, cleavage fragments, and secondary anorthosites formed from slightly crushed megacrysts surrounded by thin films of mafic material. Locally the dyke contains large amounts of clear, glassy megacrysts which make excellent museum specimens. The felspathic inclusions show an unequal distribution in the dyke, the margins may be free and the centre crowded or the inclusions may be concentrated in trains either near the centre or at one margin. There is frequently a good sorting of the

Fig. 25. Irregular distribution of insets in an alkali-gabbro. The large xenolith seen on the extreme right appears to have acted as some form of barrier so that there is a comparatively inclusion-poor zone behind it. The hammer-shaft lies parallel with the margin of the dyke. There is a slight preferred orientation of the more tabula plagioclase crystals parallel with the margin. The suggested direction of flow is towards the NE. Dyke 6, Eqalugssuit taserssuat.

fragments which is purely dependent on their size and not their textural variety (Fig. 24). This may lead to the larger fragments being concentrated near the centre and the smaller near the margin but frequently there appear to have been several phases of injection and the distribution of a particular size of inclusion in the dyke may be complex. No internal chilling has been noted between the different fractions found in the dyke, although many changes either between felspathic and non-felspathic or between hosts bearing one size of inclusion and hosts bearing another

Fig. 26. Anorthosite inclusions in olivine gabbro apophysis, locality 7, east of Kobberminebugt. Note the apparent attack by the host on the xenoliths releasing irregular inclusions (bottom right-hand corner of photograph). The attack seems to be concentrated on the interstitial mafic material in the xenoliths. Late xenolith-free dolerite apophysis cut the inclusions and their host. The pencil is 7 mm wide.

size of inclusion may be sharp and planar. Frequently particularly dense accumulations of inclusions occur in patches surrounded by comparatively inclusion-poor host rock (Fig. 23). Most of the fragments are approximately equidimensional and therefore show no preferred orientation, though there is some local evidence that lateral movement took place within the fissure. Fig. 25 shows a large xenolith which has apparently acted as a barrier round which small fragments have streamed. The sense of movement of the small fragments relative to the large appears to have been towards the north-east.

Locality 7 is from one of a small group of felspathic dykes which outcrop along the ice margin between the eastern end of Kobberminebugt and Eqalugssuit taserssuat. It is of particular note as it is one of the few places from which hypersthene has been found in a primary anorthosite xenolith. The sample (G.G.U. 61080) was taken from the mafic interstitial material found in a granular anorthosite the plagioclases of which show a poorly developed laminated texture (Fig. 3, p. 22). The mafic material occurs in pockets and consists dominantly of hypersthene, olivine, a little Fe-Ti oxide and possibly some clinopyroxene. The xenoliths are concentrated

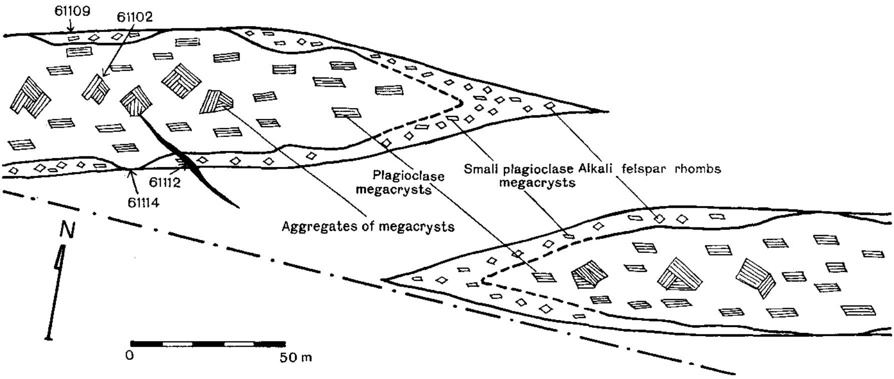


Fig. 27. Sketch to show the distribution of rock types found in dyke 8, NE corner of the nunatak 10 km east of Kobberminebugt. Note the early discontinuous margin of trachydolerite, the inclusion-rich centre of the dyke chilled against the trachydolerite margin, and the late inclusion-poor veins which can be followed out from the centre of the intrusion into the country rock.

in a 2 m apophysis of an olivine gabbro thought to have been intruded earlier than the majority of felspathic dykes in the area. There are few felspathic inclusions in the main dyke while the apophysis is choked with an ill-assorted mass of anorthosite xenoliths, plagioclase megacrysts, and occasional blocks of Julianehåb Granite. Many of the megacrysts are black or mauve while the xenoliths are white. Thin inclusion-free dolerite veins from the main dyke cut and chill against the apophysis suggesting that the felspar-bearing part of the dyke was intruded early. Some of the xenoliths show signs of chemical attack by the host. This is particularly strong in the vicinity of the granite inclusions suggesting that the host possessed greater power to attack anorthosite inclusions when it was contaminated by less basic material. Away from the granite inclusions the anorthosite xenoliths are generally fresh although the interstitial mafic material may be attacked by the host, releasing some plagioclase xenocrysts (Fig. 26). An analysis of an uncontaminated part of the apophysis is given in Table 4, No. 2. It is the most basic host analysed from the Kobberminebugt area.

Dyke 8 is composed of three en échelon dykes cutting the granite on the northern margins of the nunatak 10 km to the east of Kobberminebugt. It shows many features in common with dyke 6 although it is unlikely that there was ever direct connection between them, at least at the level of the present surface. Movement along the E–W transcurrent fault, separating the rocks on the nunatak from those further south, has destroyed the exact relationship between the two dykes.

Dyke 8 reaches a maximum width of approximately 80 m tapering out to 1 or 2 m at the end of each of the three en échelon components. Each component dyke has a discontinuous margin of microsyenite-trachydolerite with rhombs of alkali felspar and some plagioclase megacrysts. These margins vary between 2 m to a few centimetres. They are especially well developed where the main dykes thin out at the en échelon breaks. In a few places the rhomb-porphyry forms independent dykes close to the margins of the main body. The central gabbroic zones of the rock are demonstrably younger than the marginal porphyry as the basic rock sends out veins through the more alkali rock into the surrounding granite. In many exposures it appears that the rhomb-porphyry was partially stoped away when the main dyke

Fig. 28. Anorthosite xenolith from dyke 8 composed of an aggregate of crushed fragments. Some of the crushing appears to have taken place since the fragments were compacted together since there is a cleavage direction common to many of the fragments which does not follow a rational crystallographic plane in individual felspars.

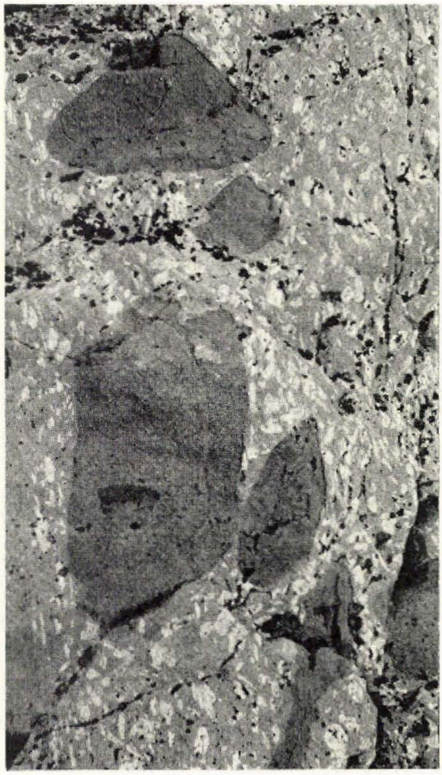


Fig. 29. Rounded xenoliths of an early non-felspathic phase of an olivine dolerite surrounded by the late felspathic phase found in the centre of the dyke. Note the protruding augite crystals in the triangular block. The outcrop shown measures 50 cm across.

was emplaced. A diagram of the relationships seen at the eastern end of the central unit is shown in Fig. 27. Three analyses of this dyke are given in Table 4, Nos. 3, 4 and 5.

Much of the outcrop of the dyke is covered in rubble of broken down host rock. The felspathic fragments tend to be less affected and large blocks of felspar are scattered along the strike of the dyke. They form between 20 and 50 % of the area of most fresh outcrops. Many of these felspars are the clear, glassy type described from dyke 6; other blocks show a variety of secondary anorthositic textures. Blocks composed of large, glassy felspar with a very small amount of gabbroic material between the individual plagioclases are common. Many show some signs of crushing (Fig. 28). Compacted fragments of felspathic material apparently derived by the mechanical breakdown of large plagioclase megacrysts are common (Fig. 8, p. 27).

Approximately 200 m north of dyke 8, close to the north-west corner of the nunatak, there is a series of thin dolerite dykes which contain scattered plagioclase

Fig. 30. 60 cm dolerite dyke (the photograph is taken looking along the trend of the dyke) in which there is a sharp junction between an inclusion-poor rock in the foreground and an inclusion-rich area further away in the picture. No chilling is seen between the two rock types which interfinger and which were apparently emplaced in the dyke fissure at the same time.

megacrysts. The main interest of the dykes lies in their mechanics of intrusion which appears to have been complex. The main dyke is approximately 5 m wide. It consists of a chilled, trachydoleritic, glassy contact zone 10–20 cm wide with a local spherulitic texture and some flow alignment of microlites. From 20 cm up to about 1.5 m from the margin the dyke is a moderately fine-grained olivine dolerite. The centre consists of a 2–3 m zone rich in plagioclase megacrysts. The felspathic part of the dyke has sharp contacts with the outer zones although no chilling is seen. In the central zone there are numerous rounded fragments of dolerite which resemble the surrounding non-felspathic part of the dyke except that they contain prominent augite crystals which protrude from the weathered surface (Fig. 29).

Surrounding the main dyke there is a swarm of thinner dykelets most of which are glassy and resemble the margins of the main dyke. One of the dykelets, which is approximately 0.6 m wide, contains chilled dolerite along half of its length and porphyritic dolerite along the other half. The junction between the two rock types is shown in Fig. 30. Neither rock type is chilled against the other and it seems that the most plausible explanation is that they were both introduced into the fissure at the same time but at slightly different places. Where they met the two magmas were only slightly miscible and interdigitated in the manner seen in the photograph.

Dyke 9 runs parallel and 200 m to the south of the western end of dyke 6. It seems highly likely that it was intruded during the same phase of dyke activity as where it has not been contaminated by the assimilation of granitic material it shows many of the same feature as dyke 6. It is probable that the dyke represents one of the dolerites running north-eastwards from the Bangs Havn complex. The dyke varies from 50–100 m wide in the west to a maximum of 500 m in the east. Felspathic inclusions are rare in the western half of the dyke (mapped by J. S. WATTERSON) but Julianehåb Granite xenoliths are abundant and differ from the basement granite seen adjacent to the dyke. Since all the granite xenoliths in the dykes described by WATTERSON from the **Ilordloq area** are similar in type they have probably been derived from one extensive deep-seated mass. In dyke 6 they are usually 20–50 cm wide and rounded in outline. They comprise about 10–20 % of the dyke, are leucocratic, medium- to coarse-grained and seldom foliated, although some are highly foliated and presumably derived from mylonite zones.

The granite xenoliths show various degrees of assimilation by the dolerite, which preferentially attacked the felspar of the granite. Quartz, forming 20–40 % of the granite, has not been dissolved significantly—if at all—by the dolerite magma. Where assimilation was most intense the dolerite is comparatively pale in colour.

An inclusion of anorthosite occurs in this assimilation zone but, in contrast to those described on p. 46, shows no marked evidence of reaction with its host.

The various types of felspathic inclusion are haphazardly distributed and all can be closely associated. The majority of xenoliths are even-grained and massive with felspar forming about 95 % of the volume of the rocks. A few xenoliths occur with tabular felspars 5–8 cm long, 0.5 cm thick, aligned in parallelism. In two blocks the mafic minerals were aligned in trains giving a pronounced banding to the xenoliths. Secondary anorthosites, similar to those described from Bangs Havn, have been noted and a few xenoliths in which magnetite forms over 20 % of the rock may be of secondary origin. Some xenoliths contain plagioclase crystals up to 40 cm long; these are thought to be of the same type as the aggregates of large megacrysts described from dyke 8.

At the eastern end of the ground mapped by WATTERSON the dyke broadens out to approximately 150 m. As the dyke widens there is a gradual increase in the size and volume of felspathic inclusions, and an increase in the proportion of xenoliths to megacrysts. WATTERSON has estimated that in places the xenolithic content exceeds 85 % of the dyke volume with some of the xenoliths measuring 25 m long. Felspathic inclusions may occur throughout the width of the dyke or be confined to a central zone abruptly separated from a marginal zone of dolerite almost devoid of inclusions.

Directly west of the neck of land separating **Augiâta tasia** from the arm of the sea the dyke contains a central granite zone 80 m wide intruding the gabbro. The granite can be followed for 50 m along the strike of the dyke then, after a concealed tract of 40 m, dolerite is seen for 20 m highly contaminated by assimilation of felspar from granite xenoliths.

East of the sea inlet **Augia** the dyke broadens out gradually until it reaches a maximum of 450–500 m after which it decreases rapidly, finally disappearing beneath a lake. Unlike the major felspathic dykes described in the preceeding paragraphs it does not continue eastwards, the shape is thus roughly that of a match stick. The bulbous north-east end of the dyke, which measures approximately 3 km by 450 m, appears to have formed by the choking of a gabbro with large xenoliths so that further lateral extension of the dyke was impossible.

The relationship between the host dyke and the anorthositic xenoliths in the bulbous end of the dyke is complex. The core consists mainly of massive granular anorthosite composed of closely packed 2–3 cm plagioclase grains with a negligible mafic content. The margins consist of a fresh gabbro which often shows some lamination in the host felspars. This lamination dips at a low angle inwards from the margins of the dyke. The gabbroic margins vein into the centre. Detached fragments from the anorthosite core occur within the host and form the normal xenolithic material seen in the westerly extension of the body. In some places the anorthositic core has been thrust against the country rock, in others there is a gabbroic mantle which may locally be free of xenolithic felspar. Evidence for the method of emplacement of this body is scanty; the anorthosite appears to have been brought up as one or two massive "plugs" which have broken down during transport. There is nothing to suggest that the present relationships are due to an accumulation of smaller fragments within the host itself.

Like many other of the felspathic dykes, dyke 9 is associated with several fine-grained microsyenites or fine-grained alkali trachydolerites which contain rhombs of heavily zoned alkali felspar. These are seen close to the northern contact of the bulbous mass; their exact field relationships are unexposed.

The only other xenolith approaching the size of the blocks of anorthosite from Augiâta tasia occurs in a dyke approximately 5 km to the east (**locality 10,** Plate 4). Here a single block of fresh granular anorthosite measuring approximately 20 by 30 m is found in an uralitised dolerite belonging to the mid-Gardar swarm. The occurrence is unusual as the dyke contains no other felspathic xenoliths (apart from small fragments obviously derived from the large block). The block resembles the masses from dyke 9 both in texture and composition. The dyke sends anastomosing veins into the anorthosite block, apparently assisted by a chemical attack on the interstitial mafic material.

Dyke 11 outcrops for 17 km in the areas on either side of **Isortoq**. It may continue further to the west than shown on Plate 4 but it loses its characteristic plagioclase megacrysts in the west of its outcrop and cannot be traced with certainty across the sea inlet Augia. It contains a smaller proportion of felspathic material than many dykes and is only used as an example of one of the many transitional dykes between the felspathic rocks and Gardar intrusions with no insets. If it were not for the presence of the more spectacular felspathic rocks packed with inclusions, the megacrysts could be regarded as phenocrysts without discussion.

Where it is typically exposed on the shores of Isortoq the dyke is 20 m wide. It chills against an older gabbro. It contains scattered plagioclase megacrysts up to 30 cm long, often with a moderately well-developed tabular form. The outer 2 m of the dyke are free of felspar insets. At the margins of the central zone the insets are generally less than 1–2 cm in length and form about 1 $^0/_0$ of the dyke rock, nearer the centre the insets increase in size and total volume until they average about 5 $^0/_0$ of the rock. Many of the felspars show a distinct alignment with the dyke margins suggesting that their distribution is controlled by flow within the dyke fissure. Most of the felspars are clear but scattered examples show glassy, black cores rimmed by clear material. The majority of the felspar insets are megacrysts or cleavage fragments, which often contain mafic inclusions along the (010) plane. A few of the felspathic inclusions are anorthosite xenoliths; these average about the same size as the megacrysts while their constituent plagioclase grains are considerably smaller than the megacrysts.

Dyke 12 is the only major microsyenite-trachydolerite dyke mapped in the ground between Kobberminebugt and Qaersuarssuk. The field relations are described in some detail as it appears that it can be used as a type example for the numerous occurrences of this group of felspathic rocks which occur further east. It extends for

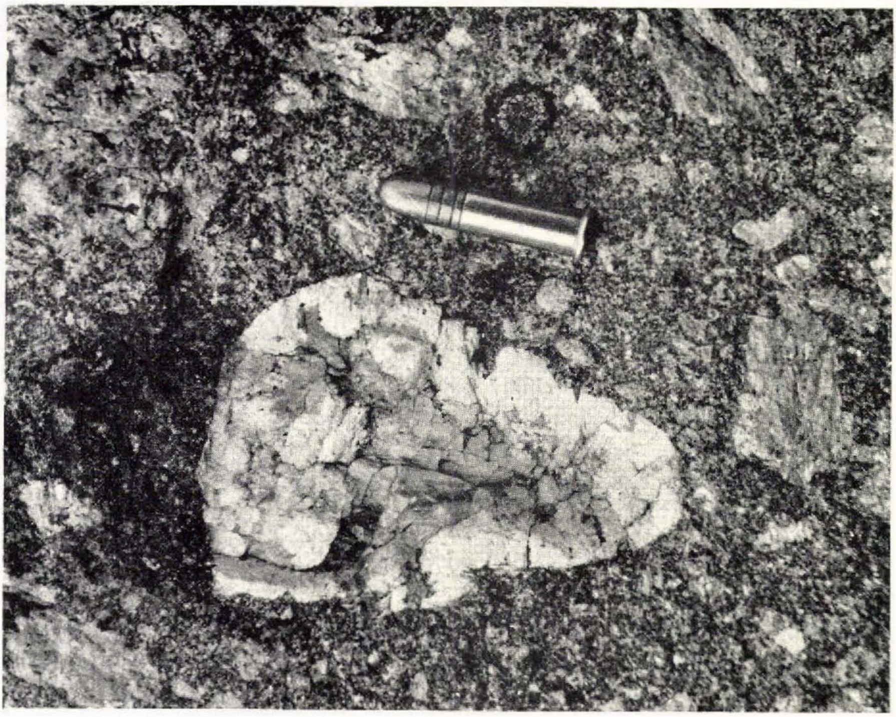


Fig. 31. A megacryst of plagioclase in a trachydolerite dyke. Note the fresh labradoritic centre and the altered rim. The bullet is 6 mm in diameter.

30 km from Usua to the east of Kûtsiaq where it disappears under the Inland Ice. Several en échelon breaks have been mapped. However, the petrology of the host rock is so distinctive that it may be treated as one unit.

The host consists of two main components; an outer marginal facies of fine-grained gray-green to black microsyenite set with scattered white or pink rhombs of alkali felspar and very few felspathic inclusions; and a central zone generally occupying at least 80 % of the dyke fissure consisting of a felspathic, highly alkaline gabbro. No chilling is seen between the two components although the boundary between them is generally abrupt. The marginal microsyenites are generally equal in width on either side of the central zone; their width shows no fixed relationship to the total width of the dyke, so that in the west where the dyke thins to under 10 m they occupy a third or more of the fissure while they are often only 1–2 m wide where the dyke reaches its maximum width of 35 m to the west of Isortoq. The marginal microsyenite sends veins and apophyses into the surrounding granites and may continue for some way into the country rock without a central felspathic core where the dyke thins at en échelon breaks. Although there is no chilling between the two components it is clear that the margins solidified before the emplacement of the dyke centres. Occasionally the centre intruded through the margins and chills against the country rock and there are some inclusions of the marginal facies in the more basic centre. At a few localities the marginal rocks appear slightly sheared before the intrusion of the centre; this is thought to be due to effects during the emplacement of the dyke material rather than due to a regional movement affecting the early part of the dyke. Chemical analyses of the margins and the central parts of the dyke

Fig. 32. A sample of trachydolerite with altered plagioclase megacrysts (light) and alkali felspar rhombs (dark, fretted crystals with light margins). Loose block, ice-margin near dyke 12. The bullet is 6 mm in diameter.

are given in Table 4, Nos. 12–21 inclusive. A block diagram, based on the relationships seen in this dyke is shown in Plate 3.

The distribution pattern of felspathic inclusions within dyke 12 is moderately constant along its strike although both the average size and total volume of the inclusions vary from place to place. Typically the marginal zones are between 1 and 2 m wide although in places they may reach 4 m. They are free of felspathic material, except for a few scattered patches of plagioclase crystals which are often larger and differ in general appearance from the inclusions in the central zone.

The central zone shows a gradual increase in the amount and size of the inclusions away from the margins; there is commonly a moderate preferred orientation of the fragments so that the tabular megacrysts are aligned parallel to the dyke walls. The average size of the inclusions in the central zone varies along the strike; there is a tendency for larger, fresher inclusion in the centre of the dyke where it reaches its maximum width. In the areas where the average size of inclusion is greater than 10 cm there is a larger proportion of xenoliths to megacrysts than elsewhere.

The felspathic inclusions from dyke 12 have a characteristic appearance which distinguishes them from the inclusions in other dykes from the same area. Compared to the wide range of xenoliths from basic hosts the inclusions from microsyenite-trachydolerite dykes tend to be smaller, better sorted, and to contain a higher pro-

Fig. 33. Loosely-packed plagioclase megacrysts forming a pseudo-ophitic aggregate. The interstices between the plagioclase crystals are filled with material from the host dyke, augite, olivine, and Fe-Ti oxides. Locality 13, ice margin 15 km east of Kobberminebugt.

portion of megacrysts to xenoliths. Many of the fragments are altered to a chalky white sericitised plagioclase which stands out against the host rock. This is particularly noticeable if the felspars have non-altered centres (Fig. 31). Although alteration is common in inclusions from basic hosts it is rarely so distinct as that seen in the microsyenite-trachydolerite hosts. The alteration is particularly marked in the small subhedral plagioclase megacrysts characteristic of the outer parts of the central zone. Rhombs of alkali felspar, some of them with centres of sericitised plagioclase are often found in the outer parts of the central felspathic component of the dyke (Fig. 32).

Two felspathic dykes are found in close proximity at **locality 13**. These trend parallel to each other WSW from the ice margin at the head of **Kûtsiaq**. The northern dyke of the pair is a normal dolerite with one large mass of anorthosite over 10 m long and few other xenoliths. The dyke is heavily sheared in a NE–SW fault zone along the ice margin.

The southern dyke runs south of the fault zone; it possesses a very distinctive appearance over much of its north-eastern outcrop due to the partial or complete assimilation of blocks of Julianehåb Granite which have contaminated the dolerite. To the south-west the dyke can be followed through the ground mapped by K. Coe as a normal dolerite without inclusions of either granite or anorthosite. At the mouth of Isortoq the dyke runs in close proximity to a giant syeno-gabbro dyke which sends alkali veins into it. It cuts an early Gardar NW-trending dolerite and both dykes were sheared by a NE fault before the intrusion of the syeno-gabbro. Anorthosite fragments are concentrated in two sections of the dyke; within 200 m of the ice margin and 1.5 km to the south-west. In both these sections the contamination of

Fig. 34. Detail of pseudo-ophitic texture, locality 13. A single interstice is filled with host dyke rock, sub-ophitic olivines and Fe-Ti oxides moulded onto the plagioclase. Note the inclusions of host dyke and dark minerals along the 010 plane of the felspars.

the host rock with inclusions of Julianehåb Granite is lower than in the surrounding anorthosite-free sections of the dyke. This would suggest that the dyke picked up either anorthosite or granite but not both at any particular point along its strike. Near the ice margin felspathic inclusions account for approximately 30 $^0/_0$ of the exposed dyke. These felspathic inclusions show complicated relationships with the host rock which have an important bearing on the interpretation of textures seen in the inclusions from other dykes. The felspathic material occurs as loosely compacted subhedral plagioclase grains between which the normal dolerite of the host has solidified giving a coarse-grained "pseudo-ophitic" texture (Fig. 33). Some of the interstices between the grains contain large crystals of ilmeno-magnetite, olivine and augite all of which may surround the plagioclases ophitically or subophitically. In many instances a single interstice is filled partly with a single large grain of one of the above minerals and partly by the host dolerite (Fig. 34). The mafic minerals may show signs of instability in their host. The plagioclase grains contain trains of mafic mineral or dolerite aligned parallel to the (010) twin plane.

The formation of this texture from the granular anorthosites, which are regarded as representing the nearest approach to the original anorthosite preserved in the xenoliths, can be seen in Fig. 11, p. 28. This photographs show a slightly laminated granular anorthosite block composed almost entirely of tabular plagioclase crystals averaging approximately 0.3×1 cm aligned approximately parallel to the pen. The granular anorthosite is overgrown by subhedral plagioclase crystals projecting into the host dolerite. There seems little doubt that these crystals have grown

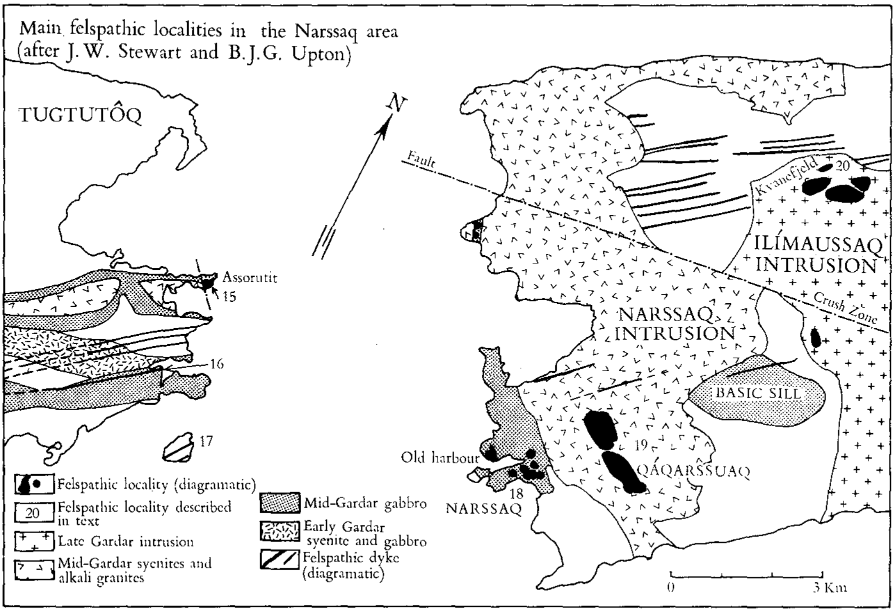


Fig. 35. Sketch map of the important anorthosite-bearing rocks in the Narssaq area.

outwards from the earlier block due to the precipitation of new material from the host dyke. Some of the overgrowth crystals project back into the laminated block which may suggest either a partial recrystallisation of the earlier texture or that some individuals have acted as preferential sites of plagioclase crystallisation in the new environment.

iii) Localities between Tuĝtutôq and Ilímaussaq

A sketch map of the important localities in the **Narssaq-Tugtutôq** area is shown in Fig. 35, compiled from maps drawn by B. G. J. Upton and J. W. Stewart with a few extra observations made by one of us (D.B.). The regional geology is described by Upton (1961, 1962 and 1964). The most important locality described by Upton is at **Assorutit** on the eastern end of Tugtutôq (locality 15, Fig. 35). This has already been described on p. 24 as it is the type locality of the laminated anorthosites. Further information is to be found in the petrological section of this paper and in the papers by Upton cited above.

The swarm of microsyenite-trachydolerite dykes which pass through **Tugtutôq** on to the **Narssaq peninsula** provide considerable information about the mode of emplacement of these dykes and the relationship between the anorthosites and their hosts. One of them, a 16 m ENE–WSW dyke is described on p. 141 in the section on the petrology of the Gardar hosts (**locality 16**, Fig. 35). The second, a small dyke on an island to the south of Tugtutôq (**locality 17**, Fig. 35), is of particular interest because of the large variety of felspathic fragments included in the central zone. These are illustrated in Fig. 36 and by Upton (1962, fig. 12, p. 34; 1964, fig. 12, p. 32). The host consists of a 10 m wide central zone of trachydolerite choked with inclusions and a narrow discontinuous marginal zone of microsyenite set with alkali felspar rhombs. The central zone is more clearly divided from the margin than in most members of this group of dykes and has broken through the margin to chill against

Fig. 36. Accumulation of felspathic fragments in a trachydoleritic matrix, locality 17, Fig. 35. Note the variable textures shown by the xenoliths. Above the hammer to the left there is an irregular accumulation of plagioclase-rich granular fragments surrounded by a slightly more mafic anorthosite. Above the hammer to the right there is an overgrowth of megacrysts on a loosely-packed aggregate.

the country rock in several places. The inclusions range from xenoliths of fresh laminated anorthosite similar to those seen at Assorutit, masses of granular anorthosite, single megacrysts up to 20 cm long with inclusions of gabbroic material along the (010) plane, and compound blocks made up of comminuted felspar grains set in an indeterminate matrix. Blocks made up of loosely packed megacrysts set pseudo-ophitically in the host gabbro are quite common. Some instances of large felspars radiating from a finer grained inclusion of earlier anorthosite of the type described from locality 13 have been noted. The majority of fragments, with the exception of the scattered xenoliths of laminated anorthosite, have been severely altered. Fresh centres are found in some of the megacrysts.

The **Narssaq gabbro** (**locality 18**, Fig. 35) contains a vast amount of felspathic material varying from the almost ubiquitous plagioclase megacrysts to masses of granular anorthosite measuring tens of metres in diameter similar to those described from Kobberminebugt. In many places the smaller fragments have been incorporated into the fabric of the host and it becomes very difficult to be certain whether some of the felspars are derived from the breakdown of anorthosite or whether they are true phenocrysts separated from the host magma. In many cases the megacrysts and blocks of granular anorthosite are surrounded by a rim of chlorite believed to have been derived by the alteration of olivine (Upton, 1964, plate 3, fig. 2).

The distribution of felspathic material within the Narssaq gabbro is highly irregular. Between Narssaq old harbour and the southern contact of the body there is a general concentration of material while to the north-west there is comparatively little. The south-east contact of the gabbro against the over-lying Gardar sandstones shows a 8 m marginal facies in which there is a fairly high percentage of megacrysts, the actual contact of the dyke is rather sheared and apparently devoid of felspathic material.

One of the commonest features seen within the gabbro is the formation of secondary anorthosites similar to those described from Bangs Havn (p. 39). A particularly well-preserved example of this phenomena is seen on the north shore of Narssaq old harbour where a large anorthositic gabbro mass is surrounded by gabbro with few other xenoliths. The block is made up of two components: broken masses of plagioclase megacrysts and xenoliths, and a surrounding host of very coarse-grained gabbro in which the felspars are distinctly tabular and show local lamination similar to that seen in the Assorutit mass. An analysis of this composite block is given in Table 1, p. 72, sample 49320. Similar secondary blocks are found in many places in the gabbro and there is every gradation between blocks made of granular anorthosite fragments compacted together with little or no interstitial host material and more mafic "anorthositic gabbros" in which megacrysts and blocks of granular anorthosite are incorporated into a secondary host. In the latter case it is quite common to see augite and olivine enclose a derived plagioclase megacryst subophitically. In many places the coarse gabbro forming the host in these secondary xenoliths resembles the coarser grained felspathic areas within the Narssaq gabbro itself. Some of the composite blocks are surrounded by a fine-grained variety of the host suggesting that they were cold when they were included; other blocks show gradational contacts in which the inclusions from the secondary anorthosite are released into their new host rock. Many of the composite blocks show signs of plastic deformation. It appears that the matrix of the secondary anorthosite was formed as an early phase of the Narssaq gabbro and was locally in an unconsolidated state before it was included in the main mass of the intrusion. It is reasonable to suggest that the two rocks are closely connected genetically; however, this does not necessarily imply that the granular anorthosites themselves are likewise closely related to their present host.

In addition to the Narssaq gabbros there are several other significant localities of felspathic rocks in the area mapped by J. W. Stewart on the western end of the Narssaq peninsula. The most prominent of these is a group of outliers of inclusion-rich gabbroic rocks capping the hills to the east of Narssaq (**locality 19**, Fig. 35). Ussing (1912, p. 206) noted the presence of these rocks which he termed the **Qáqarssuaq** (Kakarsuak) essexite-porphyrite. Ussing's analysis of a sample from one of these masses is given in Table 1, p. 72. The relationship between the anorthosite-bearing rocks and the nearby basalts of the Gardar continental series is obscure although it is generally presumed that the felspathic gabbros are later as they are not conformable with the regional dip of the basalts.

The main outcrop of the felspathic rock occurs on the 681 m peak east of Narssaq. The exact inclination of the body, which appears to be a sheet, is unknown as the top has been removed while the base forms the roof of the Narssaq intrusion. The present outcrop suggests that the body dips to the north-west at approximately 30°. It has a minimum thickness of 100 m. The base of the body has been heavily attacked by the syenite and alkali granite and blocks of anorthosite and plagioclase megacrysts are found for a few tens of metres within the younger rocks. The younger rocks develop a porphyritic marginal facies close to the contact which might be

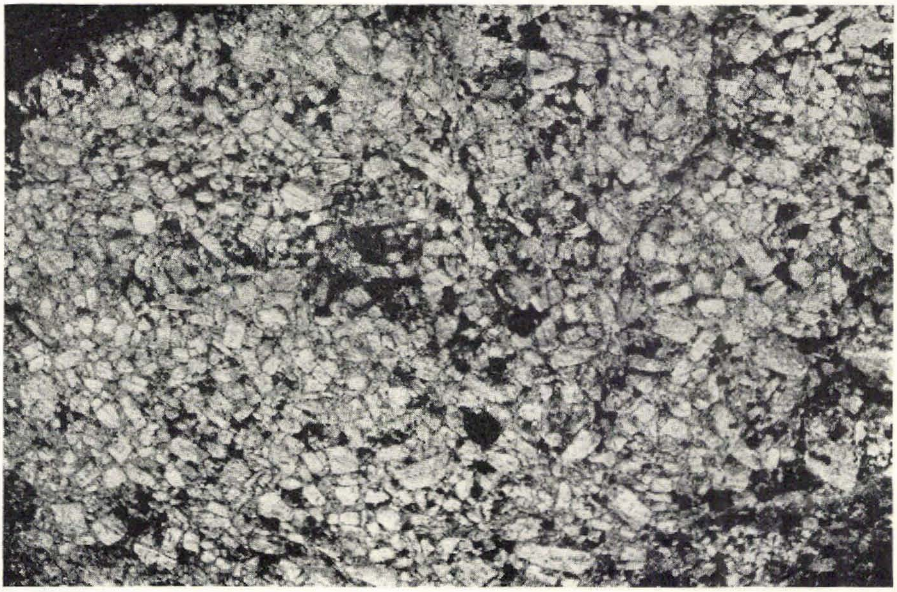


Fig. 37. Texture of an anorthosite block from Kvanefjeld (locality 20, Fig. 35). The anorthosite is clearly a secondary accumulation of megacrysts and derived fragments in an altered gabbroic matrix. It resembles the secondary anorthosites at Bangs Havn (Fig. 21) and Narssaq. The lens cap measures 5 cm in diameter.

caused by the partial assimilation of plagioclase and the seeding of rhombic felspars round small xenocrysts of plagioclase, or might reflect the low density of the alkali felspars which rose to the top of the syenite before it solidified. Similar features are found in the microsyenite-trachydolerite dykes.

It seems quite likely that the present boundary with the younger rocks is close to the original base of the sill as there is a characteristic basal facies developed. This consists of closely packed plagioclase megacrysts generally less than 1 cm in diameter. Away from the contact the megacrysts are larger and there is a gradual increase in the proportion of xenoliths of granular anorthosite upwards. At the top of the present exposure the rock consists of large masses of granular anorthosite and secondary anorthosite similar to those described from the Narssaq gabbro. The ratio of groundmass to included material does not change appreciably from bottom to top suggesting that the sorting is due to segregation of the coarse material towards the centre of the original sheet rather than gravity flotation of felspars. The groundmass has been altered too much to be certain of the original composition; USSING's analysis suggests that it might have been fairly alkaline although this could be due to late contamination. The secondary anorthosites contain prominent crystals of augite, and olivine now generally altered to chlorite.

Two smaller occurrences of xenolith-rich material similar to the Qáqarssuaq sheet are found as rafts within the lujavrites on **Kvanefjeld** where lujavrite penetrates the augite syenite margin of the late Gardar **Ilímaussaq intrusion** and invades the overlying basalts (**locality 20**, Fig. 35).

The felspathic material in the rafts resembles the inclusions in the Narssaq gabbro and it seems quite likely that the rafts and the "sill" at Qáqarssuaq represent

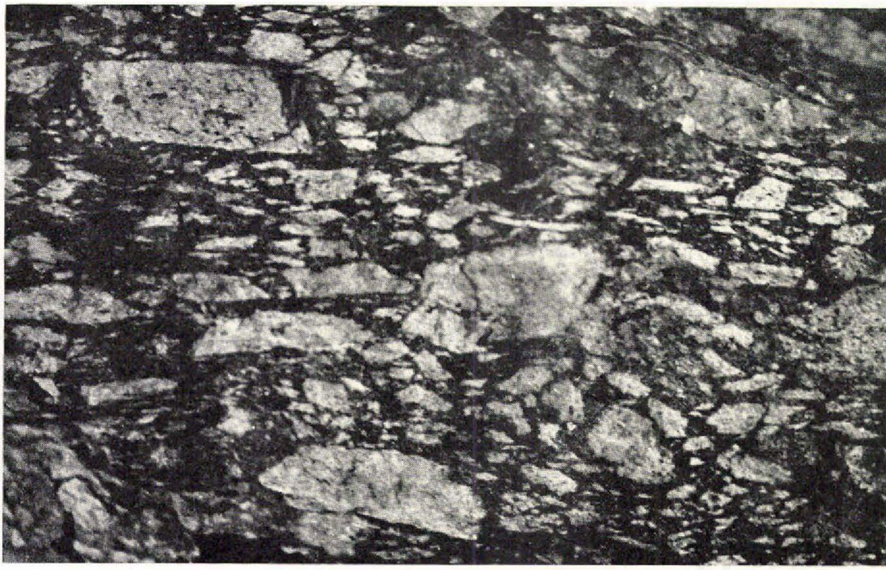


Fig. 38. Felspathic fragments in the lujavrites on Kvanefjeld. The majority of fragments are broken megacrysts, apparently derived from an anorthosite block of the same type as illustrated in Fig. 37. The gabbroic matrix has been completely replaced and the megacrysts have been aligned by movement in the lujavrites. The megacrysts are strongly sericitised. The largest crystal measures approximately 6×2 cm.

easterly extensions of the gabbro which may have lost its dyke form completely further east. However, the later intrusions have so broken up the early, and early mid-Gardar rocks in this region that strict correlations are not possible. Secondary anorthosites, laminated anorthosites, and a few blocks of granular anorthosite have all been noted; however, the majority of fragments are single broken plagioclase megacrysts averaging 5×2 cm set in an altered gabbroic matrix (Fig. 37). The lujavrites below the raft contain numerous plagioclase xenocrysts frequently together with other inclusions from older rocks in the Ilimaussaq complex (for example, the augite syenite and naujaite). The plagioclase xenocrysts appear to be relatively stable physically and characteristic fragments are found at a considerable distance away from the raft from which they were probably derived. Chemically, however, the xenocrysts show a complex sequence of processes in which they are gradually replaced by albite. Close to the margin of the Ilímaussaq intrusion the plagioclase xenocrysts are frequently involved in the late movements which affect the lujavrite apparently while it was still in a semi-magmatic state (see Ferguson, 1964). The first effect is seen in Fig. 38; the xenocrysts are aligned so that they are approximately parallel with the local schistosity in the lujavrites. This is followed by a stretching out of the fragments (Figs. 39 & 40) so that they resemble augen in augen gneiss. The final stage is the production of a rock resembling a banded gneiss in which material derived from plagioclase xenocrysts forms layers within the lujavrite.

Isolated fragments of anorthosite are found within the augite syenite margin of the **Ilímaussaq intrusion** (Hamilton, 1964, p. 20) and in the lujavrites down to 400 m below the roof of the intrusion. It is possible that some of these fragments

Fig. 39. An adjacent outcrop to Fig. 38 showing the progressive mechanical breakdown of felspar fragments in the lujavrites. The largest fragment measures 10×3 cm.

Fig. 40. A further stage in the mechanical breakdown of plagioclase megacrysts in lujavrite. About 0.5 × natural size.

may have been brought up from depth by the alkaline rocks. STEWART suggests (1961, personal communication) that the rafts found on Kvanefjeld represent stoped roof blocks or pendants which have only been transported a short way within their present host and this seems the most likely explanation for the position of these large blocks since they are relatively unaffected by their host compared to xenoliths found further away from the contacts of the Ilimaussaq intrusion. A similar explanation is used by HAMILTON (1964) for inclusions close to the margin of the augite syenite.

iv) Localities in the vicinity of the Igaliko complex

At **Narssârssuk (locality 22**, Plate 4) considerable felspathic material is found in the marginal rocks of the youngest phase of the Igdlerfigssalik centre, the most westerly of the four intrusion centres comprising the Igaliko syenite. The felspathic inclusions are concentrated in the marginal, dark, biotite-rich variety of the syenite; in a locally developed layered syenite outcropping on the shores of Tunugdliarfik; and especially in a porphyritic microsyenite sheet which intrudes most of the other marginal rocks.

The inclusions in the marginal, dark, biotite-rich syenite are mainly plagioclase megacrysts which average 3–5×2–3 cm and may reach 10 cm long. In common with the plagioclase megacrysts in the nearby syenites and in the marginal alkali gabbro of the Klokken intrusion these felspars show very regular albite and Carlsbad-albite twinning which may die out towards the edges of the crystal. In some cases the plagioclase megacrysts are included direct in the biotite-rich syenite while in other cases they were first included in a gabbroic rock which was itself caught up in the syenite. The gabbroic host retains its texture although all the plagioclases are strongly altered and contain no anorthite. Occasionally the felspar crystals in the altered gabbro show lamella twinning. It is possible that a gabbroic marginal facies was developed at Narssârssuk which contained copious plagioclase megacrysts similar to that described below from the Klokken intrusion. The altered gabbro commonly contains copious epidote presumably derived from the breakdown of calcic plagioclase. The plagioclase megacrysts appear to have been more resistant to alteration than the surrounding rock. Many of the crystals show a zoned structure with a centre of mat black felspar surrounded in turn by glassy black and white felspar mantles and a rim of altered white sericitised material in which the twinning of the original plagioclase persists although all the calcium has been removed. Most of the smaller megacrysts of plagioclase in the syenite itself are glassy and black. Some are apparently unaffected by their host while others are surrounded by a rim of nepheline and alkali felspar which appears to be replacing the earlier material. Large blocks have a mat black lustre, a feature which is peculiar to the locality and the felspars collected from the marginal facies of the Klokken intrusion.

The layered and laminated syenite in the cliff overlooking Tunugdliarfik contains a xenolith of anorthosite over 6 m long round which the host felspars are aligned. Smaller inclusions consisting, among other rocks, of anorthosite and material derived from felspathic dykes, are found in the same syenite variety at the mouth of Qôroq where they disturb layering in the host rock below them but are overlain by undisturbed material.

The porphyritic microsyenite, which intruded the slightly earlier syenites at Narssârssuk, contains copious inclusions of anorthosite and plagioclase megacrysts. The largest anorthosite mass, which measures 50 m long by 20–30 m wide, lies with its long axis parallel to the trend of the microsyenite sheet although the exposures are not sufficient to show whether it is an inclusion in the latter. It consists of undeformed rectangular white to pale gray plagioclase crystals 1×0.2 to 7×2 cm occasionally 10×3 cm in size, sometimes with cores of black felspar with a vitreous lustre. The gravel covered surface which surrounds Narssârssuk is scattered with many large blocks of plagioclase apparently released by preferential weathering of their syenitic host. Individual fragments of single crystals measure up to 30×10×10 cm. Many of the plagioclase crystals show signs of brecciation before their inclusion into the syenite host. Although they cannot be regarded as cognate inclusions in the syenite there is the possibility that they are not from the same source as the plagioclase-rich rocks brought up by the felspathic dykes and may be closely genetically connected with their host.

A further large block of felspathic material has been found more centrally in the younger syenites of the Igdlerfigssalik centre. Here, rounded or oval inclusions of coarse-grained rock composed largely of white plagioclase crystals are found in syenite surrounded by plagioclase megacrysts apparently derived from the larger masses. Some of these megacrysts are black with a dull lustre.

The spectacular layered syenite intrusion of **Klokken** (Plate 4, **locality 23**) lies approximately 5 km south-west of the margin of the Igdlerfigssalik intrusion centre. It contains a considerable amount of megacrystic plagioclase which closely resembles the material seen in the syenitic marginal members of the Igdlerfigssalik centre at Narssârssuk.

The plagioclase insets are almost entirely confined to the marginal alkali gabbro in the south and west exposures of the body where they form an arcuate mass parallel to the contact. Within this mass the plagioclase insets may exceed 80 % of the total rock set in a subsidiary gabbroic groundmass. Individual insets average 10–20 cm long although a few measure up to 1 m. Most of the insets are cleavage fragments which show local signs of brecciation or straining. Typically the plagioclase is dark gray or black with a mat lustre and evenly spaced regular albite twinning. Many of the fragments are surrounded by white material; in some cases this is due to bleaching without much effect on the composition of the mineral, in other cases it is due to the formation of secondary products.

K. Ellitsgaard-Rasmussen, who mapped the Klokken intrusion, has also noted the presence of large blocks of anorthositic material which show good lamination of the felspars and a distinct layering included in the margin gabbro. These xenoliths have many characters in common with the laminated anorthosite from Assorutit. A few anorthosite inclusions and plagioclase insets are found towards the outer contact of the syenite centre of the intrusion; these are generally more altered than those found in the gabbro.

In common with the material described from the Igdlerfigssalik centre it is not known whether the plagioclase insets in the Klokken intrusion have any genetic connection with the felspar xenoliths in the Gardar dykes.

On the south-east coast of **Qôroq** (Plate 4, **locality 24**) one of the early members of the Igdlerfigssalik centre is cut by a felspathic dyke trending 005°. The dyke is 8 m wide at fjord level but narrows upwards. The margins are aphyric while the centre of the dyke contains an assortment of felspathic material including anorthosite xenoliths and occasional large plagioclase crystals up to 0.5 m in length. There are also loose aggregates of moderately euhedral 5–10 cm plagioclase crystals set pseudo-ophitically in material indistinguishable from the surrounding dyke.

The majority of felspathic dykes cutting the older centres of the Igaliko syenite belong to the microsyenite-trachydolerite group. Several of the dykes cutting the **Qôroq syenites** (Plate 4, **localities 25 & 26**) show changes in the type of felspathic material at different places along their strike. Megacrysts are commoner than xeno-

Fig. 41. Polished specimen of megacrysts in a trachydolerite host from Jespersen Dal (locality 29, Plate 4). Note the pseudo-ophitic texture with the host chilling locally against megacryst and the inclusions of fine-grained host rock with the megacryst. The sample figured measures 12 cm across.

liths although rounded felspar aggregates with little interstitial mafic material may be quite common. One sample of a felspathic inclusion from this group of dykes is illustrated in Fig. 55, p. 97.

One of the two felspathic dykes cutting the western end of the **Motzfeldt centre** (Plate 4, **locality 27**) is composite. It has a total width of 12 m consisting of a 3 m central zone of felspathic, contaminated olivine dolerite surrounded by symmetrical margins of fine-grained porphyritic syenite. The contact between the two rocks is sharp but with no chilling and with no xenoliths of one rock type in the other. The number of alkali felspar phenocrysts in the marginal zone increases inwards.

Two felspathic dykes striking parallel with the regional trend at 050° cut the syenites north of **Motzfeldt iptasia** (Plate 4, **locality 28**). One of these, which is approximately 7 m wide, is composite with 1 m dark gray-green trachytic margins symmetrically surrounding a felspathic centre composed of contaminated olivine dolerite. The felspathic material is almost entirely composed of plagioclase megacrysts averaging 3–5 cm long with occasional crystals reaching 15 cm. Alkali felspar rhombs showing violent zoning, and resembling the phenocrysts found in the larvikite syenite dyke (Fig. 66, p. 131) are found in the central contaminated olivine dolerite.

South of the Igaliko complex, in the ground surrounding **Jespersen Dal** mapped by Steen Andersen, many of the dykes carry inclusions of felspathic material. The dyke swarm trends in general NE, most of the dykes are microsyenites or trachydolerites. Many are probably the continuation of the swarm mapped by J. H. Allaart on the Igaliko peninsula.

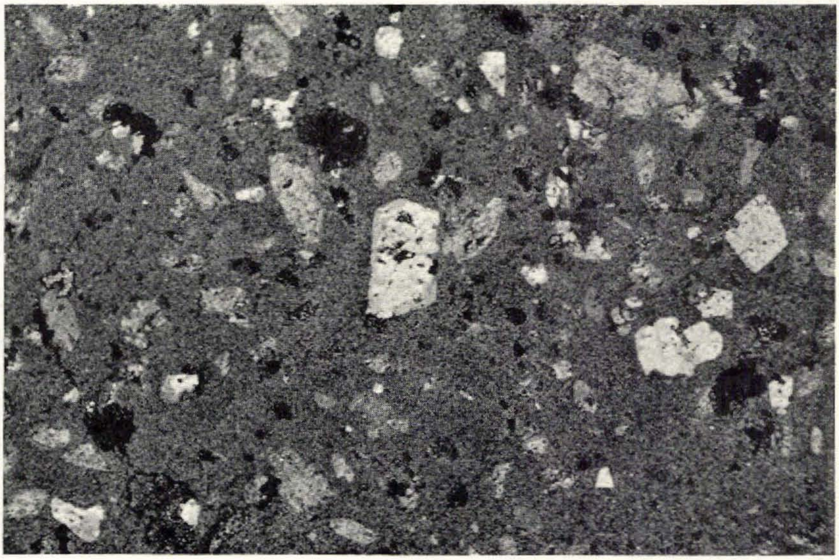


Fig. 42. Microsyenite dyke from Narssarssuaq. The dyke contains alkali felspar rhombs and scattered sericitised plagioclase megacrysts rimmed by fresh alkali felspar. This is a typical example of the least-felspathic end members of the dykes described in this paper. The megacryst measures 2×1 cm.

Fig. 43. Felspathic trachydolerite Narssarssuaq. Note the good flow orientation of the megacrysts. The outcrop figured measures approximately 25 cm across.

The most interesting of the dykes is a 3 m wide trachydolerite member of the NE swarm (Plate 4, **locality 29**). This dyke contains numerous inclusions of several rocks among which xenoliths of an older rhomb-porphyry dyke, a fine-grained trachytic rock, and anorthosite fragments are particularly abundant. The anorthosite is composed mainly of 10–20 cm subhedral crystals of altered plagioclase forming loose aggregates up to 1 × 1.5 m diameter set pseudo-ophitically in a trachydolerite groundmass. The felspars commonly contain inclusions of host rock sub-parallel with the (010) cleavage direction (Fig. 41). In some cases these inclusions of host rock are in continuity with the surrounding dyke material, in others they appear to have been completely enclosed in plagioclase. The plagioclase megacrysts occasionally show dark centres with vitreous lustre. The glassy felspar is generally surrounded by a rim of sericitised material.

Several other more alkali members of the same dyke suite carry small plagioclase megacrysts often crowded together near the centre of the dykes and the distinction between a felspathic dyke and a porphyritic dyke is arbitrary. This is also true of the series of trachydolerites and microsyenites which were emplaced following the main group of felspathic dykes in the area north of Narssarssuaq. Many of the dykes carry either small inclusions of foreign material rimmed by a younger overgrowth of felspar (Fig. 42) or are packed with phenocrysts which are only distinguishable from megacrysts in felspathic dykes by their more euhedral form and smaller size (Fig. 43).

III. PRE-GARDAR FELSPATHIC DYKES

Several periods of basic dyking earlier than the Gardar activity are recognised in South Greenland. The youngest of these is the swarm which separates the main formation of the Julianehåb Granite from its later reactivation (WATTERSON, 1965). This period of dyking is known to be older than granites which were emplaced around 1630 m.y. and it has been suggested by one of us that the dykes were probably emplaced between 1700 and 1650 m.y. (BRIDGWATER, 1965).

A few of these dykes contain plagioclase megacrysts resembling the scattered insets found in the early Gardar dolerites. These megacrysts measure up to 5×3 cm, are moderately euhedral, and contain inclusions of basic material parallel to the (010) face. They have never been found sufficiently fresh for determinative work.

North of **Ivigtut** a larger group of dykes cut the gneisses but are apparently affected by Ketilidian folding near to the base of the Ketilidian supracrustal rocks. These dykes show a progressive decrease in metamorphism north-westwards so that in the **Neria** region some are fresh olivine dolerites and can only be distinguished from Gardar dykes by their correlation with metadolerites cutting the Ivigtut gneisses. Most geologists now working in the region north of Sermiligârssuk regard these dykes as older than the Ketilidian plutonism and thus not contemporaneous with the metadolerites cutting the Julianehåb Granite further south (in contrast to the relationship shown by BRIDGWATER, 1965, fig. 1). One particular dyke described in detail by BONDESEN and HENRIKSEN (1965) can be followed for 40 km ESE from Tôrnârssuk to Arsuk Fjord. It varies in width between 25–100 m and frequently contains a well-defined zone bearing plagioclase megacrysts and locally glomeroporphyritic aggregates of plagioclase resembling the pseudo-ophitic texture described from Gardar dykes. This zone may be confined to one margin or it may occur centrally in the dyke. Glomeroporphyritic aggregates are reported in a dyke belonging to the same group found north of Sermiligârssuk.

On **Neria island** and the nearby mainland fresh olivine dolerites belonging to this group of dykes contain megacrysts of plagioclase. Some of these megacrysts are shiny, black felspars indistinguishable in the field from those in the Gardar dykes further south. These megacrysts have been studied in the same way as the Gardar inclusions and the results are given later under the appropriate headings. In general the megacrysts from the dykes in the Neria region are slightly more calcic than the corresponding inclusions in Gardar rocks, ranging from An_{60} to An_{65}. They often show extreme zoning against the host rocks and the strontium and barium content is generally lower than that in Gardar xenoliths. The optical determinations and X-ray powder diagrams suggest that they do not show the same marked anomalous properties as the inclusions in Gardar dykes although they do not fall exactly onto the standard determinative curves.

Plagioclase insets occur in other dykes along the west coast of Greenland, for example the Blåbær dyke of the Fiskefjord area (Berthelsen and Bridgwater, 1960). This dyke contains felspar with black centres provisionally identified as zoisite in the field. However, the abundance of black felspar in the dykes further south suggested that these should be re-examined and they have now been found to contain black felspar similar to that described on p. 109.

PART TWO

By D. BRIDGWATER

IV. PETROLOGY AND MINERALOGY OF FELSPATHIC INCLUSIONS IN GARDAR DYKES

a) Petrology

The xenoliths consist essentially of plagioclase, olivine and Fe-Ti oxides. Orthopyroxene has been noted in samples from three localities and may form a widespread minor constituent which has been overlooked because sampling was concentrated on the larger grains of mafic material. Clinopyroxene has only been found as scattered grains of doubtful primary origin in the typical granular anorthosite xenoliths. One sample, from the extreme limits of the occurrence of felspathic inclusions, has been found to contain clinopyroxene as an abundant primary mineral. Potash felspar, hornblende, clinopyroxene and apatite have been reported by UPTON (1964) from the Assorutit laminated anorthosite. Secondary xenoliths, composed partly of derived felspar and partly of mineral derived from the host rock or an early phase of the host contain a variety of minerals, especially augite, olivine, and Fe-Ti oxides, which are indistinguishable from those in Gardar gabbroic rocks. Alteration product minerals such as hornblende, biotite, chlorite, sphene, epidote, haematite and sericite are common. In some samples the alteration occurred before the inclusion of the xenoliths in their present hosts, in others it appears to be due to the interaction between host and inclusion. In many areas, particularly in the vicinity of the Nunarssuit and Ilímaussaq intrusions, both the host rock and the inclusions have been heavily attacked by later processes.

The plagioclase megacrysts include thin lamellae of olivine, magnetite, ilmenite, and occasionally gabbroic material. Many of the megacrysts are black in hand specimen, resembling obsidian. This is seen as a faint smoky discolouration in thin section. The colouration is often restricted to relic areas surrounded by white, clear or pale yellow plagioclase. Many xenoliths and megacrysts which appeared sericitised in the field were found to be fresh in thin section. The lack of vitreous lustre proved no guide to the amount of alteration suffered by the felspars. It is thought that minute fractures in many of the insets caused them to appear chalky white in the field.

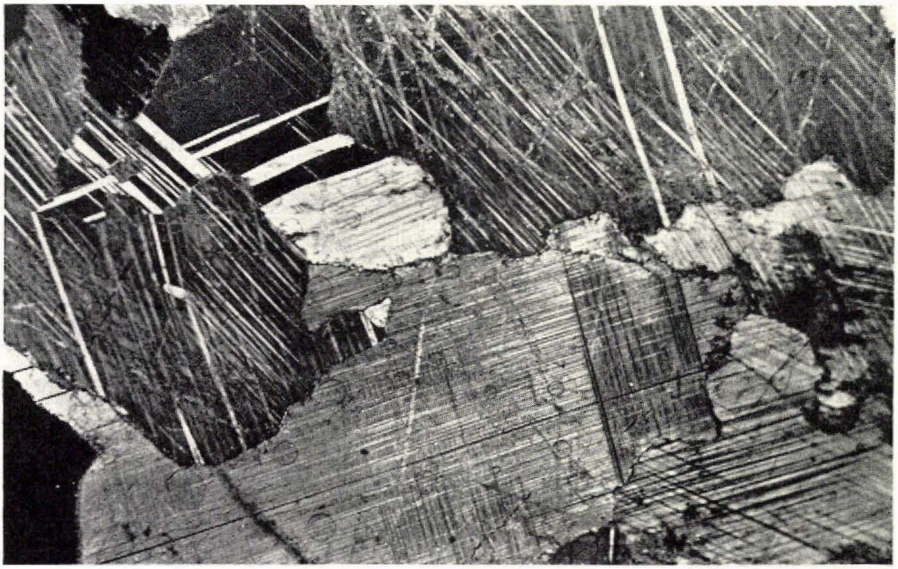


Fig. 44. Photomicrograph of granular anorthosite from Augiata tasia (dyke 9). Crossed nicols 5 ×. G.G.U. 25226. Note the closely spaced albite and pericline twinning which are almost equally developed, the irregular grain boundaries, and the slight distortion of some of the twin lamellae.

The main textural features of the anorthosites have been described in Section II (p. 20 to p. 29) as these rocks are so coarse-grained that their dominant characters can be seen in the field.

Under the microscope the granular anorthosites are seen to be composed of randomly orientated, generally equant plagioclase grains averaging less than 5 cm in diameter (Fig. 44). Identifiable mafic minerals have only been seen in thin sections from two localities; most of the material studied was collected from interstices between the plagioclase crystals by using a hammer and chisel.

Several of the larger blocks of granular anorthosite show slightly distorted plagioclase lamellae which might suggest deformation prior to their inclusion in Gardar hosts. However, no granulation is seen round the plagioclase crystals and the amount of deformation appears to have been small.

The small gabbro anorthositic xenoliths from **Tôrnârssuk** (32028) contain considerable olivine and a little hypersthene. No clinopyroxene was noted in the xenoliths although it is common in the surrounding dyke rock. Some hypersthene grains were enclosed by olivine, others enclosed small olivine crystals; both partly enclose plagioclase crystals. Although the grain size and mineralogy of the xenoliths differs markedly from the surrounding rock the margins are often indistinct; large plagio-

Fig. 45. Macrophotograph of the margin of an olivine-hypersthene-bytownite xenolith from an olivine dolerite on Tôrnârssuk (p. 33). Crossed nicols 10 ×. G.G.U. 38028. Note the way that the host dyke material interfingers with the inclusion.

clase and olivine crystals forming an integral part of the xenolith may enclose small islands of host dyke (Fig. 45), and veins of finer grained host penetrate between crystals in the xenolith. This suggests that although the xenoliths were already solid when included in their hosts they may have been rather loosely packed accumulations of early formed minerals.

The second gabbro anorthositic sample examined (57264) from an E–W, probably early Gardar, dolerite on **Akia** south of Julianehåb consists of plagioclase crystals set subophitically in orthopyroxene. This mineral forms approximately 30 $^0/_0$ of the total inclusion (Fig. 46). Occasional larger plagioclase fragments are seen in the xenolith and the slightly broken nature of many of the plagioclase grains suggests that the rock is composed of a loosely packed aggregate of plagioclase round which orthopyroxene crystallised. Veins of dyke material cut the xenolith and it is difficult to decide whether small crystals of altered olivine and a few clinopyroxene rims round the orthopyroxene are part of the original inclusion or whether they are derived from the host.

In thin section the laminated anorthosites from **Assorutit** resemble normal olivine dolerites more closely than they resemble the granular anorthosites. Part of this resemblance is due to the twin pattern of the plagioclases which is indistinguishable from the twin pattern of plagioclases taken from any normal gabbroic dyke. The Assorutit block is

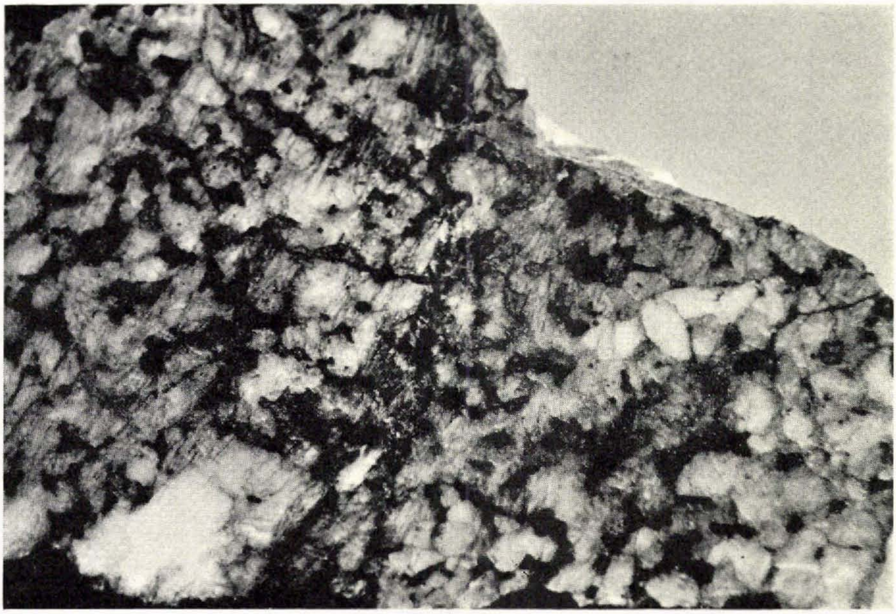


Fig. 46. Close up of a polished block of norite from a gabbroic host dyke on the island of Akia south of Julianehåb. 1.5 × natural size. G.G.U. 57264.

coarse-grained, the felspars average between 5–10 cm long and 2 cm in width. The interstitial olivine is correspondingly coarse and local pockets may reach ten or more centimetres across. The felspars show a distinct zoning in contrast to the almost complete lack of compositional variation in the granular anorthosites. Olivine which fills the major part of the interstices between the plagioclase is generally fresh, however, the olivine is cracked and contains minute magnetite grains where it is found close to the small pockets of biotite, ilmenite, augite, hornblende, and alkali felspar which apparently crystallised from material trapped between the olivine and the plagioclase. In the vicinity of these late crystallisation products the plagioclase grains may show a mottled zoning due to the partial replacement of the labradorite by alkali felspar with the formation of small grains of clinozoisite. Some chlorite is formed surrounding the margins of the olivine.

Two analyses of the Assorutit anorthosite (kindly supplied by B. G. J. Upton) are given in Table 1. Upton suggests, (1961, p. 20) that the anorthosites were formed as an accumulative rock from a magma temporarily precipitating labradorite alone. The lamination was caused by periodic currents which orientated the felspar tabulae parallel to the direction of flow. The mafic units represent periods of quiescence in which the percentage of intercumulus olivine is higher due to the less well-developed packing of the felspar.

Table 1. *Anorthosite analyses.*

	Primary		Secondary	
	Assorutit (B. G. J. Upton)		Qáqarssuaq (Ussing, 1912)	Narssaq
	50221	40523	20	49320
SiO_2	52.78	52.54	50.98	48.1
TiO_2	0.42	0.49	1.38	2.1
Al_2O_3	27.08	25.20	22.15	21.9
Fe_2O_3	0.64	0.79	1.04	8.5†
FeO	1.35	2.49	4.25	
MnO	0.03	0.05	tr	
MgO	1.03	1.60	0.79	2.6
CaO	10.77	10.45	7.90	8.5
Na_2O	4.50	4.50	6.84	4.4
K_2O	0.64	0.64	2.71	1.3
P_2O_5	0.23	0.25	0.38	0.64
CO_2	—	—	tr	
H_2O^+	0.47	0.72	1.22	
H_2O^-	*	*	0.12	
Cl	—	—	tr	
	99.94	99.72	99.76	Partial analysis
Q	0.06	—	—	
or	3.94	3.8	16.0	
ab	37.19	3.8	25.1	
an	52.24	46.64	21.7	
ne	—	—	17.7	
C	0.21	—	—	
di	—	4.3	12.8	
hy	3.97	0.8	—	
ol	—	3.1	0.1	
mt	0.94	1.1	1.5	
il	0.77	0.9	2.6	
ap	0.68	—	—	
Norms Watt (1966)				
Analyst	B. I. Borgen	B. I. Borgen	C. Winther	B. I. Borgen

* Sample dried at 110° for 2 hrs.
† Total iron as Fe_2O_3.

The commonest texture found in the felspathic xenoliths from alkali gabbro and trachydoleritic hosts is an aggregate of closely packed plagioclase fragments with little or no interstitial material. However, a second type of felspar aggregate in which subhedral plagioclase grains are intergrown together to form solid plagioclase masses up to 50 cm across is quite common. In the field these intergrowths appear simple; neighbouring plagioclase crystals with approximately plane faces are seen packed together with a little mafic material trapped between them. However, under the microscope the relationship is seen to be more complex; inclusions of one plagioclase are found in the neighbouring crystal and when the optical directions of a set of twin lamellae in one crystal are plotted on a stereogram with the optical directions of a corresponding set in the other crystal it is found that the "individual" plagioclases forming the intergrowth have a logical crystallographic relationship to each other. Commonly the two are found to be albite-pericline twins. It is not known how these aggregates form; they appear to be a special case of the giant felspars, which are common in the same dykes but which normally show a more regular crystal outline. A particular example of plagioclase intergrowth is described in the section on felspar twinning (p. 96).

b) The mafic minerals associated with the felspathic inclusions in Gardar dykes

Comparatively little work has been done on the non-felspathic components of the inclusions in Gardar dykes as fresh mafic material has only been collected from a few localities. This is partly because the anorthosites themselves consist dominantly of felspar, with the mafic material as interstitial pockets, and partly because the mafic minerals are more subject to alteration. A further complication is added because it is often difficult to decide whether the non-felspathic minerals are derived from the original anorthosite or whether they were precipitated from later Gardar magmas which surrounded the felspathic fragments.

i) Olivine

Olivine is the most common of the mafic minerals found as primary constituents of the anorthosites. It is particularly abundant in the Assorutit laminated anorthosite and in the large masses of granular anorthosite from Kobberminebugt. Many of the larger plagioclase megacrysts contain thin olivine strips parallel to the (010) twin plane. These are generally altered. In most xenoliths the olivine is restricted to small interstitial pockets where it is moulded ophitically onto the terminations of the plagioclase crystals. Since the mineral lacks a persistent cleavage, the size attained by individual crystals cannot be measured in the field. However, interstitial masses up to 20–30 cm in diameter have been seen in the Kobberminebugt anorthosites and slightly smaller masses in the Assorutit xenolith, and there is no reason to suppose that these are made of many individual olivine crystals.

Isolated megacrysts of olivine have been noted in several Gardar dykes. Some of these are found in association with plagioclase megacrysts or anorthosite xenoliths. On Tôrnârssuk they are found together with large hypersthene megacrysts. They are thought to represent basic material formed at the same time as the associated plagioclase megacrysts, their comparative rareness being due to their high specific gravity which leads them to sink in the majority of hosts. The compositions of three olivine megacrysts from different hosts are given in Table 2. It appears that the composition of both the olivine and the plagioclase megacrysts (p. 81) is related to the composition of the host.

Olivine also occurs in the secondary anorthosites, often as rims surrounding plagioclase megacrysts. In some instances, such as that illustrated in Figs. 33 & 34, p. 53, the olivine surrounds euhedral plagioclase subophitically.

Apparent twinning is seen in many of the olivine crystals from the Assorutit xenoliths and is a moderately common feature in olivines from Gardar gabbroic rocks. Apparent twinning occurs where two or more olivine crystals develop in the same interstice between plagioclase crystals. The common surface dividing the crystals is often planar and may correspond approximately to either the {110}, the {001}, or the {010} crystal planes of olivine. There is generally a slight angular discrepancy between the observed position of the face and its theoretical position with respect to the principal optical directions determined with a universal stage. It seems probable that the apparent twinning is a similar feature to the imperfect twins described from plagioclase (p. 94) which formed when two unrelated crystals set together separated by a common plane.

The composition of the olivine ranges from Fa_{27} in a sample of granular anorthosite to Fa_{57} in a megacryst from a trachydolerite. The majority of olivines from the secondary anorthosites contain more iron than those from primary anorthosites and resemble the olivines from the Gardar gabbros. Olivines from lamellae in giant felspars and material trapped between felspar aggregates in alkali gabbroic hosts is intermediate between the two, ranging from Fa_{33} to Fa_{40}. Olivine compositions from different types of anorthosites and megacrysts, together with the composition of the co-existing pyroxenes and felspar is given in Table 2.

ii) Orthopyroxene

Orthopyroxene has been identified as forming the principal mafic mineral in inclusions from two localities, one from an olivine gabbro dyke 5 km to the east of Kobberminebugt (61080, locality 7, Plate 4), and the second in one of the rare xenoliths found in an early Gardar dolerite cutting the granite on Akia south of Julianehåb (57264). It is found as a major constituent in the xenoliths and as megacrysts in the felspathic dyke on the east of Tôrnârssuk (32028). Neither the grains from 61080 nor the orthopyroxene seen in thin sections from 32028 and 57264 contained exsolution lamellae of other pyroxenes. In thin section

Table 2. *Composition of co-existing minerals (mol. %).*

G.G.U. No.	Type of Anorthosite	Locality	Olivine % Fa	Felspar % An	Pyroxene composition
49324	Layered	Assorutit	29–33	58	$Ca_{40\cdot5}Mg_{30}Fe_{29\cdot5}$
61071	Granular	Kobberminebugt Locality 2	34–38	61	—
61121	Crushed granular	10 km E of Kobberminebugt	28	altered	—
61080	Granular-laminated	3.5 km E of Kobberminebugt	27	60	Fs_{28-30}
57264	Gabbro anorthosite	Akia	—	56*	Fs_{38}
25171	Giant felspar	Isortoq (dykes)	32 ± 3†	57	—
55209	Comminuted aggregate	Narssarssuaq	40	56	—
61135	Secondary	18 km E of Kobberminebugt	37–42	49 to	$Ca_{38\cdot5}Mg_{37\cdot5}Fe_{24}$ $Ca_{36}Mg_{35}Fe_{29}$
49323	Secondary	Narssaq	48	56	$Ca_{39}Mg_{31\cdot5}Fe_{29\cdot5}$
61014	Secondary	Bangs Havn	—	58	$Ca_{35\cdot5}Mg_{36\cdot5}Fe_{28}$
61023	Secondary	Bangs Havn	—	58	$Ca_{35}Mg_{35\cdot5}Fe_{29\cdot5}$
61183	Secondary	18 km E of Kobberminebugt	—	47* to	$Ca_{37}Mg_{35}Fe_{28}$ $Ca_{33\cdot5}Mg_{36\cdot5}Fe_{30}$
Mean of several samples	Megacrysts	Eqaloqarfia	29–30	58–64†† (majority An_{58})	
32028	Gabbro anorthosite	Tôrnârssuk	27	65–80	Fs_{22}
32028	Megacrysts in olivine gabbro	Tôrnârssuk	27	—	Fs_{22}§
49331	Megacrysts in trachy-dolerite	Narssarssuaq	57	52	—

† Composition determined optically.
* Felspar altered.
†† Universal stage measurement (PULVERTAFT, 1965).
§ Partial chemical analysis. The pyroxene contained no detectable Ca.

the orthopyroxene was practically colourless although it showed moderate pleochroism as small grains. The fresh minerals co-existing with the orthopyroxenes are among the most basic found in the anorthosite inclusions (Table 2). The composition was determined using measurements of N_z (HESS, 1960). A partial chemical analysis of a single crystal from

Tôrnârssuk gave the following results: FeO 13.34 %, MgO 27.75 %, MnO 0.30 %, TiO_2 0.57 %, Al_2O_3 2.30 %, Na_2O 0.23 % and P_2O_5 0.05 %. No calcium was detected. This analysis corresponds to an orthopyroxene with a composition of $Fs_{21.5}$. The lack of calcium and the high alumina content suggest a deep-seated origin.

iii) Clinopyroxene

The only anorthosite xenolith collected which contains clinopyroxene in appreciable amounts was found by A. K. HIGGINS in a small ENE-trending dolerite on Midternæs (71724). This is one of the most northerly of the felspathic dykes noted. The xenolith has a subophitic texture resembling the orthopyroxene-bearing anorthosite illustrated in Fig. 46. The plagioclases are slightly altered while the clinopyroxene is fresh; some of the plagioclases are slightly crushed. It appears probable that the rock formed as an accumulation of small plagioclase grains later surrounded by clinopyroxene. The composition of the pyroxene is approximately $Ca_{37}Mg_{35}Fe_{28}$ (determined optically from 2V 47; and N_Y 1.703, using the curves published by HESS, 1949).

Single grains of a mineral thought to be clinopyroxene occur in the sample of orthopyroxene-bearing anorthosite (61080). This has a well-developed cleavage, oblique extinction and lacks the pleochroism seen in the orthopyroxene from the same sample. Refractive index measurements gave N_X' 1.700 and N_Z' 1.719; universal stage measurements were impossible.

The Assorutit laminated anorthosite contains one of the few clinopyroxenes positively identified from a primary xenolith. The mineral forms less than 1 % of the rock and is restricted to groups of small grains in the interstices between plagioclase crystals. In thin section the grains show some zoning. Grains extracted for refractive index determinations show a moderately pronounced mushroom colour with slight pleochroism similar to clinopyroxene from many of the alkali basic rocks of the Gardar province. Optical determination of the composition (2V 52°, N_Y 1.706) gave $Ca_{40.5}Mg_{30}Fe_{29.5}$. As the material occurs in such small amounts the results can only be regarded as approximate. However, it is worth noting that the pyroxenes co-existing with Fa_{30} olivine and An_{60} plagioclase in other Gardar gabbros are less rich in Ca and Fe than the Assorutit examples. This is thought to be due to the comparatively late formation of the Assorutit pyroxene in the crystallisation sequence of the xenolith.

In contrast to the rare occurrence of clinopyroxene within the primary xenoliths this mineral is the commonest coloured constituent of the secondary anorthosites. Generally the clinopyroxene encloses or partially encloses plagioclase. In many cases the individual pyroxene crystals exceed 10 cm in diameter. It is quite common for the pyroxene to be intergrown symplectically with Fe-Ti oxides. All the pyroxenes from secondary anorthosites studied in thin section show a marked purple colouration with weak pleochroism suggesting a high Ti content. N_Y measurements varied between 1.699 and 1.703 while 2V measurements varied between 41° and 48°. This corresponds to a fairly narrow range in composition in the augite range which is rather surprising since the host rocks in contact with the clinopyroxenes vary considerably. Some of the clinopyroxenes show a slight patchy zoning and appear unstable in their hosts.

iv) Iron-titanium oxides

Opaque oxides form a relatively minor constituent of both primary and secondary anorthosites. Fresh samples of oxides from primary anorthosites have only been collected from the Assorutit xenolith and from a xenolith in a nepheline syenite dyke south of Igaliko Fjord (p. 142). Many large masses from Kobberminebugt contain Fe-Ti oxides forming on average between 1 and 2 $^0/_0$ of the rock but in places exceeding 20 $^0/_0$. These opaque minerals are generally too altered to determine their original mineralogy. In the secondary anorthosites the opaque minerals are fresh and form quite large grains partially enclosing plagioclase.

The scarcity of samples from the primary anorthosites allied to the fact that the oxide grains appear to have accomodated themselves, at least partially, to the changes in environment which the xenoliths passed through, means that the characters of the oxides cannot be used to support any particular theory of anorthosite formation.

1) Oxides from granular anorthosites. The oxides from granular anorthosites are only represented by one fresh sample (48318) collected by J. P. Berrangé from "Vatnahverfi" and by several altered samples from Kobberminebugt. However, although the material is limited it can be seen that the original characters of the Fe-Ti oxides from the granular anorthosites differ significantly from those of the laminated rocks. Magnetite is the main mineral present accompanied by an exsolution trellis of several generations of ilmenite lamellae. Both the fresh sample from "Vatnahverfi" and the altered samples from Kobberminebugt show a much more complex history than the ilmenite from Assorutit described below; involving crystallisation of magnetite, exsolution of ilmenite, alteration under oxidation conditions with the formation of rutile and haematite, and recrystallisation under high temperature, more reducing conditions. Local late alteration at lower temperature took place probably during the cooling of the host rock. A detailed description of the mineralogy and probable sequence of mineral formation is given in the Appendix (p. 200).

2) Oxides from laminated anorthosites. The Assorutit laminated xenolith contains less than 1 % of opaque minerals as independent grains, generally smaller than 1 mm in diameter, in the interstices between the plagioclases. The relationships between these grains and the other minerals forming the xenolith suggest that the opaque minerals were formed later than the plagioclase and olivine but before the final interstitial crystallisation of alkali felspar and the fibrous chloritic minerals. Many of the opaque grains are intergrown with biotite which appears to have continued growth after their formation. In polished section the opaque minerals are seen to consist almost entirely of ilmenite as elongate infillings between the plagioclases and occasionally as strips along the (010) felspar twin planes. Some fine twinning is visible in the ilmenite when examined under high magnification ($\times$ 600 or above). The ilmenite grains contain minute inclusions of poorly-reflecting material and a few crystals of late opaque minerals which appear to be replacing the ilmenite. These secondary minerals are developed to a more marked degree from ilmenite grains in contact with the interstitial alkali felspar, suggesting that ilmenite was no longer stable in the last stages of crystallisation of the anorthosite. There are a few sulphide grains associated with the alteration products of ilmenite; these are too small for complete identification.

Olivine crystals in the Assorutit anorthosite have small crystals of magnetite developed along cracks and near crystal margins.

3) Oxides from secondary anorthosites and oxide megacrysts in Gardar hosts. These two categories may be treated together as they show similar features in polished section. Some of the oxide grains in both the secondary anorthosites and among the scattered megacrysts may have originated as constituents of primary anorthosites but their present properties appear to be dominated by the effects of a later host. The majority of grains studied are magnetite with a variable amount of ilmenite exsolved as a $\{111\}$ trellis or as separate grains surrounded by, or more rarely on the margin of, the magnetite. The relations between magnetite and ilmenite in these oxides show progressive changes in character which were apparently controlled by the effect of the surrounding host rock. A summary of these is given below; details are described in the Appendix. A few independent grains of ilmenite are found in the secondary anorthosites from Narssaq; these are regarded as being more closely related to the host than the original xenoliths.

Oxide grains from the pseudo-ophitic xenoliths (**locality 13**, Plate 4; Fig. 34, p. 54) consist of subhedral magnetite grains up to 5 cm in diameter which appear slightly unstable in their present host. The magnetite shows the exsolution of three generations of ilmenite controlled by the $\{111\}$ crystallographic direction of the magnetite (Fig. 47). This exsolution pattern resembles that found in magnetite

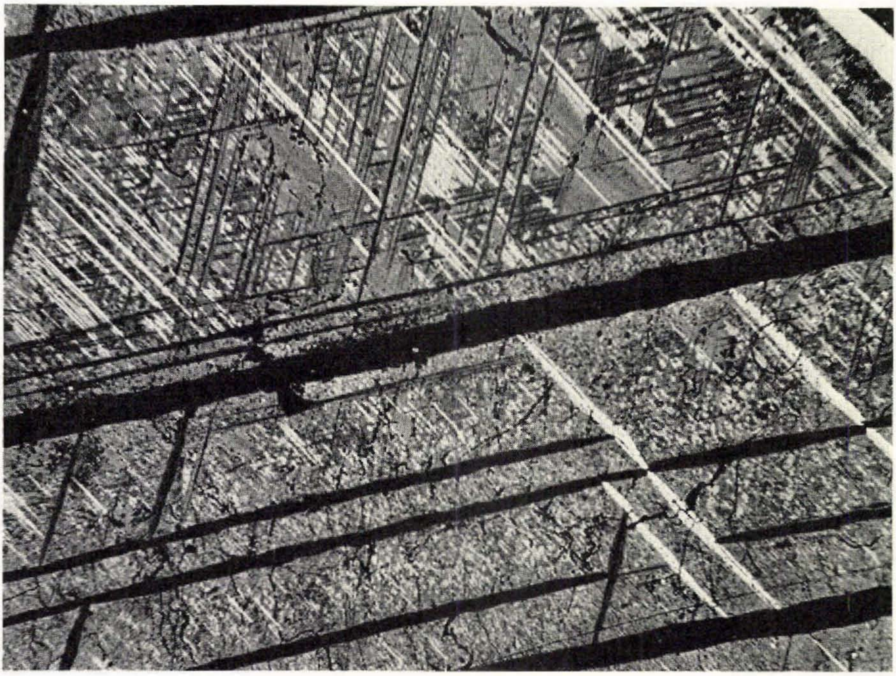


Fig. 47. Photomicrograph of a polished sample of a Fe-Ti oxide crystal from dyke 13. Partially crossed nicols 50 ×. G.G.U. 61184. Note the three generations of ilmenite lamellae. The large primary lamellae (mainly dark in this photograph) are the site of incipient alteration. The secondary lamellae (particularly well developed in the area at the top of the photograph) are irregularly distributed. The tertiary lamellae cause the patchwork effect seen in the lower half of the photograph.

crystallising direct from gabbroic magma and is the simplest relationship seen between the two minerals in the secondary anorthosites. The three generations of ilmenite are thought to represent different stages in the exsolution of this mineral from a magnetite-ulvospinel solid solution which became progressively more unstable during the cooling history of the oxides. The two earliest generations of ilmenite form characteristic lamellae parallel to {111} in the magnetite while the third generation consists of an almost submicroscopic {111} controlled dendritic intergrowth between magnetite and ilmenite in the groundmass of the oxide grains. A fourth stage in the exsolution of ilmenite is seen in which strings of ilmenite grains cut across the magnetite crystal along irregular cracks. This phase of exsolution was accompanied by the formation of scattered ilmenite grains along the margins of the magnetite crystals and by renewed growth of ilmenite along the margins of the larger primary lamellae. The magnetite appears to have been less stable than the ilmenite in the present host resulting in the local preservation of the exsolution trellis in an indeterminate groundmass of iron hydroxides.

Oxide megacrysts from the trachydolerite centre of **dyke 12** show many of the features seen in the oxides from pseudo-ophitic anorthosites. However, there is a noticeable increase in the number of ilmenite grains formed along the margins of the magnetite crystals and the dendritic groundmass exsolution of ilmenite is much more pronounced. Many grains of ilmenite within the magnetite are surrounded by

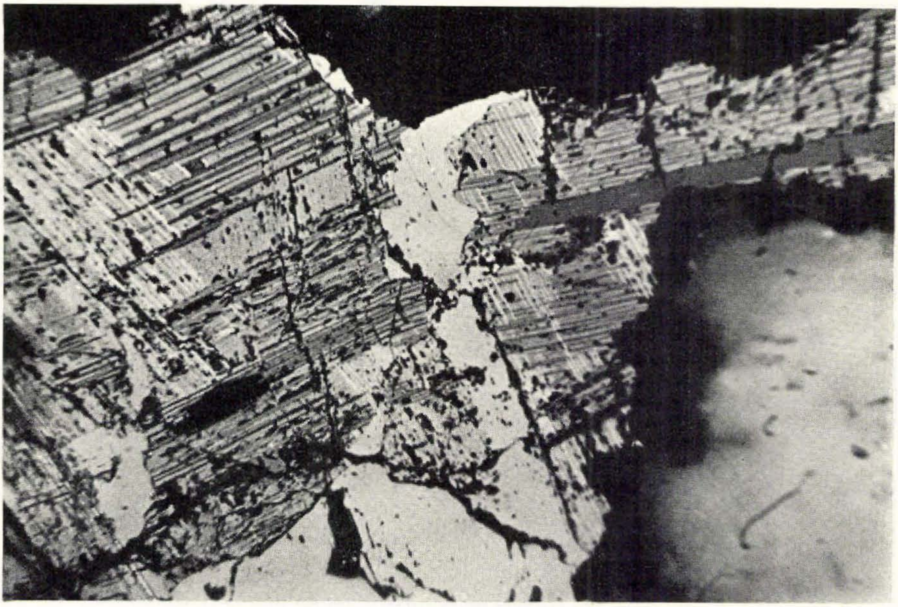


Fig. 48. Fe-Ti oxide grain from secondary anorthosite xenolith at Narssaq. Partially crossed nicols 50 ×. G.G.U. 49323. The main part of the grain is a magnetite crystal with two directions of ilmenite lamellae visible in the photograph. This intergrowth has been broken and veined by a new generation of ilmenite (light coloured in the photograph).

a coarser development of the dendritic intergrowth suggesting that the formation of the ilmenite grains within the magnetite is due to segregation of the two minerals into separate patches. Some of the earlier lamellae have been partly replaced by rutile.

Oxide grains from the secondary anorthosites at **Bangs Havn** show a further development of the same features; there is a marked segregation of ilmenite along the margins of the magnetite grains and the groundmass is composed of a patchy graphic intergrowth of ilmenite and magnetite which becomes quite coarse in the neighbourhood of the ilmenite segregations.

The oxides examined from the secondary anorthosites at **Narssaq** differ slightly from those seen elsewhere. The majority of grains occur as rather irregular magnetite crystals which appear to have been broken prior to their inclusion in their present host. These show two distinct sizes of ilmenite lamellae suggesting that at least two generations of ilmenite exsolution took place similar to those seen in the oxides from the pseudo-ophitic anorthosites. However, the lamellae in the Narssaq oxides have coalesced so that they appear to have formed at one time. There is no sign of exsolution of ilmenite from the groundmass. It is thought that the coalescence of the ilmenite lamellae and the lack of groundmass exsolution reflect a secondary segregation of ilmenite onto a pre-existing trellis of ilmenite. In several places the lamellae have been severely distorted. A new generation of lamellae has formed after the distortion of the original oxide grains and veins the magnetite (Fig. 48). This younger ilmenite shows a higher reflectivity than the lamellae and resembles independent ilmenite grains in the secondary anorthosite which are regarded as crystallised from the host. The higher reflectivity may possibly be due to a higher content of unexsolved magnetite in the younger ilmenite.

V. MINERALOGY OF THE FELSPARS

The mineralogy of the plagioclase felspars, which form the dominant mineral in the anorthosite xenoliths, was examined in some detail in the hope that some indication of the origin of the xenoliths might be suggested. Many of the investigations raised mineralogical questions of considerable interest and these are described in this paper. However, the studies were not carried through to completion as it became apparent that they would not lead to further information about the genesis of the original rock.

a) Composition

The composition of the plagioclase in the xenoliths and megacrysts varies between albite and bytownite with the majority between An_{45} and An_{62}. Preliminary work showed that there was considerable difference between the results obtained by universal stage extinction methods and those obtained by refractive index measurements. It also became apparent that the range in composition in plagioclases from different textural types was so small that precise measurements were necessary if any significance was to be placed on the results. The refractive index measurement method, described in the Appendix of this paper, was therefore devised in order that moderately rapid accurate determinations were possible. Fig. 49 shows the range of composition found in plagioclases from anorthosites with different textural features and plagioclase megacrysts from different hosts.

The two main factors influencing the anorthite content of the smaller anorthosite inclusions and the plagioclase insets are the degree of sericitisation of the felspars and the composition of the host.

It is commonly difficult to be certain when a particular xenolith or megacryst was sericitised as in some places this process occurred earlier than inclusion in the present Gardar host while in others at least part of the alteration is due to chemical instability of the inclusion in the host rock. In a few areas alteration of both host and inclusion took place at a late stage either as a result of autometamorphism or due to nearby intrusion of a major Gardar intrusion. However, there is

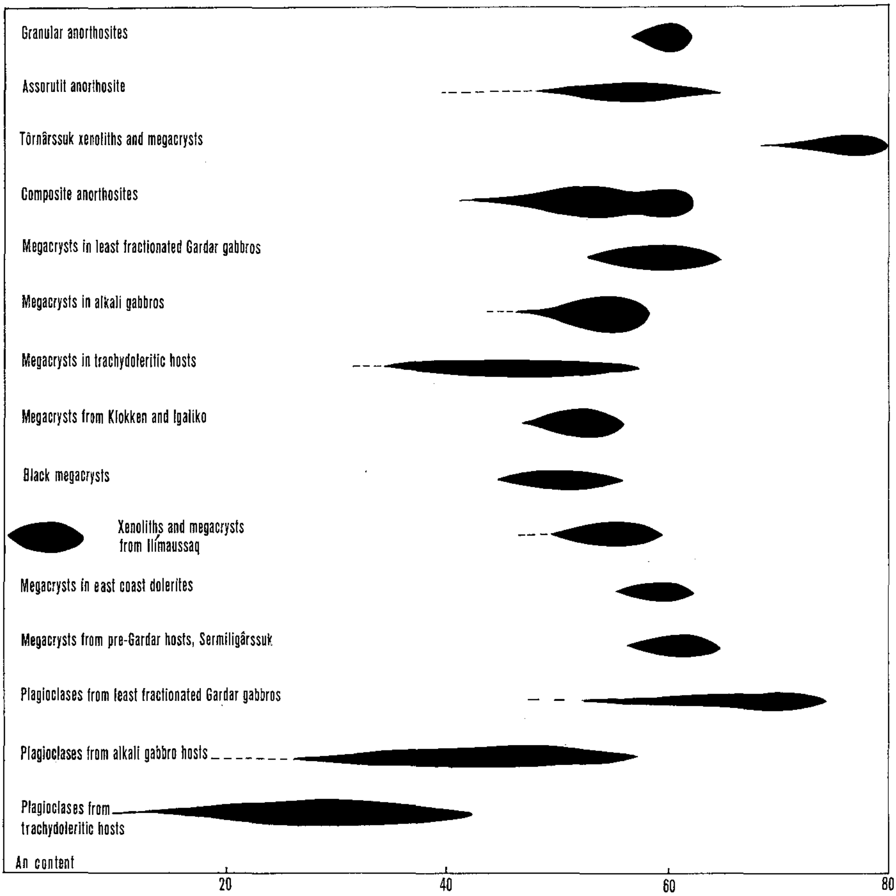


Fig. 49. Composition of plagioclases taken from various felspathic inclusions in intrusive rocks from South Greenland. The width of the line gives an impression of the frequency of plagioclases determined with a particular composition. No statistical accuracy is implied.

considerable evidence that the anorthosites were subjected to a widespread, if somewhat patchy, metamorphism which caused a sericitisation of the felspars, a general breakdown of the coloured minerals and oxidation of the Fe-Ti oxides. The best examples of this alteration are seen at Kobberminebugt and Augiâta tasia where sericitised felspar masses are found in fresh Gardar dykes the mineralogical composition of which is not very different to that of the fresh anorthosites. The effect of this alteration on the bulk composition of the anorthosite is unknown, however, a local loss of anorthite is often seen surrounding patchy sericitisation in some of the plagioclase crystals.

The effect of the Gardar host rocks on this partially altered material is complex. In some cases there appears to have been a recrystallisation of the material so that the fragments approach their original composition, in others, especially in the less basic host rocks, there has been an increase in the amount of sericitisation, followed by some recrystallisation near the margins of the felspar so that it is more in equilibrium with the host.

The only material which can be used to determine the composition of the plagioclase in the granular anorthosite are the large blocks which occur in dykes to the south and east of Kobberminebugt. Fresh felspars from these granular anorthosites showed no zoning and a composition range between An_{58} and An_{62}. No difference was detected between the composition of the coarser grained blocks within the anorthosite and the surrounding granular material.

Felspars from the gabbro anorthosite xenoliths from **Tôrnârssuk** are fresh and glassy. They range from An_{65} to An_{80}; single crystals show moderate zoning. The majority of fragments examined from the felspar centres were approximately An_{78}. This is considerably more calcic than any other felspars examined from the anorthosite inclusions in South Greenland and is the most calcic felspar found in the Gardar igneous province. The difference in plagioclase composition, allied to the relatively high mafic content of these xenoliths may suggest that they were not formed at the same time as the other felspathic inclusions and their relationship with other anorthosite xenoliths and the Gardar gabbroic rocks is discussed later (p. 188). There is no evidence to suggest that these xenoliths were derived from the local gneisses some of which contain relics of an old anorthositic complex which has passed through several metamorphic episodes (WINDLEY, 1965). The twin pattern and general texture shown in Fig. 45 suggest an igneous origin with no history of metamorphism between initial formation and inclusion.

Felspars from the **Assorutit** xenolith show moderate compositional zoning which is quite marked close to the interstitial pockets of alkali felspar. There also appears to have been some replacement of the plagioclase by the alkali felspar which forms irregular intergrowths cutting across the primary twinning. This zoning makes optical determination of the plagioclases rather less certain than in the more homogeneous granular anorthosites. Fragments taken from the centres of the plagioclases average $An_{62\text{-}63}$, one fragment reaching An_{65}. Marginal fragments averaged $An_{52\text{-}56}$, although this need not reflect the total range of plagioclase composition in the rock. The average composition of the plagioclase appears to be between An_{54} and An_{59}, slightly less calcic than that found in the granular anorthosite.

With the exception of the insets described below from Tôrnârssuk the composition of the plagioclase megacrysts and the smaller xenoliths in gabbroic hosts varies between An_{40} and An_{65} with the majority be-

6*

tween An_{52} and An_{60}. Occasionally there is some variation between granular anorthosite xenoliths and plagioclase megacrysts in the same sample. Where this occurs the xenoliths are generally about An_{60} while the megacrysts are less calcic and approach the composition of the groundmass felspars of the host rock. These may be as low in anorthite as An_{40-50} reflecting the alkali affinities of many of the Gardar gabbros.

Sporadic megacrysts in relatively undifferentiated gabbroic host are occasionally found to be more calcic than the plagioclases from the granular anorthosite xenoliths. For example, megacrysts collected by T. C. R. PULVERTAFT from an olivine gabbro on Nunarssuit had an average composition of An_{64}, while the centre of some of the megacrysts collected from the **Eqaloqarfia dyke** (PULVERTAFT, 1965) reach An_{64} although they average An_{58}.

Megacrysts from the olivine dolerite host dykes on **Tôrnârssuk** are the most calcic found in Gardar dykes (An_{60-80}). The majority are slightly less calcic than the plagioclases from the xenoliths in one of the felspathic dykes from the same locality. Apart from their composition they appear identical with megacrysts from other Gardar gabbro hosts, they are moderate in size, averaging between 3–5 cm, they are often slightly rounded, and they contain inclusions of mafic material parallel to the (010) cleavage direction.

The large glassy megacrysts and the black felspars are less calcic than the granular anorthosites, ranging between An_{48} and An_{60}. Samples taken from adjacent megacrysts are usually remarkably similar in composition (within 3–4 $^0/_0$ An). They are generally fairly close to the composition of the centres of felspars in the groundmass of the host. The large crystals enclose pockets of rock in which the host felspar is considerably more sodic than in the main dyke mass. The megacrysts zone violently against this interstitial material although they are otherwise homogeneous.

The black felspars show a very similar composition from all the dykes from which they have been collected, ranging between An_{49} and An_{56}. The black felspars from the margin of the Klokken intrusion and the Igdlerfigssalik centre of the Igaliko syenite are similar although some in the last named intrusion reach An_{57}. There is a slight but persistent difference between black and white felspar from the same crystal: the white felspar, which is formed at the expense of the black at a slightly later stage, is consistently 3–4 $^0/_0$ lower in anorthite. Anorthosite xenoliths composed of comminuted fragments derived from plagioclase megacrysts show the same anorthite content as the whole megacrysts from the same host.

The plagioclase megacrysts collected from host gabbros which have been contaminated by the assimilation of granitic material (see for example locality 13, p. 53) are often sericitised and show signs of chemical instability in their host. Relic areas of fresh material are rather variable in composition, commonly between An_{40} and An_{50}. This variability is presumably due to different degrees of sericitisation and adjustment to a new chemical environment.

A similar but even more marked variation is seen in the felspathic inclusions from less basic dykes, for example the microsyenite-trachydolerite group of minor intrusions. The majority of insets in these rather alkali hosts are sericitised subhedral megacrysts and small altered anorthosite xenoliths. They show a range in composition from An_{35}-An_{58}. Individual megacrysts are generally uniform in composition, apart from heavy marginal zoning against the host and the patchy loss of anorthite surrounding sericitised areas. Adjacent insets may vary by 10–20 % anorthite.

As a rule the smaller fragments are less calcic than the larger and it is difficult to decide whether some of the subhedral grains in the groundmass originated as comminuted fragments or whether they crystallised direct from the host rock. Recognisable plagioclase inclusions sometimes form the centre to rhombic alkali felspars which zone outwards to albite-oligoclase often with a perthitic alkali felspar intergrowth. Similar overgrowths derived from the host are found surrounding larger megacrysts; in some cases they are especially prominent as they form clear margins to sericitised centres. No megacrysts with a composition lower in calcium than An_{35} have been found in alkali dyke hosts; the albite-oligoclase rims are clearly derived from the surrounding rock. The lower anorthite content of megacrysts from the more alkali hosts is thought, at least in part, to be an original feature. Evidence from the twin patterns developed by these megacrysts (p. 99) suggests that they have not recrystallised to any great extent, at least in their centres. The change in composition seen in megacrysts from different hosts suggests that there may be a genetic link between the host and the felspathic inclusions. Some hosts show an apparent loss of calcium and aluminium in the groundmass which suggests that part of the material making up the plagioclase inclusions may have been derived from the host magma. It is impossible to draw a sharp boundary between felspathic dykes with a large percentage of "foreign plagioclase" and some of the syenitic dykes which contain a variable content of small plagioclase phenocrysts rimmed by alkali felspar.

There are relatively few occurrences of anorthosite inclusions or plagioclase megacrysts in the major Gardar intrusions. This may be in part due to the greater assimilation of foreign material in the slower cooling plutons and it may be in part due to the sinking of plagioclase insets in the syenitic magmas which formed most of the larger bodies.

The porphyritic syenite of the **Narssaq intrusion** contains fragments derived from the overlying anorthosite-bearing sills. The inclusions become smaller and less

Fig. 50. Photomicrograph of small plagioclase megacrysts in an alkaline matrix (dyke 12, north-east end). Crossed nicols 45 ×, G.G.U. 61173. In one individual the plagioclase has been sericitised; in the second it is fresh. The andesine-labradorite cores are rimmed by oligoclase and alkali felspar. The rims may take on a rhombic form.

calcic away from the contact with the original host and commonly form centres to alkali felspar rhombs. Similar twinning to that seen in fragments in alkali dykes is developed (Fig. 50). The fragments range from An_{50-60} (the composition of xenoliths in the secondary anorthosite of the overlying sill) down to An_{35-40}.

Although the black felspars found in the marginal gabbro of the **Klokken intrusion** differ in many important respects from those found in dyke rocks, they have compositions (between An_{49} and An_{54}) corresponding to black insets in the alkali gabbro dykes. Similar black felspars (An_{52-57}) are found in the marginal syenites of the **Igdlerfigssalik centre** of the Igaliko complex. The reaction between host and inset is more marked at this locality than in the majority of dykes. Some of the megacrysts show practically no effect of the surrounding rock apart from the development of a vitreous lustre and a tendency for concoidal fracture. Other megacrysts show successive zones from a mat black centre surrounded by vitreous black and translucent colourless mantles to rims of white sericitised felspar. The sericitised rims are often intergrown with nepheline or potash felspar and appear to have lost all their original anorthite; the primary labradorite composition can only be suggested by the composition of the centre of the crystals and the presence of the original plagioclase twinning which can be traced across the changes in felspar type. The field relationships suggest that the inclusions at Narssârssuk were already part of a solid rock before they were enclosed in syenite.

The plagioclase insets derived from the rafts of secondary anorthosite capping the **lujavrites** on **Kvanefjeld** show a much more complete reaction with their host rocks. Apart from the mechanical breakdown described on p. 59 the felspars have also been completely recrystallised. The first stage in this process was a thorough sericitisation in which most of the calcium was lost. This was followed by formation

Fig. 51. Photomicrograph of albitised felspathic material found as inclusions in the lujavrites on Kvanefjeld. Crossed nicols 40 ×, G.G.U. 61400. The plagioclase inclusions give the impression of being single crystals with (010) twinning; however each "twin lamella" is in fact a separate albite crystal. The plane separating the albite crystals probably corresponds to the (010) plane in the original labradorite before replacement. A mica-like calcium sodium silicate of the pectolite group is found as small grains within the plagioclase, presumably derived from the breakdown of labradorite in a highly sodic environment.

of albite (An_5) within the boundaries of the old plagioclase crystals. The albite may occur as single crystals but generally it forms distinctive granular masses in which the orientation of the new material is controlled by the crystallography and twinning of the original plagioclase (Fig. 51).

The correlation between the composition of the insets and the composition of the host masks any more subtle variation which might have been present in the anorthosites before their inclusion. The megacrysts found in the pre-Gardar and early Gardar olivine dolerites tend to be slightly more calcic than those found in the later hosts, but this agrees with the general rule that the inclusions in less differentiated hosts are more basic than those in alkali hosts. Similarly, with exception of the inclusions from Tôrnârssuk, there is little evidence to suggest a regional variation in the original anorthite content of the presumed anorthosite parent at depth. Samples described by WEIDMANN (1964, p. 96), AYRTON (1963, p. 90) and C. H. EMELEUS (personal communication) from the Tigssaluk area north of Ivigtut have compositions which are relatively low in calcium (oligoclase-andesine). However, according to information supplied by EMELEUS, and the descriptions published by

WEIDMANN, the host dykes cannot be regarded as normal dolerites and appear allied to the microsyenite-trachydolerite group of hosts. Specimens of plagioclase megacrysts from gabbro dykes collected by R. BØGVAD (p. 14) have compositions ranging from An_{55} to An_{61} indistinguishable from megacrysts in comparable hosts of known Gardar age.

b) Spectrographic strontium and barium analyses

Recent work, for example that of BUTLER and SKIBA (1962), has suggested that plagioclases taken from a series of consanguineous rocks may contain a characteristic amount of strontium which can be used to separate them from similar rocks which do not form part of the same sequence. Further, in one intrusion this element often shows a systematic variation with the calcium content of the felspars. The same authors (SKIBA and BUTLER, 1963) suggest that in the rocks they have examined the strontium content of felspars remains fairly constant although the rocks containing the felspar may have passed through moderate to severe metamorphism. It was therefore thought possible that if the felspathic material was derived from a pre-Gardar source with no genetic connection to the province, the strontium content of the inclusions and plagioclases taken from the host dykes might show a significant difference. The barium content of the felspars was determined at the same time, mainly to provide background information on the distribution of this element in plagioclase from Greenland intrusions.

Fig. 52 shows a simplified diagram of the distribution of strontium in samples of felspar separated from anorthosite xenoliths, plagioclase megacrysts and a veriety of host rocks and basic dykes found in South Greenland. The full results are tabulated in the Appendix. Exchange of material with Dr. J. R. BUTLER (Imperial College, London) and determination of identical samples using X-ray fluorescence suggests that the spectrographic determinations described here are reasonably close to the absolute values. Determinations of plagioclases with 100–1500 p.p.m. strontium agree with the determinations using other methods. Spectrographic results over 2000 p.p.m. are up to 25 $^0/_0$ above the values obtained by other methods, and can only be treated as relative amounts.

Summary and interpretation of results

The most important inference to be drawn from the results is that felspars from the anorthosites and Gardar gabbros contain similar amounts of strontium and thus the distribution of this element cannot be used to separate them. Results from megacrysts from the pre-Gardar dykes north of Ivigtut (p. 66) are in some cases lower than comparable megacrysts from Gardar gabbros although there is considerable overlap

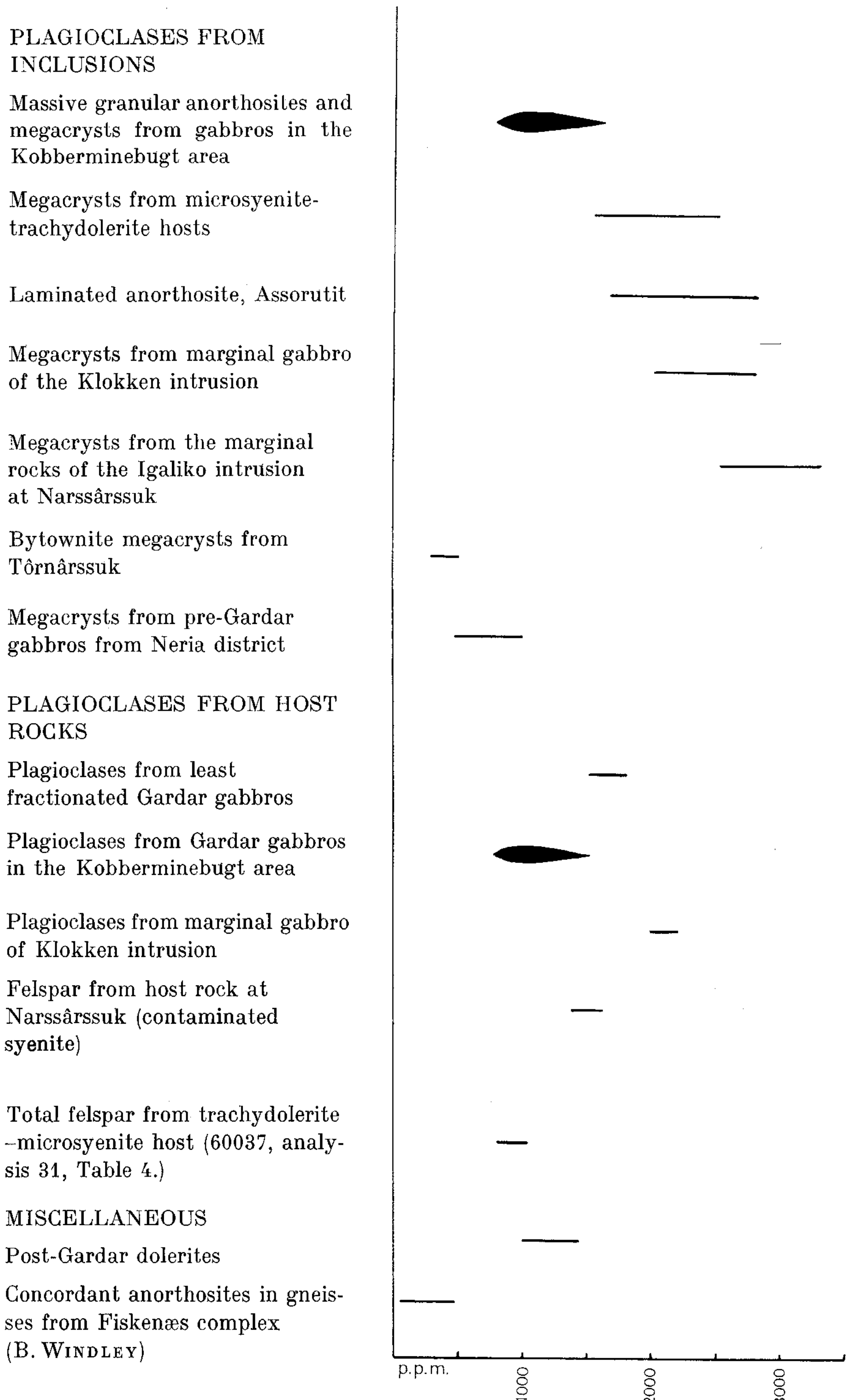


Fig. 52. Strontium content of plagioclases taken from a variety of inclusions and basic igneous rocks in South Greenland.

between them. Plagioclases separated from the post-Gardar dykes fall into the same range as those from Gardar basic rocks. Determinations on the calcic plagioclases from concordant anorthosite layers in the gneisses north of Ivigtut show that these contain less strontium.

The plagioclases separated from Gardar gabbros and the anorthosite inclusions may be divided into two groups using their strontium content. Plagioclases from the majority of Gardar gabbro dykes show strontium contents between 800 and 1800 p.p.m. which are identical to those of the inclusions of felspathic material in them. Plagioclases from the marginal gabbro of Klokken and the felspar megacrysts found in the margins of Klokken and Igaliko show a higher strontium content (between 2100 and 3400 p.p.m.). A similar high strontium content was found in the Assorutit layered anorthosite (1700–2700 p.p.m.). This gives some support to the idea that the felspathic inclusions are not all derived from the same source. Plagioclase insets from trachydoleritic hosts show a high strontium content (1600–2500 p.p.m.) throughout the area while felspar separated from the host rocks shows less than 1000 p.p.m. The bytownitic felspar from the gabbro anorthosite inclusions from Tôrnârssuk contains less strontium (400 p.p.m.) than other plagioclases from Gardar basic intrusions.

In general there was no simple correlation between the anorthite content of the plagioclase and the amount of strontium they contained; adjacent fragments with identical compositions showed the full range of strontium determined from inclusions from that particular environment. Samples were taken at intervals across one large crystal from dyke 8 (p. 46) and out into the interstitial host rock surrounding the crystal. Further samples were taken of the normal host rock of the dyke, which is slightly more basic than the material in contact with the large felspars. Strontium determinations of these samples showed no significant variation although there was a considerable difference between the plagioclase megacryst (An_{58}) and the felspars from the interstitial material (approximately An_{35-40}). The lack of a consistent variation of strontium with changes in anorthite content precluded the use of Sr/Ca ratios in this account.

A similar study was carried out on specimens from Narssârssuk and Klokken. A series of fragments from two large megacrysts set in a groundmass believed to have been formed by the "syenitisation" of earlier gabbro, were analysed from Narssârssuk. These showed no significant differences in strontium although the felspars changed in character from fresh, glassy, black labradorite in their centres to a sericitised mixture of altered plagioclase and alkali felspar at their margins and in the host rock. The high strontium content of the host-rock felspars supports the idea that the felspathic rocks along the margins of the Igdlerfigssalik centre may once have been gabbros, resembling those of Klokken, which were later altered by successive phases of syenite emplacement. A sample of fresh biotite syenite from

Narssârssuk contains 1500 p.p.m. strontium. This is higher than the strontium contents found in other Gardar alkali rocks and may be due to heavy contamination by partly replaced gabbroic material. A series of samples from the centre of plagioclase megacrysts out into the host gabbro of the Klokken intrusion showed no significant changes in strontium content.

The lack of variation in the strontium content of these megacrysts in spite of heavy marginal alteration and local complete loss of the original anorthite component of the plagioclases from Igdlerfigssalik, suggests that the values obtained from other crystals may be close to the original figure. If the strontium content of the felspathic inclusions is regarded as a secondary phenomena due to the effect of the host rock then recrystallisation must have been so complete that little can be left of other original features.

The close correlation between the strontium content of the inclusions and plagioclases from their hosts suggests that there may also be a close genetic connection. It appears significant that the highest strontium values were obtained from plagioclases thought to represent the first formed minerals of a moderately highly differentiated alkali magma (for example the plagioclases from the gabbroic margin to Klokken) while the granular anorthosites and megacrysts from less differentiated Gardar dolerites contain a relatively low strontium content. Ultimately it may be possible to use strontium as an indicator of the stage of differentiation reached in a particular basic rock when the plagioclases were precipitated, however, considerable more information is needed about the distribution of this element in the Gardar basic rocks before this can be attempted. It should be noted that plagioclases from the two gabbroic dykes which are regarded as the least fractionated of the Gardar gabbros (the Eqaloqarfia dyke and the Tûgtutôq giant gabbro; Table 4, nos 1 and 33) both contain plagioclases with a fairly high strontium content (1500–1800 p.p.m.).

Comparison between the strontium content of the felspathic inclusions with that of anorthosites and related gabbros in other parts of the world can only be tentative due to the lack of an absolute standard. However, it is worth noting that the values of 900–1500 p.p.m. given by Emmons (1953) from the Duluth gabbro and their associated felspathic inclusions correspond remarkably well with the granular anorthosites and their host gabbros.

Comparisons with the anorthosites of the Canadian Shield, which in the end may provide the best method of determining the origin of the anorthosites in Gardar rocks, is hampered because of lack of published material. Fortunately Papezik (1965) has summarised the geochemistry of many of the plutons and gives the results of a series of strontium determinations both on complete rocks and plagioclases separated from anorthosite. The average values for strontium in 25 anorthosites is 750 p.p.m. while those of 11 plagioclases separated from the Morin anorthosite is 913 p.p.m. These results are fairly close to the values found in the granular anorthosites although they are considerably lower than the average values found in the felspathic inclusions in Gardar rocks. However, it should be noted that the majority of the plagioclases analysed by Papezik are more sodic than the inclusions in Gardar rocks

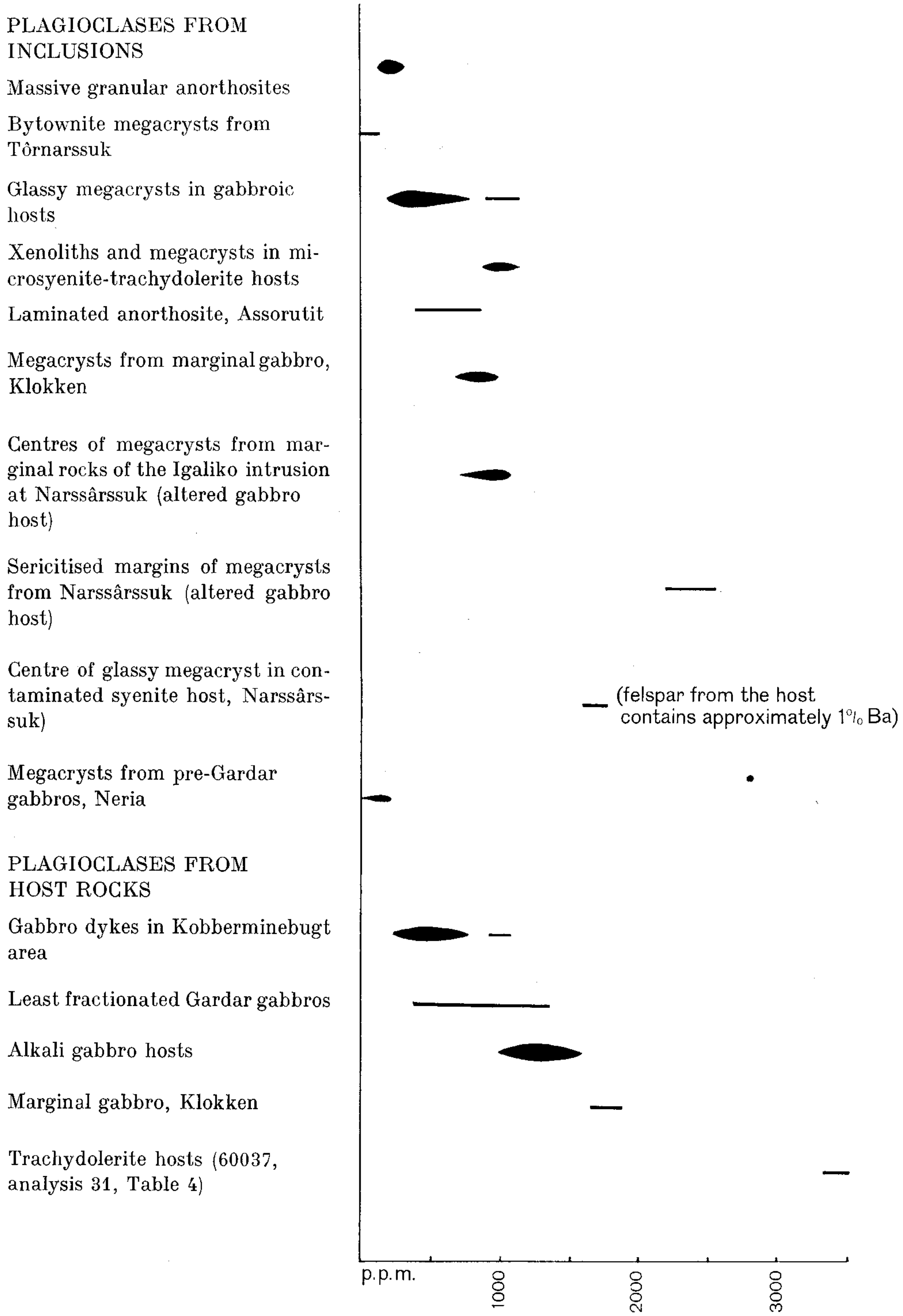


Fig. 53. Barium content of plagioclases taken from a variety of inclusions and basic igneous rocks in South Greenland.

and the strontium values for labradorite-bearing anorthosites from Canada are significantly lower than those obtained from Gardar rocks.

The strontium content of plagioclases from basic bodies in Somalia described by BUTLER and SKIBA (1962) fall into two ranges, 1000–1850 p.p.m. and 1650–2350 p.p.m. These authors conclude that the four masses were not part of the same layered intrusion on the basis of the strontium analyses. Similar arguments in the Gardar would certainly divide the granular anorthosites from the layered Assorutit block and the inclusions in the marginal rocks of the Igaliko complex.

The barium content of the felspars from the inclusions and their Gardar hosts (Fig. 53) followed the strontium fairly closely with a mean value of 300 p.p.m. in the granular anorthosites and megacrysts from the Kobberminebugt gabbros, and reaching 1700 p.p.m. in strontium-rich samples from Narssârssuk. However, there is a marked difference between the barium content of the insets and many of their host dykes and it is possible to trace a progressive increase of this element from the centre of glassy megacrysts outwards into the host rocks. This is particularly well seen in the series of samples from dyke 8 where the centre of the megacrysts show a low (200 p.p.m.) barium content while the alkali gabbro host shows an eightfold increase in this element (1700 p.p.m.). In contrast to the strontium content barium increases outwards into syenitic hosts presumably reflecting the increase in alkali felspar. This is well seen in the samples from Narssârssuk.

c) Plagioclase twinning

The study of twinning in plagioclase has received some emphasis in recent years due to the possibility that the type of twins developed gives some indication of the environment in which the plagioclase was originally formed or the processes through which it has subsequently passed (see, for example, GORAI, 1951; TURNER, 1951; EMMONS, 1953; TOBI, 1961; VANCE, 1961; RAO, 1964). At present the uncertainties attached to the interpretation of twinning patterns are too great to allow much weight to be put on them as the main criteria for the origin of particular rocks. However, they can be useful as secondary evidence to be used in conjunction with more reliable criteria.

In the study of the twin laws in the plagioclase from Gardar dykes it became obvious that in any one thin section twinning was the result of a variety of causes. Without definite criteria for distinguishing between different generations of twinning a statistical study of the twin laws is of doubtful value. The dominant types of twin were identified using the recent modification of Reinhard's method (SLEMMONS, 1962). Approximately 250 stereographic plots were made on thin sections of over a hundred samples. The sections were searched for unusual twin laws which were generally apparent due to their effect on polysynthetic albite and pericline twinning. In some cases the twins were difficult to identify with certainty due to the anomalous optical properties of the plagioclase.

The twinning seen in the felspathic inclusions varies according to the textural type of the xenolith and the composition of the plagioclase. Four main types have been recognised, corresponding to the laminated anorthosites, granular anorthosites, labradorite megacrysts and andesine-oligoclase megacrysts. Although there is some overlap between the groups the general features are sufficiently distinct to treat each type separately.

1) Twinning in the plagioclase from the Assorutit laminated anorthosite. Plagioclase from these anorthosites shows a twinning pattern indistinguishable from that in other Gardar gabbroic rocks. The (010) plane is dominant, with albite as the commonest twin law associated with some twinning on the Carlsbad and the Carlsbad-albite twin laws. Fine albite lamellae often alternate with broad albite lamellae across the crystal. There is frequently an abrupt change in the abundance of fine lamellae on either side of a particularly prominent (010) twin plane which may give the spurious impression of a Carlsbad or Carlsbad-albite twin, however, when plotted the majority of these are found to be albite twins. The majority of Carlsbad, or Carlsbad-albite twins determined occur as thin wedges separating broad albite lamellae. Rare simple twins on the Ala B law have been identified. Local subordinate patches of pericline inhibit the formation of the fine twins on (010). The pericline twins are often exceedingly fine and the boundaries of the areas in which they develop are indistinct.

Twinning on other laws is rarely seen. It is always simple, and may not be easy to identify even using the full plotting technique described by Slemmons. Some of the simple twins on a plane at right angles to (010) may be acline instead of pericline but in the composition range of these felspars it is not possible to distinguish the two with certainty. Rare apparent (010) twins do not show a twin axis when plotted while others show a large triangle of error. This may be caused by two different tabular felspars settling together on their (010) faces and being cemented by an overgrowth.

Some of the twinning is primary according to the criteria outlined by Vance (1961). Occasionally it is possible to show that broad primary twinning was formed before the plagioclase was completely crystallised. This is best seen where the compositional zoning is controlled by the original re-entrant form of the crystal (Fig. 54). Most of the finer twins are secondary according to Vance's criteria and may have developed either contemporaneous with the elimination of original zoning (Emmons, 1953), or are due to small internal stresses associated with crystal settling or contractions during the cooling of the rock. The final stage was the formation of fine tapering albite and pericline twins developed locally in response to external pressure or chemical attack on the crystals. These are seen where the plagioclase shows slight bending or dislocation, or where the margins of the crystals have been leached in contact with interstitial alteration products and alkali felspar.

2) Twinning in plagioclase from the massive granular anorthosites. The felspars in the massive anorthosites differ considerably in appearance from those in the layered rocks. They form an interlocking mass of anhedral grains with no preferred orientation. This difference in texture is emphasised by a different twinning

pattern. The (010) plane is no longer dominant and pericline and albite twins occur in approximately equal amounts. Cross-hatching is common with the pericline and albite lamellae showing the same variation in size. Twinning varies from one crystal to the next; in some crystals all the lamellae have approximately the same width, in the next thin albite and pericline twins are imposed on an untwinned background. Commonly one set of twins (either albite of pericline) is broad and rather irregular

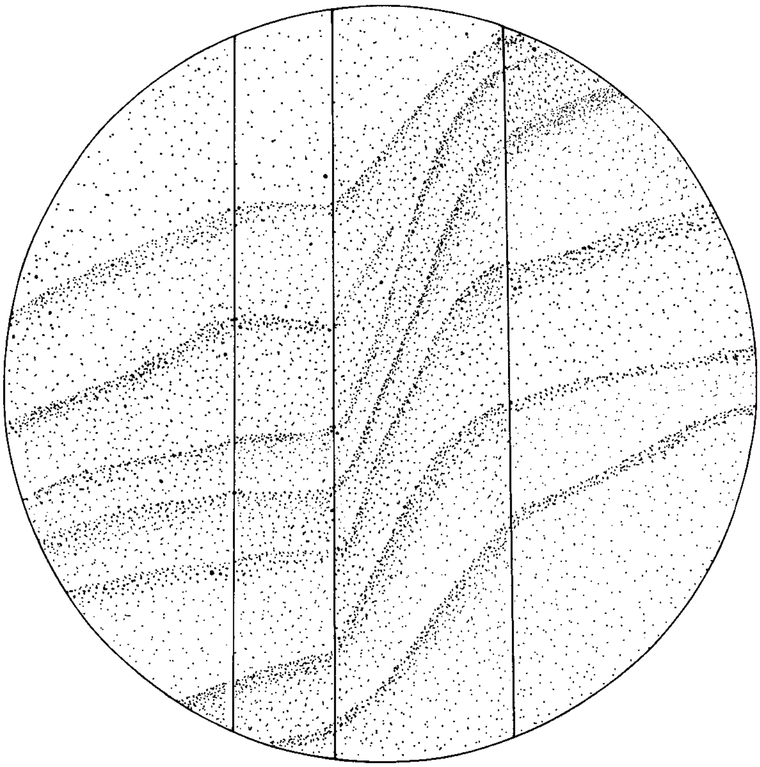


Fig. 54. Drawing of the relationship between twinning and zoning in a specimen (G.G.U. 25229) of a megacryst in the marginal gabbro surrounding the granular anorthosite block at Augiâta tasia. It can be clearly seen that the (010) twinning in this crystal controls the outer form of the crystal shown by the zoning. The twinning is therefore regarded as primary. The field of view illustrated is approximately 3.5 mm in diameter.

with a general spindle shape in section, while the set approximately at right angles is regular and has parallel sides. In general the irregular twins are the earlier.

Distortion of the twin lamellae is quite common in some thin sections and the twinning may show a sygmoidal form in the crystal. This distortion rarely reaches proportions suggesting that the granular anorthosites have been heavily deformed by external forces. Undeformed twin lamellae cutting the (presumably) earlier deformed lamellae are seen. In some crystals albite twins formed first, in others pericline. One set of twins may cut the thinner members of the other set, but where they meet wider lamellae of the second set there is commonly a diffuse area in the centre in which the orientation of the plagioclase cannot be resolved.

The twin pattern seen in plagioclase from the granular anorthosites shows some variation from xenolith to xenolith, apparently dependent on the crystal form of the constituent plagioclase grains. In the xenoliths composed almost entirely of interlocking anhedral felspars the twinning is almost equally developed on the albite and pericline laws. However, in samples taken from the partly laminated anorthosite from some of the blocks from Kobberminebugt the twinning is predominantly on the (010) plane and approaches the pattern seen in the Assorutit rocks. It would therefore seem unwise to place any genetic significance on twinning alone in these rocks. Twin laws other than albite and pericline are rare in the granular anorthosites; a few examples of Carlsbad-albite twinning and one of Manebach have been plotted.

3) Twinning in labradorite megacrysts. Twinning in the labradorite megacrysts and cleavage fragments is variable and where closely spaced can often be seen to be the result of local causes, such as strain, induced by the inclusion of foreign material with the crystals, or post-crystallisation distortion. Typically the twinning in the undistorted giant felspars is very simple with albite and pericline approximately equally developed. Other twin laws are virtually absent. A single case of albite-Ala A twinning has been noted and a few crystals show a blocky development of albite-pericline in which the composition plane corresponds to (010) in one lamella and the rhombic section in the next. Most of the giant crystals show only a few twin lamellae and extreme cases occur in which sections cut at right angles to (010) have shown no twinning. Generally narrow albite or pericline lamellae are found widely spaced on an otherwise untwinned background. The relative lack of complicated twinning may be one of the reasons why the giant crystals are often glassy clear and it is worth noting that glassy felspar megacrysts brought up in the recent volcanic activity south of Iceland (Surtsey) show similar widely spaced, narrow polysynthetic twins (material kindly supplied by S. Jakobsson). The absorbtion of light by heavily twinned plagioclase may be partly explained by the fact that the twin planes are commonly the site for slight alteration or the inclusion of foreign material. However, in some cases it may also be caused by internal reflection within the crystal itself due to the refraction of light at each twin plane.

Some of the giant crystals have developed close-set polysynthetic twinning which, in a few examples, appears to be due to strain. The individual twin lamellae may show distortion with thickening where they bend and the development of an intricate pattern of minute albite and pericline twins in the area surrounding the distorted lamellae.

Closely set polysynthetic twinning is characteristic of the felspar intergrowth texture seen in some megacrysts from alkali gabbro hosts. The intergrowth may take the form of two apparently separate plagioclase crystals each with its own polysynthetic twinning on either side of a single twin plane (commonly (010) for one individual and the rhombic section for the other). In other examples the intergrowth consists of an "inclusion" of one individual within a larger crystal so that the "inclusion" shows polysynthetic twins approximately at right angles to those in the host. The boundary between the two is frequently marked by small inclusions of olivine, chlorite, or in a few cases, gabbroic material. As the felspar on either side of the boundary are crystallographically related the intergrowth should be regarded as a single crystal. However, the appearance, both in hand sample and under the microscope, is of two interlocking individuals. A special case of this phenomena is seen in sample 58332 from the gabbroic centre of a member of the microsyenite-trachydolerite dyke suite cutting the South Qôroq centre of the Igaliko complex. The sample consists of a single crystal of plagioclase (An_{56}) which is composed of

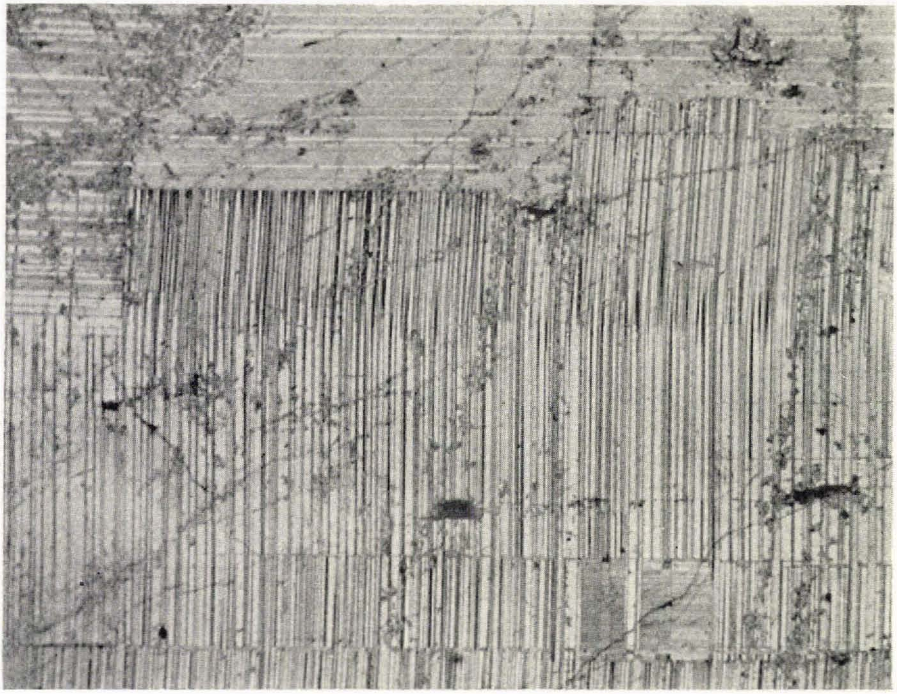


Fig. 55. Twinning in a megacryst from a trachydolerite dyke cutting the South Qôroq centre of the Igaliko complex. Crossed nicols 10 ×, G.G.U. 58332. The twin planes separating adjacent lamellae correspond to the (010) direction in one individual and the rhombic section in the next. The megacryst is divided into approximately square-sided areas defined by the (010) and pericline twin planes. Some areas in the crystal show signs of strain presumably related to the unusual type of twinning.

closely spaced twin lamellae arranged in blocks set at right angles to each other (Fig. 55). Some of the lamellae are distorted. The most remarkable feature of this crystal is not the block texture but the relationship between individual lamellae within the block. The composition face between two adjacent lamellae is the (010) face for one individual and the rhombic section for the next. This is repeated throughout the crystal.

In some of the large crystals with broader lamellae than normal the lamellae of one set of twins (either albite or pericline) are slightly spindle shaped in thin section, while the twin lamellae of the other, generally younger, set have planar composition planes. Associated with the broad lamellae there is often a later generation of fine twin lamellae parallel to the early spindle shaped set along the borders of the second set.

The twinning of the black felspar is identical with that of the white felspar from the same host and twin planes cut across the patches of black and white material indiscriminately.

Local distortion of the plagioclase crystals has resulted in the production of considerable secondary twinning. Although some of this may be due to crushing during transport it can often be shown to have developed comparatively early in the history of the megacrysts. This is best seen in those crystals that have developed late

Fig. 56. A block of plagioclase (G.G.U. 61073) which formed part of a single crystal measuring at least 50 × 40 cm. The crystal consists of dark-brown albite lamellae (approximately 1 mm wide) alternating with white lamellae which are generally less than 1 mm wide. White veins cut both the dark and the light felspar. These are composed of fresh plagioclase which does not disrupt the twin pattern.

fracture planes cutting across the crystal, often at irrational angles to the crystallographic directions. The fracturing may be so intense that the crystals break preferentially along the fracture plane instead of the cleavage plane. The crushing itself may be fairly early in the history of the crystal as it sometimes controls the sericitisation of the plagioclase or contains veinlets of material from the host dyke. However, in spite of the intenseness of this mechanical action no secondary twinning has been noted resulting from it. It seems possible that secondary twinning can only develop above a certain temperature, or when the plagioclase is recrystallised by severe chemical action. This may give some support to the ideas put forward by Laves (1965), who suggested that secondary twinning could only take place if the plagioclase was in a partially disordered state. The disorder in the case of the plagioclase megacrysts might be an original feature representing the state of the plagioclase soon after crystallisation, or it might be due to secondary effects of heating by the host dyke during transport.

The relatively stability of the plagioclase twin pattern is seen in some of the black felspars described in a following section (p. 109). Many of the crystals appear once to have been black throughout but have lost their colouration in patches. This loss in colouration took place without any change in the twin pattern. A particularly good example of this phenomena is shown in Fig. 56 where late veins of white felspar cut a single crystal of plagioclase without apparently disrupting the twinning. The crystal consists of alternating brown and white albite lamellae, the brown averaging two to three times the thickness of the white. The brown twins contain extremely closely spaced pericline twins throughout their length (Fig. 57). These pericline twins persist in the areas where the white veins "cut" the crystal but are completely absent from the white twins. Pericline twinning is so intense that the lamellae will not extinguish but show anomalous blue interference colours. While the cause of the dark colouration in many of the anorthosite felspars is not known it seems possible that the colouration in this crystal may be due to peculiar optical effects caused by the close-set pericline twins.

Fig. 57. Photomicrograph of a thin section of the sample shown in Fig. 56. Crossed nicols 25 ×, G.G.U. 61073. Note the exceedingly fine pericline twinning in one set of albite twins. This set corresponds to the dark lamellae seen in Fig. 56.

Small inclusions of plagioclase in Gardar dolerite dykes show similar twin patterns to the plagioclase of their hosts although there is every gradation between the comparatively simple twinning of the giant felspar and the complex albite, Carlsbad-albite and pericline twinning found in the smaller crystals.

Twinning in the large plagioclase megacrysts from the margins of the Klokken and Igaliko syenite complexes is generally extremely simple, consisting of regularly spaced albite twins alternating as broad and narrow lamellae. Some fine pericline twinning may be present.

The twinning seen in the felspar megacrysts from the pre-Gardar dykes of the Neria region resembles that of the giant felspars in Gardar dykes; fine albite and pericline lamellae in a generally untwinned "groundmass" are the commonest. Some of the crystals in the pre-Gardar dykes are violently zoned against their host.

4) Twinning in andesine-oligoclase megacrysts. Virtually all the fresh megacrysts with a lower anorthite content than An_{50} are from the basic centres of the microsyenite-trachydolerite host dykes. These dykes also contain some sodic labradorite megacrysts and there is no sharp boundary between the type of twinning found in, for example the giant labradorite megacrysts, and the type of twinning found in the smaller and less calcic megacrysts. As the anorthite content falls, however, there are certain characteristic changes in the twin pattern. The most noticeable of these is an increase in the number of twins per millimetre with decrease in anorthite content. The majority of large labradorite megacrysts show relatively few polysynthetic twins per millimetre while the smaller andesine-oligoclase crystals show the development of close-set albite twinning. This reaches its maximum development in the plagioclase centres to alkali felspar rhombs where the polysynthetic twinning is so fine that it is barely resolvable. Typical andesine megacrysts show properties halfway between these extremes.

In the andesine megacrysts polysynthetic twinning is well developed, especially on the (010) plane; the crystals may be divided into broad segments either by Carlsbad, Carlsbad-albite or simple albite twins and the broad segments each contain finer albite twins set in an untwinned groundmass. The broad twins can be shown to be primary as they often coincide with changes in the outer form of the crystal (VANCE, 1961). They frequently end abruptly. In thin section many of the andesine megacrysts can be seen to contain polysynthetically twinned insets of plagioclase with approximately rectangular boundaries. These insets are identical in composition with the surrounding "host" crystal. When plotted on a stereogram it is found that the (010) plane of the insets is parallel to the rhombic section of the "host" and they are thus a special type of albite-pericline twin. A few broad, simple, pericline twins and a single albite-Ala A twin pair have been noted. The broad (010) lamellae, the larger pericline twins, and the inset structure all appear to be primary and formed before the sericitisation of the plagioclases which often accompanied their inclusion in the host dykes. The majority of inclusions in the less basic hosts show marked zoning round the margins. In many cases this appears to be a secondary feature as the zoning can be seen following the fractured margins of incomplete megacrysts. The zoning is presumably due to the loss of calcium during the reaction between the megacrysts and their hosts. This loss in calcium is frequently marked by an increase in the number of albite and pericline twins, which are thus clearly secondary in origin.

Most of the fine polysynthetic twinning found in the andesine megacrysts appears to be secondary according to the criteria outlined by VANCE (1961). However, it is not known at what stage it developed, and it could either represent twinning which formed soon after the crystallisation of the plagioclase, or it might reflect a later process such as inclusion in the Gardar dykes. Most of the secondary twinning appears to have developed before the sericitisation of the plagioclase, although in a few crystals closely spaced secondary albite twinning developed in recrystallised, less calcic, plagioclase surrounding cores of sericitised plagioclase with a simpler twin pattern. Some of the secondary twinning can be seen to have formed as the result of mechanical strain, there being an increase in the number of secondary lamellae where primary lamellae have been bent or disrupted. Other crystals show a series of dislocations later than the secondary twinning.

Secondary pericline twinning is also common in the andesine megacrysts. In some cases it may be restricted to one set of primary albite twins, and it may locally inhibit the development of secondary albite twinning. In other crystals patches of secondary albite and pericline twinning alternate.

d) Relationship between composition, optical orientation. and the structural state of the plagioclase

The plagioclase from the felspathic inclusions should provide ideal material for a study of the relationship between optical orientation and chemical composition. Crystals with broad simple twin lamellae and homogeneous composition are common and it is comparatively easy to prepare thin sections in any desired plane.

The exceptional coarse grain size and the lack of compositional zoning suggests that the felspars were formed slowly under conditions of local equilibrium. It would therefore be expected that they show

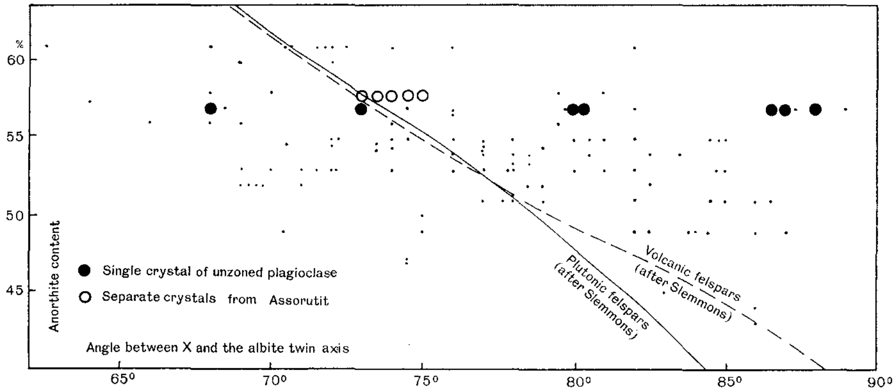


Fig. 58. The angle between X and the albite twin axis plotted against composition. Determination of the anorthite content of the plagioclases using the orientation of the optical indicatrix may be 25–30 % in error.

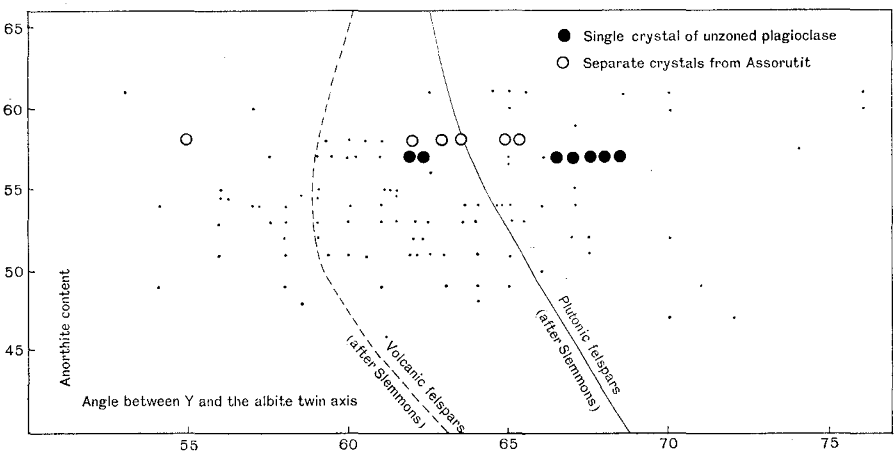


Fig. 59. The angle between Y and the albite twin axis plotted against composition.

consistent plutonic structural states (in the sense used by SLEMMONS, 1962).

Large discrepancies were found between the composition of individual plagioclase crystals determined using standard curves relating optical orientation and anorthite content and the composition of the same felspars using refractive index measurements. The variation between the angles of X and Y to the albite twin sxis for plagioclase crystals of similar compositions may be seen in Figs. 58 & 59. Similar results were obtained for Z, and for other twin laws. The discrepancies are outside the range of instrumental and personal error and resemble those found by EMMONS *et al.* (1960) in their study of less basic plagioclases from granitic rocks. W. S. WATT, working independently on megacrysts collected from Gardar dykes in the Qaersuarssuk area, obtained similar results to those described in this paper.

The most noteable features are described below:

1) The anorthite content obtained by plotting optical directions differed by as much as 25 % An from that obtained by refractive index determinations. 2V measurements agreed with the compositions obtained by refractive index determinations and gave consistent results on the curves for plutonic plagioclase. The An/(An + Ab) ratios of three felspars partially analysed were within 2 % of the An content of the same felspars given by the refractive index determinations. This is within the experimental error of the analytical method used. (The analyses suggested that the plagioclases contained approximately 3 % of the potash felspar molecule. As this is not determinable by optical methods it has not been allowed for in the anorthite percentages quoted. However, it is most unlikely that the potash felspar contained in the felspar could account for the marked discrepancies described here). Identical samples of plagioclase to those determined by refractive index measurements were analysed for sodium using neutron activation by J. R. BUTLER (Imperial College, London). The results are all within 1 % of those obtained using refractive index measurements.

2) Some crystals showed a constant discrepancy between the results obtained by refractive index determinations and those from universal stage measurements; other crystals showed no such consistancy, the optical directions in one set of twins showing a markedly different orientation to the next. In one example, a stereogram of which is shown in Fig. 60, the orientation of the optical directions with respect to the twin axis remained constant in one albite twin lamella. while in the second lamella the position of X and Z varied by 26° and Y remained constant at different positions along the length of the twin. Refractive index measurements of fragments from the same crystal showed that the composition remained constant. Adjacent crystals with similar compositions commonly show completely different optical orientations with respect to their twin axes.

3) Some crystals showed a consistently large triangle of error when the results of several measurements were compiled onto one stereogram, other crystals were variable and still others gave consistently small triangles of error. The divergence between the results obtained by refractive index and optical orientation determinations showed no relationship to the size of the triangle of error; many stereograms in which the triangle of error was negligable gave angles of X and Y to the twin axis which did not lie near to any of the standard curves. In a few cases it was impossible to identify the type of twinning present using Slemmons'

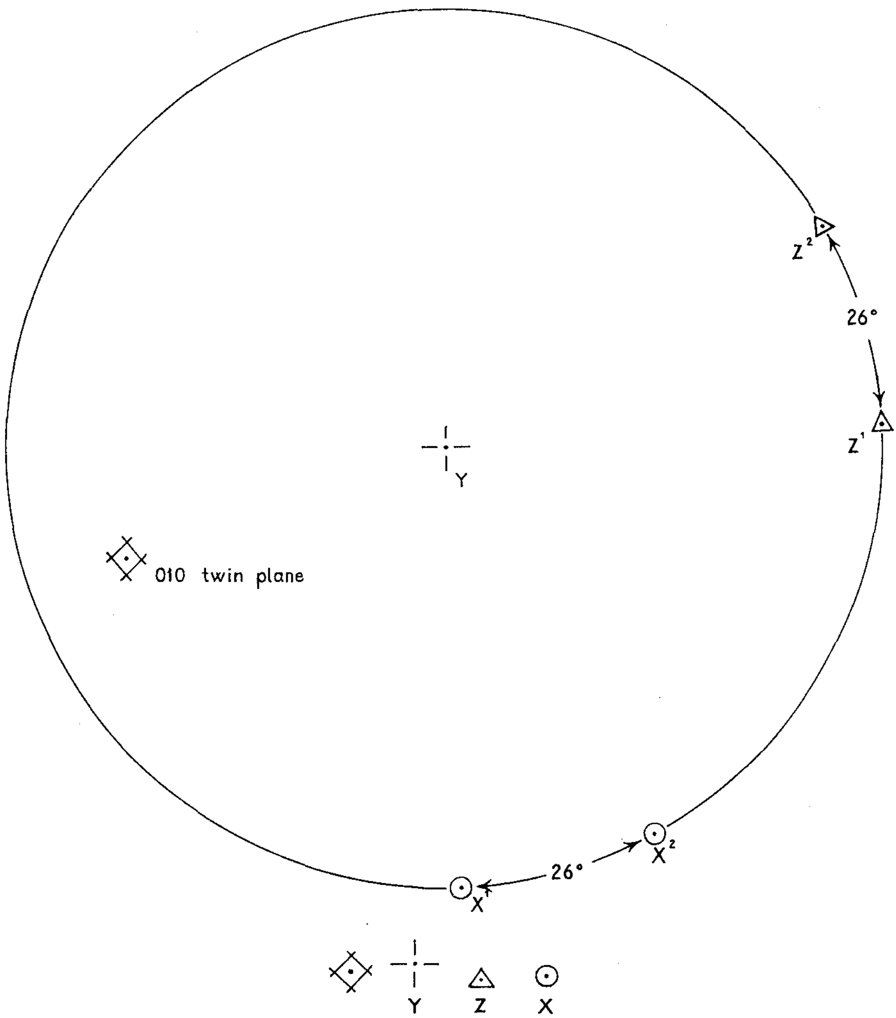


Fig. 60. The change in the position of the optical indicatrix with respect to the albite twin axis in one unzoned albite lamella from a megacryst.

method as the points lay far outside the areas enclosed by the standard curves.

4) Many of the albite twins showed marked differences (up to 10°) in the maximum extinction angle on either side of the vertical twin plane although plotting showed no particular abnormalities present and the refractive indices of the adjacent twins were constant.

5) The structural state (plutonic, intermediate or volcanic), as determined by optical measurements, was highly variable and showed little correlation with the field setting of the xenoliths. Samples taken from

the centre of coarse-grained anorthosite inclusions several tens of metres in diameter showed structural states varying from "more plutonic" than plutonic to "more volcanic" than volcanic on the standard graphs. This variation occurs from grain to grain and from twin pair to twin pair. Measurements of the structural state of felspars from the Assorutit laminated anorthosite inclusion showed more consistent results than those from granular anorthosites or megacrysts. However, even in the Assorutit felspars occasional twin pairs gave readings suggesting a volcanic structural state.

6) Both primary and secondary twins showed the same discrepancies.

7) X-ray powder diagrams and single crystal photographs (p. 107) also suggest that the structure of the felspathic inclusions is abnormal.

8) The comparatively simply twinned, giant, glassy felspars which should provide the most suitable material for this work were among those felspars with the highest discrepancies.

9) The discrepancies were not limited to particular twin laws: albite, Carlsbad-albite, Carlsbad and pericline could all show similar departures from normal curves.

10) Several pseudo-twins were found, especially in the Assorutit anorthosite. These are apparent twins in which one unit shows no rational relationship to the twin plane. Other apparent twins showed so large a triangle of error that it was impossible to identify the twin axis.

11) As a check glassy felspar from Surtsey was determined using the same techniques and instruments. The results given by SLEMMONS' curves and the refractive index measurements were within 1 $^0/_0$ of each other. As might be expected from their environment the felspars from Surtsey showed volcanic optics. Preliminary work on the Gardar gabbros suggests that some of these contain felspars with normal plutonic optics. However, some resemble the plagioclase inclusions and a thorough study would be worthwhile.

The above results suggest that the orientation of the principal optical directions of the felspar inclusions is controlled by several factors other than chemical composition, and plutonic or volcanic origin. Unless these factors are known the optical orientation can only be used as an approximate guide to composition and gives little direct information about the conditions under which the felspars crystallised. This is in agreement with the results published by EMMONS *et al.* (1960) and not with more recent work (for example, NOBLE, 1965). Plagioclases from some groups of rocks, for example the volcanic xenocrysts from

Surtsey and the calcic plagioclases found in the regionally metamorphosed gabbro anorthosites north of Ivigtut, show no discrepancies between the composition determined by refractive index measurements and those obtained by measurements of the optical orientation. In other plagioclases, such as those described here, the discrepancy is so large that the position of the principal optical directions may give little idea of the composition or structural state of the felspar. The differences are real and reflect fundamental properties of plagioclases from different geological settings.

The most interesting work on similar discrepancies is that described by Vogel (1964) as optical-crystallographic scatter. Vogel divides his observations into two groups: scatter seen in individual grains (internal scatter) and those seen in different grains in the same sample (external scatter). Vogel ascribes internal scatter to the formation of imperfect twins and states that it is found in plagioclases taken from all geological environments. It is generally restricted to a few grains per slide. External scatter, Vogel believes, is due to the twinning developing at any time during the cooling process and thus plots from the different grains may show a spread between high and low structural states. He notes that external scatter is especially common in "coarse-grained, igneous-appearing rocks such as gabbro, anorthosite, norite and pyroxenite". This description fits the felspathic inclusions in the Gardar rocks. However, Vogel's results cannot be applied *in toto* to the Greenland felspathic inclusions partly because using his definition most of the scatter observed in the Greenland inclusions is internal, as the majority of samples are single crystals and partly because the Greenland rocks appear to have passed through a more complex history.

In order to make comparisons between Vogel's results and those from the Greenland inclusions more useful it is necessary to restrict the term internal scatter to the scatter of plots from obviously imperfect twins, for example those with large triangles of error or the "pseudotwins" described under (10) above, while external scatter may be used for changes in apparent structural state between different "perfect" twin pairs from different places in the same grain. With this slight modification the following points seem of interest: Internal scatter is particularly common in the Assorutit laminated anorthosite and in samples of granular anorthosite which have moderately elongate plagioclase crystals with well-developed (010) faces. This suggests that internal scatter may be due to the combination of two crystals which settled together on their (010) faces. Although Vogel associates internal scatter with no particular geological environment it would seem probable that it might be common in igneous rocks with platey plagioclase.

External scatter is very common throughout the inclusions. It is well marked in the giant felspars. "Perfect" twins, *i.e.* those in which the adjacent lamellae show identical relationships to the twin plane and which do not give appreciable triangles of error, may show a scatter outside the limits of the high and low structural curves. This applies to all types of twinning and differs from the results published by VOGEL who found that secondary twinning and Carlsbad and Carlsbad-albite twins showed little external scatter, at least within the same sample.

Part of the explanation is probably to be found in the complex history of the felspathic inclusions. Many formed a solid rock later broken up and included in fresh gabbroic magma, which may well have led to secondary twinning forming at higher, local temperatures than the original primary twinning. Similarly, local heating effects on plagioclase lamellae forming part of primary Carlsbad twinning may have altered the structural state of the felspar in one part of the crystal but not in another. The fact that many "perfect" twins show a scatter outside the limits of the normal curves may be due to strain in the structure. These anomalies are commoner in samples taken from inclusions in the dyke rocks than they are in samples from the margins of the Klokken and Igaliko intrusions. This may be partly due to the relative speed of cooling of the two types of host. Inclusions in the dykes may have been in equilibrium at depth, either as massive rocks formed slowly under plutonic conditions, or as megacrysts suspended in a deep seated magma chamber. When brought up in the Gardar dykes they may have been reheated and then cooled too quickly to allow them to adjust fully to their new environment. In contrast, the megacrysts in the margins of the larger bodies may have either crystallised direct from a magma closely related to their present host, or they may have had longer to adjust to the new conditions. However the common occurrence of defects of this type from coarse-grained igneous rocks suggests that some of the anomalies may be original features formed at the same time as the original crystallisation of the plagioclase. (In the next section of this paper (on X-ray investigations) it is suggested that the defects are not due to twinning, at least on the microscopic scale, but are lattice defects.) A possible explanation would seem to be that the first formed centres on which the plagioclases crystallised were high temperature forms. These were followed by the crystallisation of low temperature plagioclase with different crystal constants resulting in severe strain in the structure.

Although the high incidence of external scatter found in felspathic inclusions resembles that found by VOGEL in basic rocks of igneous aspect it cannot be used to prove the igneous origin of the felspars since many of the more marked features might be the result of later effects on already crystallised material. However, none of the xenoliths show the lack of scatter seen in regionally metamorphosed rocks.

e) X-ray investigations on the structural state of the felspar

The angular separations Γ and B (SMITH and GAY, 1958) were measured on X-ray powder patterns of 36 felspar samples. Single crystal X-ray photographs were made on samples of black and clear felspar from the same megacryst. The detailed results of both studies are given in the Appendix (p. 211).

Discussion of the results.

There were two main reasons for studying the felspar using X-ray techniques; firstly to see if there was any evidence that the black colouration of some of the felspars was due to a structural defect which was not present in the clear felspar, and secondly to see whether the variation in properties found using universal stage measurements could also be detected by techniques which did not rely on the presence or absence of perfect twins.

Direct comparisons between the abnormal properties found by universal stage measurements and the X-ray powder patterns is impossible since most of the crystals examined optically show different properties from one twin pair to another. These might be expected to average out when a powder pattern is taken of felspar from many different twin lamellae. However, powder patterns taken of samples separated from different parts of the same megacrysts show similar variations to those found using optical methods. Powder patterns showed little difference between black and clear felspar; if there are any differences in structure between the two types of felspar these are masked by a much stronger scatter of properties seen in apparently identical samples from the same megacrysts. However, as shown in Fig. 76b in the Appendix, some of the black felspars show a clear separation of peaks in the 2θ range 23°–24°. This separation is lost in heated samples and in the clear rims surrounding the black megacrysts.

Individual samples, in some cases taken from apparently untwinned felspar, show a remarkable spread in the 131 reflection which leads to considerable imprecision in Γ measurements. In some samples there are apparently two 131 reflections which give values of Γ falling respectively near the plutonic and volcanic curves drawn by SMITH and GAY; in other samples the 131 peak is broad and diffuse suggesting that the crystals possess a complete spread in properties between the volcanic and plutonic forms. The values of Γ obtained had little connection with the properties which might be predicted on geological grounds. Samples from the massive granular anorthosites, which could be expected to be slow cooling and "well ordered" often gave values of Γ close to the volcanic curve of SMITH and GAY. This suggests that the variation in the position of the 131 reflection may be controlled by other factors than thermal state and chemical composition. However, it can be noted that samples of plagioclase from the Assorutit anorthosite and megacrysts from Klokken, both of which are thought to have passed through simpler histories than the majority of inclusions, showed X-ray powder patterns with well defined double 131 peaks, giving values of Γ very close to the plutonic and volcanic curves of Smith and Gay. As a control samples from the regionally metamorphosed anorthosites found as concordant layers in the gneiss north of Ivigtut and a clear glassy megacrysts from the recent volcanic activity on Surtsey were examined using the same instruments and techniques. These gave single 131 reflections and values of Γ which fall respectively on the plutonic and volcanic curves drawn by Smith and Gay. Optical measurements on the same samples gave a similar lack of anomalous properties.

When a sample of black felspar was heated until it lost its colour (between 800 and 1000°C) the original value of Γ changed until it approached the volcanic curve. However, a complete disordering of the original crystal did not take place at the temperatures used and a weak double 131 reflection persisted.

The structural state indicated by the separation B was generally fairly close to that indicated by Γ. In contrast to Γ only one value of B was measurable. This is probably due to the relatively small separation between the two reflections used in obtaining B.

The main conclusion to be drawn from the X-ray powder diagrams is that the anomalies found by the universal stage measurements are not due solely to imperfect twinning but reflect a more fundamental property of the crystals. Some of the results are probably due to a spread between high and low temperature forms of plagioclase but the problem would appear to be considerably more complex than previously suggested. In many respects the results resemble those suggested by Vogel's (1964) optical work on gabbroic rocks. However, the fact that high temperature forms of plagioclase can be seen at all by X-ray powder determinations suggests that the original structural state is locally preserved within the crystal and is not only reflected by the optical orientation with respect to the primary twin planes.

It seems likely that both the optical crystallographic scatter and the X-ray crystallographic scatter may be due in part to the complicated history through which the fragments have passed. In the Assorutit and Klokken felspathic inclusions the scatter may be due to changes between high and low temperature forms which occur in a coarse-grained gabbroic body while cooling. However, many of the smaller xenoliths and megacrysts have a more complex pattern which may be partly derived from original, possibly igneous features, and partly derived from their history as inclusions in gabbroic rocks.

The single crystal X-ray photographs (Appendix, p. 220) showed that both the black and clear varieties of felspar were structurally abnormal and provided no proof that the colouration of the black felspars was due to structural defects. It appears that only the most refined techniques which can measure the amount of any dislocation in the felspar structure would be useful in solving the origin of the colour.

The most striking result of the single crystal studies was obtained from the black fragment (An_{51}) which gave a measured angle β^* of 64°50'. This is approximately 1° larger than that given by HINTZE (1897) for sodic labradorite and is outside the range for this angle recorded for plagioclase. The white fragment studied did not give β^* by direct measurement; the average value calculated, 63°17', is significantly lower than the angle normally found in labradorite.

f) Black felspar

Glassy, black felspar megacrysts resembling obsidian are found unequally distributed in the felspathic rocks throughout the area described in this paper. They are especially abundant in Gardar dykes near the mouth of Kobberminebugt and in the marginal facies of the Klokken and Igaliko syenite intrusions. They occur in both the gabbroic and trachydoleritic host dykes and examples have been noted as megacrysts in the unmetamorphosed pre-Gardar olivine dolerite dykes of the Neria area.

Typically the black felspars are found in the same hosts as the large megacrysts of clear, glassy felspar which they resemble closely except for colour. In the majority of dykes the dark colouration appears to be unstable and the black felspars are rimmed by clear material. However, in a few examples all the megacrysts and the majority of host rock plagioclases are black and the colouration appears to have been stable during the crystallisation of the host dykes. These dykes also contain considerable primary hornblende instead of augite and olivine, a fact which may have some bearing on the genesis of the colouration. The black crystals generally occur as single megacrysts or occasionally in coarse, loosely packed aggregates forming the pseudo-ophitic texture described on p. 31. There are no records of black felspar forming major constituents of any of the granular anorthosite blocks, although rafts of plagioclase-rich material in the margins of the Igdlerfigssalik centre of the Igaliko syenite complex contain occasional fragments of dull, dark coloured felspar. The large black crystals occurring as megacrysts in the Igdlerfigssalik centre are generally less glassy in appearance than the black felspar seen elsewhere. A considerable proportion of the plagioclase megacrysts forming components of the secondary anorthosites at Bangs Havn and Narssaq are gray; this is thought to indicate that they may have passed through a black stage.

Most specimens of black felspar contain trains of coloured minerals parallel to either the (010) face or the rhombic section. Although these inclusions may have been formed at the same time as the black colour, they are not the direct cause of the colouration as they occur to approximately the same extent in the clear felspar megacrysts. Many of the black felspars are veined and rimmed by clear, colourless felspar which may follow definite crystallographic planes (especially those containing inclusions of coloured minerals) or which may form irregular anastomosing networks replacing the darker material. In some examples the younger material is clear and glassy, and it mantles the broken black megacrysts in a manner that suggests that the clear material has formed from the black felspar under the influence of the surrounding host rock (Plate 1b).

Under the microscope the colour is only faint in thin sections of normal thickness; with the naked eye a pale smoky appearance is visible when the thin sections are held up against a white card. This smokiness is quite pronounced in small fragments used for refractive index determination. Using an oil immersion lens and a total magnification of 1500 diameters the mat felspars from the Igdlerfigssalik centre and Klokken show minute rod-like and dendritic inclusions aligned parallel to (010) resembling those described in cloudy olivine (BERTHELSEN and BRIDGWATER, 1960, fig. 25). These are too small for identification but may contain both opaque and non-opaque phases (cf. WILCOX and POLDERVAART, 1958). But the black glassy material from Gardar and pre-Gardar dykes contains nothing which can be resolved as a separate mineral phase. The black felspars were found to be almost completely homogeneous when examined using phase contrast and dark field microscopy.

Preliminary investigation using electron microscopy on replicas of a fracture surface at magnifications up to $\times 8000$ has not shown a second mineral phase, but has revealed the presence of areas within the black felspar with an abnormal structure. This is seen in Fig. 61; the area to the right of the photograph is typical of the (001) surface of a normal felspar when seen under high magnification. The confused area on the left of the photograph is abnormal and according to H. MICHEELSEN (personal communication) is typical of crystals with closely spaced lattice defects.

i) Properties of the black felspar

In spite of the very obvious difference in colour between the black megacrysts and clear megacrysts from the same host there are remarkably few differences between the two types of felspar. Fresh samples of

Fig. 61. Electron micrograph (carbon replica) of an (001) fracture surface of a black felspar. (Photograph H. Micheelsen). Approximately 8000 ×, G.G.U. 49332. The area on the right of the figure is typical of a normal felspar, while the area to the left is abnormal and resembles electron micrographs of crystals with closely spaced lattice defects. The irregular dark areas are probably small felspar fragments on the surface of the specimen.

both felspars show the same glassy surface and commonly develop a conchoidal fracture which may cross the boundary between the two types without sign of a break. During manufacture of thin sections it appeared, however, that the dark felspar was considerably more liable to "plucking" than the clear felspar in the same section. This suggests that the black felspar is more brittle than the clear felspar; a property possibly allied to the presence of lattice defects.

1) Refractive index determinations. Refractive index measurements of X, Y and Z (p. 221) on a large number of samples of black

felspar and the colourless rims surrounding them showed that the black felspar had a composition in the range An_{48} to An_{57} with most between An_{50} and An_{53}. The colourless rims were generally 2–4 % less calcic than the black material. This might reflect original zoning.

2) Chemical composition. Partial chemical analyses of black and colourless material from single crystals showed no significant difference in major elements and gave anorthite contents within one percent of those deduced by refractive index measurements. In the four samples analysed, K_2O varied between 0.6 and 0.7 % equivalent to between 3 and 4 % of the potash felspar molecule in the felspar. The total iron shown by standard chemical methods varied between 0.4 and 0.45 % Fe in both the black and the white felspar. No difference was detectable in the oxidation states of the iron in the two types of felspar but at this low concentration it is doubtful whether the methods are reliable.

X-ray fluorescence trace element analyses of two pairs of black and colourless felspar showed no significant difference between the black and colourless material. The black felspar, its colourless rim and a clear felspar megacryst that was used as a standard showed an identical iron content (approx. 0.5 % Fe). A slight change was noted in the Ti content ranging from approximately 0.03 % Ti in the black felspar. 0.025 % in the colourless rim and 0.015 % in the control glassy felspar. Optical spectrographic determinations on several sample pairs of black and clear felspar taken from individual crystals showed no significant differences between the two types of felspar. The iron content of the felspars is not unusually high (cf. labradorite analyses published by Deer, Howie and Zussman, 1962–1963).

C. H. Emeleus (personal communication) reports that preliminary results using a microprobe suggest that turbid areas within the black felspars show a significant increase in iron. This is thought to be due to inclusions of optically resolvable iron-bearing minerals which are found along some of the twin planes in these felspars. In most of the samples examined optically the areas which are turbid in thin section are generally white and not glassy when seen in hand sample and appear to have been formed by secondary alteration.

The water content as determined by loss on ignition and Penfield's method on 10 mg of material is less than 0.25 % in both black and clear felspar. This means that any difference between the two is within the margin of error of the analytical methods used. Further investigations on the water content of the felspars using thermogravimetric methods are described in detail later.

3) Structural states of the black and clear felspars. The black felspars commonly plot closer to the plutonic curves than do the clear felspars from the same megacrysts. This is seen both in the X-ray powder data and in universal stage measurements of the relationship between the principal optical directions and twin axes. The felspars

from the pre-Gardar dykes and the felspars from the Igaliko and Klokken intrusions show this property particularly strongly. However, not all the black felspars show plutonic optics, and since the majority of felspars either from the anorthosite xenoliths or the plagioclase megacrysts show structural abnormalities it seems unwise to place any interpretation on this ill-defined difference between the two felspar types.

4) Twinning. In the majority of black crystals the twinning was identical to that found in the clear plagioclase megacrysts from the same environment. Where clear felspar rimmed black material the twin pattern crossed the boundary between the two varieties of felspar without a break. As described on p. 96 the twinning in the large megacrysts is generally limited to thin lamellae on either the albite or pericline twin laws set in untwinned plagioclase. The relative lack of twinning was confirmed by the single crystal X-ray photographs which showed its complete absence on a submicroscopic scale in the black and clear samples examined. The only exception to this rule is the crystal described on p. 98 and illustrated in Figs. 56 and 57. It is tentatively suggested that the close-set pericline twinning in this crystal may be an allied phenomena to the H_2O-bearing stacking faults in untwinned material, the presence of which is suggested by the thermogravimetric results described below. Late veins of chalky white fresh plagioclase cut the megacryst without disrupting the twin pattern.

5) Magnetic properties. It proved remarkably easy to separate black and colourless portions of the same plagioclase crystal using a magnetic separator. The black felspar has about the same susceptibility as augite while the clear was, for all practical purposes, non-magnetic. Generally it was possible to separate any crystal into three fractions: a highly magnetitic portion consisting of small grains of included magnetite, olivine and felspar with inclusions; a clean fraction of fresh black felspar with no inclusions; and a clean fraction of clear non-magnetic felspar. Similar separation of black felspar has been described by Herz (1951, p. 885); however, it should be emphasised that in the felspars described by Herz the magnetic fraction contained optically resolvable inclusions.

Measurement of the magnetic properties of the felspars was attempted by R. Girdler (University of Newcastle) using a sensitive magnetometer. This gave satisfactory results for the colourless material, but was unsuitable for measuring the black felspar due to the fact that fragments of crystals large enough for determination are quite likely to contain inclusions of magnetite or olivine. In order to eliminate this source of error a second series of determinations were carried out using a magnetic separator, in the way described by McAndrew (1957), to measure the susceptibility of powdered material from which all resolvable inclusions had been removed. This method proved quite sensitive enough to show the difference between black

Table 3. *Susceptability of black and clear felspar*

a) Using a sensitive magnetometer		
49332a	Black felspar (with some inclusions)	3×10^{-4}e.m.u./cm^3
49332b	Clear felspar surrounding 49332a	1×10^{-6}e.m.u./cm^3
25171	Clear glassy felspar	1×10^{-6}e.m.u./cm^3
b) Using a magnetic separator		
49332c	Powdered black felspar	9×10^{-5}e.m.u./cm^3
49332d	Powdered clear felspar	5×10^{-6}e.m.u./cm^3
49332e	Heated and powdered black felspar	4×10^{-6}e.m.u./cm^3
$758\frac{AT}{59}$	Powdered black felspar	6×10^{-5}e.m.u./cm^3
55464c	Gray felspar (from Klokken intrusion)	1×10^{-5}e.m.u./cm^3

and colourless felspar although the values obtained for the clear felspar are probably not as precise as those obtained by GIRDLER. The results given by the two methods (Table 3) are in reasonable agreement with each other; the high value obtained by GIRDLER for a fragment of black felspar is probably due to resolvable inclusions in the crystal measured.

6) Effect of heating. A powdered sample of black felspar, together with a few small fragments of black felspar, were heated in a furnace to 800°C for 336 hours. No change was apparent. The temperature was increased to 1000°C and heating continued for another 336 hours. At the end of this time the felspar was found to have become colourless. It remained glassy and showed no signs of fusion. The heated felspar showed a faint but distinct reddish colour, apparently due to ferric oxide. When placed in a magnetic separator the heated felspar proved to have lost its original magnetic properties and resembled the clear felspar from the same sample (Table 3). The effect of heating on the X-ray powder pattern is shown in Fig. 75 and discussed on p. 108. When heated moderately rapidly to 1200°C during the thermogravimetric studies the black felspar only lost part of its colour by the end of the experiment, the fine powder and material nearest to the walls of the crucible becoming white while the larger grains remained a dirty brown. This suggests that the reaction is fairly slow. Clear felspar heated to 1200°C during the same experiment took on a distinct red colouration. Both felspars became slightly fused at these temperatures forming a loose cake.

Specimens of gray felspar from Klokken, which might be regarded as black felspar which has been modified by natural heating, show a magnetic susceptability about half way between the values obtained from black and completely clear felspar.

7) Over-exposed X-ray Guinier films. X-ray Guinier films were taken of powders of black and clear material from the same felspar crystal. These were grossly over-exposed to X-rays in order to see

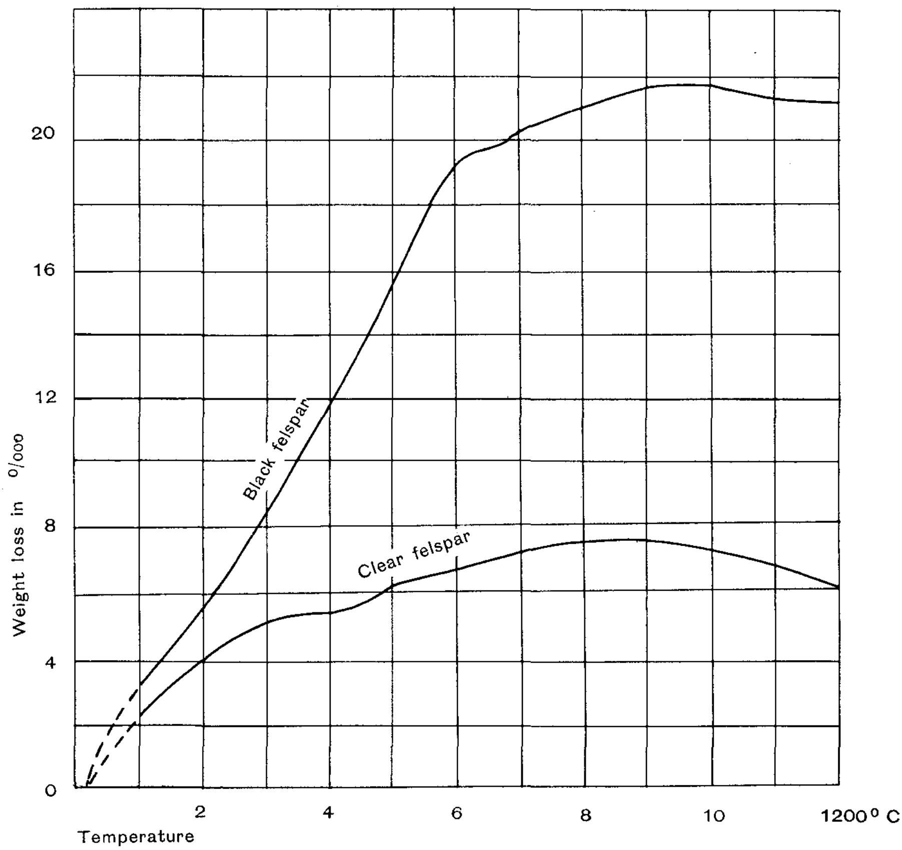


Fig. 62. Thermogravimetric curves of 10 g of black and clear felspar taken from the same crystal $\left(759\,\frac{\mathrm{AT}}{59}\right)$, illustrated in Plate 1 b.

whether characteristic patterns other than felspar could be detected. Although faint lines occurred which could possibly be interpreted as magnetite, they were equally developed in both the black and clear material, and are not thought to show the presence of any second mineral phase which could cause the difference in properties between black and clear felspar.

8) Thermogravimetric analyses. Two thermogravimetric analyses were carried out using the method described by MICHEELSEN (1966, p. 294) on 10 g of black and 10 g of clear felspar separated from the same crystal with an electromagnetic separator. The samples were heated at the rate of 4°C per minute. They were weighed at 5 minute intervals during a rise in temperature from 20°C to 1200°C. The resulting changes in weight can be seen in Fig. 62. Results below 75°C are omitted since, in this range, the rate of change in the correction curve

used for correcting the effects of convection currents in the furnace is considerably greater than the rate of change in weight of the sample. At higher temperatures the rate of change of the correction curve is much smaller.

Interpretation of results. The total corrected loss of weight on heating 10 g of black felspar to 900°C was 22.3 mg, that of clear felspar 7.9 mg. This loss in weight is attributed solely to loss in water from the felspar, thus it is apparent that the black felspar holds approximately three times as much water as the clear.

Both felspars showed a slight gain in weight above 900°C which is ascribed to the oxidation of iron. This explanation is supported by reddening of the samples after heating. In the black felspar the gain in weight is small (0.4 mg), while in the clear it is three to four times larger. Assuming that at the temperatures at which the oxidation of iron took place the rate of loss of water from the two samples was negligible, this suggests that three times as much oxygen was added to the white felspar as was added to the black, in order to oxidise the available iron. As it has been shown that the iron content is approximately the same in the black and the clear felspar in this crystal, this result suggests that the iron in the black felspar was in a different state to that in the clear felspar. It may be noted that the loss in colour in the black felspar took place above 800°C, that is after the majority of the water had been given off but within the temperature range in which the oxidation of iron takes place. It seems probable that the iron was not fully oxidised in either the black or the clear felspar at the time the experiment was stopped. Neither curve shows sign of flattening out at 1200°, and the black felspar was still brown when removed from the furnace.

Assuming that the total iron contents of the 10 g samples of black and clear felspar were identical, the 1.4 mg increase in weight in the clear felspar corresponds to an oxidation of approximately 0.1 % iron contained as FeO, while the 0.4 mg weight gain in the black felspar corresponds to oxidation of the same amount of iron were it Fe_3O_4. This does not imply that magnetite is present in the black felspar but suggests that the iron present may have the same structural state as the iron in magnetite.

The temperature at which water is lost from a mineral can give some idea of the way in which the water is held in the mineral structure and possibly some information about the structure of the mineral itself. MICHEELSEN (1966) has given a detailed account of the loss of water from flint during heating. It may be taken as a reasonable assumption that the water in felspar behaves in an analogous way to the water in flint. According to MICHEELSEN a low temperature reaction occurs in the range 20° to 250°C due to the loss of H_2O molecules. Overlapping this reaction a second reaction takes place in the range 150° to 700°C due to the break down of Si-OH groups. From the shape of the curves shown in Fig. 62 it would appear that in felspar the major part of the reaction was complete at 550° C.

The curves suggest that the two felspars contain fairly similar amounts of water as H_2O which they lose at similar rates in the range 20° to 200°. The loss in water as H_2O can be estimated to be approximately 450 p.p.m. in the case of the black felspar and 400 p.p.m. in the case of the clear felspar. Above 250° the two curves diverge markedly, the black felspar loses a further 1790 p.p.m. of water, possibly during the break down of Si-OH groups, while the clear felspar loses approximately 400 p.p.m. The great width of the thermogravimetric analysis reaction indicates that the H_2O is likely to be found on a lattice defect in the felspars. If it is assumed that these defects are similar to those described by MICHEELSEN (1966) for the Brazil law twin boundary on (0001) of quartz, and if they are parallel, then the water molecules will have a spacing of approximately 0.22 μm in the black felspar and approximately 0.25 μm in the clear felspar (see MICHEELSEN, 1966, pp. 310–312 for details of calculation).[1] It seems unlikely that these interfaces correspond to submicroscopic twin planes; the specimen selected for thermogravimetric studies was typical of the large black megacrysts and showed broad simple twin planes under the microscope. Precession photographs of an optically and chemically similar megacryst showed well defined reflections suggesting that submicroscopic twinning is not present in these felspars. According to MICHEELSEN the type of defect suggested will demand that the same weight of water is lost from Si-OH groups as is lost as H_2O. Thus, all the water lost from the clear felspar could possibly be accounted for by water which was present along a stacking fault system or similar structural defect. It is hoped that more refined thermogravimetric experiments will show whether any water exists within the structure of the clear felspar itself which might account for the anomalous properties found by optical and X-ray studies.

The black felspar, however, gives an additional weight loss of $1790 - 450 = 1340$ p.p.m. If this is assumed to derive from Si-OH groups found along a boundary similar to that described by MICHEELSEN for the Dauphiné twin boundary of quartz, then the mean spacing of Si-OH bearing interfaces would be 0.037 μm or 37 nm (370 Å). Such an interface density would in all probability make the X-ray reflections diffuse, but would be unlikely to change the lattice constants of the black felspar. However, the black felspar gives perfectly sharp X-ray diffraction patterns while the lattice constants are highly anomalous. The hypothesis of an Si-OH containing structure similar to that of the Dauphiné law must therefore be rejected and a hypothesis which accounts for a perfect, regular, spatial distribution of the hydrogen found in the thermogravimetric analysis is to be preferred. In the black felspar 1340 p.p.m. H_2O and ca. 0.1 % Fe correspond approximately to 6 H and 1 Fe in four of the "ideal unit cells" of MEGAW (1961). These impurities might, for instance, be located in a misfit subcell between intersecting stacking faults with non-parallel displacement vectors.

[1] I am greatly indebted to H. MICHEELSEN for help in interpreting the results of the thermogravimetric studies. By analogy with his studies of the distribution of water in flint (MICHEELSEN, 1966) the H_2O in the black felspar will occupy an interface arca of approximately $450 \times 10^{-4} \times 102\ m^2/cm^3$, which means 4.6×10^4 interfaces per cm or 1 per ca. 0.22 μm. Similarly the 1340 p.p.m. water thought to occur as Si-OH groups in the black felspar would occupy an interface area of $0.134 \times 10^4\ cm^2/cm^3 = 27.2 \times 10^4\ cm^2/cm^3$ if they occured along a boundary similar to that found for the Dauphiné twin boundary of quartz.

ii) Cause of the black colouration

The black felspars possess five main characteristics apart from colour which separate them from the majority of labradorites described as components of basic rocks. These are large grain size, structural defects, relatively high water contents, high paramagnetic properties and the presence of iron in a different state to that normally found in colourless felspar. Clear felspars from the same geological setting are also found as abnormally large crystals and show striking structural abnormalities and it can be argued from this that although both properties may play an important role in the formation of the colouration they are not its direct cause. The third property, high water content, is apparently confined to the black felspars and as this character has already been noted in other black felspars (Burns, 1966) it seems likely that it is a very important factor in the formation of the colour. However, the water is given off 300–400°C below the temperature at which the colour is lost during heating and it is therefore clear the water can only act as an agent in the formation of the colour. The total iron content in the black felspars is no higher than that found in many labradorites separated from igneous rocks, and there is no difference between the total iron in the black and clear felspar. However, it is thought significant that the black felspar loses its colour and magnetic properties in the same temperature range that the felspars showed an increase in weight due to an oxidation taking place. Further, it is also thought to be significant that the black felspar took on approximately a third of the weight of that taken on by a similar amount of clear felspar. This suggests that whatever substance in the felspar was oxidised controls both the colour and the magnetic properties, and that it is in a different state in the clear and black felspar. The most reasonable suggestion seems to be iron, a suggestion supported by the presence of ferric oxide at the end of heating experiments. A possible explanation of the properties seen in the black felspars would be that iron existed in the clear felspar as Fe substituting for Ca, while in the black felspar it was oxidised in the presence of water to Fe_3O_4. This suggestion has the advantage that it could be used to explain both the magnetic properties of the felspar and its dark colour. However, it is not thought to be the correct answer in the dark felspar described here since it has proved completely impossible to show the presence of magnetite as a separate mineral phase using several techniques. It is thought to be more correct to simply regard the iron in the two felspars as iron in different structural states which are themselves controlled by, or control the water content. Prolonged heating, either in the laboratory or in anhydrous conditions in the host dykes, removed the water and changed the iron to its more normal state.

iii) Comparisons between the black megacrysts and dark coloured felspars from other areas

Black colouration in felspars is generally attributed to the presence of small inclusions within the felspar whether formed by exsolution of iron-bearing material from the felspar or formed by the introduction of iron into the felspar at temperatures close to the melting point of plagioclase (POLDERVAART and GILKEY, 1954; WILCOX and POLDERVAART, 1958). The phenomenon is often called clouding and is commonly thought to be caused by thermal metamorphism (MACGREGOR, 1931). POLDERVAART and GILKEY showed, however, that clouding could occur in a variety of geological environments, including felspars from unmetamorphosed gabbroic rocks, and felspars from anorthositic layers in major layered intrusions. Using the strict definition given by MACGREGOR, and adopted by POLDERVAART and GILKEY, the Greenland megacrysts are not truly clouded since, with the exception of the felspars from the Igaliko and Klokken intrusions, they do not contain resolvable inclusions. However, it is apparent in the papers by MACGREGOR, and POLDERVAART and GILKEY that these authors meant the definition to cover the presence of "submicroscopic" or "ultrafine" inclusions. Many of the examples of clouding given in fact resemble the Greenland megacrysts closely, the clouding being described in thin section as a vague brown discolouration caused by barely discernable or submicroscopic inclusions. It is reasonable to suggest that in at least some of these examples the clouding is not caused by the presence of a second mineral phase, but is similar to the phenomenon described in this paper. As it is thought that the colour is not due to a second mineral it cannot properly be regarded as exsolution.

One example which appears to resemble the Greenland felspars closely is the clouding in the felspars from the Scourie dykes. MACGREGOR (1931, pp. 528–530) states that much of the clouding in the felspars from these dykes is "of the ultra-fine brown variety". BURNS (1966) has confirmed MACGREGOR's description and has made the important discovery that the clouded felspars from these dykes contain up to 1.2 $^0/_0$ water, considerably higher than the water found in the unclouded felspar from the same dykes.

The presence of black felspar formed by late stage alteration in the Scourie dykes might suggest a similar secondary origin for the black colouration of the megacrysts in Gardar rocks. However black felspars, the colour of which is undoubtedly original, occur in the post-Cambrian dolerites of South Sweden.

AA. JENSEN, (personal communication) has noted black tabullae in a dolerite from Hunneberg. The black felspars are three to four times the size of clear felspars from the same sample. In thin section they show no resolvable inclusions and are indistinguishable from the normal plagioclase apart from their size. The black felspars are often concentrated in the coarse pegmatite fractions of the dyke, where they are associated with quartz-felspar intergrowths and a significant increase in the amount of opaque minerals. This suggests that they were formed under late stage conditions with high concentrations of iron and water. JENSEN reports that the dark felspars are commonly intergrown symplectically with Fe-Ti oxides, in contrast to the normal dyke rock where the oxides are intergrown with pyroxenes.

Comparisons with felspars in which the clouding is due to resolvable inclusions of a second mineral phase are of less significance as it is apparent that inclusions may be found in plagioclase for a variety of reasons. However, it is of interest that the general conditions which POLDERVAART and GILKEY suggest are necessary for the development of clouding in felspars; slow crystallisation, high water content and availability of iron, are very similar to those which are thought to have given rise to the colouration in the Greenland megacrysts. POLDERVAART and GILKEY also suggest that the clouded felspars may possess "Minute surfaces of physical discontinuity" which they believe may play a role in the migration of iron into the plagioclase. It seems unlikely that migration of iron into the felspar played a large role in the development of colour in the Greenland megacrysts; they probably belong to the group of lightly clouded felspars including many examples from unmetamorphosed igneous rocks in which POLDERVAART and GILKEY believe the iron causing the colouration to be derived internally. However, it seems quite possible that migration of iron along stacking faults or similar discontinuities may account for the resolvable inclusions of Fe-Ti oxides and olivine which occur along the (010) and pericline twin planes in both the black and clear megacrysts.

The clouding seen in plagioclases from the Canadian Shield anorthosites is generally attributed to resolvable inclusions of iron- or titanium-bearing minerals, and is thus not strictly comparable to the colouration found in the Greenland megacrysts. However, the total amounts of iron held by the Canadian felspars is moderately similar to that found in the Greenland megacrysts although resolvable Fe-Ti oxides may be present in apparently large amounts. PHILPOTTS (1966) states that the felspars from the Morin anorthosites never contain more than 0.3 $^0/_0$ total iron (as Fe_2O_3). During granulation of the clouded felspar iron apparently migrates out of the felspar to form interstitial coatings of ilmenite and magnetite, in contrast to the process seen in the Greenland megacrysts where iron remains within the felspars when it becomes clear. It seems likely that more information about the type of inclusions within the felspars from the Canadian anorthosites and when they formed might throw some light on the problem whether the anorthosites crystallised from a water-rich magma (BUDDINGTON, 1960) or a water-poor magma (PHILPOTTS, 1966).

iv) Petrogenetic significance of the black colouration

The following points seem to be of interest in interpreting the geological condition under which the black felspars formed: 1) The black colour appears to be primary; there is no evidence that the colour formed as the result of a secondary process affecting older material. 2) The black colour is restricted to a particular type of felspar; it is not found in the constituent felspars from the granular anorthosites, nor in the small megacrysts from the most alkaline hosts. 3) The black felspars show a restricted range in composition, they are unzoned, and apart from their colour are practically identical to the clear megacrysts of sodic labradorite found in similar hosts throughout the Gardar province. 4) They contain a relatively high amount of water and the dark colouration is apparently stable in host dykes which themselves contain considerable water. 5) The colour disappears when the felspars were included in normal, relatively anhydrous gabbroic hosts.

From these points it is suggested that the black felspars crystallised from a relatively highly differentiated gabbroic magma in which there was a high water content. Their large size and lack of zoning suggest relatively stable conditions. The significance of these suggestions to the genesis of the megacrysts in Gardar dykes in general is discussed in the concluding sections of this paper.

VI. PETROLOGY OF GARDAR FELSPATHIC DYKES

Most of the rock types intruded during the igneous activity of the Gardar period contain anorthositic inclusions, at least locally, and a petrographic description of hosts ranging from olivine gabbros to nepheline syenites is out of place in this paper. Further details can be found in the regional accounts cited in Section II. All the analyses available of felspathic intrusions (with the exception of Ilímaussaq) are given in Table 4. No overall difference has been detected between these analyses and analyses of other Gardar intrusions, a complete list of which is found in WATT (1966). The felspathic intrusions generally show similar differentiation trends to other Gardar bodies, at least on ΣFe-Mg-alk diagrams (Figs. 63 & 64), and show no abnormalities which could be expected if the bulk of the included plagioclases was derived from the host magma. However, too little is known at present of the chemistry of the non-felspathic Gardar dykes to be certain that there has not been some loss in calcium and aluminium in the host rock of some of the dykes. It is also often extremely difficult to decide when sampling for analysis which megacrysts are derived from the dyke and which are xenocrysts.

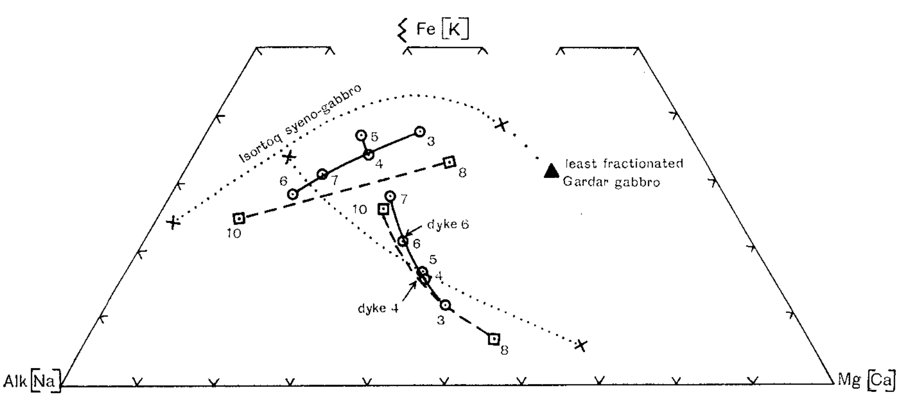


Fig. 63. ΣFe-Mg-alk and Ca-K-Na diagrams for analyses of samples from dykes 4 and 8 compared to the differentiation trends found in the Isortoq syeno-gabbro. The numbers on the diagram refer to analyses in Table 4.

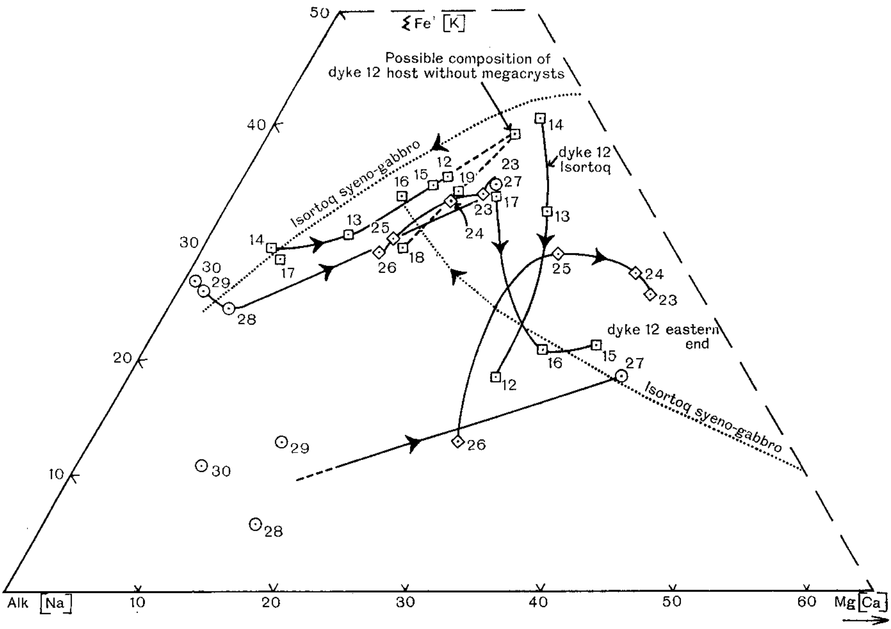


Fig. 64. Detail from the ΣFe-Mg-alk and Ca-K-Na diagrams for analyses of samples from dykes 12, 16 and 21, (microsyenites-trachydolerites) compared to the Isortoq syeno-gabbro trends. The arrows on the trend lines indicate the order of crystallisation in the four series of samples taken across the strike of the microsyenite-trachydolerite dykes. These are in the opposite direction to the sequence seen in the syeno-gabbro dyke. The numbers refer to analyses given in Table 4.

Plagioclase-rich insets are particularly abundant in two widespread groups of dykes emplaced during the first half of the mid-Gardar period. Some members of the first group, the ENE-trending gabbro dykes which are widespread in the area from Nunarssuit to Narssaq, have already been described by Upton (1964) and are only dealt with briefly in this paper. The second group of dykes, which follow the olivine gabbros in the Tugtutôq area and apparently overlap with them in time further to the west, consist of a variety of transitional rocks between gabbro and microsyenite. These dykes show complex intrusion mechanics which are discussed in Section VII.

a) Gabbros

The gabbros belong to several different generations of dykes dominantly intruded in the mid-Gardar. **The giant olivine gabbro dykes from Tugtutôq** (analysis 1, Table 4) are the most basic hosts in the Gardar to contain abundant anorthosite xenoliths and Upton (personal communication) regards them as having crystallised from one of the least fractionated magmas postulated from the province.

The presence of both laminated anorthosites of the Assorutit type and some granular fragments is therefore of considerable interest as these rocks must have

Table 4. *Host Rock Analyses*

Analysis	1	2	3	4	5	6	7	8	9	10	11
GGU No.	40452 / 30765	61055	61114	61102	61112	61109	61109§	25352	25365	25350	25183
SiO_2	45.03	46.04	47.65	51.01	50.43	52.96	52.41	49.38	56.7	58.13	58.07
TiO_2	2.48	2.24	2.48	2.00	2.40	1.62	1.98	1.11	1.3	1.28	1.30
Al_2O_3	19.02	15.85	16.30	16.27	15.25	17.40	15.14	18.85	17.1	16.70	16.34
Fe_2O_3	2.51	5.37	3.03	2.57	2.97	4.13	5.40	2.15	7.9†	2.22	2.70
FeO	10.74	8.91	10.00	8.36	9.35	4.06	4.96	7.56		4.51	4.27
MnO	0.16	0.19	0.18	0.16	0.16	0.15	0.19	0.13	—	0.14	0.16
MgO	6.46	6.04	5.29	4.03	4.14	2.59	3.13	5.86	2.4	1.72	1.42
CaO	8.10	8.57	7.33	6.71	6.51	5.85	4.95	8.40	4.6	4.54	3.53
Na_2O	3.36	3.50	3.97	4.31	4.29	4.39	4.08	3.54	4.7	4.99	4.58
K_2O	0.83	0.62	1.66	2.28	2.29	3.22	3.92	0.97	3.4	3.84	5.62
P_2O_5	0.56	0.39	0.60	0.56	0.65	0.64	0.79	0.23	0.44	0.39	0.58
CO_2	n.d.	nil	nil	nil	nil	nil	—	nil	—	nil	n.d.
H_2O^+	0.69	2.09	0.60	0.80	1.36	2.60	3.18	1.50	—	0.94	0.79
H_2O^-	0.06	*	*	*	*	*	—	*	—	*	0.08
ZrO_2	—	—	—	—	—	—	—	—	—	—	0.06
Total	100.00	99.81	99.09	99.06	99.80	99.45	100.13	99.68	Partial	99.40	99.50

† Total iron as Fe_2O_3 * Sample dried at 110°C for 2 hours § Recalculated analyses

NORMS (taken from Watt, 1966)

	1	2	3	4	5	6	7	8	9	10	11
Q	—	—	—	—	—	1.2	1.6	—	—	2.9	1.6
or	4.9	3.7	9.8	13.5	13.6	19.1	23.2	5.7	—	22.7	33.3
ab	25.8	29.6	30.6	36.4	36.3	37.1	34.5	29.9	—	42.2	38.7
an	34.3	25.7	21.7	18.3	15.5	18.2	11.3	32.6	—	11.8	7.4
ne	1.4	—	1.6	—	—	—	—	—	—	—	—
C	—	—	—	—	—	—	—	—	—	—	—
di	1.9	11.4	8.9	9.5	10.4	5.3	6.4	7.4	—	6.9	5.3
wo	—	—	—	—	—	—	—	—	—	—	—
hy	—	4.1	—	1.2	1.4	5.6	6.5	5.1	—	5.4	4.7
ol	21.3	10.1	15.4	10.6	11.0	—	—	11.8	—	—	—
mt	3.6	7.8	4.4	3.7	4.3	6.0	7.8	3.1	—	3.2	3.9
il	4.7	4.3	4.7	3.8	4.6	3.1	3.8	2.1	—	2.4	2.5
ap	1.3	0.9	1.4	1.3	1.6	1.5	1.9	—	—	0.9	1.4

Analysed samples

Analysis No:

1. Mean of two analyses of Tugtutôq and Narssaq gabbro (Upton, 1964, p. 24). Analyst: R. Soli.
2. Chilled anorthosite-bearing apophysis of gabbroic dyke (locality 7, Plate 4, p. 45 and p. 128).
3. Chill of basic centre of dyke 8 (Plate 4, p. 46 and p. 128). Eastern end of middle unit of dyke. Analyst: B. I. Borgen.

(continued)

Table 4 (cont.)

Analysis	12	13	14	15	16	17	18	19	20	21	22
GGU No.	33013	33012	33011	61172	61171	61170	44116	44116§	44898	38186	45458
SiO_2	50.44	56.20	57.93	52.20	53.75	58.04	52.98	52.18	57.57	58.85	49.40
TiO_2	2.32	1.66	1.37	2.25	2.06	1.42	1.82	2.66	1.50	1.25	2.80
Al_2O_3	15.51	15.18	14.75	15.40	14.95	15.27	17.35	13.54	14.61	15.05	14.10
Fe_2O_3	11.60†	9.58†	9.10†	11.59†	10.98†	9.18†	3.74	5.07	1.53	2.81	4.30
FeO							5.23	7.09	7.49	5.72	8.50
MnO	—	—	—	—	—	—	0.14	0.17	0.23	0.16	0.30
MgO	2.48	1.60	0.82	2.51	2.18	1.06	2.38	3.15	1.33	1.13	3.00
CaO	6.07	4.32	3.56	6.07	5.46	3.61	5.81	4.29	3.35	2.92	6.00
Na_2O	4.68	4.18	3.87	4.48	4.79	4.86	4.73	4.51	4.54	4.35	4.20
K_2O	2.36	4.69	6.03	3.10	3.10	5.12	3.10	4.16	4.92	5.55	2.90
P_2O_5	0.85	0.78	0.53	0.92	0.96	0.53	0.73	0.98	0.53	0.39	1.50
CO_2	—	—	—	—	—	—	nil	nil	n.d.	n.d.	nil
H_2O^+	—	—	—	—	—	—	1.78	2.40	1.50	1.01	2.20
H_2O^-	—	—	—	—	—	—	*	—	0.32	0.14	*
ZrO_2	—	—	—	—	—	—	—	—	0.05	0.07	—
Total	Partial	Partial	Partial	Partial	Partial	Partial	99.79	100.20	99.47	99.40	99.20

† Total iron as Fe_2O_3 * Sample dried at 110°C for 2 hours § Recalculated analyses

NORMS (taken from WATT, 1966)

Q	—	—	—	—	—	—	—	—	1.8	4.0	—
or	—	—	—	—	—	—	18.4	24.6	29.1	32.8	17.2
ab	—	—	—	—	—	—	40.0	38.1	38.4	36.8	35.5
an	—	—	—	—	—	—	16.9	4.4	4.9	5.1	11.0
ne	..	..	..	..	..	..	—	—	—	—	—
C	—	—	—	—	—	—	—	—	—	—	—
di	—	—	—	—	—	—	5.9	8.6	7.1	5.9	7.4
wo	—	—	—	—	—	—	—	—	—	—	—
hy	—	—	—	—	—	—	5.0	4.9	10.1	6.3	8.4
ol	—	—	—	—	—	—	1.3	2.6	—	—	2.5
mt	—	—	—	—	—	—	5.4	7.4	2.2	4.1	6.2
il	—	—	—	—	—	—	3.5	5.1	2.9	2.4	5.3
ap	—	—	—	—	—	—	1.7	2.3	1.2	0.9	3.6

4. Centre of dyke 8, 50 m north of (3). Analyst: B. I. BORGEN.
5. Late inclusion-poor basic dykelet cutting dyke 8, same locality as (3).
6. Early trachydoleritic margins of dyke 8 containing 19 % (by volume) of plagioclase megacrysts with an average composition of An_{48}, 2 km west of (3).
7. Recalculated analysis from (6) above after subtracting the chemical equivalent of 19 % (by volume) of plagioclase megacrysts.
8. Basic margin of larvikitic syeno-gabbro (dyke 4, Plate 4, p. 36 and p. 130). Shore of Eqalugssuit taserssuat.

(continued)

Table 4 (cont.)

Analysis	23	24	25	26	27	28	29	30	31	32	33
GGU No.	42658	42656	42655	42654	80005	80004	80003	80002	60037	48317	20632
SiO_2	49.04	49.88	51.43	51.40	48.16	60.42	61.21	66.64	51.77	56.07	44.60
TiO_2	2.68	2.69	2.35	2.20	2.99	1.28	0.95	0.71	2.43	0.68	1.49
Al_2O_3	15.10	14.95	15.73	15.35	14.70	15.55	14.32	13.08	15.35	17.50	16.46
Fe_2O_3	4.24	3.59	10.58†	2.16	12.57†	8.36†	8.96†	8.18†	3.95	3.89	2.28
FeO	6.72	7.34		7.13					6.25	3.57	9.44
MnO	0.20	0.12	—	0.12	—	—	—	—	0.19	0.18	0.16
MgO	3.39	3.20	2.44	2.43	3.56	0.86	0.30	0.20	2.49	0.60	9.77
CaO	6.29	6.12	5.03	5.19	7.08	3.04	2.86	1.50	6.16	2.77	8.68
Na_2O	3.80	3.99	4.53	6.24	4.66	8.59	8.08	7.70	5.02	6.93	2.50
K_2O	3.83	4.19	4.58	2.11	2.92	1.02	2.18	1.39	3.89	5.37	0.85
P_2O_5	1.85	1.69	1.25	1.11	1.62	0.40	0.18	0.08	0.99	0.22	0.32
CO_2	nil	nil	—	2.20	—	—	—	—	nil	nil	nil
H_2O^+	2.38	2.19	—	2.11	—	—	—	—	1.68	2.21	2.06
H_2O^-	*	*	—	*	—	—	—	—	*	*	*
ZrO_2	—	—	—	—	—	—	—	—	—	—	—
Total	99.52	99.95	Partial	99.75	Partial	Partial	Partial	Partial	100.17	99.99	98.61

† Total iron as Fe_2O_3 * Sample dried at 110°C for 2 hours § Recalculated analyses

NORMS (taken from WATT, 1966)

	23	24	25	26	27	28	29	30	31	32	33
Q	—	—	—	—	—	—	—	—	—	—	—
or	22.6	24.8	—	12.5	—	—	—	—	23.0	31.8	5.0
ab	32.1	32.6	—	52.7	—	—	—	—	34.1	36.8	20.9
an	12.8	10.5	—	4.6	—	—	—	—	7.8	0.8	31.2
ne	—	0.6	—	—	—	—	—	—	4.5	11.8	0.1
C	—	—	—	1.1	—	—	—	—	—	—	—
di	5.2	7.3	—	—	—	—	—	—	13.3	8.0	8.0
wo	—	—	—	—	—	—	—	—	—	1.4	—
hy	4.0	—	—	6.0	—	—	—	—	—	—	—
ol	4.9	7.7	—	5.7	—	—	—	—	3.1	0.1	25.5
mt	6.2	5.2	—	3.1	—	—	—	—	5.7	5.6	3.3
il	5.1	5.1	—	4.2	—	—	—	—	4.6	1.3	2.8
ap	4.4	4.0	—	2.6	—	—	—	—	2.4	—	0.8

9. Intermediate rock type from dyke 4, 2 km west of Eqalugssuit taserssuat. Analyst: B. I. BORGEN.
10. Alkali centre of dyke, 200 m north of (8). Analyst: B. I. BORGEN.
11. Chilled margin of microsyenite-trachydolerite dyke forming part of the dyke 6 complex at Eqalugssuit taserssuat (Fig. 22, p. 42). Analyst: B. I. BORGEN.
12. Sample 4 m from contact of microsyenite-trachydolerite dyke (dyke 12, Plate 4, p. 50 and p. 137) 3 km west of mouth of Isortoq.
13. 3 m from contact, same locality as (12).

(continued)

Table 4 (cont.)

14. 50 cm from contact, same locality as (12).
15. Dyke 12 at ice margin, 18 km north-east of (12) above. Sample from centre of dyke.
16. Dyke 12, same locality as (15), approximately 50 cm from contact.
17. Dyke 12, same locality as (15), contact sample.
18. Dyke 12, centre of dyke, 11 km north-east of (12) above. Sample analysed contains 31 % (by volume) of plagioclase megacrysts with an average composition of An_{48}. Collector: K. Coe.
19. Recalculated analysis from (18) above after subtracting the chemical equivalent of 31 % (by volume) of plagioclase.
20. Chilled contact of dyke 12, eastern shore of Isortoq. Collector: K. Coe. Analyst: B. I. Borgen.
21. Chilled contact of dyke 12, western end, 11 km west of (12) above. Analyst B. I. Borgen.
22. Probable centre of felspathic dyke from south side Patdlit, N.E. Qaersuarssuk (locality 14, Plate 4, p. 140). Collector: W. S. Watt. Analyst: B. I. Borgen.
23. 11 m felspathic dyke on south coast of Tunugdliarfik (locality 21, Plate 4, p. 141). Sample 1.9 m from contact. Collector J. H. Allaart.
24. Same locality as (23), 1.3 m from contact. Collector: J. H. Allaart.
25. Same locality as (23), 80 cm from contact. Collector: J. H. Allaart.
26. Same locality as (23), 5 cm from contact. Collector: J. H. Allaart.
27. Felspathic dyke cutting southern giant gabbro, east coast of Tugtutôq, 5 m from contact (locality 16, Fig. 35, p. 55 and p. 140).
28. Same locality as (27) above, 1 m from contact. Sample analysed without phenocrysts of alkali felspar seen in hand specimen.
29. Same locality as (27), above, 20 cm from contact.
30. Same locality as (27), 5 cm from contact.
31. Trachydolerite felspathic dyke approximately 1 km up river on north side of Akuliaruseq, Igaliko Fjord (locality 29, Plate 4, p. 63 and p. 142). Collector: Steen Andersen.
32. Nepheline trachyte with anorthosite inclusions and plagioclase megacrysts, 0.5 km east of "Heste Spor Sø", south side of Kujatdleq, Igaliko Fjord (locality 30, Plate 4, p. 142). Collector: J. P. Berrangé.
33. Most basic Gardar gabbro falling on the general Gardar line on ΣFe-Mg-alk variation diagrams. Contains plagioclase and olivine megacrysts. S. W. Eqaloqarfia dyke (Pulvertaft, 1965). Analyst: B. I. Borgen.

Analyst: Ib Sørensen, unless otherwise stated. All analyses except No. 1 were carried out in the geochemical laboratory of the Geological Survey of Greenland using the methods described by Borgen (1967).

---

solidified before the intrusion of their host. In his description of the olivine gabbros of Tugtutôq and Narssaq, Upton (1964) suggest that the western end represents a lower level in the dyke than the rocks exposed around Narssaq, possibly due to subsequent tilting. The anorthosite fragments are concentrated in the more differentiated upper parts of the body, especially where it loses its dyke form at the eastern extremity. Mineralogically the dyke shows considerable differences along its trend. The earliest plagioclases at the western end are approximately An_{72}, while in the east they are An_{61-56}. Olivines show a range from Fa_{32} to Fa_{55} but are normally between Fa_{36} and Fa_{43}, while the augites, which surround both olivine and plagioclase, show a composition of $Ca_{45\cdot5}Mg_{37\cdot5}Fe_{17}$. None of these minerals correspond closely to those in the inclusions in the dyke; the anorthosites containing generally a more magnesium-rich olivine and a more sodium-rich plagioclase. The olivine in the secondary anorthosite from Narssaq resembles that in the surrounding gabbro (Fa_{48}) while the pyroxene is more iron-rich than that normally seen in the gabbro ($Ca_{39}Mg_{31\cdot5}$ $Fe_{29\cdot5}$).

Compared to the Tugtutôq gabbro the host rocks from further west are all less basic in composition and are generally less aluminous. Many contain quartz in the mode. Sample 61055 (analysis 2, Table 4), from an aphanitic, chilled apophysis of a 35 m NE-trending olivine dolerite (locality 7, p. 45), is the most basic host rock analysed from the Kobberminebugt area. The centre of the dyke is a fresh ophitic olivine-pyroxene gabbro which differs from most of the other host rocks in the area in the fact that olivine exceeds clinopyroxene in the mode and there is practically no interstitial quartz or alkali felspar.

The olivine gabbro dykes from Tôrnârssuk with scattered gabbro anorthosite xenoliths are among the most basic hosts examined although no chemical analyses are available. They contain more olivine and a more calcic plagioclase than the majority of felspathic dykes from either Kobberminebugt or the Ivigtut region.

Analyses 3, 4, 5, 6 and 7 are of samples collected from **dyke** 8 (p. 46) which show the complex intrusion history summarised below: (see Plate 3)

1) Early, discontinuous trachydoleritic margins with alkali felspar phenocrysts and sericitised plagioclase megacrysts.

2) The main body of the dyke, an alkali gabbro which breaks through the early trachydoleritic margin. The alkali dolerite centre locally develops a chilled contact against the country rock or against the trachydolerite margins. The centre contains large masses of clear glassy plagioclase megacrysts and a few anorthosite fragments.

3) An inclusion-poor, medium-grained olivine dolerite which occurs in thin dykes with clean-cut, non-chilled contacts against the earlier components. These thin dykes may vein the surrounding country rock.

The early trachydoleritic margin of the dyke is chilled against the country rock. Sample 61109, analysis 6, is a fine-grained (0.03 mm) bluish-black rock containing 19 % (by volume) of plagioclase megacrysts with an average composition of An_{48}. A photograph of the analysed sample is shown in Fig. 65. The rock also contains alkali felspar rhombs which total approximately 6 % of the volume. These are occasionally intergrown with augite and are frequently rounded and show a hollow centre filled with groundmass material with inward pointing lobes of felspar. The phenocrysts show fine polysynthetic albite twinning and in a few cases albite-pericline cross-hatching. Optical properties ($2V_X$ 50°–70°, and a refractive index close to balsam) suggest that they are potash-rich oligoclase.

The groundmass of the marginal part of the dyke consists of a fine-grained mixture of Fe-Ti oxides, felspar, and considerable indeterminate material. Away from the contact the rock is medium- to fine-grained (0.3 mm), with felspar as the main mineral forming an interlocking granitiod texture. Biotite, Fe-Ti oxide grains, elongate clinopyroxenes with indented margins and irregular masses of chlorite form the main mafic minerals. There is a little interstitial quartz. The felspars show marked zoning from labradorite centres with polysynthetic albite twinning to simple

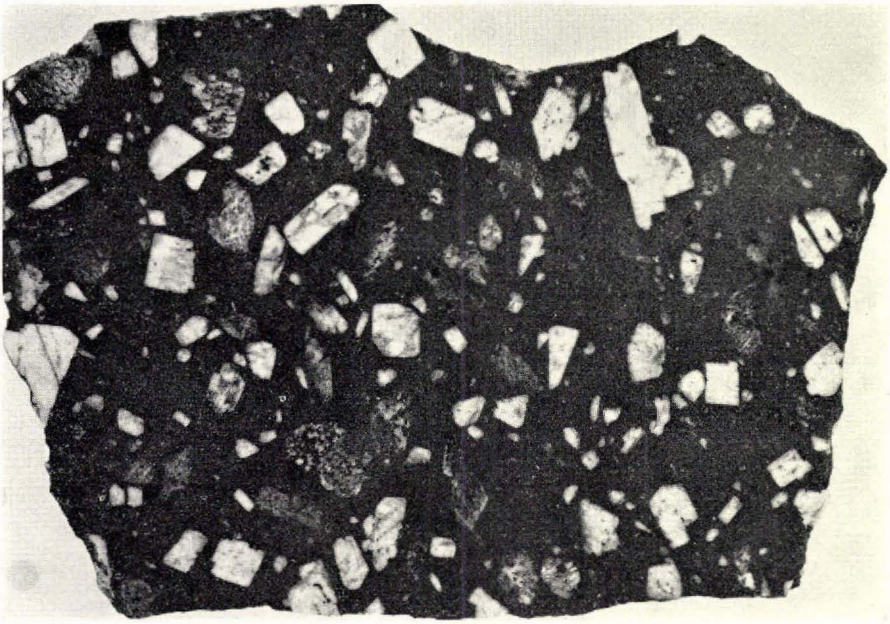


Fig. 65. Laboratory photograph (2 × natural size) of one surface of the sample (G.G.U. 61109 analysis 6 Table 4) analysed of the marginal trachydolerite of dyke 8. The sample contains 19 % by volume of plagioclase megacrysts. Note the irregular rhombs of oligoclase-alkali felspar which partially enclose fine grained matrix. Photograph S. E. Bendix-Almgreen.

albite-twinned oligoclase mantles and margins of alkali felspar. In many cases the alkali margins have replaced plagioclase, often leaving partially digested relics with heavily zoned contacts between the two types of felspar.

The analysis of the megacryst-bearing intermediate margin of the dyke (analysis 6), after the subtraction of the plagioclase megacrysts (analysis 7) resembles several of the megacryst-bearing centres of the microsyenite-trachydolerite group of dykes described later in this section. It is possible that the relatively low anorthite content seen in the norm may be due to some of the plagioclase-forming constituents being taken from the dyke magma at an early stage in its crystallisation, either to form the megacrysts, or to form plagioclase rims round original xenolithic material. There is at least no evidence that there has been addition of anorthite to a normal doleritic magma by the partial resorption of felspathic material. Only the large plagioclase megacrysts were included in the volumetric analysis used for the calculation of analysis 7. Some of the felspar in the groundmass is certainly of derived origin which suggests that the host magma at the time of crystallisation of the dyke contained an even lower percentage of plagioclase-forming material than shown by analysis 7.

The partly chilled basic margin of the central part of the dyke (analysis 3) is a medium-grained (0.3–1.0 mm), homogeneous augite-olivine dolerite. The felspars show smooth, non-oscillatory zoning from polysynthetically twinned labradorite centres to simple twinned oligoclase margins locally overgrown with alkali felspar. Slightly irregular grains of Fe-Ti oxides sometimes partially rimmed with biotite form about 5 % of the rock.

The alkali gabbro forming the centre of the dyke (analysis 4) is considerably less homogeneous than its partially chilled basic margins with irregular patches of fine-grained (0.1–0.5 mm), olivine-rich rock with stellate plagioclase crystals surrounded by coarser (up to 2 mm), ophitic augite-olivine gabbro. Locally there are areas consisting almost entirely of felspar with a little Fe-Ti oxide. The variation in grain size and constituents suggests that the rock did not reach equilibrium during crystallisation. Felspars show intense zoning from labradorite centres to alkali margins; the more calcic centres commonly show strong oscillatory zoning in contrast to the felspars in the marginal rocks. Alkali felspar rims most of the plagioclase and is especially noticeable in the coarser grained areas where it is intergrown with quartz. Some of the larger felspars in the groundmass (which may have been megacrysts) show strong zoning near to their margins commonly accompanied by an increase in the number of polysynthetic albite twins. Locally these large felspars are intergrown with augite. The material trapped between the giant felspar megacrysts (Fig. 9, p. 27) is considerably more differentiated than the normal dyke rock and consists mainly of andesine, some clinopyroxene, Fe-Ti oxides and chloritic material.

The late stage, megacryst-poor, part of the dyke (analysis 5) consists of olivine, and plagioclase phenocrysts up to 2 mm × 0.25 mm with violent oscillatory zoning, set in a groundmass the average grain size of which is less than 0.1 mm. The olivine phenocrysts show strong resorption. The groundmass contains strongly zoned plagioclase with andesine-labradorite centres and untwinned alkali margins. Pyroxene, granular Fe-Ti oxides, and a little biotite are the main mafic minerals in the groundmass contrasting with the more noticeable olivine phenocrysts.

When plotted on ΣFe-Mg-alk and Na-K-Ca variation diagrams the differentiation trends seen in dyke 8 are found to resemble the normal Gardar trends as exemplified by the Isortoq syeno-gabbro dyke (Fig. 63). However, although analyses from dyke 6 plot on moderately smooth curves in the variation diagram the relationship between composition and order of crystallisation is complex. The earliest rock emplaced is the trachydoleritic margin to the dyke (analyses 6 and 7) which falls at the alkali end of the differentiation curves. This is followed by the chilled margin of the main body of the dyke (analysis 3) which falls at the basic end of the curves. The coarse-grained centre of the dyke is an alkali gabbro (analysis 4) with a composition approximately half way between the early trachydolerite and its own chill. The final inclusion-poor fraction is very similar in composition to the centre apart from a slight iron enrichment. It is suggested that differentiation in the dyke was controlled by two processes, an early segregation of the more alkali components in the same manner as described in detail from the microsyenite-trachydolerite dykes (p. 133) and a second, more normal differentiation process after the main bulk of the dyke was emplaced which led to the basic margin and more alkali centre in a manner comparable to the other syeno-gabbro dykes seen in the Gardar.

Analyses 8, 9 and 10 are of samples taken from the eastern end of **dyke 4** (p. 36) and show stages between the early olivine dolerite margin and the larvikitic centre. Two other analyses from minor intrusions connected with this dyke are given in Watt (1966, Nos. 15 & 119). The petrology of the dyke is complex and can only briefly be summarised here; the olivine gabbro margins are considerably less basic than the Tugtutôq giant dykes, although the relatively high alumina content suggests that it may be chemically closer to the Tugtutôq rocks than other dykes in the Kobberminebugt area. The marginal gabbro (analysis 8) contains ophitic augite, subophitic to granular olivine and tabular plagioclase. It is relatively coarse-grained (up to 1 cm) although locally it contains small granular aggregates of olivine which

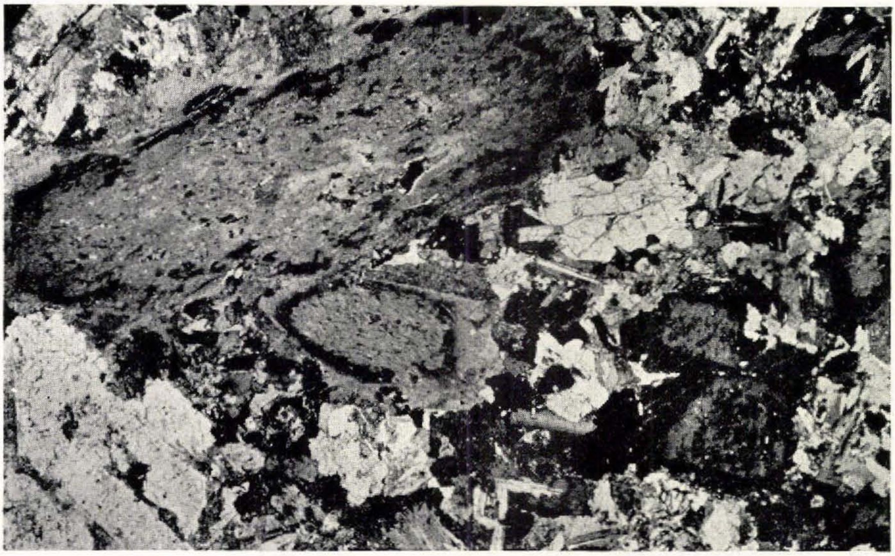


Fig. 66. Rhombs of oligoclase-alkali felspar in dyke 4. Crossed nicols × 9, G.G.U. 25365. Note the patchy intergrowth in the centre of the rhombs (albite, oligoclase and potash felspar), the intensely zoned mantle (in extinction) and the clear alkali felspar margin. The rhombs include small crystals of augite and occasionally olivine.

average less than 1 mm. The centres of the plagioclase crystals zone from An_{60-65} to approximately An_{20} at the margins. There is commonly a break marked by a change from the polysynthetic twinning in the centre to simple or untwinned felspar near the margin, this break taking place between An_{30} and An_{35}. Large felspar grains with a composition of An_{58} and little zoning near their centres are seen in many of the thin sections of the marginal gabbro. These are thought to be megacrysts. They are commonly intergrown with either olivine or augite which is identical in composition to the mafic minerals in the surrounding rock. Olivine ranges from Fa_{48} to Fa_{62}; the more magnesium-rich examples are not always found with the most basic plagioclase suggesting that conditions varied from one part of the dyke to another. Augite has a composition of about $Ca_{40}Mg_{30}Fe_{30}$ (based on 2V and refractive index measurements). Several of the gabbro samples close to the central syenitic part of the dyke show considerable development of alkali felspar and quartz.

Samples from the western outcrops of the part of the dyke exposed near Eqalugssuit taserssuat are mainly composed of a rock intermediate between gabbro and syenite both in appearance and chemistry (analysis 9). The groundmass consists of stellate clusters of plagioclase (An_{50-55}), with olivine (Fa_{60-70}), pyroxene and interstitial masses of sodic plagioclase, and alkali felspars. Many of the felspars in the stellate masses zone outwards to untwinned oligoclase margins. The rock contains up to 30 % of its volume of large felspar crystals with a distinctive form elongated parallel to the *a* axis, thus appearing as rhombs in thin section. These felspars, which are found in many of the felspathic dykes with an intermediate composition, are extremely complex intergrowths of three or more varieties of felspar arranged in a characteristic pattern (Fig. 66). The centres consist largely of untwinned felspar with a refractive index approximately equal to oligoclase and a $2V_x$ of 70–75°. The untwinned felspar in the rhomb centres contains irregular patches of andesine with extremely closely spaced polysynthetic albite twins and separate patches of alkali felspar. The central zone is surrounded by a rim of andesine-oligoclase characterised by closely spaced polysynthetic twinning often showing a "block structure". The twinned mantle often shows the typical rhomb form better than the actual edges of the crystal which tend to be intergrown either with the groundmass or with other rhombs. The mantle generally shows extremely marked zoning. The outer parts of the crystals are made up of untwinned felspar resembling that in the centre with a refractive index close to oligoclase. The optic angle changes progressively outwards from $2V_X$ 72° to $2V_X$ 52°. Patches of potash felspar with fine, cross-hatched microcline twinning and quartz intergrowths are found surrounding the rhombs. Internally they are often intergrown graphically with augite, and sometimes Fe-Ti oxides and olivine. These are identical to those in the groundmass and some of the augites can be seen to be in optical continuity with the material in the centre of the felspar rhombs.

The larvikitic centre of the dyke (analysis 10, Table 4) contains up to 30–40 % of the rhombs described above with a medium-grained groundmass of heavily zoned oligoclase-albite felspar, augite, potash felspar, and a little quartz. Biotite, Fe-Ti oxides and a little alkali amphibole are sometimes present. Occasional rectangular masses of andesine with block pericline-albite twinning are intergrown with augite. These are thought to represent the less basic megacrysts.

The differentiation trends seen when the results of analyses 8, 9 and 10 are plotted on ΣFe-Mg-alk and Na-K-Ca variation diagrams resemble the trends seen in other major syeno-gabbro dykes from the Gardar province (Fig. 63).

The gabbros surrounding the large anorthosite mass at Augiâta tasia (p. 49) resemble the Narssaq gabbros in the field with large well-laminated plagioclase tabulae and ophitic mafic minerals. However, petrographically they have more in common with the other gabbroic hosts from the area south of Kobberminebugt. Olivine is rarely seen and its presence can only be suggested by masses of chloritic material now surrounded by plagioclase and augite grains. Typically there is considerable interstitial quartz and alkali felspar present. The alkali felspar is thought to be predominantly albite which shows characteristic chequer-board twinning. The host felspar zones outwards from An_{58} to An_{25} and is thus less calcic than the anorthosites they enclose. Most of the felspars show oscillatory zoning in the range An_{50} to An_{58} followed by smooth zoning to the oligoclase margins. There is often a change in twin pattern which coincides with the break between the labrodorite centres and more sodic margins; in some cases the sodic plagioclase has attacked the central zone. The clinopyroxene is very pale mushroom to colourless, it is generally fresh and shows slight patchy zoning. It surrounds the plagioclases ophitically but not the interstitial alkali felspar and quartz. Fe-Ti oxides form less than 3 % of the rock and vary between well-formed magnetite grains and interstitial ilmenite. In most thin sections the magnetite has been replaced by a non-opaque amorphous substance leaving the crystal outlined by two generations of ilmenite lamellae.

Dyke 11 from the shores of Isortoq may be regarded as typical of the Gardar gabbros in the area which contain scattered, well-formed, plagioclase megacrysts. Petrologically it resembles the host rock at Augiâta tasia fairly closely apart from grain size (in this case plagioclase crystals are up to 2 mm and ophitic augite up to 5 mm).

Olivine is rarely preserved, the plagioclases show wide outer zones of oligoclase and alkali felspar, while the ophitic augite is found in contact with the plagioclases zoned from An_{55} centres to An_{25} margins but not in contact with the interstitial alkali material. In some cases the plagioclase of the host is replaced by a sericite, sodic plagioclase, alkali felspar mass, and biotite surrounds Fe-Ti oxide grains. Clinopyroxene is left completely fresh. This process is thought to be due to a form of autometamorphism as it does not affect the older olivine gabbro which dyke 11 cuts. It resembles the "syenitisation" seen in the mixed rock between the gabbro margins and the syenite centres of the giant dykes found at the mouth of Isortoq fjord.

The two dykes from locality 1 (p. 34) at the mouth of Kobberminebugt and the majority of megacryst-bearing dykes from the Ivigtut region show enough petrological features in common to allow them to be described together although they may not be of the same age. They show many features similar to those seen in the less basic felspathic dykes south of Kobberminebugt and are petrographically closer to the microsyenite-trachydolerite host rocks than the Tugtutôq olivine gabbros.

Plagioclase in these dykes is generally found as stellate aggregates up to 2 mm in diameter but averaging 0.5 mm. Some variation is seen from dyke to dyke; centres of plagioclase grains from the Kobberminebugt dykes average An_{50} zoning outwards to approximately An_{20}, while the rocks described by WEIDMANN (1964) and C. H. EMELEUS (personal communication) from further north have less calcic plagioclases. Some of the felspars are dark and appear smoky in thin section resembling the black megacrysts. The dark felspars are particularly noticeable in rocks containing primary hornblende. Rims of alkali felspar intergrown with quartz are almost ubiquitous round the plagioclases and in some of the rocks pockets of chequer-board twinned albite, quartz and potash felspar have segregated out and formed slightly coarser grained areas.

Clinopyroxene is subophitic to granular, often showing marked zoning from reddish brown centres to colourless margins. In one dyke cutting the Grønnedal complex, EMELEUS reports hypersthene as well as the more typical clinopyroxene. This mineral has been reported from very few Gardar dykes. Fe-Ti oxides, biotite and indeterminate masses of chlorite, fibrous amphibole, calcite, epidote and irregular grains of interstitial quartz make up the bulk of the groundmass surrounding the felspar grains.

b) Microsyenite-trachydolerite dykes

The rocks which make up the second group of megacryst-bearing dykes show a range from quartz microsyenite to alkali gabbro. The majority of dykes are less than 20 m wide and as the gradation between quartz syenite to gabbro takes place in the first 5 m from the contact in most dykes, samples collected within a few centimetres of each other may differ significantly in composition. The rocks are generally rather fine grained and contain considerable amounts of chlorite and undetermined micaceous minerals which makes them rather unsuitable for study in thin section, and a large number of analyses are necessary to give an impression of their petrology. These analyses are listed in Table 4, ΣFe-Mg-alk and Na-K-Ca variation diagrams are shown in Fig. 64.

Chemically the dykes show considerable differences and do not belong to exactly the same suite. In general the syenitic portions of the majority of the dykes in the area west of Tugtutôq are potassic while the syenitic parts of the dykes east of Tugtutôq are sodic. In areas where there is more than one generation of the microsyenite-trachydolerite dykes the later dykes are generally the most sodic. In spite of these chemical difference it is convenient to treat the dykes as one group since they show several characters which separate them from other dykes described from the Gardar.

The main distinguishing feature possessed by this group of dykes is their characteristic mode of emplacement with early alkali margins and late megacryst-rich basic centres (Plate 3). The mechanics of emplacement are discussed later (p. 144), however, it can be noted at this juncture that although the sequence of emplacement of syenitic and gabbroic rocks is the opposite to that normally expected there is considerable evidence to suggest that both are differentiated from the same basaltic magma. Thin sections from individual samples show normal crystallisation sequences with alkali felspar following plagioclase and iron-rich pyroxenes following more magnesic pyroxenes and occasionally olivines. The chemical variations between contact and centre of each dyke are most accurately seen in the analyses given in Table 4 and the

Table 5. *Changes in physical and chemical properties of samples taken from margins to centres of microsyenite-trachydolerite host dykes.*

Dyke 12, Isortoq (35 m broad at this point)			
Distance from contact	Specific gravity	SiO_2 content	100 Fe/Fe + Mg
0	2.69	—	—
50 cm	2.72	57.9	85
3 m	2.77	56.2	75
4 m	2.85	50.5	70
Dyke 12, Ice margin (20 m wide at this point)			
0	—	58	81
50 cm	—	53.8	71
10 m	—	52.2	70
Locality 16, Tugtutôq, 15 m microsyenite-trachydolerite			
5 cm	—	66.5	95
20 cm	—	61.2	93
1 m	—	60.4	98
5 m	—	48.1	64
Locality 21, Tunugdliarfik, 11 m microsyenite-trachydolerite			
5 cm	—	51.5	69
80 cm	—	51.4	68
1.3 m	—	50.0	66
1.9 m	—	49.0	65

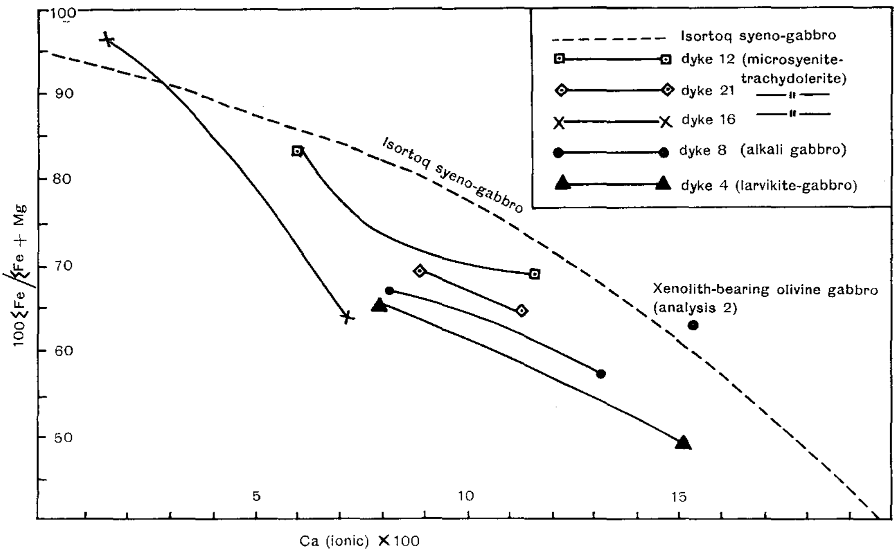


Fig. 67. Diagram showing the relative proportion of calcium in the analyses of the felspathic dykes compared to that found in the Isortoq syeno-gabbro using the $100\,\Sigma Fe/\Sigma Fe+Mg$ ratio as a standard. The results support the idea that the host rocks may be impoverished in calcium.

variation diagrams in Fig. 64; however, they are perhaps more readily appreciated in Table 5 in which the specific gravity, silica content, and $\Sigma Fe/(\Sigma Fe+Mg)$ ratios are listed against the distance from the contact for four series of samples taken at right angles across the trends of three of the dykes.

Chemically the microsyenite-trachydolerite dykes, in common with the other megacryst-bearing hosts analysed (dykes 4 and 6), are all slightly less calcic than the majority of Gardar rocks. This is best illustrated by plotting the calcium content against the $\Sigma Fe/(\Sigma Fe+Mg)$ ratios which can be used as an approximate index of the stage of differentitation a particular rock has reached. It can be seen that the megacryst-bearing dykes contain about 1.5 $^0/_0$ less CaO than comparable rocks from the Isortoq syeno-gabbro used as a standard (Fig. 67). On the other hand, the sample analysed from the olivine dolerite at locality 7 (p. 45), which contains anorthosite xenoliths and no plagioclase megacrysts, falls on the same curve as samples from the Isortoq dyke. The relative lack in calcium in the megacryst-bearing dykes could possibly be ascribed to the removal of calcium to form labradoritic plagioclase megacrysts early in the crystallisation of the rocks. However, there are so many other variables present that it cannot be taken as proof of the close genetic link between host and inclusion. There is at least no evidence that the host rocks have been contaminated by resorption of plagioclase derived from a source unconnected to the dyke magma.

ΣFe-Mg-alk variation diagrams (Fig. 64) show that although the sequence of formation of the different rock types is the reverse to that normally found in syeno-gabbro intrusions, the actual differentiation trends are remarkably similar to the Isortoq giant dyke. It can also be seen that apart from a very small variation in iron content the trends from the three dykes studied are virtually identical.

In contrast, the trends shown by the Na-K-Ca plots are highly variable, even within one dyke. Normal trends from the Gardar show a fairly constant decrease in calcium and a small increase in the K/Na ratio as differentiation proceeds from gabbro to syenite. The trends shown by the microsyenite-trachydolerite dykes show no such consistency, there is an overall decrease in calcium from the more basic to the more alkali rocks, but the ratio of K/Na varies considerably within comparable samples taken from different traverses across a single dyke. The precise reason for this divergence in differentiation trends is uncertain, though it seems probable that the mechanics of differentiation play a large part in determining the shape of the curves.

In the majority of Gardar igneous bodies studied the dominant mechanism of differentiation is crystal fractionation which results in differentiation trends such as those illustrated from the Isortoq dykes. This mechanism probably plays the dominant role in determining the relative proportions of most of the major elements in the microsyenite-trachydolerite dykes and results in similar differentiation trends to those normally found in the Gardar rocks. However, in the case of the alkalies it is suggested that a second process, diffusion within the magma, plays the dominant role in determining the distribution of these elements. In some cases this may lead to trends which resemble normal differentiation trends; in other cases, particularly where there is a high water content in the original magma, there may be a marked increase in the proportion of sodium to potassium in the rocks formed from the magma which was furthest away from the original source at the time of crystallisation.

The interaction between the two processes, crystal fractionation and alkali diffusion, is thought to be a reasonable explanation of the marked changes in the differentiation trends found on Na-K-Ca variation diagrams. The role of alkali diffusion in determining the mechanics of differentiation of the microsyenite-trachydolerite dykes is discussed in a later section (p. 147). If the megacrysts are derived from the same original magma as the host rocks, this process may play a subsidiary role in determining the irregular differentiation trends and account for the relatively small decrease in calcium from the gabbroic centres of the dykes to the alkali margins.

Individual dykes

Dyke 12 (p. 50) is the only independent member of the group described from west of Qagssimiut. The megacryst-free margins are chilled against the country rocks forming a dark blue-black cryptocrystalline border set with pink irregular alkali felspar phenocrysts. There is no sign of preferred orientation of the phenocrysts which make up between 3–5 % of the rock. In thin section the rock shows a slightly patchy development of crystalline material often centred around small (0.05 mm) quartz grains. A slight layering is sometimes developed parallel to the contact caused by alternating units of cryptocrystalline and fine-grained material. Sample 38186 (analysis 21), taken approximately 5 cm away from the margin of the dyke at its western end, is fine-grained (0.05 mm), homogeneous, and contains 3 % by volume of the alkali felspar phenocrysts. These are rounded and sometimes show a rhombic shape in thin section. They locally form aggregates which have interlocking, irregular grain boundaries. Some of the centres of the phenocrysts show fine polysynthetic twinning, generally on the albite law but occasionally cross-hatched pericline and albite. There is generally an untwinned area near the borders of the phenocrysts. Their optical properties ($2V_X$ 70° and average refractive index slightly less than balsam) suggest that they are similar to the rhombs described from dyke 4 (p. 131), only in this case complete exsolution of the different components has not taken place. Under high magnification plagioclase, biotite, pink clinopyroxene, amphibole, Fe-Ti oxides, and a few irregular grains of quartz may be distinguished in the groundmass. The plagioclases are zoned with polysynthetically twinned centres and untwinned alkali felspar margins.

No significant differences can be seen between sample 38186 (analysis 21) and the three other samples taken close to the contacts (analyses 14, 17 and 20). Scattered phenocrysts up to 1 mm in size of amphibole, clinopyroxene, Fe-Ti oxide and a few composite grains of quartz, biotite, amphibole and calcite occur in samples taken within 10 cm of the margin. Finely disseminated pyrite is found in several of the contact samples. It appears to have a rather irregular distribution.

Away from the contacts the alkali rocks become coarser grained, in some examples reaching an average of 0.5 mm. Local poorly-developed trachytic textures are found 20–30 cm in from the contact but most of the groundmass consists of equidimentional felspar grains with irregular, slightly poikilitic, elongate clinopyroxene, prominant apatite needles and finer grained rather irregular masses of Fe-Ti oxide, biotite, quartz and amphibole. There are occasional vesicular areas of chlorite. The groundmass felspars are heavily zoned and the plagioclase centres are more prominent in samples taken close to the inner margin of the megacryst-free zone of the dyke. The clinopyroxenes frequently form symplectic intergrowths either with other minerals in the groundmass or more occasionally with the alkali felspar phenocrysts.

Alkali felspar phenocrysts locally form rafts in the dyke between the chilled margin and the megacryst-rich centre. There is a progressive increase in the degree of exsolution seen in the phenocrysts from margin to centre of the dykes; those collected close to the dyke contacts are virtually unexsolved while the rarer alkali felspar phenocrysts from the basic centre of the dyke appear identical to the rhombs seen in the larger larvikitic dykes (p. 131).

Analyses 12, 13 and 16, which are of samples taken from the inner part of the inclusion-free zone in the dyke, show less SiO_2 and total alkalies, a smaller $\Sigma Fe/\Sigma Fe+Mg$ ratio and more CaO than samples taken nearer to the contacts at the same localities.

The petrography and chemistry of the basic centre of dyke 12 is more difficult to study than that of the marginal rocks because of the large amount of included material. The matrix of the host is medium-grained and may be classified as a trachydolerite or trachyandesite. It is made up of similar minerals to those seen in the marginal rocks except that plagioclase forms a greater proportion of the total groundmass felspars. In thin section the host rock plagioclases are usually subhedral approximately equant rhombs with extreme zoning from andesine-labradorite cores to alkali felspar margins. The plagioclase cores frequently show oscillatory zoning. In places the marginal alkali felspar has partially replaced the plagioclase cores, veining into the zoned labradoritic centres. The plagioclase megacrysts are frequently surrounded by a clear rim of untwinned felspar with refractive indices similar to oligoclase and a $2V_X$ changing from approximately 70° in the centres to 50° at the margins. These are regarded as potassic oligoclase zoning towards a less calcic alkali felspar. The rims are probably close in composition to the alkali felspar rhombs described earlier from this dyke and dyke 4 (p. 131). Where these rims form onto large broken megacrysts or the rare anorthosite fragments it is easy to distinguish included material from crystals formed from the dyke magma itself. However, there is a complete gradation between recognisable megacrysts with homogeneous centres and fiercely zoned late margins and the groundmass felspars. The majority of the groundmass felspars have a core of andesitic or labradoritic plagioclase which may either represent finely comminuted xenolithic or xenocrystic material, or which may have formed as early crystallisation products of the dyke itself. Detailed optical and chemical studies on the pyroxenes and other iron- and magnesium-bearing minerals has not been made due to their relatively fine grain size and heavy zoning. The pyroxenes show a general increase in 2V from the centre towards their margins and a slight change in colour from mushroom to pale pink.

The inability to distinguish host rock felspar from foreign felspathic material leads to considerable difficulties in selecting material for analysis and interpreting the final results. Where the dyke narrows close to the Inland Ice the centre contains comparatively little included material and it is possible to select samples with no demonstrable "foreign" material. An analysis of one of these samples (analysis 15) confirms the trachydoleritic nature of the host and shows that the trend from syenite towards gabbro continues in to the centre of the dyke. However, it has been noted throughout the microsyenite-trachydolerite group of dykes that there is a close correlation between the amount of foreign material held and the basisity of the host. It could be expected that where the dyke reaches its maximum width towards the south-west that the host would be more basic than that shown by analysis 15. As it is impossible to separate host and megacrysts a total rock sample was analysed (analysis 18) after making an estimate of the amount of included material in the actual block analysed (Fig. 68). By defining included material as all felspars which were twice or more the average grain size of the groundmass it was shown that at least 31 % by volume of the sample was foreign. This is almost certainly an underestimate of the true amount as it neglects all the small labradoritic centres to the groundmass felspars some of which can clearly be seen in thin section to be of derived origin. The mean value for the anorthite content of the plagioclase megacrysts from the sample was An_{48} and from partial analyses of similar felspars it was estimated that the megacrysts contained approximately 0.7 % K_2O. By subtracting the ap-

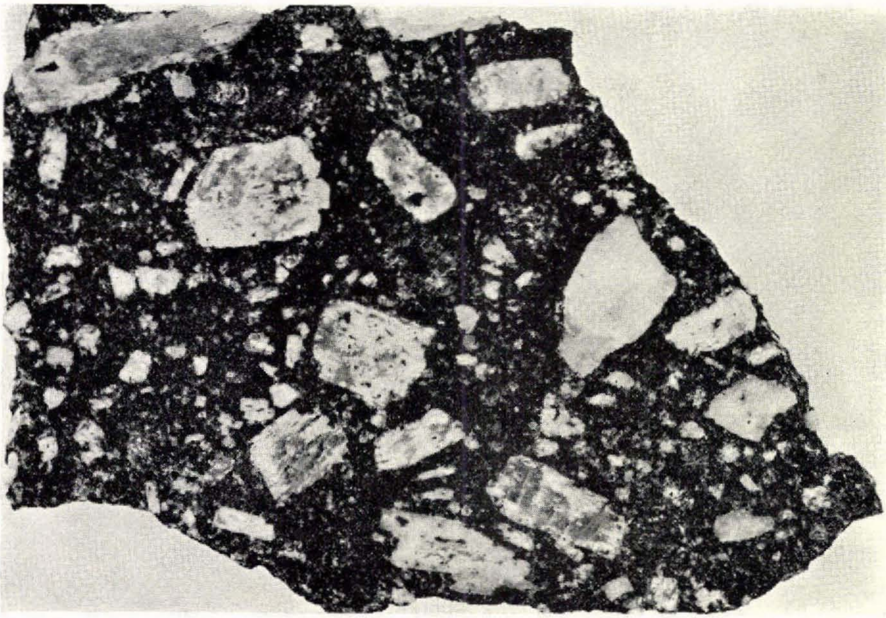


Fig. 68. Laboratory photograph × 2 natural size of one surface of the analysed sample (G.G.U. 44116 analysis 18) of the alkali gabbro centre of dyke 12. The sample contained 31 % by volume of plagioclase megacrysts. Photograph S. E. Bendix-Almgreen.

propriate amount of Na_2O, CaO, K_2O, Al_2O_3 and SiO_2 and recalculating to 100 % (analysis 19) it is possible to get an approximate value for the host without foreign material. The $\Sigma Fe/\Sigma Fe + Mg$ ratio given by this analysis (which is almost unaffected by errors in estimating the amount of included felspathic material) confirms the idea that the most basic part of the dyke contains the highest proportion of included material; however, when analysis 19 is plotted on a ΣFe-Mg-alk diagram it falls slightly to the iron-poor side of the trend found for other samples from the same dyke and between analysis 15 and the marginal samples. As all other analyses from this dyke fall on a single curve it would appear that this is due to an underestimate of the amount of included material and a truer value would be in the region of 40 % of foreign andesine-labradorite. Even with the conservative estimate used in calculating analysis 19 it can be seen that the host rock is poor in CaO compared to normal Gardar alkali gabbros, and in fact shows a smaller amount of normative anorthite than the syenitic samples from the same dyke. This peculiarity gives some support to the suggestion that some of the megacrysts may have been derived from the dyke magma at an early stage. A similar depletion in sodium is not apparent, however, it has already been noted that the distribution of the alkalies is abnormal in these dykes.

The thin alkali dykes found close to the margins of many of the large anorthosite-bearing gabbros are clearly related to the major dykes they accompany. However, many of them show considerable similarity with the microsyenite-trachydolerites and it seems likely that in the area west of Tugtutôq the major anorthosite-bearing gabbros and the dykes of the same type as dyke 12 are closely related. In the Kobberminebugt area the two groups of felspathic dyke are approximately contemporaneous and members of the microsyenite-trachydolerite suite are generally associated with major felspathic gabbros. The felspathic gabbros themselves petrologically resemble the basic components of the microsyenite-trachydolerite suite: they contain appreciable quartz and alkali felspar in the mode and often appear to have crystallised from rather water-rich magmas. It seems possible that the main difference between the two groups is the level at which the present erosion surface cuts individual bodies; the microsyenites-trachydolerites represent cross-sections close to the top of major bodies which are more gabbroic at depth but which still locally possess microsyenite or trachydolerite margins.

On Tugtutôq, and in the area to the east, the mixed dykes are clearly later than the anorthosite-bearing gabbros and there is a similar clear cut difference in the petrology of the two groups.

The similarity between alkali rocks associated with the large gabbro and the mixed group of dykes is best seen at **Eqalugssuit taserssuit** (p. 42). Sample 25183 (analysis 11) from the chilled margin of the small microsyenite dyke belonging to the **dyke 6** complex resembles the marginal rocks of dyke 12, both chemically and petrographically. It contains the same characteristic alkali felspar phenocrysts and occasional rounded clinopyroxenes set in a fine-grained (0.02 mm) groundmass of predominantly alkali felspar with a poorly developed trachytic texture. Away from the margins the dyke shows a progressive increase in plagioclase megacrysts and a decrease in the alkali felspar rhombs. This is accompanied by a change in the groundmass felspar so that in the centre of the dyke alkali felspar rims and andesine-labradorite centres are approximately equal in amount.

The microsyenite margins of the majority of felspathic dykes from **Qaersuarssuk** (**locality 14**, Plate 4, see also WATT, 1968) are generally more sodic than those described from the felspathic dykes further to the west. However, a few of the samples collected by WATT resemble the potassic microsyenite margins of felspathic dykes south of Kobberminebugt and it appears that both potassic and sodic varieties are present in the area. Many of the marginal rocks show well-developed trachytic textures. Alkali felspar rhombs of the same type as seen in dyke 12 are common. The coarser-grained rocks towards the centres of the dykes contain soda amphibole, aegerine and considerable albite which may show chequer-board twinning. Samples from the inclusion-bearing centres of the dykes show a general resemblance to samples from the centre of dyke 12 (compare analyses 12 and 22). A few of the dyke centres are more basic than the centres of microsyenite-trachydolerite dykes from further west and contain granular olivine and ophitic pyroxene. A partial analysis of felspar separated from the margin of one of the dykes showed a composition of $Or_{36.5}Ab_{63.5}$ which is more sodic than suggested by the norms of marginal rocks from dyke 12.

UPTON (1962, p. 32) has given a brief description of the petrology of some of the **microsyenite-trachydolerite host dykes from Tugtutôq**. He suggests that the dykes are dominantly olivine dolerites subjected to a varying degree of alkali contamination. Olivine is usually pseudomorphed and much of the groundmass is unresolvable. Biotite, augite and apatite have been noted while the groundmass felspar is generally oligoclase zoned outwards to antiperthite. The only material examined for the present study is from a 070°, 16 m dyke cutting the olivine gabbro at the eastern end of Tugtutôq (locality 16, Fig. 35, samples 80002–80005, analyses 27–30). This dyke has an early, fine-grained marginal facies approximately 1 m wide of a moderately soda-rich microsyenite (analyses 29–30). This is followed by 2–3 m of coarser grained soda microsyenite with inclusions of the marginal rock enclosed in it (analysis 28). No chilling is seen. The coarser grained microsyenite contains abundant alkali felspar phenocrysts often accumulated in irregular masses. Soda amphibole is the dominant mafic mineral. The 10 m central facies of the dyke (analysis 27) shows gradational contact with the margins and there is commonly a transition zone marked by abundant vesicles between the two rock types. The centre contains abundant glassy plagioclase megacrysts (An_{45-55}) and rare alkali felspar phenocrysts near the borders. The host rock is composed of a sodic plagioclase zoned from basic oligoclase to albite, augite as pink elongated crystals intergrown with the other groundmass minerals, Fe-Ti oxides, and considerable indeterminate chloritic material.

The analyses show that the dyke possesses the same gradation from early marginal alkali rocks to late central gabbroic rocks as other members of this group, only in this particular dyke the variation is extreme, ranging from a quartz-syenite with a silica content of approximately 66.5 $^0/_0$ to a central gabbro with 48 $^0/_0$ SiO_2. Samples are not available for the complete range, however, those analysed show a differentiation trend which coincides closely with ΣFe-Mg-alk variation curves from other dykes in the same suite. The marginal rocks are among the most sodic samples analysed from this suite of dykes and it is interesting to note that in the marginal syenites an amphibole is the dominant mafic mineral instead of the pyroxenes seen in the centre of this dyke and in the marginal syenites of dyke 12. This gives some support to the suggestion that there is a relative enrichment of sodium in the marginal rocks of those dykes which contained a high water content in the original magma.

Analyses 23, 24, 25 and 26 are of samples taken respectively 1,9 m, 1.3 m, 70 cm and 5 cm from the contact of a 055° **microsyenite-trachydolerite dyke on the south coast of Tunugdliarfik**, 5 km east of the Ilímaussaq intrusion (**locality 21**, Plate 4). The dyke is 10.2 m wide at this point; it shows an olive-green microsyenite megacryst-free margin approximately 1 m wide and a centre of trachydolerite with a gradual increase in plagioclase megacrysts towards the centre. The contact between the alkali margins and the more basic centre is gradational; megacrysts are first found in a rock which is already fairly basic (42656, analysis 24).

The margin of the dyke is fine-grained (0.05–0.1 mm) set with occasional irregular alkali felspar phenocrysts. The groundmass consists of simple twinned plagioclase-alkali felspar laths with some zoning, Fe-Ti oxide grains, calcite and biotite. There are occasional phenocrysts of subhedral Fe-Ti oxides. Towards the centre of the dyke the groundmass felspars are more calcic, zoning from labradorite-andesine centres to alkali felspar margins. The norms (Table 4) suggest that the felspar in the marginal rocks is predominantly sodic. This is far less marked in the centre of the dyke and in fact the analyses show an increase in potash away from the margin although the total alkali content falls. The mafic minerals from the dyke centre consist dominantly of Fe-Ti oxides and pink to colourless ophitic augite. Pseudomorphs after olivines are relatively common.

The ΣFe-Mg-alk variation curve (Fig. 64) is parallel to other dykes in this group and resembles the normal Gardar trend. However, the Na/K ratio increases rapidly in the syenitic marginal rocks in a similar manner to the dyke described from Tugtutôq. It is interesting to note that the mafic minerals in the marginal syenites are micaceous suggesting that the syenitic magma was water-rich compared to the gabbroic centre of the dyke where clinopyroxene and some olivine formed. It seems possible that both potash and soda microsyenite-trachydolerite dykes originated from the same magma at depth, the soda-rich group which are generally slightly later in local dyke chronologies may represent a regional upward migration of soda in a large magma chamber due to a gradual accumulation of water.

The most easterly sample analysed is 60037 (analysis 20, Table 4) from a 3 m NE–SW trending **trachydolerite from Jespersen Dal** on the south-west of the Igaliko intrusion (**locality 29**, Plate 4). The dyke is a fine-grained dark-brown rock with irregular masses of large glassy plagioclase megacrysts with black centres and white sericitised margins. The aggregates sometimes show a pseudo-ophitic texture with the interstices between the large felspars filled with normal dyke material. Locally the host is slightly coarser grained and felspathic around the megacrysts. In thin section the dyke is seen to consist dominantly of stubby felspar crystals with no preferred orientation. Thin needles of pale pink clinopyroxene are the most common mafic mineral. These may show a good crystal form in cross-section but a rather fretted outline when seen in sections parallel to their length. There are occasional groups of heavily resorbed olivine or pseudomorphs after olivine, in one case accompanied by a subhedral crystal of orthopyroxene which is larger than the other mafic grains in the slide. The felspars show some zoning from an andesine centre to alkaline margins. This is emphasised by incipient alteration of the cores. No sharp break can be seen between zones in the felspars; sodic and potash felspar do not form separate crystals. The felspars show simple Carlsbad or albite twins in the centre of most of the crystals. Fine, almost submicroscopic chequer-board twinning is seen nearer the margins. No nepheline was noted in thin section although the rock is moderately undersaturated. The spaces between felspars commonly contain chlorite. Under high magnification rods and skeletal crystals of Fe-Ti oxide are found to be fairly abundant. These are often surrounded by biotite. The presence of orthopyroxene in this dyke and in some other felspathic dykes is of considerable interest since this mineral is otherwise unknown from the Gardar intrusions. It is tentatively suggested that the formation of orthopyroxene instead of the normal augitic pyroxenes may be due to the comparative lack of calcium in the felspathic dykes after the formation of plagioclase megacrysts.

The two main megacryst-bearing dykes described by BERRANGÉ (1966a) from the **"Vatnahverfi"** area (**locality 32**, Plate 4) show considerable differences in petrography. The trachydolerite at Søndre Igaliko contains megacrysts across the entire width of the dyke. The margins are less basic than the centre of the dyke, containing a rather sodic plagioclase, biotite, Fe-Ti oxides, thin needles of apatite and small patches of calcite and chloritic material. The centre of the dyke, which holds a greater abundance of foreign material than the margin, contains considerable zoned pink clinopyroxene and large apatite crystals set in a groundmass of altered basic plagioclase. Some well-formed pseudomorphs after olivine are found surrounded by clinopyroxene.

The second dyke (sample 48317, analysis 32) is a nepheline trachyte with irregular patches of xenoliths and megacrysts surrounded by inclusion-free material. Sample 48317 contains no megacrysts although BERRANGÉ's field notes state that the rock is the same as that immediately surrounding the inclusions. It is a microsyenite containing phenocrysts of alkali felspar and nepheline set in a fine-grained gray groundmass. The groundmass consists of approximately equal proportions of potash and sodic alkali felspar laths with a sub-trachytic texture, small pyroxene phenocrysts mantled by aegirine, and interstitial nepheline and biotite. There are a few euhedral biotite phenocrysts visible in hand sample. The analysis shows that this dyke is interesting from two points of view: firstly, it is considerably more undersaturated than any other host rock analysed, and secondly, in spite of its undersaturated nature the dyke contains more silica than samples taken from the megacryst-bearing zones in any other of the host dykes. The fact that it is the only thoroughly undersaturated host may reflect the availability of samples rather than a particular affinity between saturated rocks and the anorthosites; the majority of the analysed samples were taken from the Kobberminebugt–Nunarssuit area where both major and minor Gardar intrusions belong to a saturated suite. The high silica and the high $\Sigma Fe/\Sigma Fe+Mg$ ratios suggest that the material surrounding the xenoliths had reached a more advanced stage of differentiation than the megacryst-bearing zones in other host dykes. The irregular distribution of megacrysts and xenoliths as patches and clumps surrounded by normal dyke matrix suggests that this dyke shows a different method of intrusion and differentiation to the majority of microsyenite-trachydolerite dykes where the insets are generally small and graded from margin to centre.

VII. EMPLACEMENT OF GARDAR FELSPATHIC DYKES

a) The mechanics of intrusion and differentiation of the micro-syenite-trachydolerite dykes

The dykes are characterised by several unusual features which suggest that their method of emplacement and differentiation was abnormal and requires explanation. The main points from the field and laboratory observations are summarised below:

1) Dykes with the same general features occur over a large area and are locally extremely abundant. No record has been kept of the exact number in different regions but the total can be estimated in hundreds rather than in tens.

2) Dykes with the same general pattern of intrusion and crystallisation show a considerable range in petrography. This paper is only concerned with felspathic dykes; however, it appears that many of the features are also present in the large swarms of trachytic dykes in the area north of Narssarssuaq and probably in some of the intermediate dykes in the Ivigtut region.

3) The dykes consist of early, generally symmetrically arranged alkali-rich margins with few megacrysts, enclosing a more basic centre with abundant megacrysts. Where the dykes thin out the margins continue for some way after the basic centres end. The width of the alkali margins is independent of the total width of the dyke (a block diagram of the probable form of a typical dyke is shown in Plate 3). The contact between the two units is generally gradational; in a few cases there is a sharp break sometimes marked by chilling of the centre against the margin. Inclusions of the early marginal rocks within the centre can be found in most dykes. Even in the dykes which show one or more sharp breaks suggesting different phases of intrusion, margin and centre each show a gradual change in petrology inwards with a decrease in silica and total alkalies and a rise in magnesia away from the margins. The change in total composition of the host rock can be seen from density measurements which range in a typical dyke from 2.70 g/cm^3

at the chilled contact to 2.85 g/cm³ in the centre. The change is gradational; even where there has been a break in the intrusion of the dykes the change in properties from alkali to basic is moderately smooth.

4) In contrast to the intrusion sequence (*i.e.* alkali followed by basic) in the dykes as a whole, individual thin sections show normal trends, with early crystallisation of relatively calcic plagioclase and magnesium pyroxene, and later crystallisation of alkali felspar and more iron-rich pyroxene. The alkali-rich rocks can therefore be regarded as more differentiated than the basic rocks.

5) The foreign material shows a characteristic distribution pattern within the dykes. There are rare irregular masses of anorthosite and rather sericitised megacrysts within the marginal syenites of a few dykes. Nearer the centre the more basic rocks carry both megacrysts and a few scattered xenoliths. There is frequently a well-developed grading with small megacrysts on the outside of the central zone and an increase in size and total amount towards the centres. Some of the centres carry a moderately high proportion of xenoliths. The average composition of the foreign material becomes more calcic towards the centre of the dykes as both the larger megacrysts and the xenoliths have a higher anorthite content than the small fragments nearer the margins. There is local evidence that the megacrysts have rafted together but none that they have been packed together. Many dykes show a moderate flow orientation of the megacrysts with their long axes parallel to the dyke margins. The host rocks appear to have crystallised in place and there is no evidence of the accumulation of early mineral phases in the centre of the dykes except for the megacrysts. The contacts are often aphanitic and show no preferred orientation of microlites; away from the contact the lath-shaped alkali felspars rarely show well-developed trachytic flow texture and are often grouped together in radiating clusters.

6) The composition of the first rock to carry abundant megacrysts varies from dyke to dyke. However, the early marginal megacryst-free phases in all the dykes analysed had a silica content exceeding 53 % while the megacryst-rich late centres all contained less than 53 % silica. The first appearance of the megacrysts is often marked by a zone of vesicular rock.

7) Alkali felspar phenocrysts show a distribution pattern somewhat similar to that seen in the megacrysts. They are rare in the most highly differentiated marginal syenites, reach a maximum concentration in the basic syenites approaching the plagioclase megacryst-bearing centres and overlap slightly with the smaller megacrysts. They may locally raft together in sieve-like masses.

8) The mineralogy of the dykes suggests that the magma contained considerable water and probably some carbon dioxide and sulphur. The dominant mafic minerals in the marginal syenites are biotite and amphibole. Pyroxene is found in the centre of the dykes but interstitial chlorite, amphibole and biotite are more abundant. Some of the aphanitic contact samples contain iron sulphides forming up to 2 % of the total rock. These may be concentrated along cooling cracks suggesting that the sulphides were formed slightly later than the chill itself. Larger crystals of pyrite are found within the microsyenite margins but are less common in the centres.

Interpretation

Minor intrusions in which a central body of rock containing accumulated xenoliths is enclosed by earlier more differentiated margins, or in which the chilled contacts do not represent the bulk composition of the body, have been described from several other igneous provinces. The main reasons put forward for the non-uniform distributions of material may be summarised under the following headings:

1) Flowage differentiation
2) Multiple intrusion
3) Liquid fractionation

1) Flowage differentiation. This idea, demonstrated experimentally by BHATTACHARJI and SMITH (1964), suggests that early crystallised material together with any xenolithic material will be concentrated in the centre of moving columns of magma by the differential flow between the wall and the centre. It has been used by BARAGER (1960) and PRINZ (1964) to explain the distribution of the felspar megacrysts in the Leopard rock dykes of the Canadian Shield. While this mechanism may play a subsidiary role in the accumulation and sorting of the megacrysts within the centre of the microsyenite-trachydolerites it is not the main cause of differentiation as it can be demonstrated conclusively that the margins were emplaced before the centres and crystallisation began at the margins and progressed inwards.

2) Multiple intrusion. UPTON (1962, p. 31–35) gives a brief factual account of the microsyenite-trachydolerite dykes on Tugtutôq which is similar in all important points to the description in the present study. In his interpretation of the phenomena UPTON suggests that the dykes were formed by multiple intrusion in which basic magma took advantage of a pre-existing fissure filled by partly consolidated microsyenites. The various transitional rocks can therefore be interpreted as hybrids between the two magmas some of which formed *in situ* and some of which may be due to alkali contamination at depth. UPTON

further explains the concentration of megacrysts in the basic centres by suggesting that the gabbro was derived from below an original anorthositic source rock while the syenites were derived above this level. There is no doubt that the microsyenite-trachydolerite dykes are the result of multiple intrusion, the occasional fragment of the margins within the centre and the local chills between centre and margin show that in some dykes there was a time lapse between the solidification of the microsyenite and the intrusion of the more basic centres. However, UPTON's interpretation needs modification before it can explain three important points: the common and widespread occurrence of the features described in this section and the fact that they occur in dykes belonging to different generations; the gradational changes in the marginal microsyenites which apparently occurred before the intrusion of the central basic masses; the presence of occasional megacrysts in the chilled microsyenite zone. A fourth point, the presence of a high proportion of megacrysts rather than anorthosite xenoliths, is discussed in a later section (p. 179). It is therefore suggested that a third mechanism, differentiation of the magma either within the dyke fissure or in a chamber feeding the dyke is the principal factor controlling the way in which these dykes developed and the distribution of megacrysts within them.

3) Liquid fractionation. The common occurrence of syenite closely followed by gabbro within the same dyke fissure suggests that the two rocks are closely related. If the features described were limited to one dyke or even a single generation of dykes of the same composition, then the intrusion of gabbro into partly consolidated syenite might be regarded as a coincidence. However, when this phenomenon is seen over a wide area in dykes formed at different times and showing a considerable range in their bulk compositions it is reasonable to suggest that the relationships are due to the same fundamental process affecting the differentiation of a series of parent magmas within separate intrusions. The parent magmas are thought to have had the composition of alkali gabbros and ranged from under- to slightly over-saturated, with possibly some initial variation in the relative proportions of potassium and sodium.

Most accounts of the production of syenitic magma from a more basic parent rely on the well-used mechanism of crystal fractionation to produce an alkali-rich residue with a relatively low melting point. However, crystal fractionation involves the progressive removal of material *as solids*, thus a series of magmas tapped from a chamber differentiated by this mechanism must always follow the same sequence from least differentiated to most highly differentiated. In the microsyenite-trachydolerite dykes, and also some of the major Gardar intrusions, however, the sequence is reversed. In the absence of any evidence suggesting that the later more basic rocks were intruded as a mush, crystal fractionation cannot be regarded as the main mechanism of differentiation. If the magmas are genetically related then they must have separated while both were in the liquid phase.

The idea that magmas can differentiate while still in the liquid is far from new in geological thought although until recently it has been overshadowed by BOWEN's views on the complete miscibility of silicate melts and his championship of crystal fractionation as the dominant mechanism of differentiation. In the last few years considerable field evidence has been presented to show that two magmas intimately associated in the same intrusion may possess such different properties that they remain immiscible (for example, the "emulsion rock", shown in BLAKE *et al.*, 1965, plate 19b), or that differentiation can take place within a magma chamber by such processes as alkali diffusion while the magma is still liquid. A good summary of recent literature is given by HAMILTON (1965) and the same author has given the process the useful term liquid fractionation which can be used without having too many implications about the actual method by which the suite of liquids separate. It is interesting to note that the ideas of liquid fractionation have often been applied to provinces with some resemblences to the Gardar, for example the peralkaline rocks of Oslo (SÆTHER, 1950), and the alkaline suites of the Scottish Midland Valley, Oslo and the Kola peninsula (TOMKEIEFF, 1937, 1961).

The ideas which have the most application to the processes seen in the microsyenite-trachydolerite dykes are those published by KENNEDY (1955). He suggests that the upward diffusion of alkalies and sulphides in the presence of water might be expected to produce moderate differentiation in a chamber several miles high.

It can be seen from the mineralogy of the microsyenite-trachydolerite dykes that the original magma from which they formed was fairly rich in water and probably other volatiles. Biotite, and chlorite are common in both the marginal syenites and the central gabbroic rocks while calcite and fibrous amphibole fill many of the vesicles which occur in the transitional zone between margin and centre. Plagioclase insets are commonly heavily sericitised. Sulphides occur in the marginal alkali-rich parts of the dykes.

Differentiation of the type proposed by KENNEDY could occur either in a chamber feeding the dykes, in which the change in composition from early margins to late centre would represent the successive tapping of lower levels in a fractionated magma at depth, or it could take place within the dykes themselves. In either case the results during the injection of the dykes would be similar: before solidification the fissures would be filled with a column of magma graded from syenite at the top to alkali gabbro at the bottom. Some crystal fractionation could be expected in the column which might assist the differentiation process already started by liquid fractionation but there is no evidence that it became the dominant mechanism for the separation of the two end products. As more material was added from below, or as the dykes moved upward the alkali fractions at the top of the fissures chilled against the country rock leaving margins of syenite. Continued movement in the centre of the dykes brought up successively more basic material which chilled against the already solidified contact zones. In some dykes the emplacement appears to have progressed smoothly and there is a complete transition from margin to centre, in others movement was probably in a series of pulses and there may be several breaks in the change from syenite to alkali gabbro.

The theory that the dykes differentiated by liquid fractionation does not depend on any fixed ideas about where this process took place. However, there are several reasons to suggest that at least some differentiation could have occurred within the dyke fissures themselves whether or not there was any change in the magma type introduced into the dyke fissure from a hypothetical chamber at depth. Dyke fissures are practically the ideal form put forward by KENNEDY; they probably extend to a considerable depth, and their narrowness precludes any strong circulatory movements of the magma which might counteract the considerable changes in temperature and pressure from top to bottom of the chamber. It is therefore rather surprising that the features found in the microsyenite-trachydolerite dykes are not more common in other dyke swarms. A possible explanation may be found in the relative rate of alkali diffusion compared with the speed of dyke emplacement; a rapid forceful injection of basic magma into a dyke fissure would have little chance to differentiate before solidifying while a slowly moving column may differentiate as it ascends. The rate at which alkali diffusion takes place is probably controlled by the amount of volatiles present in the original magma; the mineralogy suggests that this was high in the microsyenite-trachydolerite dykes. Several features suggest that the rate of flow was small in the microsyenite-trachydolerite dykes. The most important of these is the fact that the margins are preserved at all, for if a large quantity of fast-moving basic magma was introduced into a narrow fissure with partly consolidated syenite along the margins it is unlikely that the earlier rock would survive. Some of the dykes appear to have been injected in pulses, in which case, although the actual speed at which the magma moved may have been relatively fast, there was probably a chance for differentiation to take place during the pauses between successive pulses. The alkali felspar laths of the marginal rocks show a poorly developed alignment and although the felspar megacrysts in the centres of most of the dykes show some preferential orientation it is not enough to suggest violent streaming of material.

One point of interest which can be seen in the analyses listed in Table 4 is that many of the microsyenite-trachydolerites and some of the other anorthosite-bearing dykes have silica contents in the range 48–54 $^0/_0$ SiO_2. UPTON (1964, p. 45) notes that "in spite of the large number of variegated intrusions on and around Tugtutôq there would appear to be a gap between rocks with a silica content 40–48 $^0/_0$ and those with silica 54–75 $^0/_0$". It can be seen from the analyses listed in Table 4 that the microsyenite-trachydolerite dykes show a less well defined separation into gabbroic and syenitic fractions. In some cases the gradation found could be explained by hybridisation of two distinct magma types. However, it appears more likely, in view of the common occurrence of the phenomena, that a different mechanism controlled differentiation in the dykes, or at least if the same processes were involved they did not reach completion.

WATTERSON (1968) describes a series of dykes intruded under plutonic conditions in which the emplacement of basic magma was preceeded by a streaming of alkalies along the same fissures which later controlled the dyke emplacement. The alkalies replace the surrounding country rock and form adinoles. The method by which the alkali material differentiated from the basic magma in these dykes probably differs considerably from the method suggested in the microsyenite-trachydolerite dykes; however, they provide a useful illustration of the fact that alkali material can in some cases migrate at a greater rate than that of the intrusion of basic magma along a fissure. It is interesting to record that one particular group of the plutonic dykes from the mouth of Isortoq has very similar features to the microsyenite-trachydolerite dykes. The dykes show a central accumulation of oligoclase-andesine phenocrysts and fine-grained margins significantly richer in silica and alkalies than the groundmass of the centres of the dykes.

b) The distribution of the plagioclase insets within the microsyenite-trachydolerite dykes

If it is accepted that these dykes were formed by differentiation of a single parent magma, whether in the dyke fissure itself or in a single chamber at depth, then the concentration of the megacrysts in the centres cannot be due to the derivation of the basic and syenitic components from different levels. However, a more attractive theory can be suggested by considering the relative densities of the plagioclase megacrysts and those of syenitic and gabbroic magma.

According to PULVERTAFT (1965) the density of a basic Gardar olivine gabbro magma (the Eqaloqarfia dyke) was approximately 2.65 g/cm^3 at 1150°, its temperature of crystallisation. That of the gabbroic hybrids and the syenites seen in the present dykes cannot be calculated with any precision because they are highly variable rocks and are thought to have contained considerable volatile material. The mean values for density determinations on a series of samples across dyke 8 are given in Table 5 (p. 134). These show a consistent increase in density from 2.70 g/cm^3 at the margin to 2.85 in the centre of the dyke. From these it can be estimated that the gabbroic end of the series could be expected to have had a density between 2.55 g/cm^3 and 2.65 at its melting point, while the syenites and intermediate rocks probably ranged between 2.40 g/cm^3 and 2.55 depending on the stage of differentiation reached and the amount of volatiles present in solution. The density of the plagioclase megacrysts (An_{50}) may be calculated to be 2.60 g/cm^3 at 1160° and slightly higher if they were included in a magma with a lower temperature. Although these figures can only be rough approximations it can be seen that sodic labradorite crystals should sink in a syenite magma. They would probably remain suspended in a moderately differentiated alkali gabbro magma and rise in a basic gabbro magma. If some allowance is made for the buoyancy effect of an upward moving magma stream, labradorite crystals may be expected to have remained suspended in magmas which had a density between 2.55 and 2.60 g/cm^3.

A corresponding argument can be used for the alkali felspar rhombs found in the dykes. These probably had a density of about 2.52 g/cm^3 while in the magma. They would therefore be expected to sink in the most differentiated of the syenites but to have remained suspended or risen in the other magmas. This idea agrees well with the distribution of the rhombs in the microsyenite-trachydolerite dykes; they form a very small part of the most differentiated marginal rocks, rise to a maximum in zones of moderately basic syenites and become progressively rarer towards the centre of the megacryst-bearing alkali gabbroic parts of the dykes. Some overlap between plagioclase megacrysts and alkali felspar phenocrysts is to be expected especially in the smaller dykes where a complete sorting probably could not take place before solidification.

It has been shown that the microsyenite-trachydolerite dykes crystallised from the margins inwards and that material continued to move up the centre of the fissure after the margins had solidified. Thus, a series of samples taken from margin to centre at the present erosion surface represent successively deeper levels in the original magma column. The distribution of the alkali felspar phenocrysts, the plagioclase megacrysts and the anorthosite xenoliths in these dykes can thus be explained as the result of settling of solid particles in a density graded column. This mechanism neglects the effects of a difference in rate of flow between margin and centre, as it has been suggested earlier that this plays only a minor role in the differentiation of the dykes. However, it can be noted that if the mechanism suggested by Bhattacharji and Smith (1964) is at all applicable to the microsyenite-trachydolerite dykes then it will provide an additional reason for the segregation of the megacrysts away from the margins. The few fragments caught up in the margins can be explained as foreign material which was carried up by the early phases of the dyke and trapped in the solidifying syenites before they could sink to their own density level. This is supported by the observation that the fragments in the marginal syenites are often larger than those in the graded centre of the dykes and are commonly xenoliths, not megacrysts.

c) The emplacement of dykes with a high proportion of foreign material

The formation of the fractures filled by the dyke presents a tectonic problem. Many appear to be simple tensional openings into which magma was emplaced, often in more than one phase. Several dykes, especially those in the Nunarssuit-Kobberminebugt area (see, for example, Harry and Pulvertaft, 1963) show signs that the dykes were emplaced along fractures in which horizontal movement was effective.

If, during dyke intrusion, the dyke fissures were not simply enlarged by tension normal to their walls they might conceivably have fluctuated in width and this could explain a puzzling fact. Some relatively thin dykes (dyke 2, p. 34, for example) are in places so choked from wall to wall with large anorthosite xenoliths that their emplacement can hardly be envisaged. It is tempting to accept this impression at its face value and to suppose that the assemblage could not have been intruded in its present state. Then the only logical course is to assume that after initial expansion the dyke fissure contracted before the dyke consolidated and the xenoliths thus were concentrated by expression of interstitial magma. Conceivably some monomineralic dykes described in geological literature might be explained by such filter-press action.

d) Incorporation of anorthositic material in Gardar intrusions

Highly xenolithic minor intrusions have been described from various countries. Elders (1957) has described from Norway a dyke with abundant inclusions of metamorphic rocks not seen *in situ* near the dyke and has referred to other minor intrusions which contain so many xenoliths that they resemble conglomerates. The problem of emplacement of these dykes may not be unrelated to that of the xenolithic dykes in Greenland, but a striking feature of the latter is that although most of them intrude Julianehåb Granite their inclusions are almost wholly anorthositic. Granite xenoliths are virtually confined to certain tracts of a few dykes and these tracts often lack anorthositic fragments.

The general absence of granitic xenoliths can hardly be due merely to their chemical instability in the basic dyke magmas. Nor does it seem likely that granite fragments were in fact incorporated in the dykes and then swept upwards, to levels now eroded, in advance of xenolithic material of deeper origin. For some reason anorthositic rocks were preferentially brecciated and incorporated in the intrusions.

Perhaps this simply reflects their physical properties. GROUT and SCHWARTZ (1939, p. 6) cite POWERS' (1915) suggestion that an unusual amount of shattering of that very compact rock anorthosite accounts for the presence of many anorthosite fragments, but few others, in a dyke that has probably crossed several formations in Minnesota.

However, more precise reasons might be sought for the strikingly wholesale disintegration of anorthosite to blocks, crystals, and crystal cleavage fragments in the South Greenland dykes. DALY (1933) has invoked strain due to heating of country rocks as a possible factor in overhead stoping by plutons. Could strain due to heating in magma have assisted disintegration of the anorthosites in Greenland? From the thermal expansion coefficients of felspar, it would seem to have been an unimportant factor.

Did the anorthosite disintegrate because it was incoherent? It might have been invaded by Gardar magmas after the consolidation of the massive granular anorthosite which forms shattered irregular xenoliths in the Gardar hosts, but before the complete consolidation of the later parts of the body. This would suggest a close genetic connection between the inclusions and the Gardar gabbros. The preferential inclusion of anorthosite in the Gardar dykes could then easily be explained by suggesting that the anorthosites formed a partially consolidated roof to the developing gabbros. The gradation seen between the slightly more calcic and magnesic granular anorthosites, through the many varieties of secondary anorthosite to the andesine megacrysts in the trachydolerites, might be regarded as the result of the same differentiation which gave rise to the range of Gardar rocks from gabbro to syenite.

It could be argued that the anorthosite series is due to the shattering of an earlier body and that all the gradational characters are due to secondary processes impressed on the insets by the Gardar magmas.

HESS (1960) has suggested that resorption of plagioclase may have been important in the formation of anorthosites, and YODER and TILLEY (1962) do not dismiss the possibility of the extensive resorption of plagioclase in basic melts. However, such a process, if it were the main cause in producing the series from granular anorthosite to megacrysts, was so extensive that the original character of the majority of insets would have been completely destroyed and any argument about their original characters must be pure speculation. Resorption can be seen in many fragments but it is generally limited to the replacement of felspathic and mafic material which had become altered to sericite and uralite before the rocks were included in their present host. Recrystallisation of single blocks detached from an earlier body and suspended in a deep Gardar magma chamber is an obvious mechanism for producing some of the features seen in the giant crystals with their gradational properties between laminated blocks and single megacrysts. However, there is considerable evidence that many of the fragments have retained earlier features although suspended in magmas with which they were markedly out of equilibrium. It seems reasonable therefore to suggest that although resorption and partial recrystallisation may have produced many of the minor features seen in the insets, they are unlikely to have been the dominant processes linking the characters of the insets with their hosts.

The break-up of the granular anorthosites could be ascribed simply to the same, predominantly tensional, stresses that controlled the intrusion of the main swarms of mid-Gardar dykes. However, it appears quite likely that many of the anorthosites were already fractured for some time before they were included in their present hosts, and many of the larger xenoliths show signs of several stages starting with a granular anorthosite which was broken up, included in fresh gabbroic material, and then broken up again in some cases several times before their final inclusion in the dykes. This suggests that some process was already operating at depth before the intrusion of the mid-Gardar dyke swarms. A possible explanation is that suggested by HIGGS (1954) for the San Gabriel anorthosite in which he suggests that the cataclastic textures are due to the release of high vapour pressures built up in anorthosite just consolidated from anorthositic gabbro magma. There is certainly considerable evidence to suggest that the anorthosites were affected by considerable action of volatiles prior to their inclusion in the Gardar host; many of the xenoliths are heavily sericitised and the mafic minerals in the inclusions are often uralitised while those in their hosts are fresh. However, it is not necessary to postulate that volatiles which caused the alteration of the anorthosites and possibly their brecciation were derived internally. If the idea that the anorthosite formed a layer over developing Gardar magma is correct, then they may have formed an impervious cap which was periodically brecciated. It may be noted in this context that many of the host dykes, especially in the Kobberminebugt area, show signs of crystallising from a water-rich magma and it is possible that the release of dyke magma into the fissure may have been initiated by explosive disruption of an anorthosite cover. It has also been suggested (p. 120) that the formation of black felspar is linked to the accumulation of water. This apparently took place during the main phase of megacryst formation which occurred later than the formation of the granular anorthosites (p. 179).

e) Evidence of lateral flow in Gardar dykes provided by the felspathic inclusions

The main direction of magmatic flow during the intrusion of the dykes appears to have been vertically upwards but it seems that some lateral movement of material may also have taken place. Ayrton (personal communication) has observed that a dyke from the Tigssaluk area which changes course in plan contains a concentration of plagioclase insets along the outer margin of the bend just as shingle piles up at bends in rivers. East of Kobberminebugt a horizontal section through a felspathic dyke shows a narrow zone relatively free of felspathic material terminating at a large anorthosite xenolith as if the latter had deflected inclusions moving relative to the xenolith in a horizontal direction (Fig. 33). Several dykes in the Nunarssuit-Kobberminebugt area show a decrease in the size and abundance of fragments towards the north-east suggestive of flow in that direction, while in the giant dyke of Tugtutôq the inclusions decrease in size and abundance towards the south-west. An interesting change in the abundance of megacrysts is seen in an early Gardar dolerite which can be traced east from the islands just south of Julianehåb to Ũnartoq Fjord, a total distance of about 60 km. In the west the dyke overlaps with the area intruded by later Gardar dykes containing abundant xenoliths. In this area it contains a moderately high proportion of felspathic matter, some of which has accumulated as glomero-porphyritic masses. Eastwards the fragments become smaller and less abundant although they continue outside the area in which plagioclase insets are normally found. Windley (1963) reports that at one place, where the dyke forks into a main eastward trending branch and two smaller apophyses which die out a short distance from the fork, the megacrysts are concentrated in the apophyses, suggesting that they were packed by lateral movement from the west.

Where the dyke is last seen on the shores of Ûnartoq Fjord it is completely free of xenolithic material.

The observations on lateral movement in the dykes are too scattered to draw any firm conclusion about the size or shape of the parent anorthosite body. However, it should be noted that in the examples cited from the mid-Gardar dyke swarms, the dominant movement was away from the areas later intruded by syenite. This observation coupled to the fact that the anorthosites show a distinct concentration in dykes close to the syenite centres may have considerable genetic significance.

VIII. DISCUSSION AND COMPARISONS

a) The "foreign" origin of the inclusions

For several reasons the **anorthosite inclusions** cannot have formed directly by crystallisation of basic magma in the dykes themselves. In the first place the amount of dolerite and gabbro in the dykes is too small in relation to the volume of inclusions. Some of the dykes are so rich in felspathic inclusions that their total composition is that of a gabbroic anorthosite or anorthositic gabbro. However, the composition of the host itself does not differ significantly from nearby dykes of the same generation with no inclusions. It might be argued that this is due to the level of erosion and that all the dykes in South Greenland contained masses of felspar either at depth or now eroded. There is no direct method of proving this statement incorrect. The analyses listed by WATT (1966) do not suggest that the Gardar rocks are in any way unique. Further, most of the large xenoliths show features which suggest that they were solid at the time of incorporation into their present hosts and must therefore be regarded as xenoliths whether or not they are genetically connected with their hosts. The minority of xenoliths which show local gradational contacts with their hosts, such as those seen at Narssaq and Bangs Havn, have already been explained as secondary anorthosites, formed by a mixture of derived fragments and an early phase of the host magma (p. 20).

Although the anorthosites may be regarded as xenoliths it is apparent from the descriptions in the earlier sections that a complete gradation exists between the obviously "foreign" masses of granular anorthosite choking the Kobberminebugt gabbros and the single felspar megacrysts which are present in a large variety of Gardar hosts. This series could either represent a reworking of disintegrated anorthosite fragments derived from an older source, or it might represent a series of rocks genetically connected with the Gardar igneous activity, some of which formed solid anorthosite before their inclusion in their present hosts and some of which were unconsolidated or partially consolidated plagioclase-rich masses at depth.

The origin of the **felspar megacrysts** is clearly closely connected to the anorthosites. They occur together in all proportions and combinations and their mineralogical characters are similar. However, the foreign origin of the megacrysts is much more difficult to establish. Some are derived from the break down of coarse-grained anorthosite but the majority show characters which are intermediate between the felspars from the granular anorthosite and the felspars from the host rocks, and would be regarded as cognate were it not for their association with demonstrable xenoliths. WATT (1968) uses the term "cognate xenocryst" to describe the large felspars found in the microsyenite-trachydolerite dykes from the Qaersuarssuk area and the term appears highly suitable.

b) Extent, structure and age of the parental anorthositic mass

i) Extent and structure

The area in which the inclusions occur in appreciable amounts is known to be at least 200 km long and up to 150 km wide and may well prove to be larger. Moreover, this area is bounded on one side by the sea, on the other by the Inland Ice. It would be a surprising coincidence if these geographical boundaries were also the limits of the area in which the inclusions occur. Thus, even allowing for some lateral transport in the dykes, the parental anorthositic mass must be of considerable size: 250–300 km long does not seem excessive as a minimum estimate and this would be greatly increased if the occurrences noted by R. BØGVAD (unpublished diaries) on the east coast of Greenland are pertinent. The area over which the anorthosite fragments occur appears fairly closely limited to the part of South Greenland affected by the E–W system of transcurrent faulting which forms a prominent feature of the Gardar. USSING (1912) drew attention to the graben-like nature of the Gardar faulting and more recently several attempts have been made to compare the Gardar alkali magmatism with that found in rift valleys in other parts of the world (see, for example, BRIDGWATER and WALTON, 1964; SØRENSEN, 1966). Whether the faulting marks the approximate margins of the parental anorthosite mass or whether the distribution of the xenoliths within this block is due to the brecciation of a larger body within the faulted zone is unknown.

There is no evidence to show whether the parental anorthosite is made up of a series of unconnected bodies at depth or whether it forms a more or less continuous layer. There is a marked concentration of xenolith-bearing dykes close to the major alkali intrusions which might suggest that the parental anorthosite forms a series of disconnected bodies each of which is genetically connected to a syenite centre. However, this could also be explained by suggesting that the syenites were intruded in areas of crustal weakness round which an earlier continuous mass of anorthosite had been brecciated at depth.

The parental mass is completely concealed, for its fragments cannot be matched with any other anorthosites outcroping along the western coast of Greenland (see Section V (b)). Since it is nowhere exposed and yet has great extent there is reason to suppose that, in places at least, it lies at considerable depth. The general lack of other inclusions in the felspathic dykes suggests that the parental mass may have lain fairly close to the source of the dyke rocks. It has been argued by Upton (1964) from his work on Tugtutôq that the relative absence of anorthosite xenoliths in the syenitic rocks on that island suggests that the parental mass lies at a level between that of the source of the basic dykes and that of the syenitic rocks. However, it has been suggested on p. 151 that the anorthosites probably occurred above the level at which the syenites differentiated and that their absence in syenites is largely controlled by the relative densities of labradorite and syenite magma. The intimate association of gabbro, syenite and felspathic fragments in a large number of dykes and the rarity of other inclusions suggests that anorthositic material occurred in the roof of the magma chambers from which Gardar magmas were released. This need not, however, mean that the original anorthosite lay at this level as it is possible that the unfractionated Gardar magmas carried large masses of anorthosite with them. These may have remained as rafts at the top of the chambers in which the syenites and other rocks were evolved until they were brought up by successive pulses of material released from the differentiating magma.

Compositional banding or preferred orientation of felspar crystals in a few xenoliths might suggest that the parental mass was layered. But such evidence is rather scanty. Moreover, xenoliths that could represent ultrabasic differentiates (peridotite, pyroxenite) complementary to anorthosite in a layered complex are lacking. The large xenoliths lie solely within the range anorthosite-felspathic gabbro, and many of the less felspathic members of this series are "secondary" rocks formed by mixture of basic Gardar magma with anorthosite. Biotite pyroxenites do occur at Narssaq (Ussing 1912, Upton 1966) but belong to a monchiquite-carbonatite suite and are only relevant as an example of a highly fractionated Gardar rock crystallising from a volatile-rich magma. A few isolated "phenocrysts" of olivine and occasionally hypersthene and magnetite occur in Gardar dykes and it is possible that these may represent the mafic equivalent of the plagioclase megacrysts. Daly (1933) suggests that the rate of sinking of solids in a magma depends on their size, and thus larger fragments of ultrabasic layers, detached from an original layered body above the source of the Gardar host rocks, would be expected to sink while single crystals might remain suspended. However, it is equally probable that if the formation of the anorthosites is related to Gardar magmatism then a mafic counterpart of the felspathic rocks did not develop above the source of the present hosts but was formed at a greater depth, at the bottom of the (hypothetical) magma chamber from which the Gardar rocks were emplaced.

ii) Age

Most, if not all the parental mass must be older than mid-Gardar for the majority of the xenoliths occur in the mid-Gardar dyke swarm. From the available age determinations (Bridgwater, 1965) it seems likely that the mid-Gardar dykes were emplaced at about 1200 m.y. at a time of considerable crustal tension in South Greenland. A secondary anorthosite xenolith found in the early mid-Gardar gabbro at Narssaq (Upton, 1964), has been dated by Geochron Laboratories as 1025 ± 70 m.y. (biotite) using potassium-argon methods. Presumably this date reflects the nearby Ilímaussaq intrusion (1020 m.y., modified from Moorbath, Webster and Morgan, 1960).

A lower limit to the age of the anorthosites is less easy to establish; a few fragments resembling the main anorthosite xenoliths are found in early Gardar dykes suggesting that the felspathic material was available from the beginning of the Gardar onwards. This need not however imply that the whole parent body formed prior to the early Gardar dolerite intrusions as it is perfectly feasible that a gradual accumulation and solidification of felspathic rock took place at some level above the ultimate source of the Gardar magma at the same time as dykes were intruded. Sporadic felspathic inclusions also occur in the pre-Gardar dykes north of Ivigtut (see Section III) overlapping the northern extension of the Gardar occurrences. These dykes are demonstrably older than the plutonic episode dated between 1500 and 1650 m.y. and are generally regarded as either contemporaneous with or older than the Ketilidian lavas. For these to be genetically connected with the inclusions in Gardar rocks necessitates the existence of a felspathic body at depth under South Greenland throughout at least one mountain-building episode. The relatively undeformed nature of the fragments found in Gardar dykes is most easily explained by suggesting that they were formed after the last orogenic activity in South Greenland which would link them closely with the Gardar. Alternative suggestions based on ideas either that the fragments were derived from a source at such a depth that they escaped strong deformation while the overlying less plastic rocks were folded, or that they represent earlier material reworked and recrystallised at least partly as magmatic rocks in Gardar magmas at depth, are imaginable but would seem almost impossible to prove using techniques available at the moment.

c) Review of the main varieties of anorthosite found in the crust

The abundance and size of the anorthosite xenoliths described permits some comparison between them and other anorthositic bodies. To do this, a brief general review of anorthosites would seem useful.

The anorthosites of the world can be classified using a modification of the schemes proposed by BOULANGER (1957, 1959) and HARPUM (1957). As this classification is essentially descriptive, one particular body may belong to two of the groups described below. depending whether primary features, formed when the anorthosite first crystallised, or secondary features, formed during subsequent events, are emphasised.

i) Sakeny-type: (pre-tectonic conformable anorthosites interstratified with metamorphic rocks)

These rocks are characterised by metamorphic textures. Where they are associated with high-grade gneisses, for example at the type locality in Madagascar, the conformable anorthosites contain anorthite or bytownitic plagioclase; however where they are associated with lower grade surrounding rocks, for example in the Bergen region of Norway (KOLDERUP and KOLDERUP, 1940), the felspars are commonly less calcic. The origin of the Sakeny-type anorthosites is controversial. Most of the larger complexes, for example the Bergen arcs and the Sittampundi complex (SUBRAMANIAM, 1956), and some of the smaller bodies, such as those of South Harris (DAVIDSON, 1943), have been interpreted as original stratified igneous bodies subsequently folded and metamorphosed so that their original features have been destroyed. A second school of thought, for example LACROIX (1939) and BOULANGER (1959), has suggested that many of the Sakeny-type anorthosites are derived by the recrystallisation of marly sediments. Other authors, for example BERTELSEN (1957), have invoked "anorthositisation" a metamorphic migration of material leaving some layers in the gneiss complex enriched in calcic plagioclase. Part of the difficulty in explaining this type of anorthosite undoubtably lies in the inability of many workers to distinguish between primary characters and those which may have formed at a later date. In particular it should be noted that the present conformable nature of these bodies in a highly folded gneiss terrain is no evidence of their original attitude and cannot be used to prove their sedimentary origin.

Sakeny-type anorthosites are known to occur in considerable amounts along the west coast of Greenland and have been used by BERTHELSEN (1961) as stratigraphic marker horizons. These rocks were originally described as metamorphic in origin and have been used by several authors discussing the origin of the intercalated anorthositic horizons in gneisses as evidence of the general metamorphic origin of the Sakeny-type. However, recent work by WINDLEY (1965), shows that they are derived from an immense igneous body, comparable in size and mineralogy to the Bushveld complex. Descriptions of these rocks have been published from four localities:

1) Ivigtut region. Gabbro anorthosites in the Ivigtut region have been described by, among others, BONDAM (1955), BONDESEN (1960), BERTHELSEN (1960a) and AYRTON (1963). They form disrupted bands conformable in the gneisses and comprise plagioclase (An_{35-80}) hornblende and minor amounts of sphene and apatite. They generally exhibit a crude augen structure: small plagioclase augen in a network of hornblende. The presence of plagioclase more sodic than that typical of Sakeny type anorthosites has been explained by BONDAM (1955) as a result of retrograde metamorphism. The general retrograde nature of the metamorphism of the Ivigtut gneisses has been confirmed by OEN (1962).

2) Tovqussap nunâ. In this complexly folded region gabbro anorthosites form conspicuous, continuous layers up to 200 m wide with boudins and disrupted layers resembling those described from the Ivigtut gneisses. They have been described by BERTHELSEN (1957 and 1960b) who suggests that they were either formed by anorthitisation *in situ* or were of magmatic origin representing former flows and ash bands. Some pass laterally into skarns interpreted by BERTHELSEN as metamorphosed limestones.

3) Buksefjorden. Anorthosite occurs in this region as a synform layer conformably enclosed by amphibolite facies paragneiss. SØRENSEN (1955) suggests that these rocks originally belonged to the granulite facies but have undergone retrograde metamorphism. This is true of many of the Sakeny-type anorthosites so far described from West Greenland and suggests that they are older than the migmatisation which affected this area. A date of 2700 ± 130 m.y. (ARMSTRONG, 1963) from biotite gneiss at Godthåb therefore suggests that the anorthosites may be even older.

The anorthosite varies in composition from almost pure plagioclase (labradorite to bytownite) rock to types transitional to hornblende diorite. Calcium expelled from felspar during metamorphism has been incorporated in clinozoisite.

4) Qaqortorssuaq, Søndre Strømfjord. Here a mass of anorthosite, with an estimated volume of at least three cubic kilometres, occurs in gneisses belonging to the amphibolite facies (ELLITSGAARD-RASMUSSEN and MOURITZEN, 1954). The body consists almost wholly of bytownite; there is a little clinozoisite. Its contacts with the surrounding gneisses are gradational. This body is of unknown origin.

ii) Adirondack-type (plutonic) anorthosites

Much of the recent information about the anorthosites of the Canadian Shield, where plutonic anorthosites reach their most impressive proportions, is unfortunately not readily available. The writer is therefore particularly indebted to J. P. BERRANGÉ for access to his (1962) thesis on one of these massifs and for mutually valuable discussion on the 1964 draft of this paper.

General features of large anorthosite masses in the Adirondacks and similar bodies elsewhere are as follows:

1) The bodies vary in size from small to very large. KRANCK's (1961) map of anorthosites in eastern Canada indicates their extent. Some bodies have been termed batholiths by several authors, a term that conveys the magnitude of their outcrop though, like many granitic so-called batholiths, many are probably thick sheets rather than stocks.

2) Adirondack-type anorthosites are only exposed in deeply eroded terrain. Thus they are almost completely confined to Precambrian areas.

3) They can generally be referred to either high amphibolite facies or granulite facies like the high grade gneisses that surround them. BOULANGER (1959) suggests that Madagascan anorthosites can be classified as a special type of enderbite and several writers have noted the conjunction of anorthosite and charnockite in high metamorphic terrains. GOLDSCHMIDT (1922) suggested a magmatic series from anorthosite to charnockite.

4) Many of them show features regarded as igneous in origin *e.g.* ophitic texture, flow banding, autoxenoliths—"block structure"—and sometimes igneous layering and intrusive contacts. Several of the features suggest an accumulation of plagioclase by flotation rather than by settling.

5) They are often very coarse-grained.

6) Most of them have been highly deformed. Evidence of crushing is generally present and the margins of the bodies often display a foliation in parallelism with that of adjacent gneisses. Many felspar crystals show deformed secondary albite twinning. It is controversial whether the plutonic anorthosites were deformed during or after emplacement. KRANCK (1961) and BOULANGER (1959) conclude that the plutonic anorthosites were formed prior to their deformation, while MICHOT (1960), BERRANGÉ (1962) and PHILPOTTS (1966) suggest that at least some plutonic anorthosites were emplaced synkinematically. Whatever the exact relationship between the anorthosites and the tectonic processes in the surrounding rocks it is generally clear that they are associated with considerable plutonic metamorphism such as found in the centre of fold belts.

7) There is often an age sequence: anorthosite – gabbro-anorthosite – norite – pyroxenite. This is seen, for example, in the enclosure of anorthosite by anorthosite gabbro, while mafic, later rocks are generally concentrated near the margins of the bodies as in the Whiteface facies of the Adirondacks (BUDDINGTON, 1939).

8) The plutonic anorthosites are generally associated with pyroxene-bearing salic rocks of the enderbite-mangerite-charnockite suite, which

are usually marginal and present features suggesting that they were formed during the last stages of consolidation of the anorthosite mass. The charnockites may in turn be followed by rapakivi granites. The whole suite from anorthosite to granite may be termed calc-alkaline. In less eroded terrains, for example the Andes and the "Rockies", only the more volcanic members are normally exposed (andesitic and rhyolitic lavas).

9) Cryptic variation and compositional banding are rare. Over large areas crystals of one mineral species vary but slightly in composition. In some individual bodies felspar compositions vary by no more than 10 % anorthite.

10) Although the plagioclase can range from An_{35} to An_{80} in composition, it is remarkable how often it lies solely within the range An_{40-50}. It is often clouded. Antiperthite is common, especially in the less calcic plagioclase. Hypersthene is the typical dark mineral, clinopyroxene is subordinate and olivine is a possible minor constituent. Garnet is common in the marginal facies of the bodies. Ilmenite frequently forms associated ore bodies.

11) Major associated bodies of mafic rock that could represent complementary differentiates of anorthosite derived from basaltic magma are not exposed. Moreover geophysical surveys of anorthositic areas have revealed negative gravity anomalies suggesting that such mafic rocks are lacking at depth (SIMMONS, 1964).

12) Some of the plutonic anorthosites are associated with younger post-tectonic layered intrusions (MORSE, 1964) with some similarities to the alkali gabbro-syenite intrusions of the Gardar.

The origin of plutonic anorthosites is highly controversial and discussion of the problem in this paper will be limited to the points which might have relevance to the origin of the inclusions in Gardar rocks. Most of the theories are based on observations on the anorthosites in the Canadian Shield. However, there appears to be some confusion in North American papers about what constitutes a plutonic anorthosite. The confusion springs from the attempt to place all the anorthositic bodies from the Adirondacks to the Labrador coast in one category; inspite of the considerable evidence that they were formed over a considerable time span under different geological conditions and have subsequently passed through widely different histories. In the descriptive classification used in this paper the term plutonic implies either that the anorthosites formed under plutonic conditions (corresponding to the plutonic-orogenic anorthosites of BERRANGÉ, 1966) or that they have passed through a plutonic phase subsequent to their initial formation. The plutonic anorthosites thus include the Adirondack anorthosite (BUD-

DINGTON, 1939, the Morin anorthosite (PHILPOTTS, 1966), the Saquenay anorthosite (BERRANGÉ, 1962, 1963) and the Allard Lake anorthosite (HARGRAVES, 1962); all of which were affected by the Grenville plutonic episode. The Nain anorthosite (WHEELER, 1960; MORSE, 1964), which predates Grenville plutonism by 300–400 m.y., has several features in common with other Canadian plutonic anorthosites and may represent rocks formed in the dying phases of the Hudsonian plutonic episode. However, it is apparently related to an undeformed layered suite of intrusive rocks (the Kiglapait intrusion) which must be regarded as essentially post-tectonic in origin. The Michikamau intrusion (EMSLIE, 1965) was intruded at approximately the same time as the Kiglapait intrusion and shows the same post-tectonic features.

Many of the features of plutonic anorthosites, particularly where they are relatively unaffected by later movements, suggest crystallisation from a magma. The origin and chemistry of this magma is controversial: it could be derived from the mantle, for example as a primary basalt; it could be derived by the mobilisation of a deep seated crustal layer such as might be left after a granitic fraction had migrated (RAMBERG, 1952, pp. 262 and 264); or it could have formed by anatexis of crustal rocks involved in deep-seated metamorphism during the formation of a fold belt (MICHOT, 1955). Accepting the magmatic origin of plutonic anorthosites there are two main schools of thought: the first, who suggest that the plutonic anorthosites were differentiated from a highly aluminous basic magma approximating to a water-rich gabbro anorthosite melt (BUDDINGTON, 1939, 1960); and a second who suggest that the anorthosites and the surrounding pyroxene-bearing salic rocks were both differentiated from acid dioritic magma (BALK, 1931; PHILPOTTS, 1966). According to PHILPOTTS this magma was probably water-poor, accounting for differentiation trends which approach those of tholeiitic suites. A variant of the BUDDINGTON school of thought is put forward by HARGRAVES (1962) and BERRANGÉ (1962) who suggest that the salic rocks associated with the anorthosites formed by anatexis of the surrounding gneisses when a hydrous gabbro anorthositic magma was emplaced under regional metamorphic conditions.

The main drawback to the gabbro anorthosite magma hypothesis is the exceedingly high temperature needed to keep a rock consisting of approximately 90 % labradorite in a molten state. BUDDINGTON (1960) attempted to minimise this difficulty by postulating a water-rich magma. The evidence for a water-rich magma is slight in the Adirondacks and non-existent in the Morin area described by PHILPOTTS. It might be suggested that the clouding seen in the felspars could be ascribed to their formation under water-rich conditions (see p. 120), however little is known about the time at which the clouding took place or the actual minerals present, and it may be noted that charnockitic rocks in general show clouded felspars without positive evidence of a high water content at the time of crystallisation. Furthermore the formation of the anorthosite – gabbro-anorthosite – iron-rich norite series characteristic of plutonic anorthosites requires the early crystallisation of a large percentage of the plagioclase. The writer knows of no experimental work which has demonstrated that the presence of a high percentage of water in a labradorite-orthopyroxene-ilmenite-water melt would give rise to this sequence and in fact the work described by YODER and TILLEY (1962) on natural basalt melts shows that the crystallisation of plagioclase is delayed with increasing water content.

The main objection against the idea that the rocks are derived from a dioritic magma is that in many bodies the genetic link between anorthosite and norites on the one hand and the pyroxene-bearing salic rocks on the other, is far from distinct. The basic rocks are clearly derived from a magma while the charnockitic members of the suite are often intercalated with gneisses. Apart from local areas of mixed rock they often appear unrelated to the main anorthositic mass. PHILPOTTS (1966) has shown a clear transition between anorthosite through mafic diorite to pyroxene syenite exists in the Morin rocks and has used considerable ingenuity in explaining the foliated nature of the late, more acid members of the series. However, chemical trends such as those given by PHILPOTTS cannot be used without supporting field evidence to prove that a particular series of rock types was derived from a common magma. Similar sequences could equally well form as a mixture of an original basic magma with material derived from salic country rocks. In fact there is considerable field evidence to support the second idea; the charnockitic members of the series are found on the margins of the anorthosite intrusions where they commonly show a mineralogical and structural gradation into the surrounding country rocks. Anatexis and mixing of the kind suggested could give rise to a whole series of secondary magmas ranging from alumina-rich basalts to granodiorites. Each of these secondary magmas might be capable of considerable movement so that in some instances evidence for their origin would be lost. Perhaps the most significant point in this discussion is the fact that the charnockitic suite is associated with anorthosites thought to have been emplaced during the formation of a fold belt or in the period immediately following tectonism. In these conditions the country rocks could be expected to be at a moderately high temperature and the introduction of a comparatively small amount of primary basic magma could be expected to give rise to a disproportionate mobilisation of the country rocks.

iii) Anorthosites in layered intrusions (volcanic anorthosites)

These anorthosites have been termed Bushveld-type. Other layered intrusions are more typical of the class than the Bushveld complex and the term seems therefore unsatisfactory. The majority of major layered bodies were formed under cratogenic conditions and following Kennedy and Anderson's (1938) distinction between plutonic and volcanic associations could be regarded as volcanic in origin in contrast to the anorthosite-charnockite suite which appears essentially plutonic.

Pertinent features of layered intrusions are as follows:

1) As their name implies layered intrusions are characterised by the alternation of different rock types generally in well-defined units. They are generally formed by some form of crystal fractionation, for example crystal settling under relatively stable tectonic conditions.

2) The rock types formed embrace a wide compositional range. Adjacent units frequently consist of contrasting practically monomineralic rocks. Delicate structures ascribed to crystal settling resemble features seen in sedimentary rocks.

3) The intrusions may lie conformably within their country rocks, but they were formed under conditions quite different from those responsible for the metamorphism of their country rocks. Any similarity in metamorphic facies between an intrusion and its surroundings is either coincidental or due to processes affecting both the country rocks and the intrusion after the emplacement of the latter.

4) Zoned crystals are common. The late products of crystallisation may differ markedly from the main constituents suggesting that the rocks were formed over considerable ranges in temperature and pressure.

5) Anorthosites are only minor components of the complexes, and, if they accumulated from a normal basaltic parent magma, basic and ultrabasic rocks are complementary to them. Even complexes ascribed to a highly aluminous basalt parent magma contain such cumulates complementary to anorthosite.

6) The anorthosites vary with the composition of the initial magma and the stage reached by this magma when they were precipitated. In most of them plagioclase lies in the labradorite-bytownite range; orthopyroxene is common but not essential. The parent magmas of most of the largest layered intrusions in which anorthosites occur appear to have been tholeiitic basalts. Barth (1962, p. 224) points out that plagioclase in accummulative anorthosite is commonly more calcic than An_{60}, whereas in the large plutonic anorthosite bodies it is more sodic than An_{65}.

7) Layered anorthosites are not confined to the Precambrian.

The sharp division of anorthosites into groups is artificial and there are many examples of transitional types. Anorthosites emplaced towards the end of major orogenic periods may show features of both volcanic and plutonic types, while volcanic anorthosites intruded at great depth may possess original features resembling those found in plutonic anorthosites. For the purposes of this paper the classification is descriptive; however, the differences between the different types, particularly between plutonic and volcanic anorthosites, suggests that there are important differences in the mode of origin of these rocks. In the writer's opinion the most important factor influencing the type of anorthosite formed is the relative density of the plagioclase and the magma from which it forms. In layered anorthosites the plagioclases form alternating layers with the mafic minerals and both must have accumulated at the bottom of a magma chamber. It appears that plagioclases more calcic than An_{70} sink in basaltic magmas. On the other hand the less calcic plagioclases of plutonic anorthosites often show features suggesting that they formed a top cumulate or at least hung suspended in their parent magma as this was emplaced. Plagioclases more sodic than An_{60} appear less dense than the majority of magmas from which they are precipitated. This provides an attractive explanation of the apparent lack of mafic counterparts to the large quantities of plagioclase found in the plutonic anorthosites. It is possible to imagine their presence at considerable depth under the bodies of plutonic anorthosite. In the cases where gravity surveys suggest the complete absence of the mafic rocks it is possible that the plagioclase-rich fractions became separated either as crystal-charged fractions intruded at a higher level in the crust or at a later stage during tectonic activity. Once formed a large accumulate of plagioclase of the size found in the Canadian Shield might be expected to act very differently to less homogeneous mafic rocks if both were subjected to a later orogeny. It would seem probable that these large plagioclase masses would be virtually indestructable. The fact that considerable plagioclase forms in the composition range An_{40-50} suggests that the parent magma must differ from that which normally gives rise to layered anorthosites. This may suggest large scale contamination of original basaltic magma under orogenic conditions or perhaps partial remelting of crustal material.

d) Comparisons between the parental anorthosite underlying South Greenland and major anorthosites in other parts of the world

The three fold classification given above can be used as a starting point when comparing the xenoliths in Gardar rocks with major anorthosites *in situ* in other parts of the world. The xenoliths are obviously not of Sakeny-type. They cannot be matched with any of the anorthosites described from the west coast of Greenland. Relics of gabbro anorthosite do exist in the granites and gneisses of the Julianehåb area (ALLAART, 1967), but these have been so broken down by three or more plutonic episodes that any original features they may have had in common with the xenoliths in Gardar rocks have been destroyed. In the Ivigtut region, where the gabbro anorthosite layers are better preserved, their textures, mineralogy and composition are so different to anything seen in the xenoliths that they cannot be regarded as a source rock unless they were completely destroyed and reconstituted in Gardar magmas. If the xenoliths were derived from concordant layers in gneisses then there would seem no reason why amphibolitic bands derived from the same complex should not form xenoliths in the Gardar dykes. These have never been observed.

On the other hand the xenoliths show many features of both plutonic (Adirondack-type) and volcanic (Bushveld-type) anorthosites. The composition of the unaltered granular xenoliths is intermediate between typical plutonic and layered volcanic anorthosites. Their plagioclase lies in the range An_{55}–An_{63}. Most of the plagioclase in Adirondack-type anorthosites is somewhat less basic and in the Adirondacks themselves lies typically in the range An_{30}–An_{50} (BUDDINGTON, 1939). Plagioclase in volcanic anorthosites tends to be more basic than the xenoliths of South Greenland, but some of the xenoliths, chiefly those in Tugtutôq, texturally resemble volcanic anorthosites. The Assorutit felspars are euhedral, thinly tabular, often well-laminated, and ophitically enclosed by olivine. A twin pattern characteristic of magmatic crystallisation is developed and deformation is absent or slight. In contrast the felspars from granular xenoliths form roughly equant xenomorphic grains with simple albite and pericline twinning while some evidence of strain is generally present. These are common features of plutonic anorthosites and it might be suggested that there are two distinct types of inclusion in Gardar rocks: the first derived from an older plutonic anorthosite underlying South Greenland with no close genetic connection to Gardar magmatism; the second a younger layered anorthosite of volcanic type developed on a relatively small scale. However the detailed petrology of the granular blocks shows a closer resemblance to Gardar gabbroic rocks than it does to any plutonic anorthosite described from the Canadian Shield. The minerals found in the granular anorthosites are practically identical in composition to the minerals of the more basic Gardar rocks, and the plagioclases in particular show many peculiarities, such as abnormal structural properties, which they have in common with plagioclases from other coarse-grained Gardar rocks and which have not been described from rocks affected by regional metamorphic conditions. It has been noted that the plagioclases are less calcic than normally found in volcanic anorthosites; however, this is also true of the plagioclases in the Gardar gabbros in general and reflects one of the overall chemical characteristics of the province. The texture of the granular anorthosite xenoliths resembles that of the plutonic anorthosites; however, this feature is not confined to rocks involved in the formation of fold belts and appears to reflect the mechanics of plagioclase accumulation rather than the effects of external forces.

UPTON (1964) has suggested that the South Greenland anorthosite may resemble the body described by SIMPSON and OTTO (1960) from **Angola**. South of the Kunene river this complex consists of a basement of granular anorthosite with a variable coarse-grained texture and consisting dominantly of labradorite-bytownite and a little orthopyroxene, overlain by an olivine-bearing layered series. UPTON's original comparison was based entirely on the similarities between the Assorutit layered anorthosite and the upper series in Angola. The comparison becomes more applicable now that it is shown that most of the Gardar xenoliths are granular anorthosite and that these had already solidified before the formation of the layered rocks.

The lower granular anorthosites are generally concordant with the surrounding gneisses. SIMPSON and OTTO describe concordant layers of "calcareous chert" within the anorthosites and attach considerable genetic significance to them; however, more recent work by D. K. TOERIEN (personal communication, 1963) has shown these layers to be a late feature. The surrounding gneisses show a local granulite facies metamorphism apparently connected to the intrusion of the anorthosites. Veins of mangerite, thought to be formed by rheomorphism, cut the contact zone of the granular anorthosite in a manner which is reminiscent of the high grade rocks surrounding the plutonic anorthosites of the Canadian Shield. The plagioclases in the granular anorthosite may reach 20 cm or more in size; they are often chalky-white and have been moderately severely saussuritised. Hypersthene is the dominant mafic mineral while olivine is typically absent.

In contrast, the overlying layered olivine-bearing anorthosite series is characterised by tabular laminated plagioclase crystals which do not reach the extreme grain size found in the granular rocks. Many of the felspars are dark, a feature ascribed by SIMPSON and OTTO to "metamorphic clouding" although they give no other information which might suggest that the rocks have been affected by secondary processes. The minerals are otherwise fresh and do not show the alteration seen in the granular rocks. The olivine-bearing anorthosites show large scale layered structures which are markedly discordant to the structures in the surrounding gneisses while the granular anorthosites show structures concordant to those in the country rocks. This might suggest that the granular rocks belong to an earlier, pretectonic, phase of emplacement than the overlying layered rocks, however, SIMPSON (personal communication) reports that in some areas north of the Kunene river the two types are interlayered.

Chemically the Angola anorthosites are considerably more basic than the Greenland xenoliths although there is some overlap. Plagioclases range from An_{80} in the most basic granular anorthosites to An_{50} in the more highly fractionated layered rocks. Olivines range from Fa_{16} to Fa_{35} (the An content quoted by SIMPSON and OTTO, p. 225 should be reversed so that sample ZM. 56 reads An_{67}, E. S. SIMPSON, personal communication). It seems quite probable that the Angola anorthosites are linked genetically with an alkali province but this relationship is not as clear as that found in the Gardar igneous province.

By using the Angola complex as a model UPTON (1964) suggested that the anorthosite complex underlying the Gardar province in South Greenland might be elongate in form, perhaps resembling the giant dyke of Rhodesia. This idea has some support in the distribution of both the anothosite-bearing dykes and the alkali intrusions. These are confined to a relatively narrow elongate strip associated with wrench faulting. It agrees with the suggestion by BERTHELSEN and NOE-NYGAARD (1965) that Gardar differentiation may have taken place in a series of elongate chambers of considerable vertical extent controlled by the E–W transcurrent faulting.

The Greenland rocks are probably closer chemically and mineralogically to the anorthosites of the **Dzugdzhursk massif** which are thought to extend for 250 km and occupy an area of 5000 sq. km. This body has been referred to the plutonic anorthosites by LEBEDEV and PAVLOV (1957) who compare it with the Adirondacks and the Pribaikal complex. In contrast to most anorthosites of this group the Dzugdzhursk massif occurs in medium to low-grade metamorphic rocks and has been little if at all affected by later deformation. It consists almost wholly of anorthositic rocks ranging from pure labradorite rock to gabbroic anorthosites ("gabbro norites" of LEBEDEV and PAVLOV) with about 80 % plagioclase and a chemical composition somewhat comparable to that of the analysed secondary anorthosite from Narssaq (Table 1). Large bodies of gabbroic or ultrabasic composition are lacking. The anorthositic rocks are coarse-grained (average grain size of plagioclase 1–4 cm) and either even-grained or containing megacrysts up to 20 cm. Some pegmatitic anorthosite rocks are present locally. The plagioclase crystals are similar in composition to most of those in the Greenland xenoliths. They tend to be equant and anhedral; albite twinning is dominant. Both rhombic and monoclinic pyroxenes occur. Minerals of the epidote group, biotite, alkali felspar, apatite, chlorite and Fe-Ti oxides can be present.

The alkali complexes of the **Kola peninsula** show many similarities to those of South Greenland. One complex, the **Gremyakha-Vyrmes pluton** (POLKANOV and ELISEEV, 1941), is of particular interest since it contains layers of both laminated and granular anorthosite which resemble the two textural types of anorthosite xenolith found in Gardar rocks. The Gremyakha-Vyrmes pluton is an oval body, approximately 20×6 km in diameter, 45 km SSW of Murmansk. It was emplaced in an active fault zone approximately 1500 m.y. ago during the second phase of alkali activity in the eastern Scandinavian-Karelian Shield. This period of magmatism also includes the carbonatitic rocks of Alnö in Sweden.

Three intrusive phases are recognised in the Gremyakha-Vyrmes pluton: 1) Hortonalitites, peridotites, anorthosites, gabbros, alkali gabbros, syenites and alkali granites. 2) A jacupirangite-urtite series, with some rock types in common with the carbonatitic suite found at the eastern end of the Narssaq peninsula, accompanied by various syenites such as malignites and foyaites locally resembling the rocks of the Ilímaussaq intrusion. 3) A nordmarkite-alkali granite series similar to the rocks found in the Narssaq intrusion. The gabbroic rocks of the oldest intrusive phase are the most important for comparisons with the anorthosite xenoliths and their Gardar hosts. This intrusive phase has been subdivided into three associations of rock types: a) A layered andesine gabbroic body consisting of alternating units of hortonolite, peridotite, pyroxenite, gabbro anorthosite, troctolite and anorthosite (see POLKANOV and ELISEEV, fig. 5, p. 30). The layered units wedge out into leucocratic gabbro. b) A layered oligoclase gabbroic body grading into akerites and pulaskites. c) A series of syenitic and pegmatitic veins cutting other rocks in the first intrusion phase.

The relationship between a) and b) is unknown. POLKANOV and ELISEEV suggest that the two groups differentiated respectively from gabbroic and oligoclase gabbroic magma which were intruded independently.

Both the gabbroic series and the oligoclase gabbro-pulaskite series show excellent layered structures with laminated minerals suggesting that the layering was formed by crystal settling. The mineralogical composition from contrasting melanocratic and leucocratic layers from the same series are essentially the same. In the gabbroic series the felspars average approximately An_{41} (occasionally showing zoning from An_{55} to An_{30} or less). The co-existing olivine is fairly iron-rich ($Fa_{45\text{-}60}$), it is accompanied by magnetite-ilmenite, titan-augite (2V 44°–51°), kaersutite and what POLKANOV and ELISEEV term diopside. This mineral is pale green, slightly pleochroic, and has a 2V of 56°–68° and $Z \wedge c$ of 47°. It has properties resembling iron-rich clinopyroxenes in Gardar rocks. Some hypersthene replaces olivine while apatite is a common accessory. The felspars in the gabbroic rocks are commonly clouded with an indeterminate, evenly distributed, fine dust which disappears in felspars which have been subject to cataclasis.

The basic members of the oligoclase gabbro – pulaskite series contain less calcic plagioclase ($An_{23\text{-}25}$), antiperthite, and iron-rich olivine (Fa_{80}).

Two leucocratic rock types in the gabbroic series are of interest for comparison with the xenoliths in Gardar rocks. Firstly the gabbroic anorthosites and troctolites described by POLKANOV and ELISEEV appear remarkably similar to the laminated anorthosite block from Assorutit. The felspars are platey and show a good lamination parallel to the layering. The mafic minerals are found in the interstices between the felspar plates where they partially enclose the subhedral felspars. Secondly the coarse-grained anorthosites figured by POLKANOV and ELISEEV in plate 2, no. 9, show textures which are identical with the granular anorthosites described in this paper (Fig. 44, p. 69). When fresh the granular anorthosite in the Gremyakha-Vyrmes pluton contains practically no other mineral than plagioclase (An_{40}). It is found as concordant layers, lenses and irregular masses in the layered gabbroic sequence. Its occurrence as a primary igneous rock in a gabbroic series which also contains laminated anorthosites is of considerable interest and supports the idea that the granular anorthosite xenoliths in Gardar rocks are a primary plagioclase accumulation from similar magmas to those which gave rise to the laminated rocks. The association of both leucocratic types with ultrabasic layers is also of interest since it may suggest that such rocks were formed from Gardar gabbroic magmas at depth but have never been brought up as inclusions.

In their conclusions POLKANOV and ELISEEV suggest that crystal fractionation was the main process controlling the formation of the main rock types found in the Gremyakha-Vyrmes pluton. They conclude that the layered structures were undoubtedly formed by gravity settling. However, while POLKANOV and ELISEEV accept crystal fractionation as the main process controlling differentiation they find considerable difficulty in accepting it as the sole mechanism and they state (§ 403, p. 231) "this forced us to admit a separation not only of the solid phases from the liquid but also migration of the liquids heavy as well as light". As has already been mentioned on p. 147 liquid fractionation is thought to occur in some Gardar dykes. The application of this mechanism to differentiation on a much larger scale is briefly mentioned in the concluding remarks of the present paper.

As the Precambrian of Greenland is clearly an extension of the Canadian Shield the most satisfactory anorthosites with which to compare the Greenland rocks are probably those which occur in the belt of country between the Adirondacks and the Labrador coast. However, as noted earlier, although the **Canadian anorthosites** have been worked on for over a century there is still considerable doubt about their origin.

ANDERSON and MORIN (1964) divide the anorthosites into three groups which contain the following mineral assemblages: 1) Lac St. Jean; labradorite, clinopyroxene, olivine, ilmeno-magnetite and ilmenite. 2) St. Urbain; sodic andesine, antiperthite, clinopyroxene, orthopyroxene, apatite, haemo-ilmenite, ilmeno-magnetite, biotite, and zircon. 3) Unnamed; calcic labradorite – sodic andesine, antiperthite, clinopyroxene, orthopyroxene, olivine, ilmenite, ilmeno-magnetite, and allanite. Unfortunately these assemblages are not diagnostic enough to classify the granular xenoliths from South Greenland which resemble the first group in the lack of antiperthite but which contain some orthopyroxene and probably more olivine in proportion to the other mafic minerals than any of the groups.

ANDERSON and MORIN (1964) suggest that one group of the Grenville anorthosites, represented by the Lac St. Jean body, may represent "local primordial crust". This is of considerable interest for any discussion on the ultimate origin of plagioclase-rich xenoliths in igneous rocks. However, it transfers the setting of the problem of the genetic relationship between host and inclusion to regions not likely to be subject to direct observation either in South Greenland or in any of the bodies with characters resembling the xenoliths.

The two areas in the Canadian Shield which contain rocks with the closest similarites to the anorthosite xenoliths and their Gardar hosts are those of **Michikamau** (EMSLIE, 1964, 1965) and **Nain** (WHEELER, 1960, MORSE, 1962, 1963, 1964). In both areas, which lie to the north of the Greenville fold belt, the anorthosites were emplaced at about 1400–1500 m.y., that is later than the main Hudsonian—Ketilidian plutonic activity of the Canadian—Greenlandic Shield but earlier than the date now suggested for the first Gardar igneous activity (approximately 1250–1300 m.y.). Both areas are within the belt of country ascribed to the Elsonian "orogeny" by the Canadian Survey. It appears from the general descriptions of the rocks dated that this "orogeny" is probably a local late plutonic phase of the Hudsonian with the intrusion of large post-tectonic granites and basic rocks, rather than a full-scale independant fold belt. The anorthosites found in the Elsonian confirm this impression; the Nain anorthosite itself possesses many of the characters found in the typical plutonic anorthosites, though the closely related Kiglapait layered intrusion is a characteristic cratogenic body. The Michikamau intrusion shows features which belong to both groups; in the main the characters are those of a cratogenic intrusion, however, many of the felspars are chatoyant, a feature generally ascribed to formation under plutonic conditions. It seems probable that the period of emplacement of these rocks coincided with the change from plutonic to cratogenic conditions in that part of the Canadian Shield.

The **Michikamau** intrusion covers an area of approximately 800 square miles (2250 km^2), that is about the size of one of the larger Gardar intrusions. Anorthosites themselves account for approximately 25 $^0/_0$ of the intrusion at its present exposure level. The intrusion consists of a gabbroic border zone with chilled contacts surrounding a troctolitic and anorthositic core. A suite of ferrogranodioritic rocks of small areal extent form the last phase of the intrusion. The rocks grade into one another and appear to be comagmatic. The early rocks as shown by analyses 1–4 (Emslie, table II, p. 394) are close to the least fractionated Gardar gabbros in composition. They show neither strong alkali nor strong tholeiitic tendencies. They are moderately high in alumina. The anorthositic rocks are remarkably similar in composition and textures to the Assorutit block (analyses 1 and 2, Table 1, p. 72). In spite of these similarities the differentiation trends shown by the late rocks in the Michikamau intrusion are completely different to those seen in the Gardar intrusions and show a closer resemblance to the final trends seen in the Skaergaard intrusion (Wager and Deer, 1939).

One of the most interesting features recorded by Emslie is the fact that the contact rocks are less basic than the slightly younger troctolitic group. The transition is apparently gradational and can be seen in the composition of the constituent minerals which vary from an average of An_{51-56} and Fa_{42-44} in the border facies, to An_{54-72} and Fa_{20-40} in the troctolites. Emslie ascribes at least part of this change to thermal gradients. However, as the transition takes place over 300–1000 m it would seem that a more plausable explanation might be to suggest that the magma itself was partially differentiated at the time of emplacement. In the absence of evidence suggesting the emplacement of a crystal-mush, at least during the early part of the intrusion, it seems reasonable to suggest that liquid fractionation took place in the magma before intrusion in the same way as suggested for the microsyenite-trachydolerite dykes described in the present paper. If this suggestion is correct, and appreciable differentiation can take place before intrusion in magma chambers large enough to form an intrusion of this size, then it is possible that similar arguments may also be applied in the Gardar igneous province, a factor which has considerable bearing on the origin of both the anorthosites and the syenites.

Emslie produces good textural evidence to show that the dominant mechanism of felspar accumulation was one of crystal settling which produced layered and laminated rocks similar to the Assorutit rocks. It is interesting to contrast this to the majority of examples seen in the Gardar where plagioclases of similar composition to those described by Emslie have risen, apparently in magmas closely resembling those once present in the Michikamau intrusion. Emslie notes that if the plagioclases accumulated by settling then their concentration in the youngest major rock type found in the intrusion is anomalous; however this difficulty is removed if the differentiation is thought to have taken place at least partly before the injection of the magma. The last major phase of intrusion would represent the lowest magma in the hypothetical chamber feeding the intrusion. This magma could be expected to be highly charged with plagioclases which had settled from higher levels.

In many respects the sequence of events seen in the Michikamau intrusion resembles that described from dyke 8 (p. 128): an early, relatively differentiated border phase, a basic second phase and a third phase of more differentiated rock probably produced more or less *in situ* by crystal fractionation. In both cases "ideal" differentiation trends are difficult to draw because differentiation is controlled by two partly opposing mechanism: liquid and crystal fractionation. In the Michikamau intrusion a third factor, crystal accumulation, also played an important role.

Upton (1964) has already called attention to the similarity between the Gardar rocks and the igneous bodies found in the **Nain region** of Canada (Wheeler, 1960; Morse, 1962, 1963, 1964). The anorthosites of this region, consist of two main units: an older massive anorthosite of plutonic aspect **(The Nain anorthosite)**, intruded by a younger complex of layered rocks varying from troctolites to larvikites **(the Kiglapait intrusion)**. According to Morse the Nain anorthosite was still mobile in some areas when the younger intrusion was emplaced although there are abundant inclusions of the massive anorthosite in the layered sequence. From the descriptions given it seems likely that the Nain anorthosite resembles the majority of the plutonic anorthosites of the Canadian Shield. It is mainly comprised of a chatoyant, intermediate to sodic labradorite and shows some layering especially near to the contacts. The country rocks are medium- to high-grade gneisses, the formation of which may in part be connected with the emplacement of the anorthosite. Younger adamellites are associated with the anorthosite and there appears locally to be some gradation between the rock types. The dark facies described by Wheeler (p. 1757 and analysis 1, p. 1758) resembles the granular anorthosite xenoliths closely. It contains 82.7 $^0/_0$ (by volume) plagioclase (An_{59}), 11.5 $^0/_0$ olivine (Fa_{31}), 0.8 $^0/_0$ orthopyroxene and

0.6 % clinopyroxene. The anorthosites are typically coarse to extremely coarse in grain size with protoclastic textures.

The basal rocks of the layered **Kiglapait intrusion** bear a striking resemblance to the Assorutit layered xenolith. Plagioclases range from An_{62} to An_{54} while the olivines range from Fa_{28} to Fa_{36} which is remarkably close to the Assorutit rocks.

The most important feature for comparison with the Greenland xenoliths found in both the Angola and Nain complexes is the presence of an anorthositic basement resembling the major plutonic anorthosites which are either overlain or intruded by a layered series. The genetic relationships between the two types has not been worked out in either of the complexes; however, it appears to be more than a chance association and may ultimately provide the most satisfactory evidence to explain the association of granular and layered xenoliths in the Gardar rocks.

Comparison between the Greenland xenoliths and **anorthosites in major layered igneous intrusions** such as the Stillwater, Bushveld and Freetown complexes are less satisfactory since although anorthosite layers in these bodies may be several hundred of metres thick they still only form a small percentage of the total. It is of course possible that the granular anorthosites are themselves underlain by considerable ultrabasic or basic material but as little has been brought up as inclusions in Gardar dykes they cannot be used for comparison. Most of the anorthosite layers found in the layered gabbroic complexes are considerably more calcic than the xenoliths, with the exception of those described from Freetown (Wells, 1964). The structure of the Freetown body however, does not resemble anything seen in the Gardar.

e) The origin of the megacrysts

Many of the small fragments found in Gardar dykes are obviously derived from larger masses of granular felspathic material by a combination of mechanical and chemical breakdown. However, a significant number of single plagioclase crystals show distinctive characters. These characters are transitional between those of felspars seen in the massive anorthosites and felspars from the Gardar hosts. The megacrysts are found closely associated with fragments of granular anorthosite and it seems most unlikely that this relationship is fortuitous. The most reasonable explanation is that the two are genetically connected but have passed through different histories.

Many dykes, particularly the more alkaline gabbros, contain a preponderance of large megacrysts which may be grouped together to form loosely compacted masses of secondary anorthosite cemented with material derived from the host. The interstitial material may either be gabbroic or single "ophitic" crystals of augite, olivine or Fe-Ti oxides. In many instances this mafic material forms lamellae within the large felspars controlled by the plagioclase crystallography.

The megacrysts show considerable variation in their physical and chemical properties which appears in part to be related to the composition of the surrounding host. For example, the most basic gabbros contain a high proportion of xenoliths and cleavage fragments directly derived from xenoliths together with a few medium sized megacrysts with a composition close to the granular anorthosite and showing twinning similar to that seen in the host's plagioclase. The proportion of megacrysts to xenoliths increases in the more alkali gabbros and some reach exceptional sizes. They are generally less calcic than the granular anorthosites, with compositions between An_{50} and An_{58}. They show simple twinning and may clump together in rafts. The megacrysts of black felspar show a narrow compositional range and are only stable in hosts with a high water content. The megacrysts from the microsyenite-trachydolerite hosts are smaller, frequently sericitised, and show complex twinning. They generally have an anorthite content between An_{40} and An_{50}.

The variation in chemical properties seen in the plagioclase megacrysts from hosts with different compositions suggests that host and inclusion may be closely connected genetically and the crystals should be termed phenocrysts. This is especially true in the microsyenite-trachydolerite group of host dykes, the groundmass of some of which appear to contain lower amounts of calcium than might normally be expected. However, the term megacryst is preferred in this paper as a general term since some at least of the crystals may be shown to be derived and nearly all were out of equilibrium with the magma surrounding them at the time of solidification of their hosts. In a few places, for example the giant crystals from the marginal gabbro at Klokken, the megacrysts show so close a resemblance to the felspars in their host that they can be regarded as true phenocrysts. However, it would seem unnecessarily pedantic to use the term phenocryst for the plagioclases from Klokken and the term megacrysts for the identical plagioclases from Narssârssuk where the host has been altered by late syenitisation.

The genetic relationship between the giant megacrysts and the granular anorthosites is one of the most important subjects for discussion since in many respects the megacrysts show transitional characters between the xenoliths and the plagioclases in the Gardar hosts. Occasional large crystals are found within the granular anorthosites and these have previously been taken as evidence that the megacrysts were derived by breakdown of massive rock. However, the fact that the megacrysts are coarser grained than the associated granular rocks suggests that few originated in this way. Although there appears to have been some overlap, the majority of megacrysts are thought to have formed later than the granular rocks. The best direct evidence for this suggestion is seen in dyke 13 (Fig. 11 p. 28) where large megacrysts may be seen growing out from a granular block and in dyke 12 (Fig. 10 p. 28) where an euhedral rim surrounds a granular mass at the centre of a megacryst. Their less calcic composition makes it extremely unlikely that the megacrysts in general represent an early uncompacted phase of the granular anorthosite. The majority of large megacrysts show no evidence of forming part of an earlier massive rock, at least in their present form, and they appear to have been free-floating crystals until the solidification of the surrounding host rock. Their chemical uniformity, general sparsity of twinning and giant size suggest that they must have formed under remarkably stable physical and chemical conditions which extended for a considerable distance at depth under South Greenland. These conditions were modified when the felspars were carried up in their hosts resulting in fierce marginal zoning and in some cases loss of the black colouration. The cause of the structural abnormalities found in the megacrysts is unknown but it appears more likely to be an original feature than one caused during the complete recrystallisation of earlier material.

The relationships seen between the xenoliths and megacrysts can be explained by suggesting either that the large glassy crystals represent younger anorthositic material which never consolidated to form a massive rock or that they were formed by the recrystallisation of xenoliths of granular anorthosite suspended in Gardar magma at depth. The majority of megacrysts show no features which might suggest they formed by recrystallisation of earlier material; they are clear, homogeneous, simple twinned crystals which resemble phenocrysts formed under remarkably stable conditions. If recrystallisation has taken place then it has been so thorough that all traces of an earlier history have been obliterated. The presence of occasional large crystals of plagioclase in the granular anorthosites, apparently as an original magmatic feature, (Fig. 4 p. 23) suggests that at least some of the megacrysts are primary and that a few were formed at the same time as the granular rocks. The majority of observations, however, suggest that the megacrysts continued formation after the granular rocks; some contain inclusions of granular rock surrounded by an overgrowth of euhedral felspar. When "rimmed aggregates" of this kind are found in gabbroic rocks there is little difference in the composition of the granular centre and the mar-

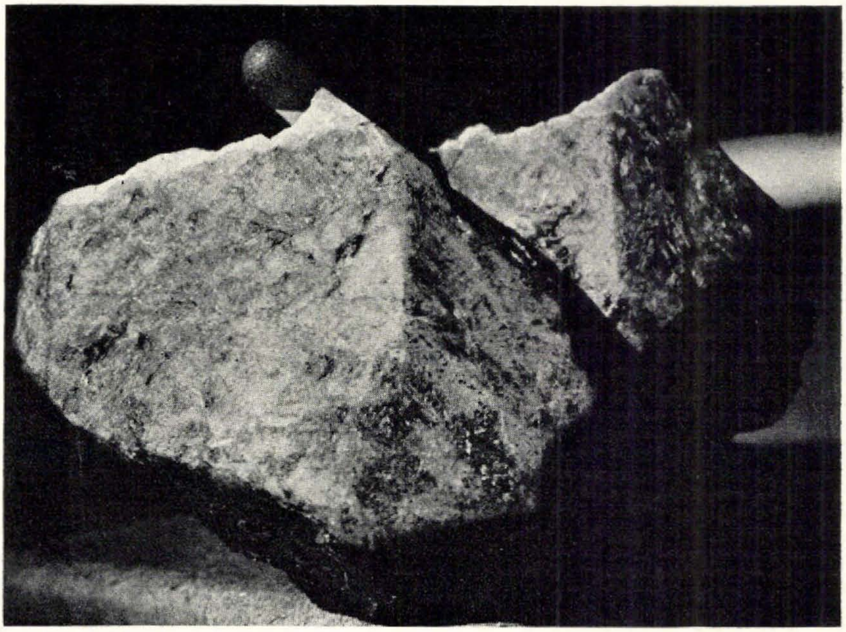


Fig. 69. Laboratory photograph of a fragment of felspar from the centre of a large megacryst × 3 natural size, G.G.U. 38137. Note the euhedral faces.

ginal plagioclase. However, in the more alkali gabbroic hosts there may be a significant fall in the anorthite content from centre to margin. In several instances it is difficult to decide whether particular inclusions are aggregates made up of several grains with a common orientation or whether they are single crystals. In one example the centre of a 1.5 m felspathic inclusion was seen to consist of closely packed plagioclase crystals with a common orientation which had intergrown with each other in a complex dove-tailed pattern. The (010) crystal planes were orientated in the same direction throughout the centre, individual crystals being separated by thin films of chloritic material. When broken the individual crystals were found to be euhedral (Fig. 69). Towards the margins the mass became much more homogeneous and it was impossible to distinguish individual crystals. The (010) plane remained in a constant position. In this example it is very difficult to decide whether the inclusion should be called a xenolith or a megacryst; individual grains can be seen in some places but the texture differs very little from the felspar intergrowth described on p. 96 where plagioclase grains separated by chloritic and gabbroic material are found to be twin pairs. Some of these complex textures, particularly the rimmed aggregates, appear to have formed when early blocks of granular anorthosite were suspended in a magma from which plagioclase megacrysts were crystallising. Others, for example the interlocking twin pairs, are probably an original feature.

It appears from the textural and compositional evidence that the granular anorthosites and the megacrysts form a series. The oldest and most calcic are the massive granular anorthosites which are generally found in basic hosts. Next there are the granular anorthosites with scattered plagioclase megacrysts set in a finer grained matrix. These are followed by "rimmed aggregates" and overgrowths of megacrysts on cores of granular anorthosite. These are less calcic and are generally found in alkali gabbro hosts. Finally, there are the range of simple megacrysts which appear to have crystallised direct from magmas related to their present hosts. The significance of this series is discussed in the final section.

f) Comparison between the inclusions in the Gardar and felspathic inclusions in other igneous provinces

Plagioclase megacrysts, glomeroporphyritic aggregates and, more rarely, anorthosite xenoliths are reported from a large number of igneous provinces, ranging from the modern Hawaiian and Icelandic basaltic lavas to Precambrian dykes, lavas and major intrusions of several contrasting igneous suites. Porphyritic minor intrusions are one of the commonest rock types found in many igneous provinces and while in many cases the phenocrysts are clearly related to the host, in others the insets should properly be called megacrysts.

There are two reasons for comparing members of this rather heterogeneous group of rocks with the Gardar felspathic intrusions. Firstly felspathic inclusions in igneous provinces which are structurally and petrologically similar to the Gardar province may show genetic links between the host and inclusions which are not apparent in the Gardar rocks. Secondly examples from the more diverse group without close petrological affinities to the Gardar may illustrate particular points in common with the inclusions in Gardar rocks and thus give some ideas about the mechanics of formation of the felspathic insets.

The most suitable felspathic rocks for comparison with the Gardar are those from other alkali basaltic suites, for example those of Oslo and possibly the rocks in the Carboniferous igneous suite of the Scottish Midland Valley (Richey and Thomas, 1930; Tomkeieff, 1937). The Oslo province does not show the same concentration of anorthositic material as that found in Gardar rocks and the few xenoliths found (Oftedahl, 1953) are so problematical that they can be of little help in understanding the Greenland rocks. Anorthositic inclusions have not been described in the Scottish rocks although local concentrations of "phenocrysts" may pack together towards the upper margins of some trachydolerite sills. These resemble the felspathic aggregates found in host rocks of comparable composition in the Gardar but have never been described in sufficient detail to allow more than a superficial comparison. The main interest of these occurrences in the present context is to note the association of felspathic inclusions with provinces showing trends from gabbro to syenite. This may support the idea that there is a genetic link between insets and hosts in Gardar rocks.

Anorthosite xenoliths and plagioclase megacrysts are found abundantly in the Duluth gabbro and associated rocks and in the Precambrian intrusions of central Sweden. Both provinces were formed at approximately the same time as the Gardar (c. 1100 m.y.). However, petrographically they are not particularly close to the South Greenland rocks as they are both less alkaline, trending from calcic gabbroic rocks to granophyres. In both areas, however, alkali rocks occur with some resemblances to the Gardar, for example the syeno-gabbro dyke of central Sweden (ECKERMANN, 1936) and the Coldwell gabbro-larvikite-nepheline syenite complex on the northern shores of Lake Superior (FAIRBAIN, BULLWINKEL, PINSON and HURLEY, 1959; and H. V. TOUMINEN, personal communication).

The **Eskilstuna dykes** from central Sweden (GORBATSCHEV, 1961) vary from olivine-augite-hypersthene dolerites with calcic plagioclase and moderately iron-rich olivines to granophyric quartz dolerites. Plagioclase insets are found in both olivine and non-olivine bearing host rocks. They are often confined to one part of the dyke similar to the distribution seen in the Gardar dykes. GORBATSCHEV distinguishes three types of inset in the dykes, anorthosite fragments, plagioclase phenocrysts and glomeroporphyritic plagioclase accumulations. Although the terminology used differs slightly with the present paper the features described by GORBATSCHEV correspond closely to many noted in the Greenland felspathic dykes and several of his ideas may apply equally well to both provinces. Texturally the anorthosite fragments illustrated by GORBATSCHEV 1961, (fig. 18) are indistinguishable from the aggregates of crushed plagioclase described from dyke 8, p. 46, while the "dolerite porphyrite" (fig. 14) and the "glomeroporphyritic rock" (fig. 16) could be matched with many of the megacryst-bearing Gardar dykes. Many of the plagioclases from the Eskilstuna insets show inclusions of olivine and pyroxene parallel with the main twin directions, similar to the mafic lamellae in Gardar inclusions.

The first plagioclase to crystallise from the olivine dolerites in the Eskilstuna dykes was approximately An_{82} while in the majority of "dolerite porphyrites" the groundmass plagioclase averages An_{55-60}. The xenoliths range from An_{69-73} and the megacrysts between An_{62-75}

disregarding the sharply zoned outer zones which are close to the felspar seen in the host rock. Although the composition range differs the compositional relationship between host and insets is the same as in the Gardar.

Gorbatschev regards the anorthosites as comagmatic with their hosts, representing a top accumulation of early plagioclase by flotation in deep-seated magma chambers under tranquil conditions. These were then broken up when the residual magma was injected into dykes. The origin of the "phenocrysts" is less explicity stated although Gorbatschev implies that they may represent the unconsolidated stages in the formation of the anorthosites. He notes that in some "phenocryst"-bearing dykes there is a marked impoverishment in plagioclase in the groundmass compared with normal dolerites suggesting a direct genetic link between host and inset. This feature has never been observed in individual Gardar dykes although it is quite possible that the apparent low calcium contents found in all the analyses of megacryst-bearing dykes may be due to similar causes.

A second group of anorthosites and associated rocks in the Swedish Precambrian has been described by Bergström (1963). The anorthosites occur as large inclusions in norites intruded during Dalslandian (≡ Grenville) plutonic episode at approximately 1000 m.y. The basic suite of rocks show calc-alkaline affinities and are characteristic late orogenic intrusions. Petrologically they show little resemblance to the Gardar rocks, the plagioclases are less calcic and the dominant mafic mineral is hypersthene. According to Bergström the anorthosites were formed by upward accumulation of plagioclase so that they formed a solid roof over the developing noritic magma chamber. Local movements broke up the roof and crushed the anorthosites so that they developed a slightly strained texture. The broken blocks of anorthosite were included in the norite which was released from the chamber during the movements. Large phenocrysts of plagioclase are found within the norite, these may aggregate together to form anorthositic clumps and all stages are seen between single phenocrysts and the formation of solid anorthosites. The phenocrysts show approximately the same composition as the felspar in the anorthosite and the felspars in the noritic rocks. In spite of the difference in geological setting and chemistry the rocks described by Bergström show several features which are remarkably similar to those in the Gardar; for example the gradation between solid anorthosite and free-floating megacrysts and the upward accumulation of anorthosite blocks and felspar crystals in a basic magma.

The Duluth Gabbro and the associated lavas and minor intrusions on the north-west shore of Lake Superior are among the most widely cited basic suites of rocks from the Precambrian. Anorthosite xenoliths and plagioclase megacrysts occur in the main gabbro at Duluth and more sporadically in both the lavas and some of the minor intrusions along the coast of the lake to the north-east of Duluth. The large plagioclase megacrysts have been used as chemical and crystallographic standards due to the purity of the felspar and the extremely coarse grain size.

The Keweenawan igneous rocks, to which the Duluth Gabbro belongs, are generally less alkaline than the Gardar and were emplaced during a period of tectonic activity (TAYLOR, 1956). The earliest rocks are a series of basic and rhyolitic flows forming a series up to 21,000 ft (6500 m) thick which form the main country rock into which the later intrusions were emplaced. The largest body, the Duluth Gabbro itself, is a lopolith up to 20,000 ft thick consisting of two main units, an older anorthositic gabbro which forms the roof of the body (TAYLOR, 1964) and a younger layered intrusion ranging from troctolite to syeno-gabbro. There is a late series of ferrogranodiorites and granophyres which are thought to be genetically connected with the layered series.

The anorthositic gabbro is texturally almost identical with the secondary anorthosites described from Bangs Havn and Narssaq. It consists of a gabbroic host with inclusions of a large number of different textural types of anorthosite ranging from crudely layered blocks with some mafic minerals to xenoliths of pure felspar with fluxion laminated felspar or massive anorthosite made up of brecciated plagioclase fragments. Several of the blocks described are surrounded by rims of augite in coarsely crystalline masses believed, by TAYLOR, to be the result of additions from the surrounding magma. The main difference between the rocks illustrated by TAYLOR and those in the Gardar is that there seems to be a higher proportion of partly consolidated anorthositic material present in the Duluth Gabbro. It is therefore important to note that TAYLOR (1964, p. 11–12) states that "True anorthosite with more than 95 % plagioclase occurs only as discrete inclusions". This suggests a similar process to that put forward at Narssaq and Bangs Havn where rigid blocks of primary anorthosite were included in early gabbro, itself found as partially consolidated inclusions in the xenolith-free late stages of the complexes. At both Duluth, and at Narssaq and Bangs Havn, the anorthosite fragments have accumulated close to the roof of the bodies suggesting that they rose in gabbroic magma.

The anorthosites were originally regarded as an early phase of the Duluth Gabbro (see for example GROUT and SCHWARTZ, 1939) with a similar relationship between the main body and the inclusions as that proposed by UPTON (1964) between the Tugtutôq gabbros and the Assorutit xenoliths. However, more recent papers are more cautious. TAYLOR (1965) states that although most of the inclusions are cognate the origin of the rounded anorthosite xenoliths is unknown. One of the main lines of evidence suggesting that anorthosite may have formed an early mass at depth is the presence of occasional xenoliths in the lavas (GROUT, SHARP and SCHWARTZ, 1959) and in the pre-Gabbro Logan sill. It seems probable that the structure of the Duluth area at depth may resemble the Angola and Nain bodies with an early granular anorthosite overlain by a later complex of dominantly layered rocks.

Apart from the main Duluth Gabbro anorthosites occur in many of the smaller Keweenawan intrusions of the Lake Superior region. The largest inclusions described by GROUT and SCHWARTZ (1939) reach 0,25 miles (0.4 km) across, other fragments range down to single megacrysts little larger than the felspars in the host rocks. They are sometimes found in association with inclusions of gabbro. Texturally all the types described from the Gardar occur, varying from granular granitiod textures and crude layering seen in the large blocks to single, glassy crystals and glomeroporphyritic accumulations of megacrysts. GROUT and SCHWARTZ note that the crystals in the larger masses are commonly coarser grained than those seen in the anorthositic facies of the Duluth Gabbro and suggest that this may be due to recrystallisation of the anorthosite. Now that it is thought that the anorthosites at Duluth are secondary it would seem more likely that it reflects different stages in brecciation. However, it is also apparent that the single felspar megacrysts found at such localities as Beaver Bay and Grand Marais, which have become famous because of their size and purity, are larger than the majority of crystals found in the anorthositic masses. This is similar to the relationship seen in the Gardar and it is possible that the Keweenawan megacrysts may represent slightly younger unconsolidated anorthositic material which remained suspended in a gabbroic magma for a considerable time. Results of X-ray powder determinations (SMITH and YODER, 1956) and optical-crystallographic scatter measurements (VOGEL, 1964) on some of these large glassy megacrysts show similar anomalous properties to those of the Gardar megacrysts.

Mineralogically both the primary and secondary anorthosites from Minnesota are fairly close to their Gardar equivalents although the plagioclases are on average approximately 5 to 10 $^0/_0$ higher in anorthite. The primary anorthosites consist (in order of abundance) of labradorite, augite, olivine, magnetite and rarely orthopyroxene. Some of the plagioclase crystals are dark. GROUT and SCHWARTZ note that the augite is late crystallising and it seems possible that some may be formed by addition from the host similar to the Greenland secondary anorthosites. No information is available on the composition of the plagioclase from different textural types of anorthosite and they cannot be directly compared with the xenoliths and megacrysts in Gardar rocks. The inclusions are more calcic than the felspar in their hosts but they are less calcic on average than the first felspars to crystallise from the more basic members of the province. The composition of the Duluth anorthositic gabbro given by Taylor (1964, table 2) is remarkably close both in bulk composition and in mineralogy to the secondary anorthosites from Narssaq and Bangs Havn though the general chemistry of the rocks found in Minnesota is not particularly close to the Gardar. Of the gabbroic rocks analysed the olivine gabbro from Bardon Peak (Taylor, 1964, table 8, p. 29) is perhaps the closest to a normal Gardar gabbro. The differentiation trends shown by Taylor differ significantly from Gardar rocks, resembling those described by Emslie (1965).

The suggestion by Powers (1915) that the comparative lack of other xenoliths in the Minnesota rocks is due to the peculiar physical properties of anorthosite has already been mentioned (p. 153). It seems more likely that the association basic intrusion—anorthosite xenoliths—plagioclase megacrysts reflects a more fundamental genetic relationship.

The most plausible explanation would seem to be that the original anorthosite formed a roof to the zone in which the host magmas and megacrysts were developed. This could either be in the mantle itself or at some intermediate level below which the magmas differentiated.

IX. CONCLUSIONS

The form and origin of the parental anorthosite mass, and the relationship between the anorthosites, the plagioclase megacrysts and the Gardar igneous rocks

The South Greenland anorthosites appear only as fragments which have often been modified by their Gardar hosts. Discussion of their origin can therefore be little more than speculation or argument by analogy, but it seems justified since it raises issues of magnitude. Any final statement about the relationship of the anorthosites and the Gardar igneous rocks must wait until considerably more is known about the province as a whole and particularly the chemistry and evolution of the parent magma (or parent magmas). This, however, presents a paradox because it is unwise to discuss the evolution of the Gardar igneous rocks without knowing the origin of the anorthosites which probably form one of the major rocks types at depth.

The present distribution of anorthosite fragments suggests that the parent body is an elongate mass in the order of 250–500 km long, and 50–100 km broad, comparable in form to the Angola anorthosite. It is not known whether there is more than one body present or whether the mass forms a continuous layer under South Greenland. The main rock type is a granular anorthosite with a slightly crushed texture, composed of labradorite (An_{60}) anhedra with a considerable range in grain size but remarkably uniform composition. The main mafic minerals (in order of abundance) are olivine ($Fa_{27\text{-}40}$), Fe-Ti oxides, a little orthopyroxene ($Fs_{28\text{-}40}$) and a little clinopyroxene.

The mafic minerals generally form interstitial patches between the felspars. There is no evidence to suggest the presence of massive ultrabasic or gabbroic counterparts to the granular anorthosite. A few fragments of olivine- or Fe-Ti oxide-rich xenoliths with a strong foliation have been found which might suggest the local presence of a dark sheared border facies corresponding to the marginal rocks of some of the plutonic anorthosites of the Canadian Shield. The significance of the rare basic inclusions of bytownite, hypersthene and olivine found in a Gardar olivine gabbro from Tôrnârssuk is uncertain. It seems possible that they represent a lower level in the hypothetical parent body. There are no signs of strong cryptic variation in the large anorthosite fragments – the felspars show a uniform composition – and while there is some variation in the olivine composition it is probable that the iron-rich members are derived from later host rocks. Some slight variation in texture is found with local changes in the form of the felspars and the total content of mafic minerals which may lead to a crude layering. The most common variant of the normal granular rock is a poorly laminated gabbro anorthosite with subhedral plagioclase crystals surrounded by olivine or Fe-Ti oxides.

The granular anorthosite is locally associated with a group of layered anorthositic gabbros (the Assorutit anorthosite) which show a superficial mineralogical similarity to the granular rocks. However, the differences in texture suggest that the two rocks were formed by different mechanisms; the layered anorthosites appear to have resulted from a felspar cumulate at the bottom of a magma chamber (Upton, 1961) while the granular rocks appear to have been the result of the upward flotation of plagioclase. The first-formed felspar in the layered anorthosites is more basic than the felspars found in the granular rocks and the dominant Fe-Ti oxide is ilmenite, not magnetite. The genetic relationship between the two is uncertain. The layered anorthosites contain numerous inclusions of granular rock and do not appear to have passed through so complex a history. Layered anorthosite xenoliths show a much more restricted distribution than the granular rocks and are practically confined within a 10 km radius of Narssaq, though there are possibly a few examples close to the Klokken and Igaliko intrusions. By analogy with the bodies described from Nain and Angola it seems likely that the Assorutit type of xenoliths represents a layered gabbroic body overlying a granular massive anorthosite. Upton (1961 and 1964) has shown that there is in all likelihood a close genetic link between the layered anorthosites and the associated Gardar gabbros and syenites and it seems significant that there is a close spacial association between the layered anorthosite xenoliths and the mid-Gardar syenites of Narssaq and the east of Tugtutôq. If the comparison with the Kiglapait intrusion is valid then the layered anorthosites probably represent the basic part of a syeno-gabbro layered intrusion. The large felspars found in the margin of Klokken and Igaliko, which show several chemical resemblances to the Assorutit rocks, could then be regarded as true phenocrysts.

The relationship between the granular anorthosites and the remaining Gardar igneous activity is much more difficult to define. The following points seem pertinent:

1) The anorthosite xenoliths are found within the same fault-controlled zone as the major and minor alkali intrusions. Two particularly marked concentrations of anorthosite-bearing dykes occur north-east of Nunarssuit and in the area between Tugtutôq and Igaliko. Both areas show considerable alkali magmatic activity.

2) Xenoliths of granular anorthosite are practically confined to mid-Gardar dykes. Megacrysts and glomeroporphyritic aggregates of plagioclase are found in pre-Gardar and early Gardar dolerites but may not necessarily belong to the same story.

3) The Gardar rocks exposed at the present erosion level show a marked preponderance of syenites and alkali granites compared to gabbro (see WATT, 1966), recalling the Oslo province.

4) There appears to be little or no direct genetic connection between the host dykes and the xenoliths. The large plagioclase megacrysts associated with the anorthosites are, however, particularly abundant in the more alkali gabbros and show a range of properties which are in part dependent on the composition of their hosts. A few of the megacrysts appear to have formed by recrystallisation of detached blocks of granular anorthosite; the majority are thought never to have formed solid rock prior to inclusion in their present hosts and appear to have crystallised slowly. The majority of megacrysts are less calcic than the granular rock and are thus unlikely to represent early stages in its formation. They might however represent later partly unconsolidated material. The formation of at least some of the megacrysts appears to have taken place in water-rich surroundings.

5) The majority of fragments show textures which suggest accumulation by flotation rather than accumulation at the bottom of magma chambers. The xenoliths appear to have been less dense than Gardar gabbro magmas but were probably denser than the syenites.

6) The xenoliths form a variety of secondary anorthosites with Gardar hosts which solidified at depth but which were themselves broken up by later pulses of basic magma.

7) The mineralogy of the xenoliths is moderately close to that of the least fractionated gabbros exposed at the present surface (the Eqaloqarfia dyke, analysis 33, Table 4; the giant gabbro of Tugtutôq, analysis 1, Table 4). The olivine in the anorthosites is more magnesium-rich than those in the gabbros, while the felspar is more sodic than the first formed plagioclase in the gabbros (An_{72}) but more calcic than the bulk composition (approximately An_{50}) of the felspar in the majority of these rocks. There is a marked difference between the amounts of strontium and barium found in the granular anorthosites and the less

alkali Gardar rocks compared with the amounts found in the layered anorthosites and gabbros associated with syenites.

8) Inclusions of granular anorthosite, layered anorthosite and megacrysts are found mixed together in the same dykes. There are very few dykes containing mixtures of anorthosite with non-felspathic inclusions.

9) Rare inclusions of gabbroic material with a much more basic plagioclase than that normally found in Gardar gabbros have been found. There seems no doubt that these inclusions are cognate.

10) Liquid fractionation has been shown to be a feasible process in Gardar intrusions.

Associations of syenite with anorthosite are well known from other provinces and are often taken as genetic. Two theories have been used to explain this relationship; the first suggests that the syenites are the result of regional anatexis round large intrusive anorthosites emplaced under plutonic conditions, the second more orthodox viewpoint suggests that the two rock types are derived by magmatic differentiation from the same primary magma.

The Canadian Shield has provided most of the evidence for the large scale anatexis of older gneisses surrounding plutonic anorthosites. Syenites and related rocks have long been recognised as late intrusive rocks surrounding the Adirondacks although their genetic relationship to the older anorthosites is controversial (Buddington, 1939). Similar associations have been noted surrounding the Saqueney and Allard Lake anorthosites and both Berrangé (1962) and Hargraves (1962) have suggested that there has been large-scale mobilisation of the country rocks giving rise to a charnockite-magnerite-enderbite series. The ultimate origin of the anorthosite itself is unimportant in explaining this association; it can either be derived from a gabbro-anorthositic magma as suggested by Buddington, it may have formed by some ultrametamorphic process as put forward by Michot (1955), or it may represent a plagioclase-rich differentiate from a gabbroic or dioritic magma. However it was formed, it provided a large enough heat source to mobilise the surrounding rocks. The dominance of the syenites seen in the Gardar and Oslo provinces could certainly be explained if it is assumed that a large part of the felsic material was derived in some way from the crust (see, for example, Barth, 1954). Some of the rocks surrounding the plutonic anorthosites in the Canadian Shield resemble the Greenland larvikites in gross chemical composition. However, it is dangerous to take the comparison too far; the term "syenite" has been used in the Canadian Shield to cover a whole group of rocks in the plutonic charnockite-mangerite-enderbite series, few of which resemble the rock types found in the cratogenic Gardar province. The geological setting of the Gardar igneous rocks is completely different to that of the major anorthosites in the Canadian Shield and unless it is believed that the Gardar syenites are the final products of anatexis which began in the plutonic episode 3–500 m.y. before their intrusion there seems little point in pursuing the comparison.

Strontium isotope studies on Greenland intrusions, which may in the end give direct evidence of the origin of the felsic members of the Gardar magmatic province, are limited to preliminary work on the Ivigtut granite (MOORBATH and PAULY, 1962). So far there is no evidence that substantial amounts of crustal material were involved in the formation of the granite, although the isotopic composition of the associated lead mineralisation suggests that it has been reworked from a much older source. It may also be noted that the agpaites from Ilímaussaq contain a concentration of rare earths, zirconium and niobium which HEIER and TAYLOR (1964) suggest may be diagnostic of igneous rocks differentiated from a magma heavily contaminated by crustal material.

The chief reason for suggesting an anatectic origin for the Gardar syenites is the low proportion of basic rocks seen at the present erosion surface. In order to provide the large heat source required to produce the Gardar syenitic magmas by anatexis in the absence of regional metamorphism it is necessary to suggest the presence of a large igneous body at depth. The anorthosite xenoliths might be regarded as evidence of the existance of such a body. However, if a large basic body is present at depth then the predominance of syenite over gabbro seen at the surface is no reflection of the volumetric relations between the two rock types in the crust. If this is the case the main argument in favour of an anatec tic origin of the syenites is no longer valid.

The alternative idea – that the syenites and anorthosites are associated together because they are derived from the same parent magma – has a long history. BOWEN (1917) postulated the formation of anorthositic plutons by gravitative accumulation of plagioclase crystals in basic magmas which yielded granitic or syenitic rest liquids. More recently a number of intrusive complexes containing large gabbroic masses that gave rise to syenitic rocks have been described, and several of these contain anorthositic or gabbro anorthositic layers. Examples are the Kizir pluton (LEBEDEV and BOGATIKOV, 1963) the Gremyakha-Vyrmes pluton (POLKANOV and ELISEEV, 1941) and the Kiglapait intrusion. BOWEN's ideas, or some modification of them, might therefore be applicable to the anorthosites and syenites of South Greenland, and UPTON (1961, 1962, 1964) has explained the Assorutit block as the result of the settling of plagioclase tablets in a basic or dioritic magma developed from an initial aluminous alkali basalt magma. The larvikitic syenites which UPTON suggests played a parental role to many of the other more highly differentiated alkali rocks are thought of as residual fractions produced by extensive crystal fractionation of large bodies of high alumina alkali basalt (UPTON, 1964, p. 45).

A second theory which might have considerable application in the Gardar to explain the preponderance of syenitic material is that given by BAILEY and SCHAIRER (1966). These authors suggest that the disparity in the relative volume of syenite and gabbro could be explainep in terms of partial melting of basaltic material in the mantle or lower crust and not to the concentration of residuals from an original olivine basalt magma. Although this theory may help in explaining the Gardar syenites alone, it makes it more difficult to explain the relationship between syenite and anorthosite. If it is accepted that the anorthosites represent a large basic body at depth then the apparent disparity in the ratio of syenite to basic rocks exposed at the present erosion surface is probably misleading and does not reflect the volumetric relationship throughout the crust. WATT (1966) has shown that the majority of the analyses of single Gardar intrusions fall on smooth curves when plotted on ΣFe-Mg-alk and Na-K-Ca variation diagrams, and while this is no proof that the various rock types are genetically related it at least makes such an idea reasonable. In many individual intrusions, for example the Isortoq syeno-gabbro dykes, the differentiation trends of which are shown in Fig. 63, there is an almost complete gradation between olivine gabbro and quartz syenite. As there is no sign of large scale mixture of two separate magmas or extensive contamination by country rocks it is reasonable to suggest that the rock types found in the dykes are the result of differentiation from a common magma. If orthodox ideas of magmatic differentiation are followed the gabbros would certainly be regarded as more primitive than the alkali rocks and there are thus good grounds to suggest that at least some of the Gardar syenites are formed by the differentiation of an alkali basalt magma.

Although from the broad theoretical point of view it can be argued that the gabbros (and therefore the anorthosites) could be genetically related to the Gardar syenites there are several difficulties to be explained before the idea can be accepted with any confidence. The outstanding anomaly is the composition of the anorthosites themselves. If it is suggested that the Gardar developed from a normal basaltic or alkali basaltic parent liquid then an enormous volume of ultrabasic or at least very mafic material must have been generated at the same time. With the exception of the xenoliths from Tôrnârssuk no evidence for this exists in the blocks brought up in Gardar dykes and their absence can only be explained in one of two ways: by assuming that the primary magma was abnormal in some way, or by suggesting that the basic components of an original stratified body were never brought up in Gardar dykes because of some factor such as density or depth provenance of the hosts. The second idea is convenient since it removes research beyond the reach of direct observation. However, it cannot be dismissed out of hand. Many of the blocks of anorthosite show signs of suspension at depth in Gardar magmas for a considerable time and it would be quite reasonable to suggest that only the lighter blocks would accumulate in the top portion of a magma chamber feeding the dykes. Again, if granular anorthosites are regarded as top accumulates in a differentiating primary Gardar magma, either at extremely deep levels in the crust or in the mantle itself, it could be suggested that the Gardar gabbro hosts represent partly residual liquids formed in the chamber between the top accumulation of plagioclase and a hypothetical basic fraction. In this case the Gardar dykes would not be expected to carry fragments of the chamber floor.

If it is specifically postulated that the Gardar parent magma was a high alumina basalt (Upton, 1961), quantitative difficulties are minimised but not eliminated. The Gardar basic rocks are commonly alkalic. Yoder and Tilley (1962, p. 419) believe that high alumina basalts are members of both the tholeiitic and alkali basalt suites and attribute their high alumina to their plagioclase. Removal of 22 % plagioclase (An_{64}) from high alumina basalt (Warner flow) would yield "normal" basalt (*op. cit.*). There is no proof that the primitive Gardar basalts were particularly aluminous. Upton (personal communication) regards the composite analysis of the Tugtutôq giant dyke (analysis 1, Table 4) as representing a little fractionated Gardar basalt composition and this is certainly rich in alumina (19.02 %). However, the analysis of the Eqaloqarfia dyke (analysis 33, Table 4) is not so markedly rich in alumina (16.46 % Al_2O_3) and this applies even more to many of the analyses of early Gardar gabbros quoted by Watt (1966). It might be expected that dykes derived from magmas from which at least some plagioclase had been extracted to form the anorthosites or the megacrysts might show a low alumina content. This may be true in the case of some of the analyses of host rocks given in Table 4. However, insufficient is known at present about the relationships of the inclusions to individual host dykes, or how the analyses listed in Table 4 compare with non-xenolith bearing dyke rocks at comparable stages of differentiation.

An even more marked deviation from a normal basaltic magma, such as the highly aluminous gabbro-anorthositic magma postulated by Buddington (1939, 1960) to produce the massive anorthosites of the Canadian Shield, seems unlikely in the Gardar. The formation of such a magma is not easy to envisage in cratogenic conditions and in any case would not explain the relationship between the syenites, gabbros, and anorthosites found in South Greenland. The role of water and other volatiles which BUDDINGTON suggested played an important part in the formation of the North American anorthosites is unknown in Greenland. UPTON (1961, p. 20) argues that the plagioclase in the Assorutit block crystallised early from an olivine-plagioclase melt when the water vapour pressure was reduced, thus extending the anorthite field (see YODER and TILLEY, 1962, fig. 28). However in the Equaloqarfia dyke the water content was apparently high and resulted in the formation of considerable hornblende.

Whatever the water content of the magma during the actual consolidation of the anorthosites there is considerable evidence that there was a marked accumulation of water and other volatile material close to the anorthosites before their inclusion in Gardar dykes. This caused the extensive sericitisation of the felspar and oxidation of the Fe-Ti oxides, which are common features of many of the larger xenoliths. It may also account for the brecciation and block structures seen in the larger xenoliths in a comparable manner to that suggested by HIGGS (1954) for the anorthosites of the San Gabriel Mountains, California. The most likely explanation for the accumulation of volatiles is that they were trapped by a roof of solid anorthosite overlying differentiating Gardar magma at depth.

Assuming that the anorthosites and the Gardar province are derived from a common basaltic magma (without being too specific about its chemistry), and further, assuming that Gardar rocks followed established differentiation trends so that high Mg/Fe, Ca/Na + K ratios are indicative of the more primitive rocks, there are several problems which need to be discussed.

First, the relationship between the two textural types of anorthosite and the differentiation sequence which gave rise to the rest of the rock types found in the Gardar. As the anorthosites occur as inclusions in Gardar rocks varying from some of the most primitive Gardar gabbros to highly differentiated syenites it can be assumed that they formed early in the differentiation sequence. The olivines found in the anorthosites (Fa_{27}-Fa_{40}) are among the most magnesian found in the Gardar containing about 5 % more of the forsterite molecule than the olivines found in the Tugtutôq giant dyke gabbros (Fa_{40}) and the normative olivine ($Fa_{35-37.5}$) given by PULVERTAFT (1965, table 1) from the groundmass of the Eqaloqarfia dyke. On the other hand the plagioclases from the anorthosites are less calcic than the first formed felspars in many of the Gardar gabbros which may be as basic as An_{72-75}. This might suggest that the anorthosites were formed from a Gardar magma which has already undergone considerable differentiation. Host dykes such as the giant gabbro on Tugtutôq which contain more basic felspar than is found in the anorthosites would then represent a new influx of relatively undifferentiated magma introduced after the anorthosites first crystallised. However, if the conditions under which the anorthosites and gabbro dykes crystallised are compared it can be seen that the initial composition of the dyke plagioclase need have little bearing on the relative position of the anorthosites and gabbros in the differentiation sequence. Considerable evidence suggests that the anorthosites crystallised slowly under deep-seated conditions, and that, although the felspars represent some form of crystal cumulate, they were in direct contact with the magma from which they formed for a long period. This is born out by the coarse-grained, unzoned, simple-twinned nature of the plagioclases themselves and by the remarkable lack in chemical variation between different xenoliths. Under these conditions it could be suggested that the composition of the first felspars to solidify was modified by solid diffusion and resorption so that they remained in equilibrium with the liquid surrounding them. In contrast, the felspars in the dykes crystallised in a much more "volcanic" environment are moderately to heavily zoned and did not keep in equilibrium with the surrounding magma during crystallisation. Accepting these arguments the only valid comparison which can be made between the anorthosite plagioclase and those of the dyke rocks is one of bulk composition. This is, like the olivines, more basic in the anorthosites than in the dykes. It is therefore possible that the anorthosites began crystallisation from a more primitive Gardar magma than any which reached the present surface as dykes.

However, it has been seen in the section on the petrology of the microsyenite-trachydolerite host dykes that liquid fractionation may take place in Gardar magmas. This means that the characters generally taken as indicating that a rock crystallised from a primitive or relatively undifferentiated magma may no longer be valid. If the magma from which the anorthosites accumulated was already stratified while still liquid then it is perfectly possible that labradorite formed in the sodic upper parts of the chamber while the felspar in equilibrium with the magma lower in the chamber was bytownite. Subsequent release of magma from different levels in the hypothetical chamber would lead to a variety of hosts ranging from the "more primitive" olivine gabbros derived from lower levels to the "highly differentiated" magmas trapped beneath the anorthositic roof. Although local conditions might lead to the intrusion of a suite ranging from gabbro to syenite, apparently in the order produced by normal crystal fractionation, there is no reason – given liquid fractionation as a mechanism – why the same chamber should not yield a variety of rock types in an almost random order. This is in fact what is seen in the Gardar dykes; although their characters generally conform to a differentiation trend, suggesting derivation from a common basaltic magma, the order of their intrusion suggests that quite commonly the gabbros remained liquid after the differentiation of the syenitic magmas. This is impossible if the rocks are comagmatic and crystal fractionation is the only differentiation mechanism. The possibility that labradorite formed at the top of the chamber while any plagioclase forming at a lower level would be more basic may explain the derivation of the bytownite-bearing xenoliths from Tôrnârssuk. These may now be regarded as forming contemporaneously with the normal anorthosites from the roof zone but at a deeper more calcic level in the magma chamber.

The possibility that liquid fractionation could have played a large role in the derivation of the different magma types found in the Gardar has a significance far outside that of the present problem. Firstly another variable is added to phase relationships. Trend lines on theoretical phase diagrams are drawn on the assumption that crystal fractionation is the main mechanism of differentiation. This is valid on the small scale used in the laboratory but becomes much more questionable when applied to intrusions which have crystallised over a long period of time from a magma column several kilometres in vertical extent. Secondly, liquid fractionation may help to explain the great preponderance of alkali rocks seen in the Gardar and similar provinces. If the syenitic fractions of a section of molten mantle rose to the top of the immense magma chamber formed either in the lower crust or upper mantle then it is reasonable to suggest that these would form major intrusions in the upper crust out of proportion to the actual volume of syenite and gabbroic magma produced at depth. On the other hand minor intrusions, the emplacement of which does not require the bodily removal or displacement of large amounts of crustal material, may be expected to represent magmas taken from a variety of levels in the primary magma chamber, and would therefore be expected to contain a higher proportion of basic rocks. This is in fact what is found in the Gardar, the major intrusions contain little gabbro while the minor intrusions and the lavas are dominantly gabbroic.

The relationship between the layered anorthosites and the rest of the Gardar is perhaps more complex. Mineralogically they resemble the granular anorthosites although they show less well developed plutonic features and it could be argued that since they occur in the same dykes as the granular rocks they are local late variants. However, the textural differences between the two types suggest that they formed under widely different conditions and that the magmas from which they crystallised must have possessed different properties. The granular anorthosites all appear to have formed by top accumulation; thus the magma from which they crystallised probably had a specific gravity greater than 2.63 at 1000°C. This suggests a gabbroic mother liquid. On the other hand UPTON (1961) argues convincingly that the textures in the Assorutit block are due to settling of plagioclase. Even allowing for some mechanical aid to plagioclase settling, such as downward sweeping magma currents, it seems doubtful that the magma would have been denser than 2.60–2.63 at 1000°C. This suggests a dioritic or dioritic-gabbro magma which may also have given rise to the closely associated syenites. However, the host gabbro surrounding the Assorutit block is believed to represent one of the more primitive gabbros in the Gardar. This could be used as good evidence that there was more than one pulse of little differentiated magma in the area. The possibility that liquid fractionation has taken place in a primary Gardar magma chamber may simplify the history slightly. It is no longer necessary to postulate repeated pulses of primary magma derived over a considerable period of time from the mantle. The "more primitive" hosts like the giant dykes of Tugtutôq can just be regarded as more gabbroic differentiates of the same primary magma from a lower level in the same chamber from which the anorthosites had already crystallised. The structures seen in the Assorutit anorthosite suggest that a floor may have been present between the still molten basic magmas at depth and the crystallising layered anorthosite syenite complexes nearer the surface.

Perhaps the main bulk of the Gardar gabbroic magma was trapped beneath a granular anorthosite cover. Small amounts of partly differentiated material escaped carrying with them inclusions of granular anorthosite and differentiated further to form layered bodies with labradorite-olivine cumulates at the bottom and syenite at the top. While these were developing (for example in the Narssaq and Klokken areas) a further pulse of material from a lower level was released carrying with it blocks of the original granular anorthosite roof, and the solidified lower portions of the layered bodies. In the case of the Giant dyke—anorthosite—syenite complex found on the Assorutit peninsula the syenite appears to have been in a liquid state when the body was broken up by the intrusion of a new phase of olivine gabbro. However, in the Klokken intrusion, where troctolitic rocks containing anorthosite form the lowest member of a layered syenite intrusion, the magma chamber was not disrupted by a later pulse of olivine gabbro and the resulting rocks may represent the crystallisation of an alkaline fraction which escaped from the top of an earlier Gardar chamber at depth.

If the idea that the granular anorthosites formed a roof over a large differentiating magma chamber is accepted then the relationship between the megacrysts and their host can perhaps best be understood by regarding them as unconsolidated roof material, some of which may have been derived by blocks detached from the overlying granular anorthosite, but the majority of which probably formed free-floating crystals until they were included in dykes given off from the chamber. As the magma beneath the granular anorthosite roof became less basic the flotation of felspars became a less efficient mechanism for the accumulation of anorthosite and the single crystals hung suspended. Some of them rafted together to form loose aggregates and there appears to have been a complete gradation between solid granular rock and single crystals. The megacrysts, however, are generally larger and slightly less basic than those in the granular rocks reflecting perhaps a longer period of suspension in the progressively more differentiated magma. It is possible that the formation of plagioclase megacrysts at the top of a large chamber of this type and the corresponding sinking of the mafic minerals is one of the chief processes giving rise to the syenites as a residual magma in the Gardar, although liquid fractionation on a large scale cannot be ruled out. Because of their high density the mafic minerals are rarely brought up in Gardar host dykes unless they are trapped by felspar.

When the stage of differentiation at the top of the primary magma chamber had reached a point at which the megacrysts in suspension were between An_{56} and An_{50} the water content of the magma reached a concentration high enough to change the state of the iron held within the felspar lattice. This resulted in the formation of black felspars. When these were carried upwards in hosts which themselves contained considerable water they remained in equilibrium with their hosts and retained their colour; however, when the black megacrysts were included in more anhydrous magmas, derived from a slightly lower level in the chamber, the colouration became unstable and the black felspars developed a clear rim.

If differentiation continued, a point was presumably reached at which the large plagioclase crystals would no longer hang in suspension but would sink. From the analyses of the dykes given in Table 4 it appears that this limit was reached when the silica content was around 53 $^0/_0$. Few more differentiated Gardar rocks carry anorthosite as xenolithic material, and the few cases which do can be explained as fragments brought up from the granular mass by strong flow.

The differentiating magma was tapped from time to time and gave rise to a series of intrusions, the most basic of which were probably fairly close to the magma from which the granular anorthosites were precipitated, while the final alkali residues may have resulted in some of the major late Gardar syenites released from under the anorthosite capping by block faulting. As the type of magma released from the chamber changed so did the typical inclusions of anorthosite. The more basic hosts are characterised by large irregular masses of granular anorthosite. As the hosts become more alkaline so the typical inclusions changed, first to a mixture of smaller granular anorthosite masses and large glassy megacrysts, and finally to medium to small megacrysts with a lower anorthite content (An_{40-50}) and a characteristic close-set twin pattern.

The ideas suggested in the final section of this paper are only an hypothesis built on the major features seen in the anorthosites and their hosts. Even if they are correct in general they can only be a gross over-simplification of the actual processes which took place. Although it has been suggested that some of the syenites in the Gardar may have developed by the removal of anorthosite from an earlier basic magma no ready explanation can be given for the early Gardar alkali rocks which were intruded before the majority of anorthosite-bearing dykes. Perhaps the anorthosites were formed throughout the early and mid-Gardar and their relatively sudden appearance in large amounts in the mid-Gardar is due to some change in tectonic conditions which brecciated the anorthosites at depth. It seems quite conceivable that the formation of a massive anorthosite roof over the developing Gardar magma might take a considerable time to form, with several stages of growth corresponding to the influx of new, undifferentiated material from depth. Whatever the actual mechanics of their formation it seems likely that the formation of anorthosites played a decisive role in the production of syenites in the Gardar and possibly in other alkali provinces. It can be seen that instead of being mineralogical curiosities the xenoliths and plagioclase megacrysts give valuable information about the mechanics and stages of differentiation of the Gardar province as a whole.

X. APPENDIX

a) Detailed description of textural relationships seen in Fe-Ti oxides from the anorthosites

i) Textures of Fe-Ti oxides from primary granular anorthosites.

The only fresh sample of Fe-Ti oxides from a xenolith of granular anorthosite was collected by J. P. Berrangé (48318). In this sample the oxides occur as single grains up to 3 mm median diameter grouped together in a groundmass of crushed sericitised plagioclase. The larger oxide grains are subhedral with embayed margins, smaller grains tend to be more euhedral showing the magnetite octahedral form well developed. A small amount of sulphide is found in the embayments of the larger grains and in small cracks crossing the oxides.

The opaque minerals consist of several components: magnetite, ilmenite, spinel, and a series of other iron oxides (probably hydrated) which have formed from the magnetite. The relationship between the various components suggests that the minerals have passed through a multistage history involving original crystallisation, alteration under oxidising conditions, and recrystallisation under elevated temperatures. Late alteration at lower temperatures probably took place during the cooling history of the host rock, although this is not always divisable from the first phase of alteration. These stages agree with the history of the anorthosite fragments suggested from the field setting although the evidence is too small to make a firm correlation between the two.

The dominant mineral present is magnetite showing a pale pinkish-buff colouration in reflected light which becomes quite pronounced using oil immersion objectives.

The first phase in the history of the mineral was the exsolution of spinel in minute quantities parallel to the {111} planes in the magnetite. This presumably took place early in the crystallisation history of the anorthosites.

The exsolution of spinel was followed by that of ilmenite. The temperature for this exsolution is unknown and in view of recent work on the relationship between magnetite, ulvöspinel and ilmenite it would be unwise to make any suggestions without an accurate knowledge of the oxidation and reduction history of the rock.

The {111} trellis formed by these ilmenite lamellae is moderately uniform. In some areas one or two of the lamellae directions may be dominant leading to a slightly patchy appearance under partially crossed nicols and the development of a diamond trellis instead of the more usual equilateral triangles. Individual ilmenite lamellae measure in the order of $1\ \mu m \times 10$ nm and thus their detailed structure is too small to be resolved optically. It seems possible that the pink colouration of ilmenite lamellae on a sub-microscopic scale. There is no evidence from the directions of the ilmenite lamellae that ulvöspinel developed as an independent phase at any stage although it presumably was present in solid solution with the magnetite before the formation of the ilmenite lamellae. In some of the smaller grains the exsolution of ilmenite is on a larger scale than normal and forms a complex intergrowth with altered magnetite. This intergrowth is in the form of diamond shaped dendritic individuals of ilmenite with {111} control of the dendrites making up each ilmenite crystal. These dendrites may have formed by the coalescence of independent ilmenite lamellae which took place after the original exsolution.

The second stage in the development of the Fe-Ti minerals was an oxidation and possibly hydration of the magnetite and ilmenite. This gave rise to a patchy development of a second iron oxide which differs from the magnetite by being silver -gray in reflected light. The reflectivity is slightly higher than magnetite and there is a fairly pronounced orange-yellow internal reflection. The mineral appears to be isotropic which suggests that it is not haematite, however, this may be due to the sub-microscopic division of the material which masks the properties of individual grains. The second-stage mineral is clearly later than the magnetite forming the groundmass of the Fe-Ti oxide crystals as it fills in cracks and fissures cutting the early exsolution textures. Ghosts of the original ilmenite trellis are left undisturbed within the second-stage oxide. Most of the larger grains are embayed and show signs of considerable chemical attack. Some of this undoubtedly took place at a late stage in the history of the grains, however, it is reasonable to suggest that the main embayment took place as a result of the same process which caused the widespread formation of the second stage Fe-Ti oxides. The embayment of the crystals was accompanied by the formation of discordant veins and patches of non-isotropic, poorly reflecting material believed to be an iron hydroxide.

The oxidation and possible hydration of the Fe-Ti oxides under moderately low temperature conditions, which occurred during the second stage of their development, probably correspond to the widespread alteration of the anorthosites which took place before their inclusion into their host dykes. This process is more commonly seen as sericitisation of the felspars and the breakdown of the original coloured minerals.

A third stage in the development of the Fe-Ti oxides is recognised by the regeneration of magnetite at the expense of the second-stage oxides. The new magnetite forms veins cutting both the first and second stage minerals and as rims surrounding and partially replacing isolated areas of the second-stage oxides. This new generation of magnetite is slightly paler in colour than the early magnetite, possibly due to a more complete exsolution of Ti oxides. The formation of the second generation of magnetite was also accompanied by the exsolution of a poorly-reflecting spinel seen as dark centres to ilmenite lamellae within the new magnetite. This second phase of spinel exsolution suggests that the anorthosite xenolith was heated strongly, presumably during its inclusion in its present host. The ilmenite and spinel lamellae within the second generation of magnetite were exsolved in the normal {111} trellis, however, there is commonly an angular discrepancy of up to 20° between the new trellis and the original exsolution pattern in the surrounding early magnetite. This change of direction shows that the orientation of the new magnetite is not the same as the first generation and may have been controlled by the crystallography of the minerals formed during the second stage in the development of the Fe-Ti oxides.

The last process which affected the Fe-Ti oxides was a late alteration and the precipitation of small quantities of sulphide. The sulphide grains are closely associated with the embayments seen in the outlines of the main oxide crystals and some of them may have been formed during the earlier phase of alteration. However, some sulphide grains are undoubtedly later than the regeneration of magnetite described above as they are found in cracks and fissures cutting all the earlier structures.

Most of the sulphide grains are too small for optical determination; among those identified are grains of pyrrhotite and pentlandite. Limonite commonly surrounds the sulphides.

Two samples (61071 and 30816) were collected from the primary granular anorthosite blocks found in dyke 2 at the head of Kobberminebugt. Both have been heavily altered, presumably before their inclusions in their present host. However, the earliest structures preserved in the specimens suggest that the main opaque mineral of the granular anorthosite was magnetite with ilmenite exsolved along the characteristic {111} planes. This early structure is seen in various stages of destruction and recrystallisation. In 61071 the former presence of magnetite and ilmenite can only by recognised as amorphous masses of hydrated iron oxides containing a ghost ilmenite trellis. Within these masses there are strings of fresh rounded magnetite grains often distributed along lines suggesting control by the original trellis pattern in the opaque mass.

30816 shows a more complex history which is probably due to recrystallisation of the sample under the influence of the surrounding dolerite host. As the sample was collected close to the margin of an anorthosite block some of the opaque minerals may have been derived from the host. The Fe-Ti oxides in this sample can be divided into two main categories. The first consists of irregular broken and corroded masses of intergrown magnetite, ilmenite and rutile (Fig. 70). The second consists of rounded grains of fresh magnetite with exsolved ilmenite forming an exceedingly fine trellis in the groundmass. The grains may coalesce to form irregular patches of magnetite. The large, broken grains show complex relationships between the component minerals which suggest that the grains have passed through a multistage history. In the centre the grains consist of a rather altered magnetite with ilmenite lamellae along the {111} planes. Rutile forms thin spearing plates which are generally aligned parallel to the ilmenite lamellae but which locally cut the earlier trellis discordantly. In some cases ilmenite lamellae are seen to have been altered to rutile along their length, the new material containing small relics of the older near to the boundary. It seems probable that most of the rutile was formed in place by replacement of ilmenite, limited migration of material taking place to form the discordant plates.

The complex intergrowth of magnetite, ilmenite and rutile forming the centres of the grains is surrounded by a clear rim of magnetite against which the older rutile and ilmenite often end abruptly. In contrast to the rather altered appearance of the centres of the grains this rim is clear and appears to have been formed by recrystallisation or addition of new material. A fine trellis of exsolved ilmenite is seen in the rimming magnetite. In some cases small irregular patches of ilmenite are found within the central parts of the crystal suggesting that there has been some recrystallisation of this mineral at the expense of rutile. In one crystal a new generation of ilmenite has formed across the centre cutting rutile and older ilmenite.

Many of the larger crystals show signs of mechanical disruption. This may be seen by the breaking and cracking of the crystals with the formation of small fragments surrounding the main mass or it may be seen in the distortion of the ilmenite and rutile lamellae within the magnetite (Fig. 71). The fact that the rutile is apparently broken as well as the ilmenite suggests that the disruption took place at a late stage in the history of the oxides; it is however possible that rutile replaced distorted ilmenite and retained its form.

There is no direct evidence in the relative time of formation of the outer clear rim of magnetite and the disruption of the crystals. In general the clear rims appear late but some crystals have been broken apparently after the formation of the rim. This may suggest that both disruption and recrystallisation took place during transport of the anorthosites in their Gardar host.

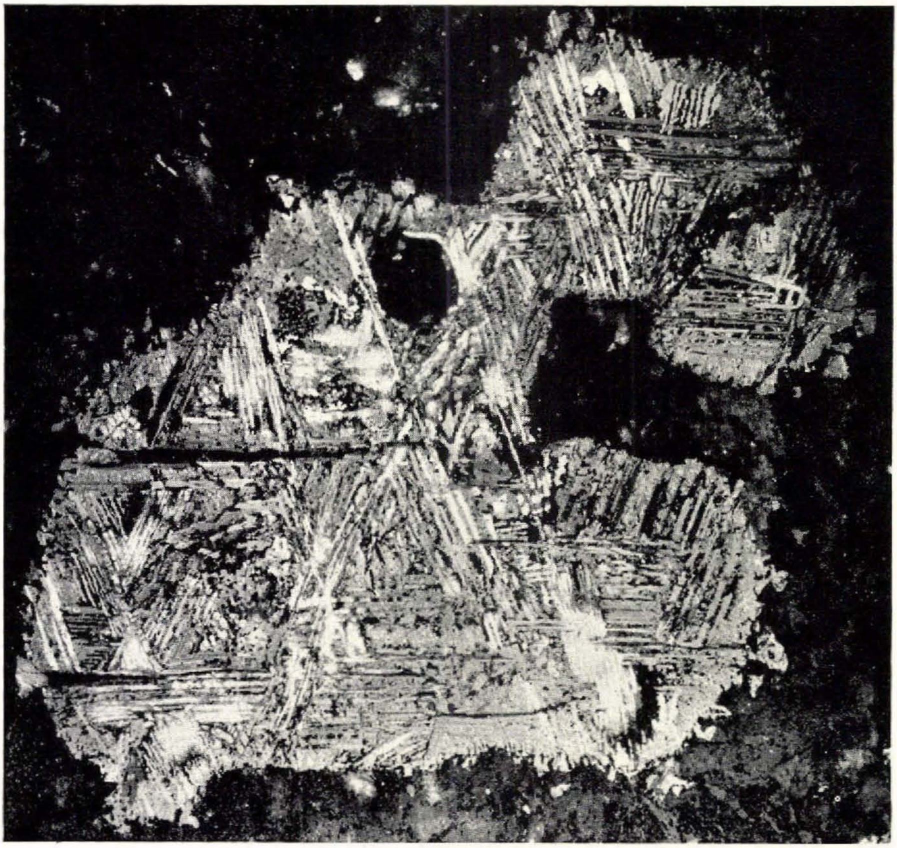


Fig. 70. Photomicrograph of Fe-Ti oxides in a polished sample of the margin of a granular anorthosite xenolith in a dolerite host (locality 2 Kobberminebugt). Partially crossed nicols, × 50, G.G.U. 30816. The oxide grain has passed through a complex history of alteration and regeneration. The centre consists of magnetite with {111} ilmenite lamellae. In many places the magnetite has been replaced by an undetermined poorly-reflecting substance while the ilmenite has been replaced by rutile (white on photograph). At a later stage both ilmenite and magnetite have been regenerated. This is particularly noticeable on the margins of the grain where the ilmenite forms irregular rod-like segregations in clear magnetite.

Photograph P. Edwards, Danmarks Tekniske Højskole.

The rounded grains of magnetite, which form the second type of Fe-Ti oxide seen in the sample, may have been formed as the result of addition of new material from the host or they may be due to a more complete recrystallisation of the magnetite-rutile-ilmenite grains in which the rutile has been converted back to ilmenite under the influence of the high temperature of the host dyke. In either case they would presumably correspond to the rim of clear magnetite surrounding the first type of oxide grain.

The last stage in the history of the opaque minerals from the Kobberminebugt anorthosites was the deposition of small amounts of pyrite and pyrrhotite along cracks in the magnetite. This affects both types of Fe-Ti oxide grains.

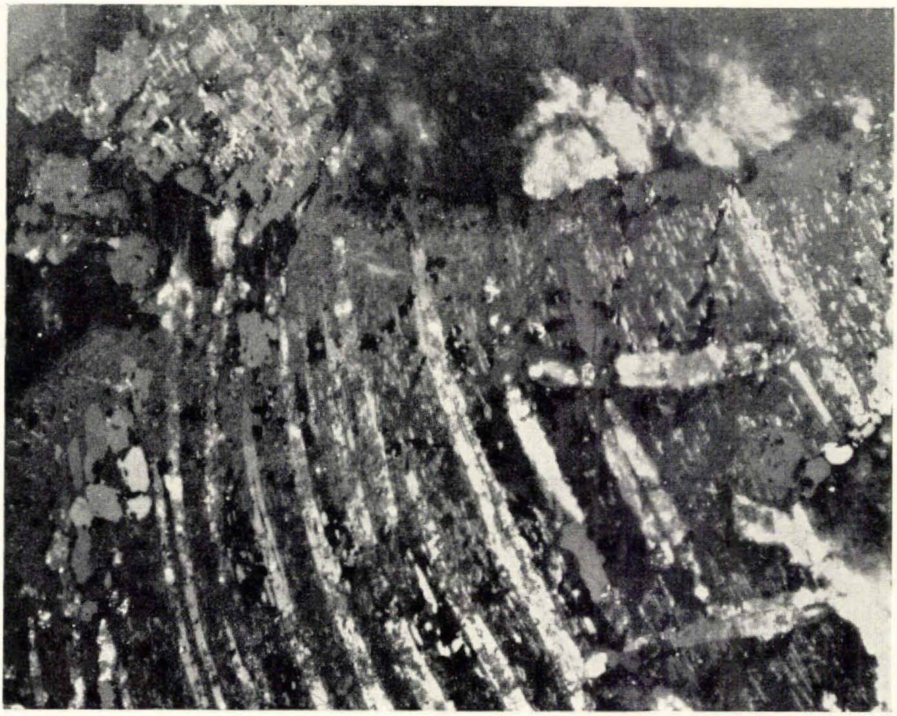


Fig. 71. Photomicrograph of a second field of view of the same polished sample shown in Fig. 70. Partially crossed nicols, × 100, G.G.U. 30816. Margin of a magnetite grain showing small detached fragments and distortion of ilmenite lamellae. The ilmenite has been partially replaced by rutile (white on photograph). A new generation of ilmenite formed as irregular grains is seen in the magnetite near the margin of the grain. Photograph P. Edwards, Danmarks Tekniske Højskole.

ii) Textures of Fe-Ti oxides from secondary anorthosites and oxide megacrysts in Gardar hosts.

The simplest textures are found in samples (61134 and 61183) of the large oxide grains which surround the plagioclase megacrysts found in Dyke 13, p. 54, Fig. 34. These are believed to represent primary crystallisation features of the oxides very little effected by secondary recrystallisation. The oxides themselves are thought to have crystallised slightly later than the megacrysts they surrounded and possibly considerably later than the granular anorthosites. The oxide grains are moderately well developed crystals up to 5 cm in diameter; they partially enclose plagioclase crystals and show euhedral or subhedral outlines against the host rocks. Many of the crystals are embayed along the contacts with the host rocks. These embayments cut across the internal structures of the grains suggesting that the grains were no longer stable in the host dyke at the time of its final crystallisation. Several of the crystals show cracks which have acted as the loci for late alteration of the magnetite. In many cases embayments of host material in the oxide crystals contain fragments of magnetite and ilmenite which appear to have been mechanically detached from the larger grains. A few grains of sulphide are present either in the host rock or along cracks cutting the oxide grains. These sulphide grains are often surrounded by an indeterminate mass of poorly reflecting material. Pyrite, chalcopyrite and pentlandite have been identified among the sulphides.

The oxide minerals consist mainly of magnetite with ilmenite lamellae exsolved within them. There are few independent ilmenite grains within the host rock but most of these appear to have been derived from larger composite oxide crystals. Ilmenite also occurs within the magnetite crystals as small irregular grains which may be joined into veins and stringers. Incipient oxidation of the magnetite is quite common, forming irregular areas with a higher reflectivity than the surrounding magnetite and showing local red internal reflections. This oxidation is particularly noticeable at the edge of the crystals and along the larger ilmenite lamellae. The mineral formed does not show the characteristic anisotropic properties of haematite; this may be due to the grain size of the new mineral which is replacing the magnetite in exceeding fine scale intergrowth with ilmenite (cf. p. 201). Some oxide crystals have been further altered to a substance which shows poor reflecting properties but strong red internal reflection. This is probably an iron hydroxide.

The ilmenite lamellae are parallel to the {111} planes in the magnetite; they range in size from sub-microscopic to 0.1 mm wide and several centimetres long. Individual magnetite crystals may contain up to three different generations of ilmenite lamellae (primary, secondary and tertiary), the initial exsolution of which is thought to represent different stages in the early thermal history of the oxides. Some of the primary lamellae have features thought to have been formed by a later stage of growth and thus the inter-relationships between the different ilmenite lamellae generations are variable.

It is impossible to estimate the proportions of ilmenite and magnetite present in a polished sample as the development of a particular generation of lamellae varies within wide limits from one part of a crystal to another and because the smallest ilmenite lamellae are beyond the resolving power of the microscope.

Primary lamellae. The primary ilmenite lamellae divide the oxide crystals up into large triangular areas. The spacing between the primary lamellae varies considerably and there is no general correlation between their thickness and their abundance at any one part of an oxide grain. In some fields of view they occupy over 30 % of the area, generally they form less than 5 % of the total oxide. Primary lamellae range between 5–100 μm wide and may be over 1 cm long, most are at least fifty times as long as they are wide although a few crystals develop short stubby primary lamellae which run en échelon across the mineral. The primary lamellae may be developed parallel to any of the {111} planes in the magnetite. However, at any one point the ilmenite lamellae are generally exsolved preferentially in one particular plane and are therefore parallel to each other. This dominant direction is not constant throughout the crystals and need not agree with the main direction of other generations of ilmenite lamellae at the same place. The primary lamellae are straight, they show a slight thickening and thinning. Thinning is especially common where two lamellae intersect. Many of the lamellae are zoned with a straight sided, clear, central zone and marginal areas which contain minute inclusions of a poorly reflecting substance, probably a spinel. These inclusions are generally parallel to {111} and are often in a direct line with secondary ilmenite lamellae in the surrounding magnetite. The outer margins of the primary lamellae are often slightly irregular while the central zone of clear ilmenite remains fairly constant in thickness (Fig. 72). It seems probable that the marginal zones were formed by a secondary addition of ilmenite after the general exsolution pattern had been established. The presence of spinel in these late overgrowths may point to elevated temperatures when the oxides were included in their present hosts. Most of the primary lamellae show tapering

Fig. 72. Polished sample of Fe-Ti oxide showing {111} controlled primary ilmenite lamellae with a secondary overgrowth containing inclusions of an unidentified substance (?spinel). Partially crossed nicols, × 150, G.G.U. 61135.

ends and it is quite common for several primary lamellae to end simultaneously within the magnetite. Some of the ilmenite lamellae, especially those which have a pronounced zoned structure, show forked and irregular endings. This is thought to have developed at a later stage than the initial formation of the primary lamellae.

The centre of the larger primary lamellae are frequently full of alteration products, including a little sulphide and possibly some haematite. This is presumably due to the lamellae acting as a structural weakness, at the margins of the oxide grains; ilmenite is almost always more stable than the surrounding magnetite. Some of the larger ilmenite lamellae contain thin magnetite veins either centrally or to one side of the ilmenite. These veins were formed after the exsolution of ilmenite as may be seen by their cross-cutting relationship to the secondary and tertiary ilmenite lamellae. They occasionally branch and displace the margins of the primary lamellae. The late magnetite is lighter in colour than that seen in the general groundmass but is distinctly buff when compared to areas of late oxidation.

Secondary lamellae. Secondary lamellae occur within the triangular trellis defined by the primary lamellae. In places where only one set of primary lamellae was developed the secondary lamellae generally formed along one of the other {111} planes resulting in a ladder-like texture. The secondary lamellae show a wide size range and it is only possible to distinguish them from tertiary lamellae with certainty when all three generations are developed in the same crystal. Typical secondary lamellae measure 1–5 μm wide and up to 100 μm long; the length is at least twenty times the breadth. The ends of the secondary lamellae commonly taper on approaching primary lamellae. The distribution of secondary lamellae is very irregular, adjacent interstices between primary lamellae may show a widely differing development of secondary lamellae. In some of the interstices between primary lamellae, secondary lamellae form a subsidiary trellis, in other interstices the secondary lammela show a local preferred orientation with development of a patchwork texture.

The relationship between the primary and secondary lamellae is complex, many primary interstices show an increase in the size of secondary lamellae towards the centre of the triangle. In a few instances the secondary lamellae are formed parallel and in close proximity to the nearest primary lamella. Other primary interstices are filled with a fine-grained tertiary exsolution intergrowth in which small secondary lamellae have developed. A few secondary lamellae show a zoned structure similar to that described from the primary lamellae. There is occasionally a marked development of secondary lamellae close to cracks and late ilmenite veins. In these cases the secondary lamellae themselves may form a grid of coalescing irregular veins still controlled by {111}. This appears to be caused by the growth of ilmenite at a late stage onto the pre-existing trellis of exsolution lamellae.

Tertiary lamellae (groundmass exsolution). Under low magnification the groundmass in which the primary and secondary ilmenite lamellae are developed appears to be isotropic and formed of pure magnetite. However, when examined under a magnification of 500 × or more the groundmass is seen to be composed either of magnetite with ilmenite needles less than 10 μm in length or of an interlocking patchwork of diamond shaped areas with a slightly different extinction position. This patchwork, which is most noticeable under partially crossed nicols, is made up of rectangular blocks measuring approximately 2×8 μm, the long axes of which are parallel to {111}. The internal structure of the majority of these small areas is not resolvable, however, by analogy with the distribution of some of the larger secondary and tertiary lamellae it seems likely that they are composed of an intergrowth of magnetite and sub-microscopic ilmenite dendrites with a preferred orientation within each area. The small scale intergrowth of magnetite and ilmenite which makes up the groundmass of the oxide grains is much more evenly distributed than either primary or secondary ilmenite exsolution lamellae and is only absent in areas where there has been exceptionally strong exsolution of early ilmenite lamellae or where there has been some recrystallisation. There is commonly some local decrease in the size of the tertiary intergrowth close to primary or secondary lamellae.

Most of the magnetite crystals are cut by **irregular veins of ilmenite** made up of strings of separate grains. Independent ilmenite grains are also found within the magnetite groundmass. The outlines of both the independent grains and the constituent grains in the veins are commonly controlled by the {111} direction of the magnetite. Both are often in optical continuity with primary lamellae which appear to have acted as a focus for further exsolution of ilmenite. Like the primary lamellae the veins commonly show a zoned nature with {111} orientated inclusions of a poorly reflecting substance concentrated near the border. The junction between the primary lamellae and the grains is sometimes complex, the primary lamellae may pass through the grain or it may pass through the contact at either side but not continue through the grain centre. In some crystals the formation of late ilmenite grains appears to accompany the oxidation of magnetite, in other grains it appears to accompany the formation of areas of magnetite free from ilmenite exsolution lamellae.

There is a slight tendency for the ilmenite to be concentrated near the margins of the magnetite as irregular grains. Some of these irregular grains appear to have been derived by the segregation of ilmenite from the exsolution lamellae while others appear to have crystallised from the surrounding dyke magma and form symplectic intergrowths with the silicate minerals. The grains of ilmenite thought to have been derived from the dyke magma are paler in colour than those which have exsolved from the magnetite. This may be due to haematite in solid solution. The pale ilmenite frequently shows no logical relationship to the trellis pattern within the magnetite and cuts the early exsolved ilmenite discordantly. Many of the oxide grains are surrounded by a sheath of alteration products containing various oxides and hydroxides of iron. Haematite occasionally occurs as thin wisps within this altered mass.

Independent oxide megacrysts which are presumed to have been formed at approximately the same time as the plagioclase megacrysts have been noted as inclusions in some of the felspathic microsyenite-trachydolerite dykes and in the secondary anorthosite xenolith from Narssaq. They may occur in other gabbroic hosts but they have not been observed presumably because they would only be distinguishable from the host's opaque minerals by careful observations on polished samples.

Most of the Fe-Ti oxides found as inclusions in microsyenite-trachydolerite hosts have a broad rim of indeterminable alteration products. A polished sample (38186) from the western end of dyke 12 (p. 50) shows many features in common with the large grains surrounding the plagioclase megacrysts in dyke 13 which are described above. These include three generations of ilmenite exsolution as lamellae, followed by the formation of grains of ilmenite within the magnetite. However, the formation of ilmenite as separate grains concentrated at the edge of the magnetite crystals is more pronounced. The most noticeable differences between the oxide grains found in the microsyenite-trachydolerite hosts and those from dyke 13 are the formation of rutile at the expense of ilmenite and the almost complete alteration of magnetite in some grains so that ilmenite and rutile are left as a comparatively stable trellis surrounded by an indeterminable mass of iron hydroxides.

The primary lamellae have rather irregular margins and partially enclose areas showing groundmass exsolution textures which suggests there has been some segregation of ilmenite after the initial crystallisation of the mineral. The zoned structure described on p. 205 is not seen. Secondary and tertiary lamellae are also developed, the secondary lamellae commonly control the exsolution pattern in the groundmass. The tertiary exsolution is developed on a larger scale than that described earlier and consists of a graphic intergrowth of ilmenite and magnetite. The ilmenite dendrites are controlled by {111}, commonly with one direction better developed within a small area. This produces a patchwork affect when seen under partially crossed nicols. The size of the two components increases close to independent grains of ilmenite within the magnetite so that there are commonly quite large areas of magnetite with no ilmenite lamellae surrounding ilmenite grains. The ilmenite grains appear to have formed by a separation and segregation of the ilmenite and magnetite normally found intergrown in the groundmass.

The secondary anorthosite xenoliths from **Narssaq and Bangs Havn** are composed of a gabbroic host together with plagioclase megacrysts. The gabbroic hosts contain plentiful Fe-Ti oxides some of which have been derived from the original anorthosite at depth but which have become so recrystallised that they are mineralogically difficult to distinguish from the opaque minerals of the host rock. In the specimens from Bangs Havn the oxide grains form symplectic intergrowths with large augite crystals which also enclose plagioclase megacrysts. In polished samples the oxides are seen to have been attacked either prior to or during the formation of the surrounding augite. This suggests that they may be derived megacrysts. The internal relationship between ilmenite and magnetite shows a further development of the processes described in the samples from microsyenite-trachydolerite hosts in which ilmenite is gradually separated as independent grains (Fig. 73). Ilmenite forms up to 30 % of the oxides as pure grains within the magnetite crystal boundaries (Fig. 74). The ilmenite grains are commonly controlled by the {111} direction in the

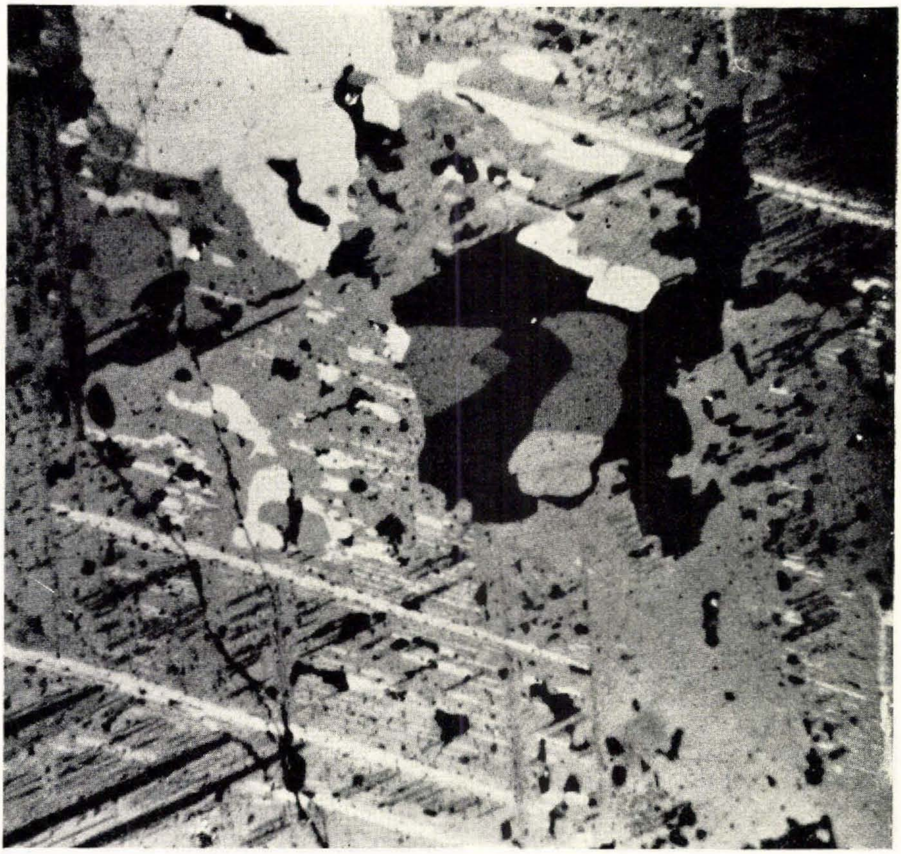


Fig. 73. Segregation of ilmenite to form irregular grains within magnetite. Note that the magnetite surrounding the ilmenite grain is comparatively free from the fine exsolution lamellae of ilmenite seen in the magnetite further away. Partially crossed nicols, × 240, G.G.U. 61014. Photograph P. Edwards, Danmarks Tekniske Højskole.

magnetite, the grains may reach 2–3 mm across and be made up of several ilmenite individuals with a slight difference in optical orientation. The texture shown is similar to large quartz grains which have been subjected to slight strain. Many of the ilmenite grains show thin twin lamellae which taper at either end. The ilmenite contains minute inclusions of an unidentified mineral with moderate reflectivity which shows a consistent orientation within one ilmenite grain. This orientation changes in the twin lamellae. Many of the ilmenite grains are surrounded by a coarse intergrowth of ilmenite and magnetite and in some areas within the magnetite incipient formation of ilmenite grains is shown by coarse segregation of ilmenite from magnetite free of ilmenite lamellae. Primary and secondary lamellae exist but they often coalesce and show irregular margins with the groundmass texture. There are no independent tertiary ilmenite lamellae, the groundmass is formed of graphic intergrowth of the two minerals controlled by the {111} magnetite planes. This intergrowth shows the patchy preferred direction of ilmenite exsolution described earlier. It is thought to have been formed as a result of recrystallisation of original tertiary ilmenite lamellae. During the recrystallisation process the ilmenite has re-

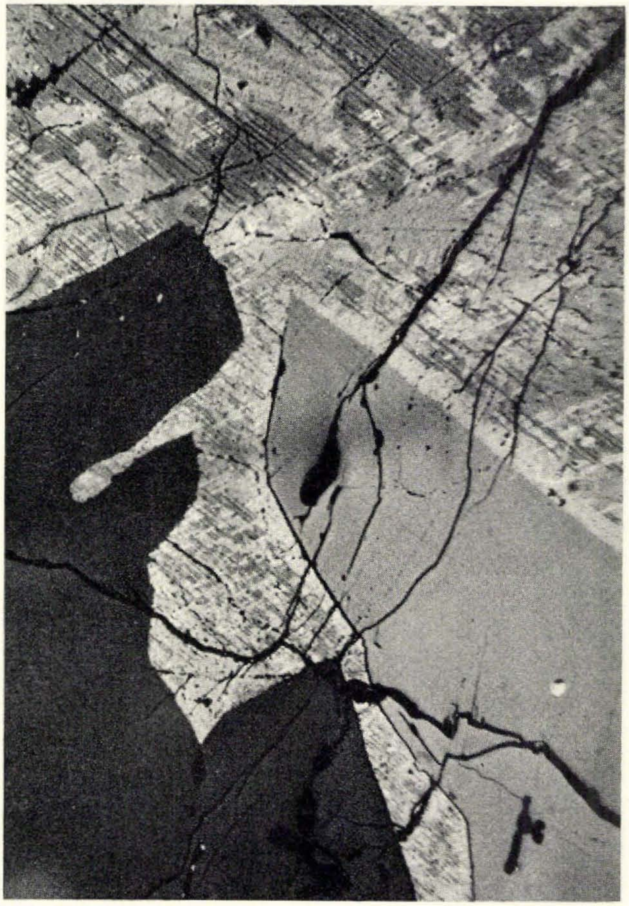


Fig. 74. Polished sample of Fe-Ti oxides from a secondary anorthosite at Bangs Havn. Segregation of ilmenite to form large masses near the margins of Fe-Ti oxide grains. Partially crossed nicols, × 45, G.G.U. 61014.

tained its original orientation so that ilmenite dendrites within one patch of intergrowth shows three distinct positions of maximum reflectivity.

The opaque minerals from Bangs Havn show an appreciable amount of sulphides including pyrite, calcopyrite and pyrrhotite. Some of the sulphides form complex intergrowths with the oxides.

Sample 49320 from the Narssaq anorthosite shows magnetite crystals with a comparatively simple trellis of primary and secondary ilmenite lamellae in a magnetite groundmass completely clear of tertiary lamellae or intergrowth between magnetite and ilmenite. The ilmenite forming the primary and secondary lamellae both show the addition of new material since their original exsolution which can be seen by the irregular margins and local coalescence of the two generations. It seems probable that the segregation of ilmenite from the magnetite groundmass took place when the xenoliths were included into their host to form the secondary gabbroic anorthosite. Apart from grains of magnetite with exsolved ilmenite the

Narssaq anorthosites contain independent grains of ilmenite which appear to have been precipitated during the crystallisation of the secondary host. This ilmenite is distinctly paler than the exsolution lamellae in the magnetite. One of the magnetite crystals shows considerable signs of mechanical deformation before the solidification of the surrounding material suggesting that it has been derived from an earlier source in a similar manner to the felspar insets. This deformation is clearly seen distorting the ilmenite trellis within the magnetite and in places completely dislocating the crystal. After this dislocation took place a new generation of ilmenite was formed which veins the older magnetite and ilmenite along the planes of dislocation. The veining ilmenite is the same colour as the independent ilmenite grains in the surrounding gabbroic matrix of the secondary anorthosite. There seems no reason to doubt that the early generation of magnetite and exsolved ilmenite represents partly recrystallised textures developed in the original anorthosite while the ilmenite grains and vein fillings were derived from their host.

b) Methods and results of strontium and barium spectrographic studies

Strontium and barium were determined by H. Bollingberg, using a Hilger "large Quartz-spectrograph" focussed onto plates set in the range 2800 Å to 5000 Å. The specimen was powdered and mixed with analytical carbon containing 0.02 % Pd as an internal standard. The intensities of the Sr line at 4607 Å and the Ba line at 4934 Å were compared to the Pd line at 3404 Å. Strontium and barium standards were prepared artificially to cover the total range from 1–0.001 % in a synthetic mixture of 55 % SiO_2, 30 % Al_2O_3 and 15 % $NaCO_3$.

The detailed results are listed in Table 6 together with the results of determinations of identical samples using other methods (J. R. Butler, Imperial College London).

c) Methods and results of X-ray powder and single crystals studies

i) X-ray powder patterns.

X-ray powder patterns of 36 felspar samples were made by M. Danø and E. Krogh Andersen using a diffractometer. CuKα radiation was used in the $2\,\theta$ range from 21° to 36°. 33 of the samples examined were taken from the natural anorthositic rocks and plagioclase megacrysts described in this paper and one of the samples was re-examined after heating (p. 114). Samples of the concordant anorthositic layers in the gneisses of the Neria district and a labradorite phenocryst from an olivine basalt lava extruded during the 1964 eruption on Surtsey were used as standards of plutonic and volcanic origin respectively. The angular separations $2\theta\,(131) + 2\,\theta\,(220) - 4\,\theta\,(1\bar{3}1)$, ($\Gamma$) and $2\,\theta\,(1\bar{1}1) - 2\,\theta\,(201)$, (B), (Smith and Gay, 1958) were determined and are listed in Table 7. Fig. 75 shows the result of plotting the angles Γ and B against the felspar composition obtained by refractive index measurements using a variety of symbols to indicate the different types of inclusion examined. The refractive index measurements were made on the same powder as the material used for X-ray determinations before final crushing and there is no reason why there should be a compositional difference of more than 2 % An between the samples used for X-ray studies and the fraction separated for refractive index measurements. Universal stage determinations of the twinning and "structural state" of the felspars (p. 100) were carried out as far as possible on material from an adjacent slice to that X-rayed.

Since one of the peculiarities of the felspars described in this paper is the apparent lack of well defined peaks in the X-ray powder patterns (especially notice-

Table 6. *Strontium and Barium analyses.*

G.G.U. No.	Type of rock	Locality	Anorthite (Mol. %)	Sr content (p.p.m.)	Ba content (p.p.m.)	Remarks
25226	Granular massive anorthosite	Augiâta tasia, Isortoq	59	1500	250	
61071 h	Granular massive anorthosite	Inner Kobberminebugt	60	800	200	
61071 i	Granular massive anorthosite	Inner Kobberminebugt	59	900	200	
61071 j	Granular massive anorthosite	Inner Kobberminebugt	59	1200	200	
32028	Clear subhedral bytownite megacryst in olivine gabbro	Tôrnârssuk	78	400	70	
758 AT/59 1	Black megacryst in alkali dolerite	West Kobberminebugt	54	1000	300	X-ray fluorescence gives 1000 p.p.m. Sr
758 AT/59 2	White margin to above	West Kobberminebugt	53	1100	700	X-ray fluorescence gives 1000 p.p.m. Sr
61073 1	Black megacryst	Inner Kobberminebugt	55	1000	600	
61073 2	White veins in above	Inner Kobberminebugt	53	1000	300	
27411	Black megacryst in alkali	West Kobberminebugt gabbro	54	1100	400	
61086 2a	Centre of white megacryst in alkali gabbro	Nunatak 10 km east of Kobberminebugt	58	1000	200	
61086 2b	Margin of megacryst in alkali gabbro	Nunatak 10 km east of Kobberminebugt	ca.54	1600	400	
61086 2c	Interstitial felspar between megacrysts	Nunatak 10 km east of Kobberminebugt	35	1000	1700	
61086 8	Centre of megacryst	Nunatak 10 km east of Kobberminebugt	58.5	1000	200	
61102	Host gabbro to 61086	Nunatak 10 km east of Kobberminebugt	36	1000	1600	

25171a	Glassy cleavage fragment in gabbro	Head of Isortoq	57	1500	250	
25171b	Glassy cleavage fragment in gabbro	Head of Isortoq	57	1200	280	
25214	Megacryst in gabbroic secondary anorthosite	Inner Kobberminebugt	60	1000	500	
61001	Megacryst in gabbroic secondary anorthosite	Mainland east of Bangs Havn	62	1200	400	
61066	Black felspar megacryst from trachy-dolerite	5 km east of Kobberminebugt	53	1600	700	
48307	Glassy megacryst from trachydolerite	"Vatnahverfi"	52	1800	800	
48319	Black megacryst from micronepheline-syenite dyke	"Vatnahverfi"	49	2000	800	
60037(1)	Black megacryst from microsyenite dyke	Igaliko fjord	53	2500	700	
60037(2)	Interstitial host rock surrounding 60037	Igaliko fjord	26	900	3500	
55209	Glomeroporphyritic mass in trachydolerite	Narssarssuaq	56	2100	900	
49324	Plagioclase from Assorutit layered anorthosite	Assorutit	56	2700	800	Isotope dilution gives 2108 p.p.m. Sr on similar specimen
49324 1	Plagioclase from Assorutit layered anorthosite	Assorutit	56	1700	400	
43749	Megacryst in marginal gabbro	Klokken	48	2200	900	
55464c	Megacryst in marginal gabbro	Klokken	52	2000	700	
55464 c 1	Megacryst in marginal gabbro	Klokken	52	2500	800	X-ray fluorescence gives 2200 p.p.m. Sr

(continued)

Table 6 (cont.).

G.G.U. No.	Type of rocks	Locality	Anorthite (Mol. %)	Sr content (p.p.m.)	Ba content (p.p.m.)	Remarks
55466 2	Megacryst in marginal gabbro	Klokken	53	2800	1000	X-ray fluorescence gives 2200 p.p.m. Sr
55466 1	Megacryst in marginal gabbro	Klokken	51	2100	800	
43751	Plagioclase separated from marginal gabbro	Klokken	52	2100	1800	
56954 a	Centre of glassy black megacryst in marginal syenite	Narssarssuk	54	2500	700	
56954 a 1	Centre of black megacryst	Narssarssuk	54	2800	1000	
56954 a 2	Clear mantle to black megacryst	Narssarssuk	54	3000	1000	
56954 a 3	Sericitised margin to black megacryst	Narssarssuk	—	3200	2300	
56954 c 1	Black centre to megacryst	Narssarssuk	54	3200	1000	
56954 c 2	White margin to megacryst	Narssarssuk	54	3300	2800	
56954	Felspar from host rock (altered gabbro)	Narssarssuk	—	3000	2400	
56961	Black felspar in syenitic host	Narssarssuk	54	3000	1700	
56961 a	Syenitic host (probably contaminated)	Narssarssuk	—	1500	ca. 1 %	Syenite highly contaminated by anorthosite
68824	Black megacryst in pre-Gardar-dyke	Neria	62	500	70	
68824 a	Black megacryst in pre-Gardar dyke	Neria	59	1000	200	X-ray fluorescence gives 1000 p.p.m. Sr
68824 b	Black megacryst in pre-Gardar dyke	Neria	61.5	1000	200	

69675	Black megacryst in pre-Gardar dyke	Neria	65	1000	150	
68508	Concordant anorthosite layer in gneisses (B. WINDLEY)	Fiskenæsset district	85	140	—	Mean of several determinations, X-ray fluorescence gives 130 p.p.m.
	Plagioclase from other Gardar dykes					
25295	110° early Gardar dolerite	Isortoq	46	1000	500	
25232	110° early Gardar dolerite	Isortoq	53	1000	300	
25176	Early Gardar gabbroic plug	Isortoq	60	1000	300	
25194	70° early Gardar gabbro dyke	Inner Kobberminebugt	56	800	260	All plagioclases separated from dyke rocks determined by fusing felspar to glass and determining the refractive index
25209	70° early Gardar gabbro dyke	Inner Kobberminebugt	59	800	300	
31213	70° early Gardar gabbro dyke	Isortoq	42	1200	700	
25238	70° early Gardar gabbro dyke	Isortoq	46	1300	600	
38639	Inclusion of layered gabbro in mid-Gardar dolerite	Inner Kobberminebugt	59	1100	300	
31272	60° mid-Gardar dolerite	Isortoq	48	1500	1000	
38142	60° mid-Gardar dolerite	Isortoq	46	1400	600	
20632	Total plagioclase separated from unfractionated gabbro	Eqaloqarfia	72.5	1500	400	
77993	Total plagioclase from felspathic layer in Tugtutôq gabbro	S.E. Tugtutôq	45	1800	1500	
77994	Total plagioclase from "normal" gabbro, Tugtutôq gabbro	S.E. Tugtutôq	55	1500	1000	
	Plagioclase from post-Gardar dolerites					
25266	150° post-Gardar dolerite	Augiâta tasia, Isortoq	46.5	1000	800	
25267	150° post-Gardar dolerite	Augiâta tasia, Isortoq	40	1400	700	

Table 7. *B and Γ value*

No.	Composition	B	Γ (main peaks)	Γ (secondary peaks)	Type of inclusion
49324	58	0.84	0.59	0.89	Assorutit layered anorthosite
61070	60	0.84	0.90	—	Kobberminebugt, massive granular anorthosite
61047	58	0.90	0.83	0.45	Granular aggregate, Bangs Havn gabbro
61099	58	0.83	0.74	0.97	Granular aggregate in alkali gabbro
61001a	60	0.87	0.42	—	Broken megacryst, Bangs Havn gabbro
61001b	55	0.93	0.25	0.55	Broken megacryst, Bangs Havn gabbro
61086_2	58.5	0.86	0.60	0.87	Glomeroporphyritic aggregate in alkali gabbro
61086_3	59	0.86	0.60	0.78	Large single clear megacryst in alkali gabbro
61086_6	54.5	0.88	0.56	0.91	Breccia of small fragments in alkali gabbro
61086_8	58.5	0.82	0.76	0.94	Large glassy megacryst in alkali gabbro
25171	58	0.86	0.74	0.98	Large glassy megacryst in alkali gabbro
61073w	53	0.91	0.48	—	Clear lamellae from large megacryst
61073b	55	0.88	50	—	Dark lamellae from large megacryst
49332w	50	0.87	0.49	—	Clear part of black and clear megacryst
49332b	51	0.92	0.29	0.75	Black part of same megacryst
$49332b_1$	51	0.91	0.36	0.68	Second sample of black felspar from same megacry
49332h	51	0.83	0.81	—	42332b after heating for 28 days (p. 114)
$758\frac{AT}{59}w$	49	0.88	0.26	—	Clear rim to black felspar (Plate 1b)
$758\frac{AT}{59}b$	53	0.93	0.30	—	Large black felspar inclusion in alkali gabbro
27411	54	0.88	0.22	0.58	Large black megacryst in hormblende gabbro
61066	52.5	0.88	0.58	0.85	Black megacryst in trachydolerite host
60037_1	52.5	0.93	0.33	0.65	Black megacryst in trachydolerite host
60037_2	52	0.91	0.29	0.59	Black megacryst in trachydolerite host
48307	52	0.87	0.66	0.99	Glassy clear megacryst in alkali dolerite
48319	49	0.89	0.65	0.95	Glassy clear felspar in nepheline trachyte
43749	48	0.92	0.32	0.69	Megacryst in margin of Klokken intrusion
55464	49	0.89	0.42	—	Black megacryst in margin of Klokken intrusion
55464_1	53	0.91	0.24	0.54	Gray megacryst in margin of Klokken intrusion
55466_1	51	0.91	0.38	0.73	Black megacryst in margin of Klokken intrusion
55466_2	54	0.90	0.37	0.77	Black megacryst in margin of Klokken intrusion
68824	62	0.86	0.63	0.88	Black megacryst from pre-Gardar olivine gabbro
68824a	59	0.88	0.37	0.60	Black megacryst from pre-Gardar olivine gabbro
68824b	61.5	0.88	0.46	0.76	Black megacryst from pre-Gardar olivine gabbro
69675	65	0.90	0.41	0.71	Black megacryst from pre-Gardar olivine gabbro
69625a	60	0.87	0.64	—	Plagioclase from anorthositic layer in gneiss, Neria
Surtsey	62	0.78	1.04	—	Phenocryst from 1964 eruption, Surtsey

rom plagioclase inclusions.

131 peak	111 & 1$\bar{3}$0 peaks	Structural state and composition from universal stage measurements
ouble	Separable	Mainly intermediate or volcanic, gives An_{55-58} consistently
lightly diffuse	Compound	Very large scatter from below plutonic to above volcanic curves
lear secondary	Compound	On plutonic curve at An_{58}
oor secondary	Compound	Considerable scatter. Y to twin axis 74° instead of ~62° expected
lear secondary	Separable	
ery clear double	Separable	
ery clear double	Double top	Extreme scatter Y to twin axis 41° instead of ~60° expected
iffuse	Compound	Considerable scatter. An determinations vary by up to 20 %, most of optical measurements closer to volcanic curves
arge secondary	Compound	
iffuse	Compound	
mall secondary	Compound	Very large scatter from plutonic to volcanic. Up to 25 % variation in An determinations
ingle peak	Compound	Fairly constant results close to volcanic curve at An_{53}
ouble peak?	Broad diffuse compound	Constant results close to plutonic curve at An_{53}
ingle peak	Double top	
arge secondary	Clear separation	Twinning too fine for accurate determinations
mall secondary	Clear separation	
ingle	Compound, sharp	
ingle	Small double top	Give fairly consistent volcanic optics with An_{52} composition
ingle	Clear separation	
oor secondary	Clear separation	
iffuse	Compound	Consistent volcanic optics at An_{60}
arge secondary	Broad compound	
mall secondary	Broad compound	
arge secondary	Compound	
iffuse	Compound	
lear secondary	Separable	Close to plutonic curve at An_{49}
ingle	Separable	Close to plutonic curve at An_{60}
mall secondary	Compound	Close to plutonic curve at An_{60} and An_{50} (some scatter)
mall secondary	Broad compound	Below plutonic curve at An_{52}
mall secondary	Separable	
mall secondary	Broad compound	
mall secondary	Separable	
mall secondary	Broad compound	
mall secondary	Broad compound	
harp single	Compound	Plutonic optics on curve at An_{60}
harp single	Sharp single	Volcanic optics on curve at An_{62}

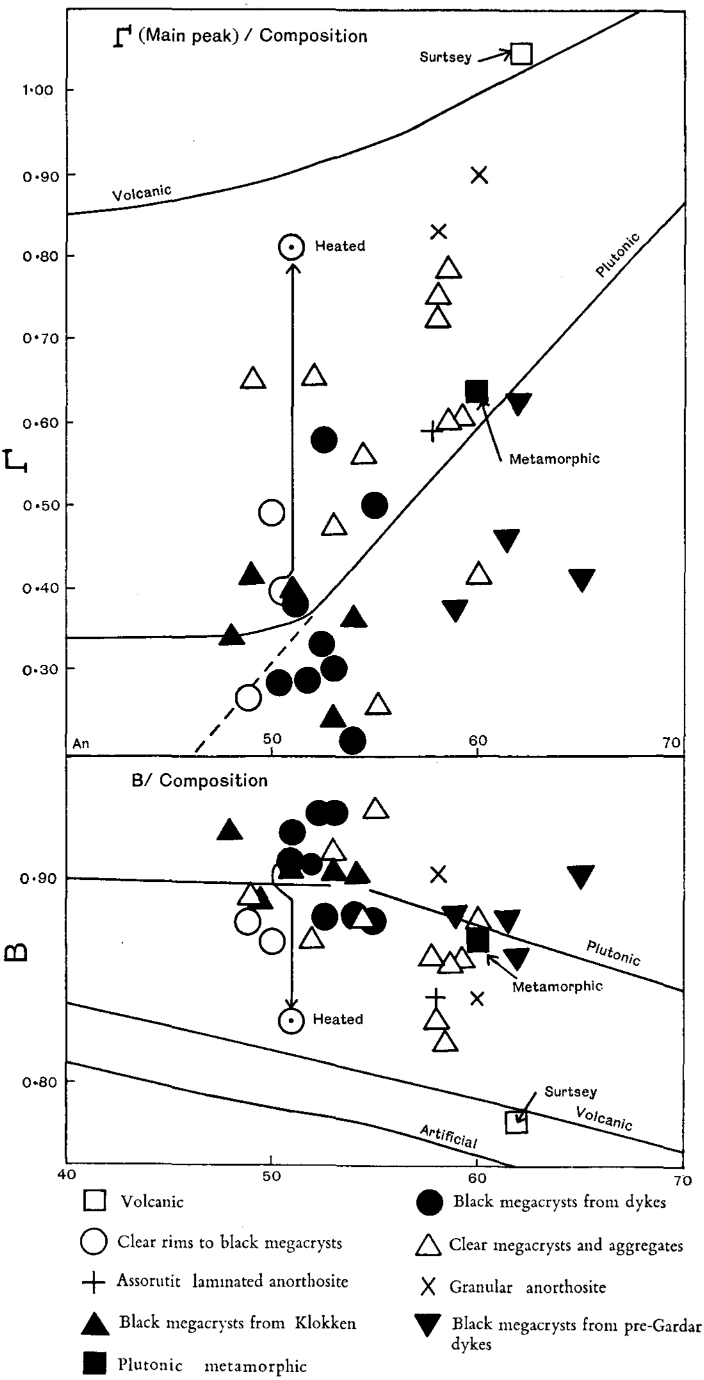


Fig. 75. Diagram to show the relationship between Γ and B, and the anorthite content of various plagioclases found as inclusions in Greenlandic intrusions. A sample from Surtsey and a plagioclase from a metamorphic rock were used as volcanic and plutonic standards.

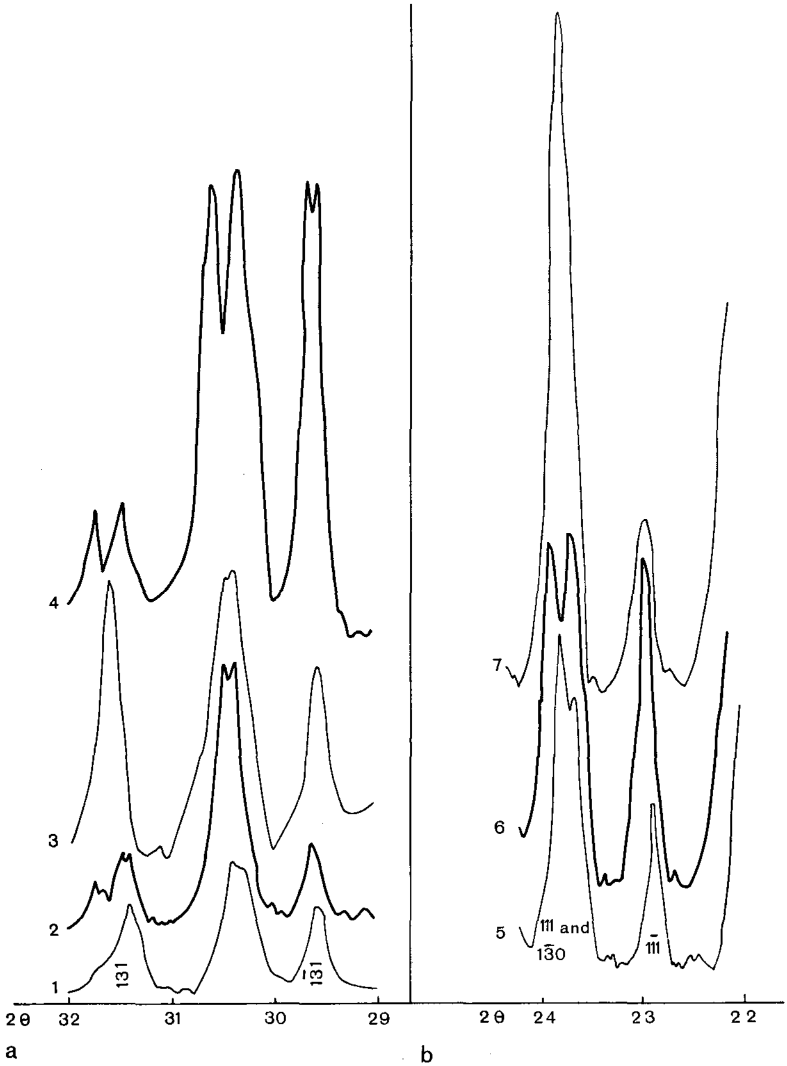


Fig. 76. Powder patterns from 7 plagioclases the properties of which are given in Table 7. All peaks are on the same scale.

Fig. 76 a. Powder patterns in the range 2 θ 29°–32° to show different types of 131 reflections:

1) G.G.U. 69625a: An_{60} plagioclase from gneisses of Neria district, single 131 reflection.
2) G.G.U. 48319: An_{49} glassy megacryst with moderately developed double 131 reflection.
3) An_{62} volcanic plagioclase from Surtsey, single sharp reflections.
4) G.G.U. 49324: An_{56} Assorutit laminated anorthosite, clear double 131 and 1$\bar{3}$1 reflections.

Fig. 76b. The effect of artificial and natural heating on 110 and 1$\bar{3}$0 reflections.

5) G.G.U. 49332w: natural clear felspar rimming black megacryst. Note slight double peak at approximately 23.7°.
6) G.G.U. 49332b_1: black megacryst, note pronounced double peak at 23.7°.
7) G.G.U. 49332h: heated black megacryst, note single peak (presumably compound of 111 and 1$\bar{3}$0).

able in 131 reflections) the values of Γ given in Table 7 were obtained by selecting the main reflection in the 2θ range 31° to 32°. The values are qualified in Table 6 by a description of the type of 131 peaks seen in the powder pattern. Where possible 2 values of Γ are given, the weakest is always given second irrespective of the size of the separation. Typical examples of powder pattern tracings are shown in Fig. 76a. Double values were not obtained for B, presumably because the separation between $1\bar{1}1$ and $\bar{2}01$ is small. In general the values obtained for B fall at approximately the same position between the volcanic and plutonic curves as the main value obtained for Γ, however, in a few examples the differences are larger than could reasonably be expected as experimental error. This may suggest that the B values obtained are averages corresponding to intermediate values of Γ between the two listed in Table 7. The large discrepancy in β^* noted below was not reflected in the values of B obtained from the powder patterns. This is presumably due to the fact that B does not depend on β^* alone and a relatively large difference in β^* need not cause a corresponding large difference in B.

Both Γ and B commonly lie outside the limits drawn by Smith and Gay, especially on the plutonic side of the standard curves. Apparently identical samples of felspar from the same large crystal did not give the same value of Γ and B. This confirms the suggestion from optical studies that although the plagioclase crystals are chemically homogeneous they show considerable differences in crystallographic properties from one part of a crystal to another. For comparison with the X-ray data the dominant characteristics found from universal stage measurements are given in the last column of Table 7. In general the optical results show a far greater scatter than the powder patterns. This can be expected because of the averaging nature of powder samples contrasting with the point determinations using a universal stage. If the universal stage measurements showed consistant values for the angles between X, Y and Z, and the twin axis within any one plagioclase crystal the values of Γ and B obtained generally fell on the plutonic curves at the same composition as given by the optical orientation. This however did not always correspond to the composition given by refractive index studies.

Powder patterns from several of the large megacrysts show a clear double peak in the 2θ range 23.5°–24°, Fig. 76b. This is particularly prevalent in the large black megacrysts although it may also be present in samples from large xenoliths, for example, the Assorutit block. The double peak is characteristically absent from the clear rims surrounding black felspars, the Surtsey felspar, and the heated sample of black felspar (49332h). According to Smith (1956) sodic felspars show reflections in the 2θ range 23.5°–24.5° corresponding to the 111, $1\bar{3}0$, and 130 faces. When the felspars are heated the 111 and $1\bar{3}0$ peaks become inseparable. In the intermediate felspars the 2θ value for 130 is over 24° and while no 2θ values are given for 111 and $1\bar{3}0$ they can be presumed to lie close together between 23° and 24°. The clear double peaks in the 2θ range 23.5–24 seen in the powder patterns of some of the felspars used in the present study may correspond to reflections from 111 and $1\bar{3}0$ faces. The fact that these double peaks are inseparable in the powder pattern of artificially heated black felspar and that they are also inseparable in the powder patterns of the clear rims to the black megacrysts gives some support to the idea that the clear felspar formed from the black felspar when the latter was heated naturally in its host dyke.

ii) Single crystal X-ray photographs.

Single crystal X-ray photographs were taken of small fragments of black and clear felspar separated from the same plagioclase crystal (49332). The hol and okl planes of the reciprocal lattice of the black felspar were photographed using CuKα

radiation. On these two photographs (Plate 2) all the reciprocal constants could be measured directly and the following values were calculated for the unit cell:

a = 8.172 Å	α = 93°27′
b = 12.84 Å	β = 115°07′
c = 14.09 Å	γ = 90°00′

Partial analysis of the crystal together with refractive index determinations gave a composition of An_{51} with a maximum variation of $\pm$ 2 % anorthite in the black felspar.

According to HINTZE (1897) β^* is 63°38′ for andesine, 63°54′ for labradorite and 63°57′ for anorthite. The angle measured (64°50′) is approximately 1° larger than would be expected in a felspar of the composition quoted. A corresponding discrepancy is found in the calculated angle β (115°07′) which HINTZE gives as 116°28′ for andesine and 116°03′ for labradorite. The angles α and γ agree closely with those given by HINTZE.

A similar set of precession photographs were taken of clear material from the crystal. Unfortunately the initial orientation of the fragment was not ideal and β^* could not be measured directly. Calculated values varied between 62°15′ and 63°40′ according to the method used. The average value for β^* was 63°17′ which differs significantly from the black felspar and is lower than the angles for andesine and labrodorite given by HINTZE. It was decided that further studies were unlikely to increase our knowledge on the origin of the anorthosite. Copious material is available should further work be of interest from a crystallographic point of view.

d) Method used in plagioclase determination

At present refractive index measurements appear to be the only widely accepted optical method of determining the composition of plagioclase. Provided that a permanent laboratory is set up for refractive index work the method is approximately as fast as universal stage determinations. Refractive index measurements are obviously ideal for the large unzoned crystals described in this paper especially as methods relying on the orientation of the optical indicatrix proved extremely unreliable. Determination of zoned plagioclase is more time consuming, however, with practice at least the outer and inner parts of a plagioclase can be taken out of a thin section using a needle and determined in the normal manner.

The method described below has been developed from the work of TSUBOI (1923) and MICHEELSEN (1957). It relies on the use of a monochromotor. This instrument considerably reduces the time taken for each determination and provided that measurements are made in the range 6000 Å–5000 Å it is thought to be more accurate than several recently described procedures.

TSUBOI's classic work on the determination of plagioclase using cleavage flakes is widely known but his equally important work on the dispersion of plagioclase is neglected. MICHEELSEN has developed an elegant immersion method of measuring the refractive indices of a mineral using optical glass as an internal standard. This has considerably increased the accuracy of refractive index determinations and has provided a rapid technique by which mineral dispersion curves can be drawn with considerable accuracy.

i) Summary of the MICHEELSEN glass method.

The most significant feature of the MICHEELSEN glass method is that the refractive index of the immersion liquid is determined at the same time as the refractive index of the mineral studied. This eliminates many of the errors inherent in refrac-

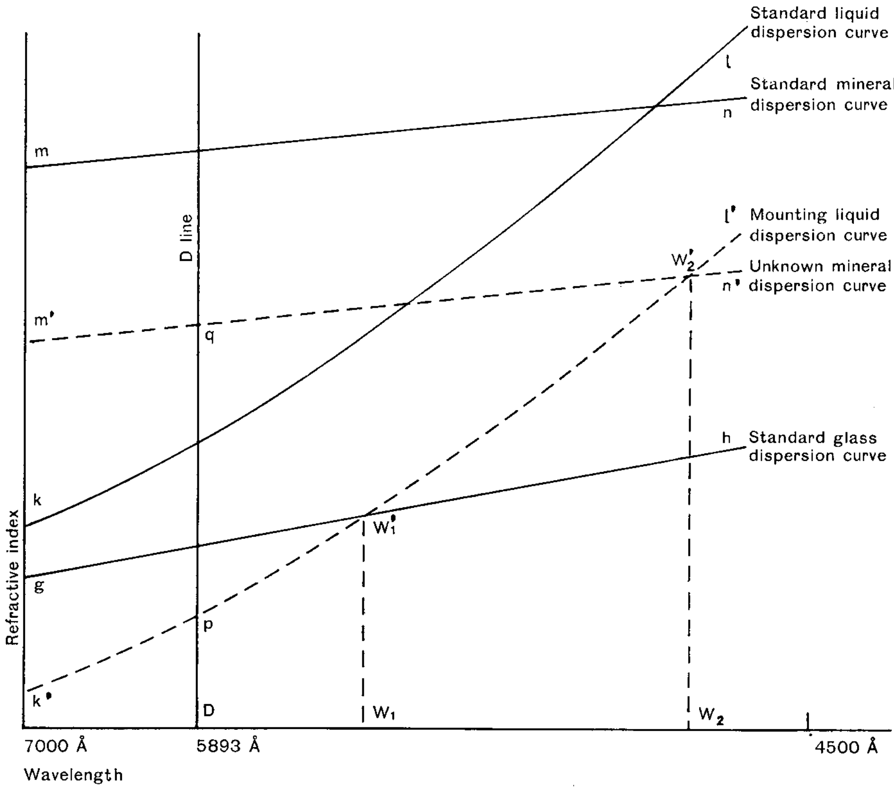


Fig. 77. Diagram to illustrate the MICHEELSEN glass method of refractive index determination.

tive index determinations due to such factors as local heating of the mount while on the microscope stage.

The mineral powder is mixed with a standard optical glass powder of known dispersion and with a refractive index close to that of the mineral. Both are immersed in a liquid in the series paraffin oil-monobromo-napthalene and are examined using light from a variable monochromatic light source. The refractive indices of the liquid and mineral are determined by noting the wavelengths at which the Becke line disappears, first between standard glass and liquid and second between liquid and mineral. Knowing these two wavelengths and the dispersion of the standard glass, the dispersion of the mounting liquid and the dispersion of a mineral similar to the mineral studied (called the standard mineral below), the refractive index of the unknown mineral can be found graphically (Fig. 77, modified from MICHEELSEN, 1957, fig. 1, p. 180).

Procedure

1) W'_1 is marked on the standard glass dispersion curve at the wavelength W_1 at which standard glass and the mounting liquid have the same refractive index.

2) The dispersion curve of the mounting liquid (k'l') is constructed through W'_1 parallel to the dispersion curve of the nearest standard (k l). The intersection (point p) of the curve k'l' with the D line gives the refractive index of the mounting

liquid. (If the difference between the refractive index of the mounting liquid and that of the standard liquid is more than 0.005 then the curve k'l' should be interpolated between two standard curves. This is because the slope of the dispersion curves of the liquids increases as their refractive index at D increases).

3) The point W'_2 is marked on the k'l' curve at the wavelength W_2 at which mounting liquid and the mineral to be determined have the same refractive index.

4) The dispersion curve of the mineral to be determined (m'n') is drawn through W'_2 parallel to the dispersion curve of the standard mineral (m n). The intersection of the curve m'n' with the D line at point q gives the refractive index for D light of the mineral to be determined.

If the dispersion of the mineral is unknown this may be found simply by determining the wavelengths W_1 and W_2 several times using the same mixture of glass and mineral and a series of liquids. This is most conveniently carried out by using a heating cell, a slight rise in temperature will lower the refractive index of the liquid appreciably without affecting those of the mineral and glass significantly. The dispersion curve of the mineral and its refractive index can be determined by drawing a curve through the points W'_2 found for each combination of glass, liquid and mineral.

The following practical details may prevent a considerable waste of time:

1) If the refractive index of the mineral is completely unknown an approximate determination should be made in white light. The mounting liquid used in the final determination shoule have a refractive index no more than 0.001 above or 0.005 below that of the mineral.

2) The optical directions in the mineral can either be selected from cleavage fragments or by orientation of a grain on the universal stage. If the second method is used it is generally necessary to mount the minerals in a suitable medium before placing on the universal stage. In the present study a mixture of shellac and alcohol was allowed to dry on a microscope slide. While still sticky plagioclase grains plus glass were shaken onto the surface. Sufficient grains remained mounted to allow manipulation on the universal stage. If a three or four axis stage is used it is vital to check that the lower polarisor is parallel to the E–W axis of the stage when the determination is made.

ii) Modification of the Micheelsen method applied to plagioclase determination.

If a large number of plagioclase determinations are to be made using glass and liquid within a narrow range of refractive indices the graphical method outlined above can be simplified so that the refractive index of the liquid can be read directly knowing the wavelength W_1 (see Plate 5). On Plate 6 the refractive index of the mineral can be determined if the wavelength W_2 and the refractive index of the liquid determined from Plate 5 are known.

From Micheelsen's method described above it can be seen that when the wavelength W_2 is known for any combination of a standard glass and a member of the paraffin oil-monobromo-napthline series then the refractive index of the liquid is also known. Plate 5 shows curves relating W_1 with the refractive index of the mounting oil using the dispersion curves of five standard glasses given by the manufacturers and the dispersion curves of 5 standard liquids.

Construction of Plate 5.

Fig. 78 shows the relationship between the dispersion curves of two standard liquids and a standard glass. The dispersion curve of the liquid that has the same

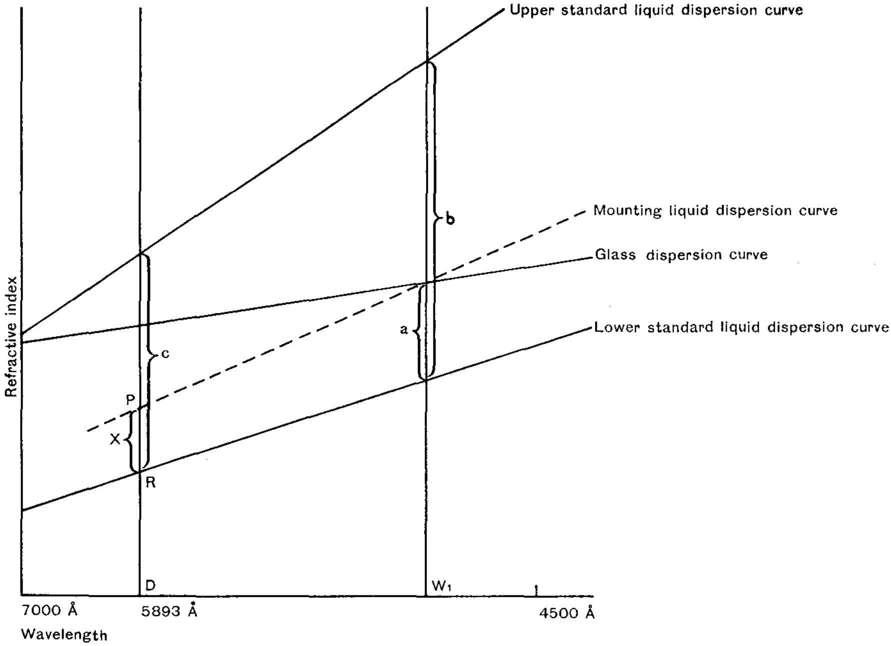


Fig. 78. Diagram to illustrate the method of construction of Plate 5.

refractive index as the standard glass at a given wavelength W_1 intersects the D line (5893 Å) at P. The refractive index of P may be expressed as R + X where R is the refractive index of the lower standard liquid at D and X is the distance R to P. The value of X can be calculated at varying values of W_1 from the relationship $X = a\frac{c}{b}$, where a is the difference between the standard glass and lower standard liquid at W_1, b is the difference between the standard liquid dispersion curves at W_1 and c is the difference between the standard liquid curves at D.

Plagioclase-liquid curves.

Similar curves can be constructed relating the wavelength at which the mineral to be identified and the now determined mounting liquid have the same refractive index (W_2) with the refractive index of the mineral at D. Plate 6 shows the result of constructing these curves using four standard liquid dispersion curves and the data publishing by TSUBOI for the dispersion of labradorite. According to TSUBOI the dispersion of X, Y and Z are not significantly different, therefore only the data for X were used to construct the curves. For convenience a range of curves corresponding to liquids at 0,001 intervals have been interpolated from the four original curves drawn with slightly heavier lines in Plate 6.

To determine the refractive index of plagioclase from Plate 6.

1) Using the result determined from Plate 5 chose the appropriate curve which cuts the D line at the refractive index of the mounting liquid.

2) The intersection of this curve with the wavelength (W_2) at which plagioclase and mineral have the same refractive index gives the refractive index of the mineral in sodium light.

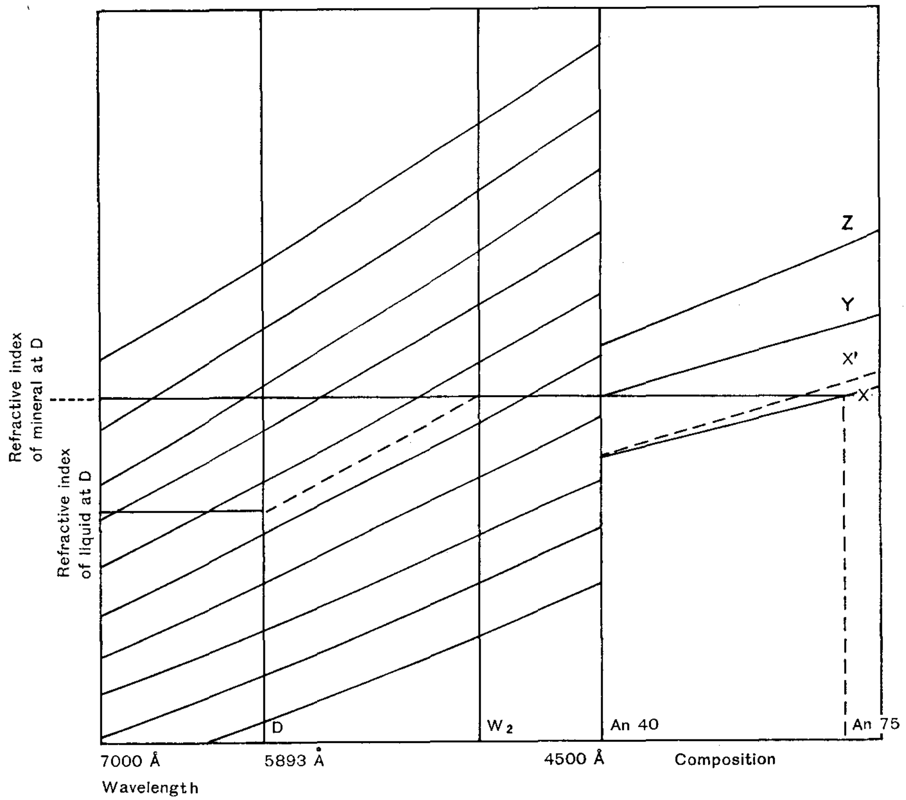


Fig. 79. Diagram to explain the use of Plate 6.

If standard glasses are not available to measure the refractive index of the liquid during the mineral determination Plate 6 can still be used with values for the refractive index of the liquid obtained by other methods. This, however, reduces the potential accuracy of the method.

Hess' curves for the determination of anorthite content are represented on the right side of Plate 6. All the felspars from the anorthosites in South Greenland produced well formed cleavage fragments when crushed by percussion and Tsuboi's cleavage fragment method was perfectly satisfactory for rapid determinations. In order to maintain consistency in the results obtained by measuring X and X' Tsuboi's original curve has been modified in Plate 6 so that the X refractive index quoted by Tsuboi for a felspar used in constructing his curves is assumed to have the same composition as a felspar with an identical X refractive index on Hess' curves.

Accuracy of the method.

The refractive index of a given grain can be determined within limits of ± 0.0002. This is equivalent to a relative change of less than 1 % anorthite and is well within the graphical accuracy of most determinative curves. Absolute error in the determination of anorthite content is probably in the region of ± 2 % An, due to the differences in refractive indices shown by felspars with identical chemical compositions and the effect of other constituents, notably potash felspar. In the present study there seems no reason why there should be a relative error of more than 1 % An between different samples.

Table 8a. *Dispersion curves of 2 labradorite felspar specimens* (determined by T. Jørgart).

G.G.U. 25171

Wavelength (Å)	R.I. X′	R.I. Z′
6070	1.5547	..
5710	..	1.5623
5540	1.5576	..
5240	..	1.5640
4940	..	1.5664

Clear margin to black felspar
G.G.U. 49322

Wavelength (Å)	R.I. X′	R.I. Z′
6110	1.5571	..
5740	1.5589	..
5360	1.5611	..
5230	..	1.5669
5130	1.5628	..
5020	..	1.5684
4910	1.5646	..
4660	..	1.5713
4540	..	1.5726

Table 8b. *Dispersion curves of plagioclase glass.*

G.G.U. 49324

Wavelength (Å)	R.I.
6500	1.5320
5890	1.5343
5150	1.5381
4680	1.5415

Data used.

1) Felspars.

The dispersion curves for labradorite were taken from Tsuboi (1923, p. 102). As Tsuboi's original work showed that the dispersion curves of X, Y and Z are essentially parallel in felspar only the figures for X were used. As the optical properties of the plagioclase from the South Greenland anorthosites are anomalous measurements of the dispersion of two specimens were carried out. The results (kindly determined by T. Jørgart) are given above (Table 8a). No significant difference can be seen between these results and those of Tsuboi.

The dispersion curves of a series of plagioclase glasses (Schairer, Smith and Chayes, 1956) were determined and a set of typical results is given in Table 8b. The plagioclase glasses were made by heating powdered felspar wrapped in platinum foil in an air-gas burner. In most cases the dispersion curves obtained for the plagioclase glass was similar to the dispersion curves of plagioclase crystal grains but a few single glass grains showed variable properties from one part of the grain to another so that it was impossible to obtain an accurate refractive index using a monochromator. This variability in properties within one grain is thought to be caused by the rapid cooling necessary to obtain a felspar glass. Quenching of this kind frequently leads to optical inhomogeneities in glasses caused by strain. The lack of consistency in the dispersion curves of plagioclase glass suggests that the method cannot be used for the most precise determination of plagioclase composition.

2) Liquids.

The dispersion curves of the paraffin oil-monobromo-napthaline series of liquids were kindly supplied by H. Micheelsen. He obtained the curves using a hollow prism and a precision goniometer and the results are believed to be accurate to $\pm$ 0.00002. A paper describing the method and the results is in preparation by Micheelsen.

3) Glasses.

The dispersion curves of the standard glasses were given by their makers. The following glasses were used:

Veb Carl Zeiss Jena.

PSK 3 70337, nd 1.55284; BAK 5 73641, nd 1.55616; SK 20 Pt 59350, nd 1.55867; BAK 7 Pt 510297, nd 1.56366.

Chance-Pilkington. Optical works, St. Asaph, Flintshire. NBC 569558, nd 1.56760.

The dispersion curves were plotted on a Hartmann dispersion net No. 397 1/2 supplied by Schleicher & Schüll, Einbeck Hannover. Permission to reproduce part of this net in Plates 5 and 6 is gratefully acknowledged. I would also like to thank Chr. Halkier (Geological Institute, Copenhagen) for his advice during the preparation of Plates 5 and 6.

DANSK RESUMÉ

Anorthosit-xenoliter og plagioklas-megakryster optræder i intrusive Gardar-bjergarter (1000–1250 millioner år) i Sydgrønland mellem 60°30' og 61°30' nordlig bredde (Plate 4). De feldspatrige inklusioner er koncentreret i gange, der findes i et ca. 70–150 km bredt øst-vestgående bælte, som fra kysten nær Nunarssuit strækker sig ca. 200 km mod øst til Indlandsisen. Endvidere er sporadiske forekomster kendt fra Grønlands østkyst. Enkelte isolerede fund af feldspatrige inklusioner kendes fra større Gardar-plutoner f. eks. fra lujavriterne i Ilímaussaq og fra grænsezonebjergarter i Igalikokomplekset. Alle hovedforekomster af feldspatrige inklusioner, findes inden for et område, der påvirkes af nogle store øst-vestgående forkastninger med horisontale forsætninger. Disse forkastninger kontrollerer formodentligt også tilstedeværelsen og fordelingen af de større alkaline Gardar-komplekser.

Størstedelen af de feldspatrige inklusioner findes i gabbroide gange, der er intruderet i en tidlig fase af mellemste-Gardar. Inklusionerne er især almindelige i ØNØ-gående gabbroer syd for Kobberminebugt, og i en gabbroid gangsværm der med ØNØ-retning kan følges fra Tugtutôq til Indlandsisen nord for Igaliko-syenitkomplekset. I begge områder efterfølges gabbrogangene af mere alkaline trakydolerit- og mikrosyenitgange. Disse igen følges af de store plutoniske alkaline intrusioner, der er dannet i sen-Gardar.

I gangene findes en distinkt sammenhæng mellem inklusionstyper og værtsbjergarter (se Fig. 49, side 82). Olivingabbro indeholder store anorthosit-xenoliter og spredte små basiske labrador-megakryster. Mindre basiske alkaligabbrogange indeholder store Na-rige labrador-megakryster samt aggregater bestående af megakryster og fragmenter af anorthosit-xenoliter. Trakydoleriter indeholder mindre andesin- megakryster og små omdannede anorthositstykker. I syenitiske gange er kun fundet få inklusioner.

Blandt anorthosit-xenoliterne findes 3 hovedvarieteter: granular anorthosit, lamineret anorthosit og breccieagtige aggregater kaldet sekundær anorthosit.

De granulare anorthosit-xenoliter er de mest udbredte. De er især almindelige i området syd for Kobberminebugt, hvor enkelte inklusioner kan nå størrelser på adskillige tusinde kvadratmeter. De er grovkornede med anhedrale plagioklaskorn og indeholder interstitielt mafisk materiale (Plate 1). Stedvis kan xenoliterne indeholde lagdelte partier, og i de mindre feldspatholdige lag har feldspaten ofte en tabular form. Sammensatte inklusioner, der viser en såkaldt blokstruktur, er hyppige. Disse består af plagioklasrige aggregater indesluttet i en xenolitvarietet, der har et lidt større indhold af mafiske bestanddele. De mafiske mineraler udgør sjældent mere end 10 $^{0}/_{0}$ af det granulare anorthositmateriale i xenoliterne. Uomdannede xenoliter er sammensat af labrador (An_{61}), olivin ($Fa_{27\text{-}40}$), magnetit med ilmenitlameller, sjældne ortopyroxenindivider ($Fs_{28\text{-}40}$) og enkelte klinopyroxenkorn.

De laminerede anorthosit-xenoliter har en mere begrænset udbredelse. De er særligt hyppige i den nordligste af de meget brede gabbrogange på Assorutit på østkysten af Tugtutôq, i gabbrointrusionen ved Narssaq og i marginalbjergarter i Igalikointrusionen. De formodes at repræsentere tidligere lagdelte dannelser, der er associeret med de gabbrosyenitkomplekser, der findes på de samme lokaliteter. De laminerede anorthositer er tydeligt lagdelte med feldspatrige bånd bestående af vellaminerede tabulare plagioklaskrystaller alternerende med mere mafiske bånd, hvori plagioklaslamineringen ikke er særligt godt udviklet. Plagioklaserne er moderat zonare fra ca. $An_{62\text{-}65}$ til ca. $An_{45\text{-}50}$. Interstitiel alkalifeldspat er normalt til stede. Det dominerende mafiske mineral er olivin ($Fa_{27\text{-}33}$), og endvidere findes lidt klinopyroxen ($Ca_{40}Mg_{30}Fe_{30}$) og interstitiel ilmenit.

De sekundære anorthositinklusioner er almindelige i store gabbrogange. Særlig tydeligt ses de ved Bangs Havn på det østlige Nunarssuit og ved Narssaq (Fig. 6, side 25). Begge steder er xenoliterne dannet ved akkumulation af 1) itubrudte dele af granular anorthosit, 2) noget lamineret anorthosit og 3) et stort antal plagioklas-megakryster fra en tidlig gabbroid fase af inklusionernes værtsbjergart. Xenolitkomponenterne blev herefter brækket i stykker i en yngre gabbroid fase, hvori de nu danner de "sammensatte xenoliter", der akkumuleredes i den øverste del af de gangformede gabbrointrusioner.

Plagioklas-megakrysterne varierer i størrelse fra få centimeter til ca. 2 meter i længde. De er almindeligvis euhedrale (Fig. 12, side 29). Deres gennemsnitlige kornstørrelse overstiger størrelsen af de plagioklasindivider, der findes i såvel den granulare som den laminerede anorthosit, og der er intet tegn på, at de skulle være dannet ved en mekanisk nedknusning af tidligere dannede bjergarter. De fleste af megakrysterne er mindre basiske end den plagioklas, der findes i xenoliterne; imidlertid kendes en undtagelse herfra, idet en Gardar-gabbro fra Tôrnârssuk indeholder subhedrale bytownit-megakryster (An_{65-80}), olivin-megakryster (Fa_{27}) og ortopyroxen-megakryster (Fs_{22}). Disse mineraler er stedvis samlet således, at de danner små aggregatiske anorthosit-xenoliter.

De største megakryster er fundet i alkalirige gabbroer, hvori plagioklas-megakrysternes sammensætning spænder fra An_{50} til An_{58}. Normalt er de ikke zonare, når der ses bort fra en meget tynd rand af feldspatmateriale (An_{35-40}), der stammer fra værtsbjergarten. Megakrysterne i trakydoleriterne er almindeligvis mindre, idet de i gennemsnit kun er få centimeter lange. De er ofte stærkt sericitiserede og omgivet af en rand af oligoklas og alkalifeldspat, der stammer fra værtsbjergarten. I nogle tilfælde danner de kerner i rhombeformede alkalifeldspatstrøkorn. Deres oprindelige sammensætning når fra An_{35} til An_{58}, og almindeligvis er de små krystaller de mindst basiske.

De fleste megakryster er lyse og enten gennemsigtige eller sericitiserede, men på nogle lokaliteter er megakrysterne sorte. Dette gælder især i alkaligabbrogangene i Kobberminebugt og i randzonebjergarter i Igalikointrusionerne. I hovedparten af gangene synes den sortfarvede feldspat at have været ustabil, da den er omkranset af en lysfarvet feldspatrand (Plate 1b), men i værtsbjergarter, der indeholder primær hornblende, er den sortfarvede feldspat stabil. Det antages, på grundlag af resultaterne fra magnetiske og thermogravimetriske undersøgelser, at farvningen skyldes den tilstand, hvori jern findes i feldspaten. Denne tilstand tænkes at være kontrolleret af den vandmængde, der var til stede i magmaet da plagioklasen krystalliserede.

Spektrografiske strontium- og bariumanalyser af feldspatinklusionerne viser ingen signifikante forskelle mellem xenoliterne, megakrysterne eller andre plagioklaser af lignende sammensætning fra gabbroide Gardarbjergarter (Fig. 52 & 53, side 89). Feldspat fra de granulare anorthositer indeholdt mellem 800–1500 p.p.m. Sr og 200–250 p.p.m. Ba. Megakryster fra gabbroide værtsbjergarter viste lignende strontium- og bariumindhold, medens de der fandtes i trakydoleritiske værtsbjergarter indeholdt mellem 1600–2500 p.p.m. Sr og 800–900 p.p.m. Ba. Den laminerede anorthosit-xenolit fra Assorutit og megakryster fra randzonebjergarter i Igalikointrusionerne indeholder noget større mængder af disse elementer, idet Sr-indholdet udgør fra 1700–3300 p.p.m., og Ba findes i 800–2800 p.p.m. Lignende værdier for feldspaternes Sr- og Ba-indhold fandtes i disse inklusioners værtsbjergarter. Strontiumindholdet holder sig konstant ud gennem omdannelseszoner i megakrysterne, medens bariumindholdet viser en markant stigning nær megakrysternes kontakter til den alkalibetonede værtsbjergart.

Både megakryster og feldspaten fra anorthosit-xenoliterne udviser anomale egenskaber, når de optiske orienteringer måles på drejebord og ved anvendelse af røntgenpulveranalyser. Enkeltkrystal presessionfotografier af både klar og sort feldspat fra samme megakryst viste anomaliteter. Dette var særlig tydeligt i en prøve af sort feldspat, i hvilken målingen af β^* gav 64°54', hvilket er udenfor det interval, der ellers er beskrevet for plagioklas. Det er umuligt at bestemme feldspaternes sammensætning ved hjælp af sædvanlige målinger af de optiske orienteringer.

Værtsbjergarterne udviser ingen markante kemiske forskelle fra andre Gardar-bjergarter. De megakrystholdige ganges Ca-indhold er en lille smule mindre end andre Gardar-bjergarters, hvilket kan antyde, at megakrysterne i det mindste delvis er dannet i værtsbjergartens magma. En særlig gruppe gange, i hvilke der findes gradvise overgange mellem de tidligt dannede inklusionsfri mikrosyenitiske randzoner og de inklusionsrige sent dannede trakydoleritiske og gabbroide centrale dele (Plate 3), tænkes at være udskilt ved differentiation fra et almindeligt basaltisk magma. Differentieringen antages at være fremkommet ved en separation af de mere alkalirige bestanddele i magmaet, endnu medens dette var flydende. Fordelingen af inklusionerne i disse gange forklares ved at antage, at den labradoritiske feldspat ville synke i et syenitmagma, men forblive flydende i et gabbroidt magma.

Anorthositerne sammenlignes med en række større anorthositlegemer i den øvrige verden. Selvom det kan vises, at mange af xenoliterne har egenskaber, der ligner de man kender fra anorthosit i prækambriske orogene bælter, så er der alligevel ingen grund til at antage, at xenoliterne stammer fra en præ-Gardar anorthositmasse i dybet under Sydgrønland. Xenoliterne betragtes som breccierede fragmenter af et anorthositisk differentiat, der samledes som et næsten monomineralt tagdække over Gardar-magmaet. Anorthositerne tænkes at være krystalliseret i et basaltisk magma, der i kemisk sammensætning står nær en udifferentieret Gardar-basalt. De udkrystalliserede labradorstrøkorn vil flyde op i det basaltiske magma, hvilket vil forårsage den omtalte plagioklasakkumulation nær magmaets top. En tilsvarende dannelse af mafiskbetonede lag kan have fundet sted, men i så fald må denne have resulteret i en ansamling af de mafiske bestanddele under det niveau, hvorfra senere Gardar-magma frigøres. Det foreslås, at Gardar-magmaet i dybet kunne have været lagdelt, endnu medens det var flydende, med koncentration af alkalier i den øvre del. Dette kunne forklare den overvægt af alkalibjergarter, som findes i forhold til basiske bjergarter i hovedintrusionerne i Gardar-provinsen.

Ved den fremadskridende differentiation i Gardar-magmaet bliver vægtfyldeforskellen mellem magma og feldspat mindre, og på et tidspunkt forbliver plagioklaskrystallerne blot svævende i stedet for at flyde opad. Dette resulterede i dannelsen af store Na-rige labradorkrystaller, der aldrig blev pakket sammen til en fast anorthotitisk bjergart. Dannelsen af sort feldspat tænkes at indikere forhold, hvor anselige mængder H_2O er blevet tilbageholdt i Gardar-magmaet under anortosittagdækket.

I et appendix præsenteres et sæt kurver (Plate 5 og 6) til bestemmelse af plagioklas, ved hjælp af præcise lysbrydningsmålinger baseret på "MICHEELSENS glasmetode".

РЕЗЮМЕ

Гардарский интрузивный комплекс Южной Гренландии (1000–1250 млн. лет), залегающий между 60°30′ и 61°30′ северной широты, характеризуется включениями ксенолитов анортозита и мегакристаллов плагиоклаза. Основная масса включений приурочена к области сдвигов восточно-западного направления, по всей вероятности повлиявших на распространение главных щелочных комплексов гардарской магматической серии. Основная масса полевошпатовых включений сосредоточена в дайках, образующих пояс шириной от 70 до 150 км. и протяженностью около 200 км. от западного побережья (район Нунарссуит) на восток до материкового ледника. Спорадически они встречаются в породах восточного побережья Гренландии и в главных гардарских интрузиях, например, в луявритах щелочного массива Илимауссак и в краевых породах комплекса Игалико.

Габбровые дайки, содержащие главную массу полевошпатовых включений, сформировались в течение ранних фаз середины гардарской магматической деятельности. Из них наиболее богаты включениями дайки ВСВ простирания (к югу от Кобберминбугт) и серия габбровых даек того же направления, протянувшаяся по всей длине острова Тугтуток до метерикового ледника, к северу от сиенитового комплекса Игалико. На обеих территориях ранние среднегардарские оливин-габбровые дайки сопровождаются более щелочными трахидолеритовыми и микросиенитовыми дайками, за которыми следуют главные щелочные интрузии позднего Гардара.

Между составом вмещающих даек и типом полевошпатовых включений существует четкая связь (фиг. 49). Оливиновые габбро содержат крупные ксенолиты анортозита и мелкие разбросанные мегакристаллы кальциевого лабрадора. Менее основные щелочные габбровые дайки содержат крупные мегакристаллы натрового лабрадора и агрегаты мегакристаллов, а также фрагменты анортозитовых ксенолитов. Для трахидолеритовых вмещающих пород характерны мегакристаллы андезина средней величины и небольшие измененные массы анортозита. Несколько полевошпатовых включений отмечены в сиенитах.

Анортозитовые ксенолиты встречаются в трех главных разновидностях: зернистые анортозиты, слоистые анортозиты и брекчиеподобные агрегаты, именуемые вторичными анортозитами.

Наиболее распространенными являются ксенолиты зернистых анортозитов. Особенно часто они встречаются к югу от Коббермин-бугт, где одиночные ксенолиты достигают нескольких тысяч квадратных метров. Это грубозернистые породы, состоящие из ангедральных кристаллов плагиоклаза и темного мезостатического материала. Изредка наблюдается слоистость, причем, слои несколько обедненные полевым шпатом обычно содержат таблитчатый плагиоклаз. Блоковая структура характерна для пород с высоким содержанием плагиоклаза и несколько повышенным содержанием темного мезостатического материала. Содержание темных минералов редко превышает 10% состава ксенолитов зернистых анортозитов. Свежие ксенолиты сложены лабрадором (Au_{61}), оливином ($Fa_{27\text{-}40}$), магнетитом с эксолвентными пластинками ильменита и реже ромбическим ($Fs_{28\text{-}40}$) и моноклинным пироксеном.

Ксенолиты слоистого анортозита имеют более ограниченное распространение. Они встречаются главным образом в районе Ассорутит в более северной из двух гигантских габбровых даек, на восточном побережье острова Тугтуток, в габбровых интрузиях района Нарссак и в краевых породах интрузий Клоккен и Игалико. Повидимому, они представляют собой рано расслоенные породы, связанные с габбро-сиенитовыми комплексами, залегающими на тех же участках. Прекрасная слоистость этих пород обусловлена чередованием слоев с направленным положением кристаллов пластинчатого плагиоклаза с более темными слоями, в которых кристаллы плагиоклаза менее упорядочены. Плагиоклаз умеренно зонален от приблизительно $An_{62\text{-}65}$ до $An_{45\text{-}50}$. Присутствует также мезостатический калиевый полевой шпат. Из темных минералов преобладает оливин ($Fa_{27\text{-}33}$), затем отмечены моноклинный пироксен ($Ca_{40.5}\ Mg_{30}Fe_{29.5}$) и мезостатический ильменит.

Вторичные анортозиты широко распространены в крупных габбровых дайках. Особенно часто они встречаются в районах Бангс Хавн (восточная часть Нунарссуита) и Нарссак (фиг. 6). На обоих участках ксенолиты образованы скоплением в ранних габбровых породах обломков масс зернистого анортозита и, в меньшей степени, пластинчатого анортозита и большого количества мегакристаллов плагиоклаза. Все это в свою очередь образует обломки сложных ксенолитов в более молодых габбро, сосредоточенные близ кровли вмещающих интрузий.

Величина мегакристаллов плагиоклаза варьирует от нескольких сантиметров до 2 метров в длину (фиг. 12). Преобладающая величина зерен больше величины отдельных кристаллов плагиоклаза в зернистых или в слоистых анортозитах. Мегакристаллы обычно идиоморфны и не содержат признаков, указывающих на их образование в результате механического разрушения ранних пород. Большинство мегакристаллов отличается меньшим содержанием кальция, чем плагиоклаз ксенолитов. Исключение составляет гардарское габбро из Торнарссука, которое содержит мегакристаллы гипидиоморфного битовнита (An_{65-80}), оливина (Fa_{27}) и ромбического пироксена (Fs_{22}). Скопления этих минералов местами образуют небольшие анортозитовые ксенолиты. Самые большие мегакристаллы найдены в богатых щелочами габбровых породах. Их состав соответствует An_{50-58}. Для них характерно отсусттвие зональности, за исключением очень тонкой каемки, имеющей состав полевых шпатов вмещающих пород (An_{35-40}). Мегакристаллы трахидолеритовых вмещающих пород обычно меньшего размера и достигают в среднем нескольких сантиметров в длину. Они часто сильно серицитизированы и обрамлены олигоклазом и щелочным полевым шпатом вмещающих пород. В некоторых случаях мегакристаллы служат центрами в ромбах щелочных полевых шпатов. Их первоначальный состав варьирует от An_{35} до An_{58}, кристаллы меньшего размера в основном менее кальциевые.

Большинство мегакристаллов либо прозрачные, стекловидные, либо серицитизированные, хотя иногда встречаются и черные мегакристаллы. Последние особенно характерны для щелочных габбровых даек залива Кобберминбугт и для краевых пород интрузий Клоккен и Игалико. Черный цвет, повидимому, непостоянен, как явствует из большинства даек, где черный полевой шпат окружен прозрачным (табл. 1в). Однако, в дайках, содержащих первичную роговую обманку, черный цвет полевошпатовых мегакристаллов устойчив. На основании магнитного и термогравиметрического анализов можно предположить, что окраска обусловлена структурным состоянием железа, содержащегося в полевом шпате. В свою очередь это состояние зависит от количества воды во время кристаллизации плагиоклаза.

Стронциевый и бариевый спектральные анализы полевошпатовых включений не выявили значительной разницы в составах ксенолитов, мегакристаллов или плагиоклазов сходной композиции, взятых из гардарских габбровых пород (фиг 52 и 53). Содержание стронция в полевом шпате зернистых анортозитов колеблется в пределах от 800 до 1500 ч.н.м., а бария – 200-250 ч.н.м. Содержание обоих элементов в мегакристаллах габбровых пород примерно такое же, а в мегакристаллах трахидолеритов содержание стронция колеблется от 1600 до 2500 ч.н.м. и бария – от 800 до 900 ч.н.м. В ксенолите слоистого анортозита из Ассорутита и в мегакристаллах краевых пород интрузий Клоккен и Игалико содержание этих элементов выше: 1700-3300 ч.н.м. стронция и 800-2800 ч.н.м. бария. Анализы вмещающих пород дали сходные результаты. Содержание стронция в мегакристаллах остается неизменным через все зоны изменения, тогда как содержание бария заметно увеличивается близ контактов мегакристаллов с щелочными вмещающими породами.

Как мегакристаллы, так и полевые шпаты анортозитовых ксенолитов показали аномальные свойства на дебаеграммах и при определении оптической ориентировки на универсальном столике Федорова. Сильная аномальность проявлена также на фотографиях прецессии кристаллической решетки единичных кристаллов прозрачного и черного полевого шпата, взятых из одного и того же мегакристалла. Определение состава полевых шпатов с помощью обычных методов измерения оптической ориентировки оказалось невозможным.

В сравнении с другими гардарскими породами описываемые вмещающие породы не проявляют заметных химических аномалий, за исключением даек с включениями мегакристаллов, которые характеризуются несколько пониженным содержанием кальция. Последнее, возможно, объясняется тем, что мегакристаллы (хотя бы частично) образовались из той же магмы, что и вмещающие породы. В частности, для одной группы вмещающих даек характерен переход от ранних, свободных от включений микросиенитовых краевых пород к богатым включениями поздним трахидолеритовым и габбровым центрам. Можно предположить, что эта дифференциация произошла из обычной базальтовой магмы путем отделения жидких фаз (табл. 3). Распределение включений в этих дайках можно объяснить предположением, что лабрадор оседал в сиенитовой магме, но оставался взвешенным в габбровой магме.

Описываемые анортозиты подобны анортозитам центров докембрийских орогенических поясов всего мира. Но несмотря на сходство, образование анортозитов в различных местах могло протекать по-разному. Нет основания считать, что рассматриваемые породы были перенесены из догардарской анортозитовой массы из глубин Южной Гренландии. В настоящей работе ксенолиты рассматриваются как брекчированные обломки анортозитовой крыши, образовавшейся на глубине поверх фракционирующей гардарской магмы. Вероятно, сами анортозиты образовались из базальтовой магмы довольно близкой по составу к нерасщепленным гардарским базальтам. Скопление плагиоклаза происходило путем флотации лабрадора к верху базальтовой магмы. Возможно, что образование темноокрашенных слоев явилось результатом дополнительного процесса, протекавшего ниже уровня, от которого выделились более поздние фазы гардарской магмы. Можно предположить, что гардарская магма была стратифицирована на глубине, еще будучи в жидкой фазе, путем направленной вверх концентрации щелочного компонента и, возможно, железа. Этим можно объяснить преобладание щелочных пород над основными в главных интрузиях Гардарской провинции.

Во время дифференциации гардарской магмы, происходившей на глубине, разница в плотности между магмой и полевым шпатом уменьшалась, и потому кристаллы плагиоклаза оставались во взвешенном состоянии. В результате образовались крупные кристаллы натрового лабрадора не уплотненные в сплошную анортозитовую массу. Образование черного полевого шпата может указывать на присутствие значительного количества воды, заключенной между анортозитовой крышей и развивающейся гардарской магмой.

В приложении к настоящей работе (табл. 5 и 6) представлены фигуры кривых по определению состава плагиоклазов на основании точного измерения показателей преломления по методу стекла Микеильсена.

REFERENCES

ALLAART, J. H. (1967) Basic and intermediate igneous activity and its relationships to the evolution of the Julianehåb granite, South Greenland. Meddr Grønland, Bd. 175, Nr. 1.

ANDERSON, A. T. and MORIN, M. (1964) Anorthosites in the Grenville of Quebec: A classification. Trans. Amer. geophys. Un., Vol. 45, 125 only.

ARMSTRONG, R. L. (1963) K-Ar dates from West Greenland. Bull. geol. Soc. Amer., Vol. 74, 1189–1192.

AYRTON, S. N. (1963) A contribution to the geological investigations in the region of Ivigtut, SW Greenland. Meddr Grønland, Bd. 167, Nr. 3.

BAILEY, D. K. and SCHAIRER, J. F. (1966) The system Na_2O-Al_2O_3-Fe_2O_3-SiO_2 at 1 atmosphere, and the petrogenesis of alkaline rocks. J. Petrol., Vol. 7, 114–170.

BALK, R. (1931) Structural geology of the Adirondack anorthosite. Miner. petrogr. Mitt., Bd. 41, 308–434.

BARAGER, W. R. A. (1960) Petrology of basaltic rocks in part of the Labrador trough. Bull. geol. Soc. Amer., Vol. 71, 1589–1643.

BARTH, T. F. W. (1954) The igneous rock complex of the Oslo region. 14: Provenance of the Oslo magmas. Skr. norske Vidensk. Akad., I. Mat.-naturv. Kl., Nr. 4.

— (1962) Theoretical Petrology. 2nd edition. New York and London: J. Wiley & Sons Inc.

BERGSTRÖM, L. (1963) Petrology of the Tjörn area in western Sweden. Sveriges geol. Unders., Ser. C, N:o 593.

BERRANGÉ, J. P. (1962) The plutonic geology of part of the Grenville province north of Lake St. John, Quebec. Unpubl. Ph.D. thesis, London University.

— (1963) *in discussion to* DEARNLY, R. The Lewisian complex of South Harris. Quart. J. geol. Soc. Lond., Vol. 119, 309 only.

— (1966a) The bedrock geology of Vatnahverfi, Julianehåb district, South Greenland. Rapp. Grønlands geol. Unders., Nr. 3.

— (1966b) Some critical differences between orogenic-plutonic and gravity-stratified anorthosites. Geol. Rdsch., Bd. 55, 617–642.

BERTHELSEN, A. (1957) The structural evolution of an ultra- and polymetamorphic gneiss-complex, West Greenland. Geol. Rdsch., Bd. 46, 173–185.

— (1960a) An example of a structural approach to the migmatite problem. Rep. 21st Intern. geol. Congr. Norden, Pt. 14, 149–157.

— (1960b) Structural studies in the pre-Cambrian of western Greenland. II. Geology of Tovqussap nunâ. Meddr Grønland, Bd. 123, Nr. 1.

— (1961) On the chronology of the Precambrian of western Greenland. *In* RAASCH, G. O. (edit.), Geology of the Arctic. Vol. 1, 329–338. Proc. 1st Intern. Symp. Arctic Geology, Toronto.

— (1962) On the geology of the country around Ivigtut, SW-Greenland. Geol. Rdsch., Bd. 52, 269–280.

BERTHELSEN, A. and BRIDGWATER, D. (1960) On the field occurrence and petrography of some basic dykes of supposed pre-Cambrian age from the southern Sukkertoppen district western Greenland. Meddr Grønland, Bd. 123, Nr. 3.

BERTHELSEN, A. and NOE-NYGAARD, A. (1965) The Precambrian of Greenland. *In* RANKAMA, K. (edit.) The Precambrian, Vol. 2, 113–262. London and New York.

BHATTACHARJI, S. and SMITH, C. H. (1964) Flowage differentiation. Science, Vol. 145, 150–153.

BLAKE, D. H., ELWELL, R. W. D., GIBSON, I. L., SKELHORN, R. R. and WALKER, G. P. L. (1965) Some relationships resulting from the intimate association of acid and basic magmas. Quart. J. geol. Soc. Lond., Vol. 121, 31–49.

BONDAM, J. (1955) Retrogradt omdannede gabbro-anorthositer fra Frederikshaab Distrikt, Sydvest Grønland. Unpubl. lecture, 2nd Nordic Geologists Winter Meeting, Oslo.

BONDESEN, E. (1960) Sermerssût-Tôrnârsuk områdets geologi. Unpubl. thesis, Copenhagen University.

BONDESEN, E. and HENRIKSEN, N. (1965) On some pre-Cambrian metadolerites from the central Ivigtut region SW Greenland. Meddr Grønland, Bd. 179, Nr. 2.

BORGEN, B. I. (1967) Analytical procedures used in the geochemical laboratory of the Survey. Rapp. Grønlands geol. Unders., Nr. 10.

BOULANGER, J. (1957) Les anorthosites de Madagascar. Comm. Coop. Tech. Afr., 2nd Meeting, Tananarive, 55–56.

— (1959) Les anorthosites de Madagascar. Annls géol. Madagascar, Vol. 26, 1–71.

BOWEN, N. L. (1917) The problem of the anorthosites. J. Geol., Vol. 25, 209–243.

BRIDGWATER, D. (1965) Isotopic age determinations from South Greenland and their geological setting. Meddr Grønland, Bd. 179, Nr. 4.

BRIDGWATER, D. and WALTON, B. J. (1964). Tectono-magmatic evolution of the Svecofennid chelogenic cycle in South Greenland. Nature, Lond., Vol. 203, 278–281.

BUDDINGTON, A. F. (1939) Adirondack igneous rocks and their metamorphism. Mem. geol. Soc. Amer., No. 7.

— (1960) The origin of anorthosites re-evaluated. Rec. geol. Surv. Ind., Vol. 86, 421–432.

BURNS, D. J. (1966) Chemical and mineralogical changes associated with the Laxford metamorphism of dolerite dykes in the Scourie-Loch Laxford area, Sutherland, Scotland. Geol. Mag., Vol. 103, 19–35.

BUTLER, J. R. and SKIBA, W. (1962) Strontium in plagioclase feldspars from four layered basic masses in Somalia. Miner. Mag., Vol. 33, 213–225.

DALY, R. A. (1933) Igneous rocks and the depths of the earth. New York: McGraw Hill.

DAVIDSON, C. F. (1943) The Archean rocks of the Rodil district, South Harris, Outer Hebrides. Trans. roy. Soc. Edinb., Vol. 61, 71–112.

DEER, W. A., HOWIE, R. A. and ZUSSMAN, J. (1962) Rock forming minerals. 5 vols. London: Longmans, Green and Co.

ECKERMANN, H. VON (1936) The Loos-Hamra region. Geol. Fören. Stockh. Förh., Bd. 58, 128–343.

ELDERS, W. A. (1957) A preliminary note on a xenolith-rich dyke from Lyngen, northern Norway. Acta boreal., A. Scient., No. 12, 1–12.

ELLITSGAARD-RASMUSSEN, K. and MOURITZEN, M. (1954) An anorthosite occurrence from West Greenland. Meddr dansk geol. Foren., Bd. 12, 436–442.

EMELEUS, C. H. (1964) The Grønnedal-Íka alkaline complex, South Greenland. The structure and geological history of the complex. Meddr Grønland, Bd. 172, Nr. 3.

EMMONS, R. C. (edit.) (1953) Selected petrogenic relationships of plagioclase. Mem. geol. Soc. Amer., No. 55.

EMMONS, R. C., CRUMP, R. M. and KETNER, K. B. (1960) High- and low-temperature plagioclase. Bull. geol. Soc. Amer., Vol. 71, 1417–1420.

EMSLIE, R. F. (1964) Potassium-argon age of the Michikamau anorthositic intrusion, Labrador. Nature, Lond., Vol. 202, 172 only.

— (1965) The Michikamau anorthositic intrusion, Labrador. Canad. Jl Earth Sci., Vol. 2, 385–399.

FAIRBAIN, H. W., BULLWINKEL, J. H., PINSON, W. H. and HURLEY, P. M. (1959) Age investigation of syenites from Coldwell, Ontario. Proc. geol. Ass. Canada, Vol. 11, 141–144.

FERGUSON, J. (1964) Geology of the Ilímaussaq alkaline intrusion, South Greenland. Description of map and structure. Meddr Grønland, Bd. 172, Nr. 4.

GOLDSCHMIDT, V. M. (1922) Stammentupen der Eruptivgesteine. Skr. Vidensk. Selsk., Oslo, I. Mat.-naturv. Kl., Nr. 10, 1–11.

GORAI, M. (1951) Petrological studies on plagioclase twins. Amer. Min., Vol. 36, 884–901.

GORBATSCHEV, R. (1961) Dolerites of the Eskilstuna region. Sveriges geol. Unders., Årsbok 55 (1961), No. 4, Ser. C, N:o 580.

GROUT, F. F. and SCHWARTZ, G. M. (1939) The geology of the anorthosites of the Minnesota coast of Lake Superior. Bull. Minn. geol. Surv., No. 28.

GROUT, F. F., SHARP, R. P. and SCHWARTZ, G. M. (1959) The geology of Cook County, Minnesota. Bull. Minn. geol. Surv., No. 39.

HAMILTON, E. I. (1964) The geochemistry of the northern part of the Ilímaussaq intrusion, S.W. Greenland. Meddr Grønland, Bd. 162, Nr. 10.

HAMILTON, W. (1965) Diabase sheets, Taylor Glacier region, Victoria Land, Antarctica. U.S. geol. Surv. Prof. Pap., 456-B.

HARGRAVES, R. B. (1962) Petrology of the Allard Lake anorthosite suite, Quebec. *In* ENGEL, A. E. J., JAMES, H. L. and LEONARD, B. F. (edits.) Petrologic studies. New York: Geological Society of America.

HARPUM, J. R. (1957) *in discussion to* BOULANGER, J. Les anorthosites de Madagascar. Comm. Coop Tech. Afr., 2nd Meeting, Tananarive, 56–60.

HARRY, W. T. and PULVERTAFT, T. C. R. (1963) The Nunarssuit intrusive complex, South Greenland. Meddr Grønland, Bd. 169, Nr. 1.

HEIER, K. S. and TAYLOR, S. R. (1964) A note on the geochemistry of alkaline rocks. Norsk geol. Tidsskr., Bd. 44, 197–203.

HENRIKSEN, N. (1960) Structural analysis of a fault in south-west Greenland. Meddr Grønland, Bd. 162, Nr. 9, 15–41.

HERZ, N. (1951) Petrology of the Baltimore Gabbro, Maryland. Bull. geol. Soc. Amer., No. 62, 976–1016.

HESS, H. H. (1949) Chemical composition and optical properties of common clinopyroxenes. Pt. I. Amer. Min., Vol. 34, 621–666.

— (1960) Stillwater igneous complex, Montana: A quantitative mineralogical study. Mem. geol. Soc. Amer., No. 80.

HIGGS, D. W. (1954) Anorthosite and related rocks of the western San Gabriel mountains, southern California. Univ. Calif. Publ. geol. Sci., Vol. 30, 171–222.

HINTZE, C. (1897) Handbuch der Mineralogie, Bd. 2. Leipzig.

KENNEDY, G. C. (1955) Some aspects of the role of water in rock melts. *In* POLDERVAART, A. (edit.) The crust of the earth. Spec. Pap. geol. Soc. Amer., No. 62, 489–503.

KENNEDY, W. Q. and ANDERSON, E. M. (1938) Crustal layers and the origin of magmas. Bull. volcan., Sér. 2, t. 3, 23–82.

KOLDERUP, C. F. and KOLDERUP, N. H. (1940) Geology of the Bergen Arc system. Bergens Mus. Skr., Nr. 20, 69–89.

KRANCK, E. H. (1961) The tectonic position of the anorthosites of eastern Canada. Bull. Comm. géol. Finl., No. 196, 299–320.

LACROIX, A. (1939) Sur un nouveau type de roches métamorphiques (sákenites) faisant partie des schistes cristallins du Sud de Madagascar. C.R. Acad. Sci., Paris, t. 209, 609–612.

LAVES, F. (1965) Mechanical twinning in acid plagioclases. Amer. Min., Vol. 50, 511–514.

LEBEDEV, A. P. and BOGATIKOV, O. A. (1963) Petrologiya Kizirskogo Gabbro – Sienitovogo Plutona (Vostochnyi ayan). Trudy Inst. Geol. rud. Mestorozhd., Petr. Miner. Geokh., 1963, vyp. 97. (in Russian).

LEBEDEV, A. P. and PAVLOV, N. V. (1957) Dzhugdzhurskii Anortozitovy Massiv. Akad. Nauk SSSR, Moscow. (in Russian)

MCANDREW, I. (1957) Calibration of a Franz isodynamic separator and its application to mineral separation. Proc. Aust. Inst. Min. Metall., No. 181, 59.

MACGREGOR, A. G. (1931) Clouded feldspars and thermal metamorphism. Miner. Mag., Vol. 22, 524–538.

MEGAW, H. D. (1961) Order and disorder. III. The structure of the intermediate plagioclase felspars. Proc. roy. Soc. Lond., Ser. A, Vol. 259, 184–202.

MICHEELSEN, H. (1957) An immersion method for exact determinations of refractive indices. The glass method. Meddr dansk geol. Foren., Bd. 13, 177–191.

— (1966) The structure of dark flint from Stevns, Denmark. Meddr dansk geol. Foren., Bd. 16, 286–359.

MICHOT, P. (1955) Anorthosites et anorthosites. Bull. Acad. roy. Belg., Cl. Sci., 5e série, t. 41, 275–294.

— (1960) La géologie de la catazone: le probleme des anorthosites, la palingenese basique et la tectonique catazonal dans le Rogaland meridional. Guide book g, Guide de l'excursion A 9, 21st Intern. geol. Congr. Norden, 1960.

MOORBATH, S. and PAULY, H. (1962) Rubidium-strontium and lead isotope studies on intrusive rocks from Ivigtut, South Greenland. *In* HURLEY, P. M. (edit.) Variations in isotopic abundance of strontium, calcium, and argon and related topics. 10th Ann. Rep. 1962, U.S. Atomic Energy Comm. Dept. Geol. Geophys. Mass. Inst. Tech., 99–102.

MOORBATH, S., WEBSTER, R. K. and MORGAN, J. W. (1960) Absolute age determination in south-west Greenland. Meddr Grønland, Bd. 162, Nr. 9, 1–13.

MORSE, S. A. (1962) The Kiglapait layered intrusion, coast of Labrador, Canada. Canad. Min., Vol. 7, 342 only.

— (1963) Mineral variation in the layered series of the Kiglapait intrusion, coast of Labrador, Canada. Spec. Pap. geol. Soc. Amer., No. 73, 208 only.

— (1964) Age of the Labrador anorthosites. Nature, Lond., Vol. 203, 509–510.

NOBLE, D. C. (1965) Determination of composition and structural state of plagioclase with the five-axis universal stage. Amer. Min., Vol. 50, 367–381.

OEN ING SOEN (1962) Hornblendic rocks and their polymetamorphic derivatives in an area NW. of Ivigtut, South Greenland. Meddr Grønland, Bd. 169, Nr. 6.

OFTEDAHL, C. (1953) Studies on the igneous rock complex of the Oslo region. 13: The Cauldrons. Skr. norske Vidensk. Akad., I. Mat.-naturv. Kl., Nr. 3.

PAPEZIK, V. S. (1965) Geochemistry of some Canadian anorthosites. Geochim. et cosmoch. Acta, Vol. 29, 673–709.

PHILPOTTS, A. R. (1966) Origin of the anorthosite-mangerite rocks in southern Quebec. J. Petrol., Vol. 7, 1–64.

POLDERVAART, A. and GILKEY, A. K. (1954) On clouded plagioclase. Amer. Min., Vol. 39, 75–91.

POLKANOV, A. A. and ELISEEV, N. A. (1941) Petrology of Gremyakha-Vyrmes pluton, Kola peninsula. Leningrad State Univ. (in Russian)

POWERS, S. (1915) The origin of inclusions in dykes. J. Geol., Vol. 23, 166–182.

PRINZ, M. (1964) Geologic evolution of the Beartooth Mountains, Montana and Wyoming. Pt. 5. Mafic dike swarms of the southern Beartooth Mountains. Bull. geol. Soc. Amer., Vol. 75, 1217–1248.

PULVERTAFT, T. C. R. (1965) The Eqaloqarfia layered dyke, Nunarssuit, South Greenland. Meddr Grønland, Bd. 169, Nr. 10.

RAO, P. N. (1964) Studies of plagioclase twinning in the anorthosites of Oddanchatram, Madurai district, south India. Proc. Ind. Acad. Sci., Vol. 60, No. 2 Sect. B, 135–144.

RICHEY, J. E., THOMAS, H. H. *et al.* (1930) The geology of Ardnamurchan, North-West Mull and Coll. Mem. geol. Surv. Scot.

SCHAIRER, J. F., SMITH, J. R. and CHAYES, F. F. (1956) Refractive indices of plagioclase glasses. Yearb. Carnegie Instn Wash., No. 55, 195–197.

SIMMONS, C. (1964) Gravity survey and geological interpretation, northern New York. Bull. geol. Soc. Amer., Vol. 75, 81–98.

SIMPSON, E. S. and OTTO, J. D. T. (1960) On the Pre-Cambrian anorthosite mass of southern Angola. Rep. 21st Intern. geol. Congr. Norden, 1960, Pt. 13, 216–227.

SKIBA, W. and BUTLER, J. R. (1963) The use of Sr-An relationships in plagioclases to distinguish between Somalian metagabbros and country rock amphibolites. J. Petrol., Vol. 4.

SLEMMONS, D. B. (1962) Determination of volcanic and plutonic plagioclases using a three- or four-axis universal stage. Spec. Pap. geol. Soc. Amer., No. 69.

SMITH, J. R. and YODER, H. S., JR. (1956) Variations in x-ray powder diffraction patterns of plagioclase feldspars. Amer. Min., Vol. 41, 632–647.

SMITH, J. V. (1956) The powder patterns and lattice parameters of plagioclase felspars. I. The soda-rich plagioclases. Miner. Mag., Vol. 31, 47–68.

SMITH, J. V. and GAY, P. (1958) The powder patterns and lattice parameters of the plagioclase felspars. II. Miner. Mag., Vol. 31, 744–762.

SUBRAMANIAM, A. P. (1956) Petrology of the anorthosite-gabbro mass at Kadavur, Madras, India. Geol. Mag., Vol. 93, 287–300.

SÆTHER, E. (1950) On the genesis of peralkaline rock provinces. Rep. 18th Intern. geol. Congr. Great Britain, 1948, Pt. 2, 123–130.

SØRENSEN, H. (1955) Anorthosite from Buksefjorden; West Greenland. Meddr dansk geol. Foren., Bd. 13, 42–45.

— (1966) På spor af sjældne metaller i Sydgrønland. II. Gardar rift-systemet. Grønland, 1966, 163–176.

TAYLOR, R. B. (1956) The Duluth gabbro complex, Duluth, Minnesota. *In* Precambrian of northeastern Minnesota. Field trip No. 1. Guidebook Series, Geol. Soc. Amer., 42–66.

— (1964) Geology of the Duluth gabbro complex near Duluth, Minnesota. Bull. Minn. geol. Surv., No. 44.

TOBI, A. C. (1961) Recognition of plagioclase twins in sections normal to the composition plane. Amer. Min., Vol. 46, 1470–1488.

TOMKEIEFF, S. I. (1937) Petrochemistry of the Scottish Carboniferous-Permian igneous rocks. Bull. volcan., Sér. 2, t. 1, 59–87.

— (1961) Alkalic ultrabasic rocks and carbonatites in the U.S.S.R. Intern. Geol. Review, Vol. 3, 739–758.

TSUBOI, S. (1923a) Optical dispersion of plagioclase. Miner. Mag., Vol. 20, 93–107.

— (1923b) A dispersion method of determining plagioclases in cleavage-flakes. Miner. Mag., Vol. 20, 108–122.

TURNER, F. J. (1951) Observations on twinning of plagioclase in metamorphic rocks. Amer. Min., Vol. 36, 581–589.

UPTON, B. G. J. (1960) The alkaline igneous complex of Kûngnât Fjeld, South Greenland. Meddr Grønland, Bd. 123, Nr. 4.

— (1961) Textural features of some contrasted igneous cumulates from South Greenland. Meddr Grønland, Bd. 123, Nr. 6, 3–31.

— (1962) Geology of Tugtutôq and neighbouring islands, South Greenland. Pt. I. Meddr Grønland, Bd. 169, Nr. 8.

— (1964) The geology of Tugtutôq and neighbouring islands, South Greenland. Pt. III. Olivine gabbros, syeno-gabbros and anorthosites. Meddr Grønland, Bd. 169, Nr. 3, 1–47.

— (1965) The petrology of a camptonite sill in South Greenland. Meddr Grønland, Bd. 169, Nr. 11.

— (1966) Ultrabasic intrusions from Narssaq and Tugtutôq. Rapp. Grønlands geol. Unders., Nr. 11, 41–44.

USSING, N. V. (1912) The geology of the country around Julianehaab, Greenland. Meddr Grønland, Bd. 38.

VANCE, J. A. (1961) Polysynthetic twinning in plagioclase. Amer. Min., Vol. 46, 1097–1119.

VOGEL, T. A. (1964) Optical-crystallographic scatter in plagioclase. Amer. Min., Vol. 49, 614–633.

WALTON, B. J. (1965) Sanerutian appinitic rocks and Gardar diatremes, north of Narssarssuaq, South Greenland. Meddr Grønland, Bd. 179, Nr. 9.

WATT, W. S. (1966) Chemical analyses from the Gardar igneous province, South Greenland. Rapp. Grønlands geol. Unders., Nr. 6.

— (1968) Petrology and geology of the Precambrian Gardar dykes on Qaersuarssuk, South Greenland. Rapp. Grønlands geol. Unders., Nr. 14.

WATTERSON, J. S. (1965) Plutonic development of the Ilordleq area, South Greenland. Pt. I. Chronology, and the occurrence and recognition of metamorphosed basic dykes. Meddr Grønland, Bd. 172, Nr. 7.

— (1968) Plutonic development of the Ilordleq area, South Greenland. Pt. II. Late-kinematic basic dykes. Meddr Grønland, Bd. 185, Nr. 3.

WEGMANN, C. E. (1938) Geological investigation in southern Greenland. I. On the structural divisions of southern Greenland. Meddr Grønland, Bd. 113, Nr. 2.

WEIDMANN, M. (1964) Géologie de la région située entre Tigssaluk fjord et Sermiligârssuk fjord (partie médiane), SW-Groenland. Meddr Grønland, Bd. 169, Nr. 5.

WELLS, M. K. (1964) Structure and petrology of the Freetown layered basic complex of Sierra Leone. Overseas Geol. Min. Resour., Bull. Supp., No. 4.

WHEELER, E. P., JR. (1960) Anorthosite-adamellite complex of Nain, Labrador. Bull. geol. Soc. Amer., Vol. 71, 1755–1762.

WILCOX, E. and POLDERVAART, A. (1958) Metadolerite dike swarms in Bukersville-Roan Mountain area, North Carolina. Bull. geol. Soc. Amer., Vol. 69, 1323–1368.

WINDLEY, B. F. (1963) The plutonic development of the Sardloq area, SW Greenland. Unpubl. Ph.D. thesis, Exeter University.

— (1965) Charnockites and pre-metamorphic anorthosites in the Fiskenaesset area of W. Greenland. Geologi, 1965, 144 only.

YODER, H. S., JR. and TILLEY, C. E. (1962) Origin of basalt magmas: An experimental study of natural and synthetic rock systems. J. Petrol., Vol. 3, 342–532.

Færdig fra trykkeriet den 7. maj 1968.

PLATES

Plate 1a

Anorthosite block from dyke 6 at Eqalugssuit taserssuat. Packed plagioclase crystals with marked preferred orientation. Note the subhedral larger crystals set in a groundmass of anhedral plagioclase.

Plate 1b

Plagioclase megacryst set in a doleritic groundmass crowded with smaller felspar fragments. The megacryst is composed of black felspar rimmed and veined by clear material. Natural size. Sample 758 $\frac{AT}{59}$.

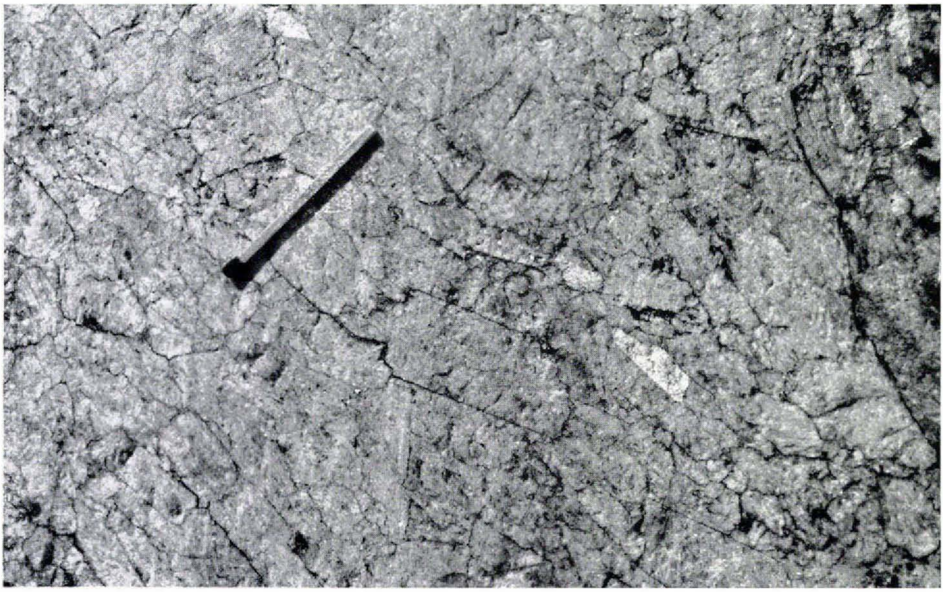


Plate 1a.

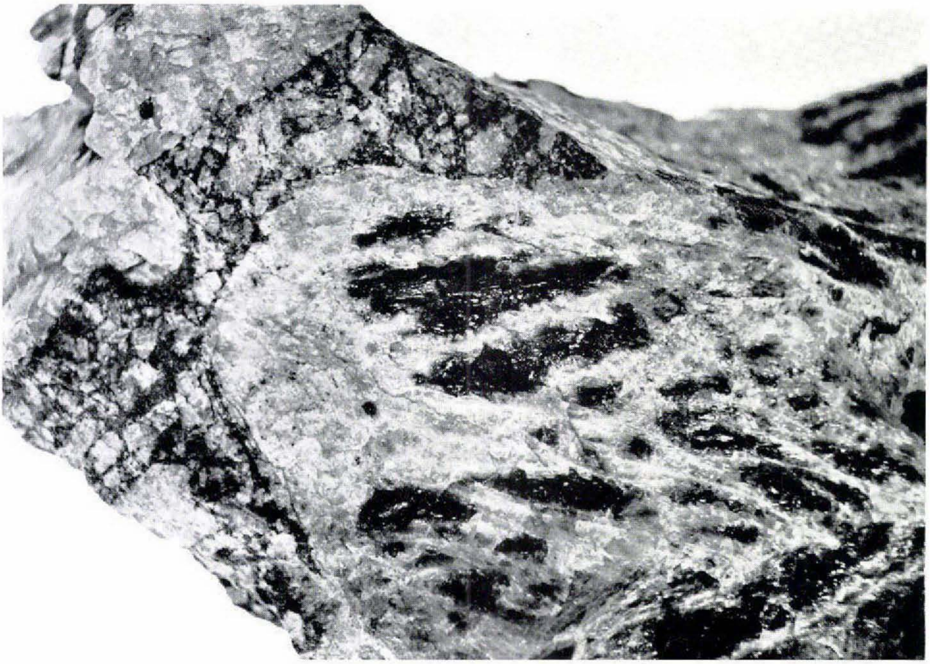


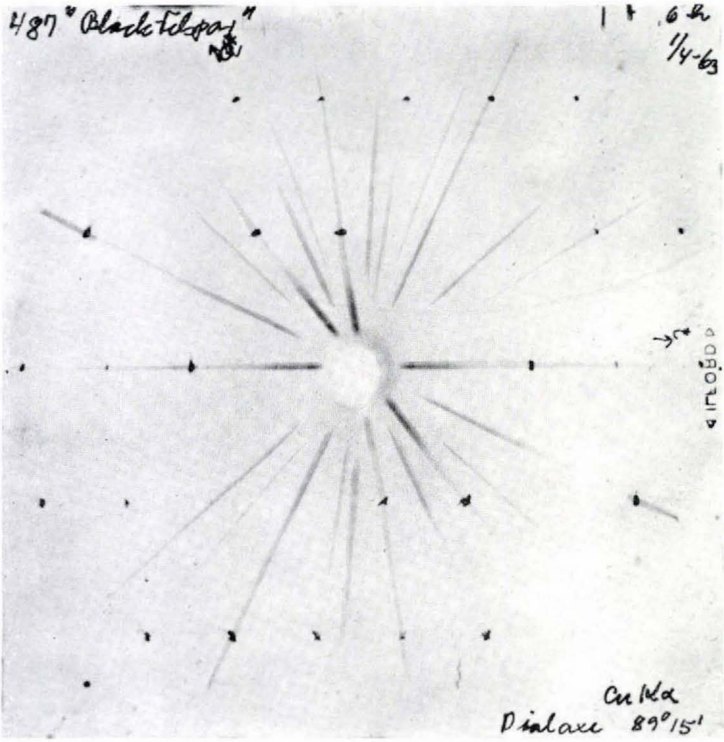
Plate 1b.

Plate 2

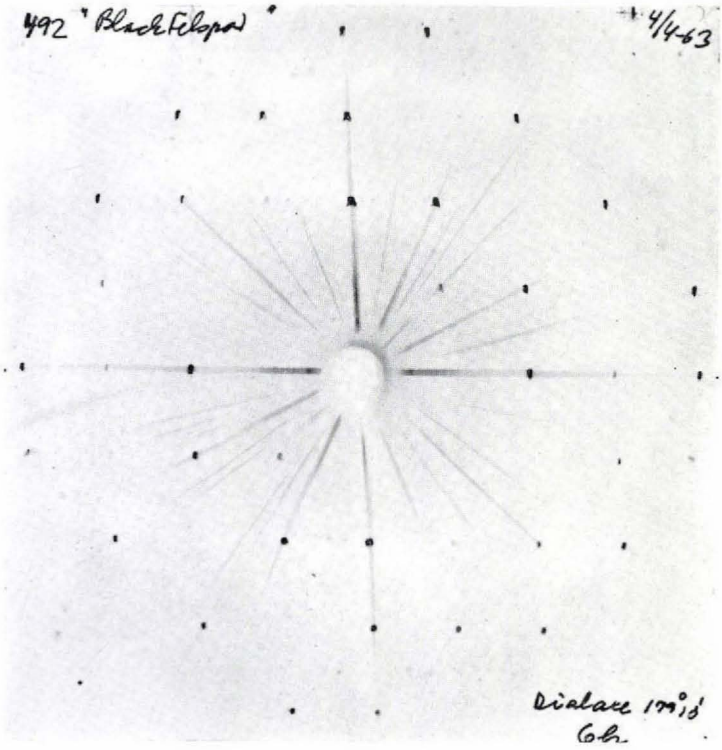
Precession camera photographs of a cleavage fragment of black felspar (sample 49332). Cu$K\alpha$ radiation.

a) a* c* Dialaxe 89°15′

b) b* c* Dialaxe 179°10′


487 "Black Felspar"
1/4-63
Cu Kα
Dialaxe 89°15'

a

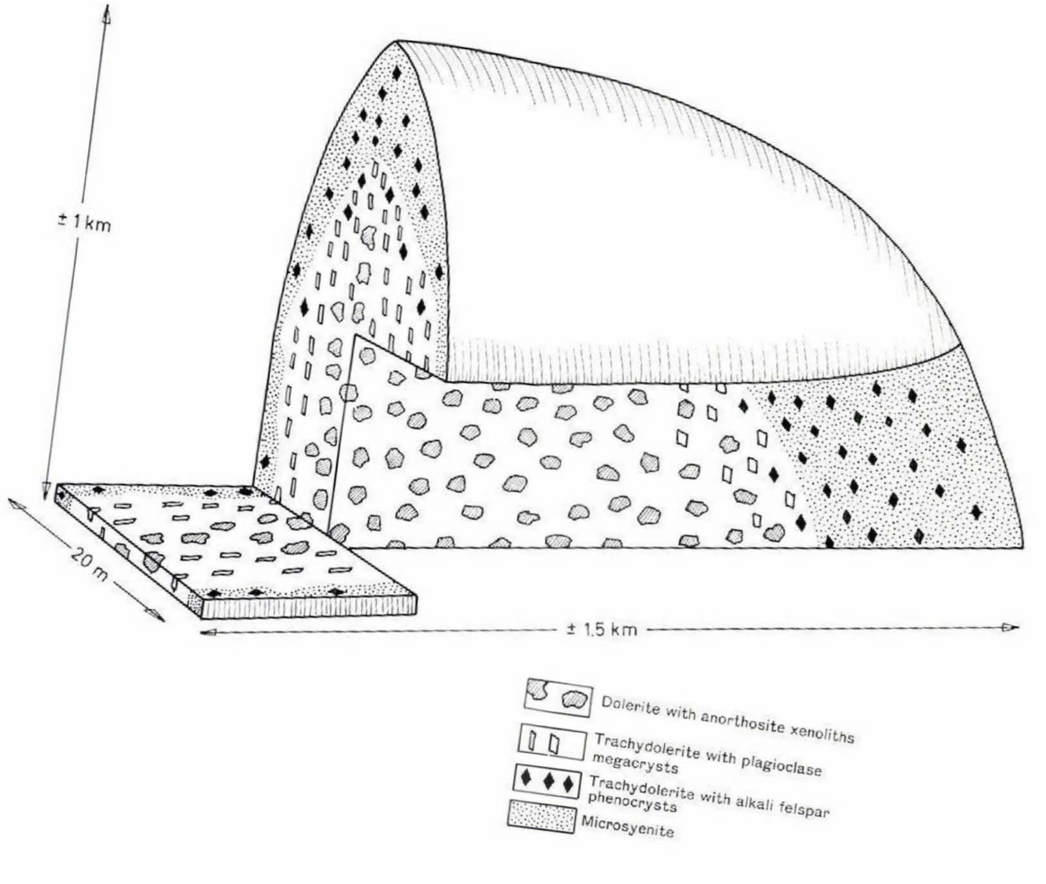

492 "Black Felspar"
4/4-63
6h

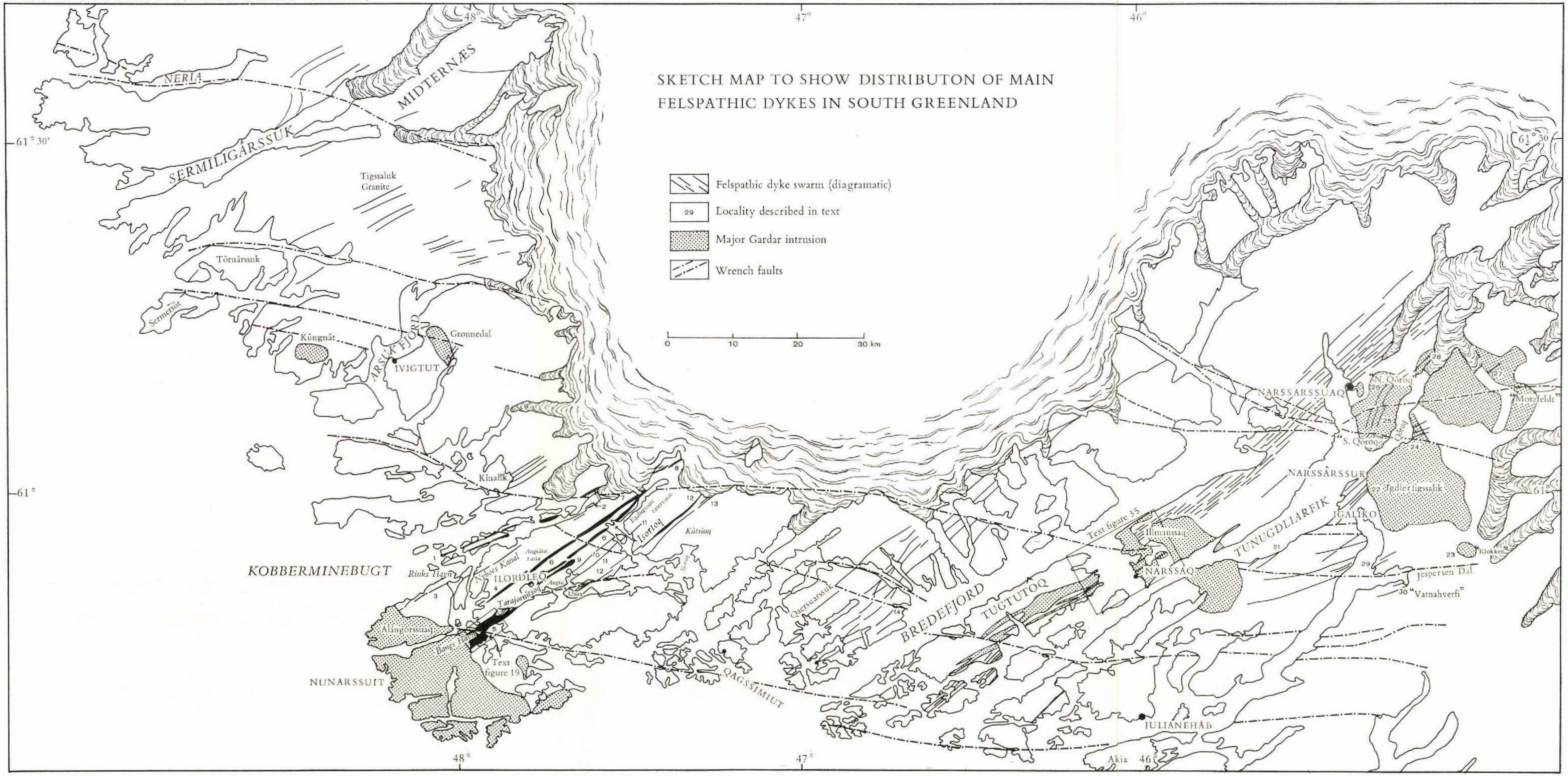
b

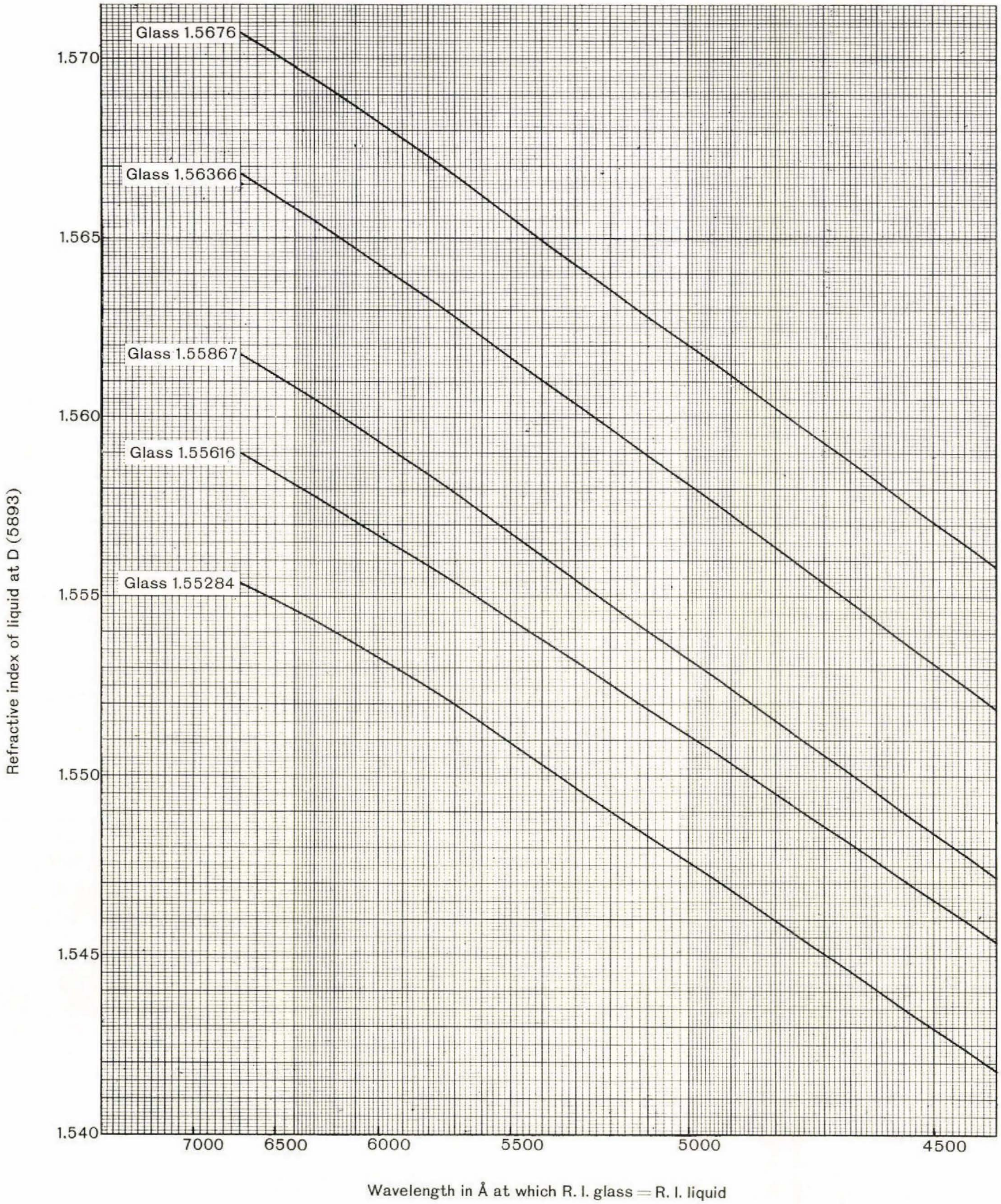
Plate 3

Block diagram to show relations between various components in a microsyenite-trachydolerite dyke. Note the early margins of microsyenite which are proportionally wider as the dyke thins and the concentration of included material in the basic centre of the dyke.

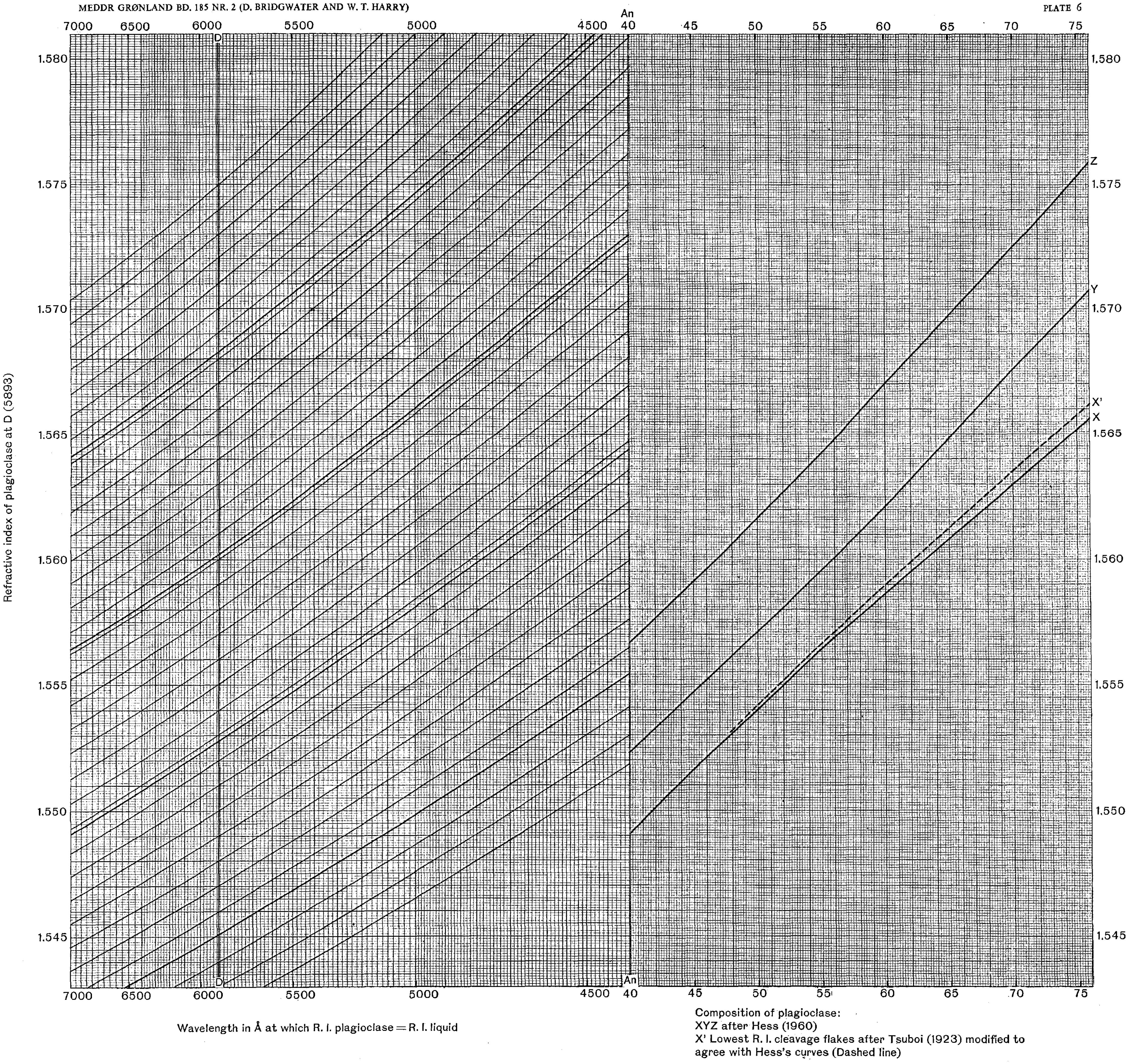
Plate 3.


± 1 km
20 m
± 1.5 km
Dolerite with anorthosite xenoliths
Trachydolerite with plagioclase megacrysts
Trachydolerite with alkali felspar phenocrysts
Microsyenite


SKETCH MAP TO SHOW DISTRIBUTON OF MAIN FELSPATHIC DYKES IN SOUTH GREENLAND
Felspathic dyke swarm (diagramatic)
29 Locality described in text
Major Gardar intrusion
Wrench faults
0 10 20 30 km
NERIA
MIDTERNÆS
SERMILIGÅRSSUK
Tigssaluk Granite
Tôrnârssuk
Sermersût
Kûngnât
ARSUK FJORD
Grønnedal
IVIGTUT
Kinalik
KOBBERMINEBUGT
Rinks Havn
Nøboes Kanal
ILORDLEQ
Tarajornitsoq
Isortoq
Kâtsiaq
Alángorssuaq
NUNARSSUIT
Text figure 19
QAGSSIMIUT
Qaersuarssuk
BREDEFJORD
TUGTUTOQ
Text figure 35
Ilímaussaq
NARSSAQ
TUNUGDLIARFIK
IGALIKO
NARSSÂRSSUK
NARSSARSSUAQ
N. Qôroq
S. Qôroq
"Motzfeldt"
Igdlerfigssalik
Klokken
Jespersen Dal
"Vatnahverfi"
JULIANEHÅB
Akia
48°
47°
46°
61° 30'
61°



Curves to determine the refractive index of the paraffin oil – monobromo-napthalene series of mounting liquids using standard glasses.



Curves to determine the refractive index and anorthite content of plagioclase grains mounted in paraffin oil – monobromo-napthalene immersion liquids.