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THE FISKENÆSSET COMPLEX, WEST GREENLAND

PART I.

A PRELIMINARY STUDY OF THE STRATIGRAPHY, PETROLOGY, AND WHOLE ROCK CHEMISTRY FROM QEQERTARSSUATSIAQ

BY

B. F. WINDLEY, R. K. HERD AND A. A. BOWDEN

WITH 40 FIGURES AND 2 TABLES IN THE TEXT, AND 2 PLATES

KØBENHAVN
1973

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Abstract

The Fiskenæsset complex, which occurs as conformable layers up to 1.5 km thick in high grade Early Archaean gneisses, has a well-preserved igneous stratigraphy. An upward sequence of lithological zones is observed in the complex as follows: 1. Pyroxene Amphibolite; 2. Ultramafic Group (hornblende-bearing spinel- and magnetite-layered dunites, peridotites, pyroxenites); 3. Layered leuco-gabbro; 4. Dark gabbro; 5. Homogeneous leuco-gabbro; 6. Anorthosite; 7. Chromite horizon; 8. Garnet Anorthosite; 9. Pyroxene Amphibolite.

Eighteen whole rock XRF analyses (for major and trace elements) indicate that this sequence is mainly the differentiation product of a basaltic magma. Cryptic layering is present within zones 2 to 8 inclusive, and shows that this part of the sequence belongs to a stratiform igneous intrusion, zone 8 being the topmost layer of the original igneous stratigraphy. There is especially a marked upward increase in the Fe/Mg ratio.

The amphibolites of zones 1 and 9 are much richer in iron than any rocks belonging to the intrusion, and are considered to have been volcanics into which it was emplaced. Metasomatic sapphirine-rich rocks are localised along the contact of zones 8 and 9.

The complex has internal sub-complexes within which its upper zones are locally confined. The stratigraphic sequence is repeated symmetrically in reverse order and the differentiation trends of the two halves mirror each other. The original body was flattened so that its layering was deformed into an isoclinal syncline. This deformation accounts for the repeated stratigraphy. The complex was subsequently double folded and metamorphosed under both granulite and amphibolite facies conditions. Mineral assemblages indicate that there may have been appreciable water in the original magma, and this viewpoint is strengthened by the observed differentiation trend.

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Fig. 1. Aerial photograph of the Fiskenæsset region of West Greenland, looking east The folded white horizons are the anorthositic layers of the Fiskenæsset complex. 26 July, 1948. Copyright, Geodetic Institute, Copenhagen, Denmark.

INTRODUCTION

Layered anorthosites are an important feature of the Early Archaean, particularly in the high grade, deep level gneissic terrains of shield regions (Windley & Bridgwater, 1971). They formed about 3000 m.y. ago, and their calcic composition (bytownite to anorthite plagioclase) and frequent association with chromite gives them an affinity with the calcic chromiferous anorthosites described from the moon, as suggested by Windley (1970).

There is a remarkable concentration of these rocks in the central Archaean block of West Greenland (Pulvertaft, 1968), where there are at least 900 km of exposed anorthositic rocks (as measured along the strike of the folded layers). They occur as conformable layers in high grade gneisses and reach a maximum thickness of 6 km (Windley, 1969 a).

The Fiskenæsset complex (Plate 1) represents the best anorthosite development in West Greenland. It comprises all the major anorthositic layers in the Fiskenæsset region which are chromite-layered (GHISLER & WINDLEY, 1967) or which can be proved to be chromite-layered when followed along strike. Most of the anorthosites of the region, especially those of the Fiskenæsfjorden–Qeqertarssuatsiaq–Taseq area (fig. 1), appear to belong to a single folded layer. GHISLER (1970) concludes that the chromite deposits belong to a single sheet of anorthosite or to only a few comagmatic sheets, because of the remarkable constancy of chromite composition throughout the area. Although anorthositic rocks are the most characteristic rocks of the complex, there are also many mafic and ultramafic rocks, all of which are mutually intercalated to form a layered complex. The fact that the anorthosites are chromite-layered and the ultramafics are magnetite-, chromite-, and spinel-layered provides the best evidence that part of the complex was of igneous stratiform-type formed by gravity differentiation. It was subsequently highly folded and metamorphosed during regional plutonism. Restricted metasomatism of part of the complex has given rise to abundant sapphirine-bearing and associated rocks with as many as thirty different minerals (HERD *et al.*, 1969). The Fiskenæsset complex has a present exposed strike-length of at least 500 km, and the maximum thickness is about 2 km.

It is emphasised that all rocks have been thoroughly recrystallised by high grade regional metamorphism. The cumbersome prefix meta- has been dropped in the following text, rock types being named according to their igneous-equivalent mineral compositions—i. e. anorthosites, gabbroic anorthosites, anorthositic gabbros, gabbros, etc. The usage of textural terms is as recommended by SPRY (1969).

This report presents field and laboratory data from that part of the complex that we have studied. A stratigraphic and geochemical framework for the complex is presented and this provides the key to its internal structure. The geochemical data provide a way-up for the stratigraphic sequence which cannot be deduced unequivocally from field observations. The petrological data allow an assessment of the physical conditions under which the intrusion was emplaced and later metamorphosed. It is hoped that this framework for the complex will provide a basis for comparison with variations in neighbouring areas.

General geology of the region

We conclude that the following sequence of events best explains presently available evidence for the history of the Fiskenæsset complex.

We assume that the Fiskenæsset intrusion was emplaced into cover rocks overlying a crystalline basement. The supracrustal rocks are now

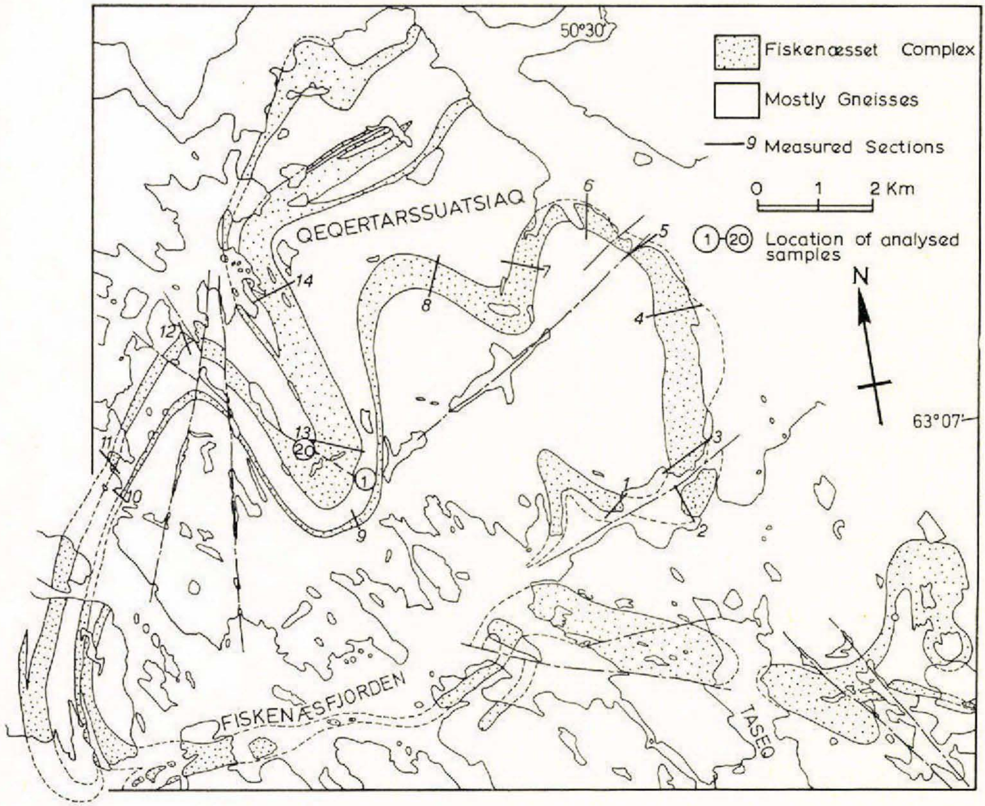

Fig. 2. Map of the Fiskenæsset complex in the Fiskenæsfjorden region showing the location of measured sections and analysed samples.
Correction: the encircled number 20 should read 18.

represented by amphibolites, minor marbles and mica schists, which border the intrusion in places, whilst the basement consists of quartzo-feldspathic gneisses. All evidence of an unconformity between cover and basement has been obliterated by deformation with the result that the intrusion and the border rocks now appear as a conformable layer in the gneisses (see later under 'Structure of the complex').

The complex was recrystallised by a hornblende granulite facies metamorphism and subsequently downgraded in some areas by an amphibolite facies metamorphism. Locally epidote amphibolite facies conditions also occurred. In areas with well-preserved granulite facies rocks, orthopyroxene is present in most rock types of the intrusion. The basement consists largely of red, brown or greenish weathered hypersthene-diopside gneisses containing agmatitic fragments of hypersthene amphibolite, while the cover amphibolites also bear hypersthene. In retrogressed areas hypersthene has been converted to hornblende particularly in felsic rocks such as the basement gneisses, but mafic and ultramafic rocks in both the cover and the intrusion commonly retain their hypersthene, although it may be partly altered to hornblende. In the basement the retrogression appears to have been stratigraphically (compositionally) controlled, so that hypersthene gneiss layers (up to about 1 km thick) alternate with those containing hornblende and/or biotite. Purple gneisses with relict hypersthene rimmed with hornblende, similar to those described from the Tovqussaq (BERTHELSEN, 1960 a) and inner Fiskefjord (WINDLEY, 1969 b) areas of West Greenland, represent an intermediate stage in the retrogressive conversion of reddish hypersthene–diopside gneisses to white or grey biotite–hornblende gneisses.

The structure of the complex is as follows: The original intrusion and its cover were flattened by deformation (F_1) so that they were compressed into an isoclinal syncline. The resultant double-layered complex was deformed by a series of NE-trending isoclinal folds (F_2) and then by a series of N–S to NNW-trending open to tight folds (F_3). The major F_2 and F_3 folds can be easily recognised on fig. 2.

ROCK TERMINOLOGY

The rock types of the complex have been divided into different classes depending on their percentage content of mafic minerals. Textural or mineralogical prefixes have been added to their names in suitable cases. A classification defined by mineral percentages is particularly useful as it allows identification to be made in the field as well as by visual estimate or modal analysis in the laboratory. The definitions of rock types used in the text follow those given by STRECKEISEN (1967). The leuco-gabbros are further subdivided after BUDDINGTON (1939) (slightly modified). The use of the names gabbro and mafic gabbro (melagabbro) was revised and leuco-gabbro introduced as the result of discussion in the field with members of the Fiskenæsset team during the summer of 1972 and agrees with that of MYERS (1973).

Rock Type		*Percentage of Mafic Minerals*
Anorthosite		0– 10
Leuco-gabbro	gabbroic anorthosite	10– 20
	anorthositic gabbro	20– 35
Gabbro		35– 65
Mafic gabbro (melagabbro)		65– 90
Ultramafic rocks		90–100

Anorthosite is the major constituent of the anorthositic horizons of the complex, although locally the amounts of mafic minerals increase so that *gabbroic anorthosite* and *anorthositic gabbro* are found. Leuco-gabbro is used as a field term to embrace these last two types following MYERS (1973). All of these leucocratic rocks have essentially the same mineralogy: a plagioclase matrix containing variable amounts of hornblende, hypersthene, diopside, biotite, and garnet, with minor epidote, muscovite, quartz, scapolite, and cummingtonite. There is a general increase in the variety of mafic minerals as the colour index of the rocks increases. Garnetiferous anorthosites, etc. are mostly confined to the *Garnet Anorthosite* marker horizon of the complex (see 'Stratigraphy, Part I'). These rocks may be termed garnet hornblende anorthosites if their mafic content is less than 10 %, or garnet gabbroic anorthosites and garnet anorthositic gabbros as the amount of hornblende, biotite and/or pyroxene increases.

The term *gabbro* is considered most applicable for a coarse-grained, plutonic rock which probably consisted of diopside and plagioclase, with minor hypersthene. Hornblende is commonly derived from the pyroxene and the rock becomes a hornblende gabbro. Hypersthene gabbro, hypersthene hornblende gabbro, and norite are local compositional variations. Gabbros and anorthositic gabbros are interlayered where the mafic content of the rock varies above and below 35 %. Gabbros contain accessory amounts of magnetite, and occasionally green magnesium aluminium spinel, and minor amounts of quartz. Chalcopyrite intergrown with pyrrhotite has been noted in one anorthositic gabbro.

Homogeneous cumulate leuco-gabbro forms a marker horizon in the complex. It consists of diopside and plagioclase, the diopside invariably being rimmed by some hornblende, and in completely downgraded areas it is a hornblende leuco-gabbro. It is a homogeneous rock (i.e. it has no pronounced layering, foliation, or lineation) with a marked cumulate texture. Euhedral to subhedral plagioclase megacrysts are set in a diopside-hornblende matrix. Where there is rather more plagioclase than mafic minerals the plagioclase megacrysts are aggregated, and only small interstitial areas of diopside and hornblende are present. These cumulate rocks are compositionally anorthositic gabbros and gabbroic anorthosites. In previous years the terms "gabbroic or gabbro anorthosite" have been applied to rocks with these textures in both the Fiskenæsset complex (GHISLER & WINDLEY, 1967), and in other anorthositic layers in West Greenland (BERTHELSEN 1960 a, 1961; WINDLEY, 1969 a). This was, however, a field usage of the terms applied without regard to a strict mineralogical definition. They were subsequently renamed 'ophitic gabbros' by WINDLEY (1971) but are best called cumulate leuco-gabbros etc. to emphasise their origin (MYERS, 1973). There may also be *augen gabbro* which is the deformed equivalent of the cumulate gabbro.

Mafic gabbros (melagabbros) are melanocratic rocks with more than 65 % mafic minerals—normally diopside, hypersthene, and hornblende. Olivine is unknown in these rocks, as it is in the gabbros in general. Melagabbros are rather uncommon, although they may be found among the layers of the *Dark Gabbro* horizon of the complex. Many ordinary gabbros appear to be melagabbros until thin-sectioned, as the dark colour of the plagioclase in hand sample gives a false impression of the true mafic content. Melanorites are also a possible variety, where hypersthene becomes the most predominant mafic mineral.

Ultramafic rocks contain more than 90 % mafic minerals. Major minerals are olivine, orthopyroxene (enstatite/hypersthene), clinopyroxene (diopside/sahlite), and hornblende (several varieties). Accessory minerals include: translucent spinellids (magnesium, iron, chromium, aluminium varieties); opaque spinellids (magnetite, chromite), and ilmenite; sulphides (pyrite, pyrrhotite, and chalcopyrite); biotite, phlogopite, and chlorite; the magnesian amphiboles anthophyllite-gedrite and cummingtonite; and minor zircon, sphene, and rutile. The ultramafic rocks are divided into seven groups, using a strictly mineralogical classification similar to that of DRYSDALL & STILLMAN (1961):

Rock Type	*Mineralogy*
I. Dunite	Olivine 90–100 %
II. Peridotite	Olivine + Pyroxene 90–100 %; Olivine < 90 %; Pyroxene < 90 %
III. Pyroxenite	Pyroxene 90–100 %
IV. Hornblende Dunite	Hornblende 10–50 %; Olivine + Hornblende 90–100 %
V. Hornblende Peridotite	Hornblende 10–50 %; Olivine + Pyroxene 50–90 %; Olivine, Pyroxene < 90 %
VI. Hornblende Pyroxenite	Hornblende 10–50 %; Pyroxene + Hornblende 90–100 %
VII. Hornblendite	Hornblende 50–100 %

Note that hornblendites are classed as ultramafic rocks. Garnet is a characteristic accessory mineral of the hornblendites, which by definition may also contain accessory to substantial amounts of olivine, orthopyroxene, diopside, and even epidote. Locally they are derived from other ultramafics, and veined ultramafic masses exist where the hornblende forms a metasomatic network (see 'Petrogenesis'). Among the ultramafic rocks in general, few types contain no hornblende. Dunites are uncommon, but bronzitites and less commonly diopsidites are found. Hornblende bronzitites, hornblende diopsidites, and the hornblende peridotites are the most common ultramafics. Several arguments lead to the conclusion that the ultramafic rocks, and indeed the other rocks of the intrusion, may have contained primary amphiboles prior to metamorphism (see later under 'Petrogenesis').

Chromitite contains more than 50 % chromite. It forms layers or seams (cf. Bushveld complex) usually up to about 10 cm thick with subsidiary hornblende, plagioclase, rutile, and biotite. Because of the predominance of hornblende amongst the accessory minerals, the term *hornblende chromitite* is more correct (GHISLER, 1970). These layers alternate with anorthositic layers of about the same thickness and together they form chromite horizons between 10 cm and 3 m thick, reaching a maximum thickness of 20 m. Where there is a smaller amount of chromite, the rocks are best called chromite- or chromite-layered anorthosites, gabbroic anorthosites, or anorthositic gabbros. There are also layers of chromite-layered bronzitite up to about 60 cm thick comprised of bronzite chromitite layers up to 1 cm thick alternating with bronzite layers of comparable thickness.

Allanite pegmatites are found as conformable bodies up to a metre thick, in the upper anorthositic rocks of the complex. Allanite crystals up to 10 cm in length may constitute as much as 40 % of these rocks. The pegmatites are acid, the major minerals being quartz, plagioclase, and potash feldspar, with accessory biotite, epidote, and blue apatite. It is possible that these pegmatites are genetically connected to the layered igneous rocks, and represent late-stage high-volatile residues of the differentiated magma of the intrusion, as they have not so far been found in the border amphibolites or basement gneisses (see further under 'Petrogenesis').

Amphibolites are foliated, lineated, medium-grained rocks with 35–90 % mafic minerals. They commonly constitute the marginal and central layers of the complex. Amphibolites are distinguishable from the compositionally similar mafic layers of the complex of obvious plutonic meta-igneous origin by their finer grain size and more prominent and penetrative foliation and lineation. The amphibolites may contain diopside, hypersthene, garnet, and biotite, but the two pyroxenes may be concentrated in different layers or occur sporadically with the result that a single hand sample may only contain one pyroxene plus an amphibole. Thus the use of the term 'pyribolite' for these rocks is actually unsuitable (cf. GHISLER & WINDLEY, 1967); according to the original definition of the term, pyribolite should contain both ortho- and clinopyroxenes in definite proportions to hornblende (BERTHELSEN, 1960 b). Further, it is implicit in the original definition of pyribolite that the contained amphibole is stable with the two pyroxenes. In the Fiskenæsset complex, many examples are known of two-pyroxene rocks in which the amphibole is being derived from one or both of the pyroxenes. To avoid such confusion of terminology, the pyroxene-bearing amphibolites of the Fiskenæsset area are termed hypersthene amphibolites, or diopside amphibolites, or two-pyroxene amphibolites, or simply pyroxene amphibolites, depending on their mineralogy. Diopside garnet amphibolites are an important subclass.

Other *mafic rocks* are present among the layers of the complex. They may form lenses, schlieren, pods, or segregations in the amphibolites, in ultramafic rocks, rarely in the anorthosites, and along the contacts of different zones in the complex. It may be impossible to decide if they are igneous in origin, or represent metasediments engulfed by the intrusion. They consist of varying amounts of garnet, hypersthene, hornblende, diopside, cummingtonite, biotite, spinel, etc., usually with plagioclase, or plagioclase and some quartz. Some sulphides may be present. One group of garnetiferous mafic rocks with the composition garnet, clinopyroxene, hornblende, epidote, scapolite, sphene, plagioclase, carbonate, quartz, and potash feldspar, is similar enough to the more carbonate-rich 'marbles' described below, so that it may be inferred that they represent metasomatised carbonate rocks. Texturally, the mafic rocks may resemble amphibolites (i.e. medium-grained and foliated), or they may be coarser-grained than the surrounding rocks and appear as conformable mafic pegmatites. They are considered separately from the amphibolites, metagabbros, etc., because of their slightly exotic mineralogies in comparison with those rocks. Some garnet, diopside, plagioclase, mafic lenses are considered by the second writer (RKH) to represent assemblages possibly formed under high pressure granulite facies conditions (see under 'Petrological Summary').

In a few areas rocks rich in carbonate have been found. These are termed *marbles* with the qualification that they could have been derived by metamorphism/metasomatism of either sedimentary (limestone, marl) or igneous (carbonatite) bodies. They have been found along anorthosite–amphibolite contacts in a similar environment to the sapphirine rocks (see below). They may contain carbonate, garnet, clinopyroxene, scapolite, hornblende, olivine, sphene, ilmenite, sulphides etc. In two instances known to the second writer (RKH) they are connected with sapphirine-bearing rocks. In one case, almost colourless forsterite and diopside, tremolite, also spinel and a brittle mica (clintonite) have been seen in hand samples. In the other case, a forsterite–diopside–hornblende marble rich in pyrrhotite is seen to be metasomatically altered to rusty hornblendite. The possible origin of some of the minor ultramafic rocks within the complex through metasomatism of impure carbonate rocks must be contemplated.

Sillimanite rocks and *mica schists* have been found as rare rusty sillimanite–biotite–garnet pelitic schists in gneiss, and as sillimanite-plagioclase nodules in an amphibolite of the marginal layers of the complex. The mica schists are usually biotite-rich with or without garnet, and occur either in the marginal rocks or scattered as schlieren throughout the anorthosites etc. of the complex. The presence of such rocks within the border amphibolites would seem to strengthen ideas that these amphibolites are supracrustal in origin, whilst garnet–biotite rocks within anorthosites could have been derived from mafic segregations of igneous origin. Boron-rich(?) kornerupine–phlogopite "schists" associated with sapphirine-bearing lenses might also be supracrustal in origin.

Sapphirine-, kornerupine-, and ruby-bearing rocks occur as lenses and layers of highly variable size (from less than a metre up to at least a kilometre in strike length) along the contact of the intrusion with the upper marginal amphibolites. Several localities have been described in a preliminary manner (HERD *et al.*, 1969), and field and laboratory work is in progress on over 30 such localities now known from within the Fiskenæsset complex. The terminology of these rocks is complicated because of their metasomatic mineral assemblages. Further, the possibility that associated rocks at the sapphirine localities may have had different origins (some original igneous rocks, some sediments) makes nomenclature even more difficult. The terms "sapphirinite" and "sakenite" (LACROIX, 1941) should not be used for these rocks. Instead a division of the sapphirine-bearing rocks on the basis of their contained ferromagnesian minerals has been effected (see HERD *et al.*, 1969, pp. 10–18). This classification has been useful both in mapping and analytical work. Detailed description of these rare mineral assemblages is beyond the scope of this paper.

Finally, rocks of similar composition and mineralogy to the rocks of the Fiskenæsset complex, but occurring outside the boundaries of the complex within the adjacent gneisses, should be mentioned. Minor cumulate gabbros and gabbroic anorthosites, amphibolites, and ultramafic bodies occur as conformable layers, or boudinaged and/or agmatised units of highly variable size and extent. Acid pegmatites, plagioclase-rich pegmatites, and quartz pegmatites intrude both the gneisses and the complex, or may indeed be found as conformable bodies within the gneisses or along lithological contacts within the complex. Several generations of fresh to slightly amphibolised dolerite dykes cut the gneisses and the rocks of the complex. Some conformable amphibolite layers in the gneisses may be derived from early dolerites. No further mention will be made of these sundry rock types which are not intimately related to the layered rocks of the complex.

MINERAL ASSEMBLAGES

The following mineral assemblages have been noted, to date, within the rocks of the Fiskenæsset complex from a study of more than 400 thin sections. As these assemblages will be referred to in later discussion, each assemblage has been numbered.

Anorthosites (0–10 % mafic minerals)

1.		Hyp		Hb						Plg		
2.			Di	Hb		(Bi)				Plg		
3.				Hb						Plg		
4.	(Gnt)			Hb						Plg		
5.				Hb		Bi				Plg		
6.	Gnt			Hb		Bi				Plg		
7.				Hb		Bi			Qtz	Plg		
8.				Hb		Bi		Ep	Qtz	Plg	Scap	
9.				Hb				Ep		Plg		
10.	Gnt					Bi				Plg		
11.	Gnt					Bi	(Musc)			Plg		
12.	Gnt					Bi	(Musc)			Plg		(K-sp)
13.								Ep		Plg		

Accessory minerals and alterations: serpentine, chlorite, epidote, quartz, scapolite, carbonate, magnetite, pyrite

Gabbroic anorthosites (10–20 % mafic minerals)

1.		Hyp	Di	Hb			(Ep)		Plg
2.		Hyp		Hb					Plg
3.	Gnt	Hyp		Hb					Plg
4.			Di	Hb					Plg
5.			Di	Hb			(Ep)		Plg
6.	Gnt		Di	Hb					Plg
7.			(Di)	Hb		(Bi)			Plg
8.				Hb					Plg
9.	(Gnt)			Hb	Cumm	Bi			Plg
10.				Hb		Bi			Plg
11.	Gnt			Hb		Bi			Plg
12.				Hb		(Bi)	Ep		Plg
13.						Bi			Plg
14.						Bi		Qtz	Plg
15.	Gnt					Bi			Plg

Accessory minerals and alterations: chlorite, epidote, quartz, scapolite, carbonate, magnetite, pyrite, apatite, Fe stain

Anorthositic gabbros (20–35 % mafic minerals)

1.		(Hyp)	Di	Hb			Plg
2.		Hyp	Di	Hb		(Ep)	Plg
3.	Gnt	Hyp	Di	Hb			Plg
4.	Gnt	Hyp		Hb	Cumm		Plg
5.		Hyp		Hb			Plg

Anorthositic gabbros (cont.)

6.		Sp	Hyp		Hb (Anth)						Plg			
7.				Di	Hb						Plg			
8.				Di	Hb				Ep		Plg			
9.				Di	Hb					Qtz	Plg			
10.					Hb						Plg			
11.					Hb						Plg	(Mag)		
12.	Gnt				Hb	(Bi)					Plg			
13.	Gnt				Hb	(Bi)				(Qtz)	Plg			
14.	Gnt				Hb	Bi		Chl			Plg	(Mag)	(Ilm)	
15.					Hb	(Bi)	(Musc)	(Chl)			Plg			
16.					Hb				Ep		Plg			
17.	Gnt					Bi				(Qtz)	Plg			

Accessory minerals and alterations: chlorite, epidote, carbonate, magnetite, ilmenite, pyrite, sphene, Fe stain

Gabbros and norites (35–65 $^0/_0$ mafic minerals)

1.			Hyp	Di	Hb						Plg			
2.			Hyp	Di	Hb					(Qtz)	Plg			
3.		Sp	Hyp	Di	Hb						Plg			
4.			Hyp		Hb						Plg			
5.				Di	Hb						Plg	(Mag)		
6.	Gnt			Di	Hb					(Qtz)	Plg	(Mag)		
7.				(Di)	Hb				Ep	(Qtz)	Plg			
8.					Hb						Plg			
9.					Hb					Qtz	Plg			
10.	(Gnt)				Hb					Qtz	Plg			

Accessory minerals and alterations: chlorite, epidote, magnetite, pyrite, pyrrhotite, zircon, apatite

Mafic gabbros and mafic norites (65–90 $^0/_0$ mafic minerals)

1.			Hyp	(Di)	Hb						Plg			
2.					Hb				Ep	Qtz	Plg			(Sphene)

(Accessories etc. as for gabbros and norites)

Ultramafic rocks (90–100 $^0/_0$ mafic minerals)

I. *Dunites*

1.	Sp	Ol	(Opx)	(Hb)			(Serp)		(Mag)	(Ilm)		
2.	Sp	Ol				Phlg		(Chl)	(Mag)			(Zircon)

II. *Peridotites*

1.	Sp	Ol	Opx									
2.	Sp	Ol	Opx	(Hb)	(Ged)							

Accessory minerals and alterations: serpentine, chlorite, magnetite, pyrite, rutile

III. *Pyroxenites*

1.	Sp	(Ol)	Opx						(Mag)	(Ilm)		
2.	Sp	(Ol)	Opx	(Hb)					(Mag)	(Ilm)		

Alterations: chlorite

IV. *Hornblende dunites*

1.	Sp	Ol		Hb			(Serp)				Pyrrh	
2.		Ol		Hb			(Serp)					

Accessory: magnetite

Ultramafic rocks (cont.)

V. *Hornblende peridotites*

1.		Ol	Opx	Di	Hb		(Serp)		(Mag)	(Pyr)	
2.	(Sp)	Ol	Opx	Di	Hb		(Serp)				
3.		Ol	Opx	Di	Hb	(Phlg)					
4.	Sp	Ol	Opx		Hb		(Talc)				
5.	Sp	Ol	Opx		Hb						(Rutile)
6.	Sp	Ol	Opx		Hb		(Serp)		(Mag)		
7.	Sp	Ol	Opx		Hb		(Serp)		(Mag)(Ilm)		
8.	Sp	Ol	Opx		Hb	(Phlg)					
9.		Ol	Opx		Hb		(Serp)				
10.		(Ol)	Opx		Hb		(Serp)	(Chl)	(Mag)		
11.	Sp	Ol		Di	Hb						(Zircon)
12.		Ol		Di	Hb				(Mag)(Ilm)		

Accessory minerals and alterations: serpentine, talc, chlorite, magnetite, ilmenite, pyrite, pyrrhotite, chalcopyrite, rutile, Fe stain

VI. *Hornblende pyroxenites*

1.		Opx	Di	Hb			
2.		Opx	Di	Hb			(Mag)
3.	Sp	Opx		Hb			
4.	Sp	Opx		Hb			(Mag)
5.		Opx		Hb			
6.		Opx		Hb			(Chromite)
7.		Opx		Hb	(Cumm)		(Mag)
8.		Opx		Hb		(Phlg)	

Accessory minerals and alterations: serpentine, chlorite, quartz, plagioclase, carbonate, magnetite, pyrite, pyrrhotite, Fe stain

VII. *Hornblendites*

1.	Gnt		(Opx)	(Di)	Hb				Ep	
2.		(Sp)	Opx		Hb					
3.	Gnt	(Sp)	(Opx)		Hb					
4.				Di	Hb					
5.		Sp			Hb		(Phlg)			
6.					Hb			(Chl)		
7.	Gnt				Hb					(Mag)(Ilm)
8.					Hb		(Bi)			
9.					Hb	Trem			Ep	(Mag)

Accessory minerals and alterations: chlorite, quartz, plagioclase, carbonate, magnetite, pyrite, chalcopyrite, sphene, zircon, Fe stain

Chromite rocks

Chromiferous ultramafic rocks

1.	Opx	Hb	(Chromite)

(cf. Ultramafic assemblage VI. 6)
Accessory minerals and alterations: talc, chlorite, epidote, pyrite

Hornblende chromitites (50–100 % chromite)

1.	Hb	(Bi)	Chromite	(Rutile)

Chromiferous leuco-gabbros, gabbros (*matrix* 10–65 % mafics)

1.	Hb					Plg	Chromite	(Rutile)
2.	Hb	(Bi)				Plg	Chromite	(Rutile)
3.	(Hb)	Bi				Plg	Chromite	
4.			Musc	Chl	Ep	Plg	Chromite	(Rutile)

Allanite pegmatites

1.								(Ep)	Qtz	Plg		Micr		Allanite		Ap
2.							Bi		Qtz	Plg		Micr		Allanite		

Alterations: muscovite, chlorite

Amphibolites (35–90 % mafic minerals)

1.	(Gnt)		Hyp	(Di)			(Bi)			Plg					
2.	Gnt			Di			(Bi)								
3.			Hyp	Di	Hb					Plg					
4.			Hyp	Di	Hb				(Qtz)	Plg					
5.			Hyp		Hb					Plg					
6.			Hyp		Hb				Qtz	Plg					
7.			Hyp		Hb		Bi			Plg					
8.				Di	Hb					Plg					
9.				Di	Hb					Plg				(Mag)	
10.	Gnt			Di	Hb					Plg					
11.				Di	Hb				Qtz	Plg					
12.	Gnt			Di	Hb				Qtz	Plg					
13.				Di	Hb			Ep							
14.					Hb					Plg					
15.	Gnt				Hb					Plg					
16.					Hb				Qtz	Plg					
17.					Hb				Qtz	Plg					(Sphene)
18.	Gnt				Hb				Qtz	Plg					
19.					Hb	Cumm	(Bi)		Qtz	Plg					
20.					Hb		Bi			Plg					
21.					Hb			Ep		Plg					
22.					Hb	Cumm	Bi		Qtz						

Accessory minerals and alterations: chlorite, epidote, potash feldspar, carbonate, magnetite, pyrite, pyrrhotite, sphene, Fe stain

Mafic rocks as minor layers and lenses (35–90 % mafic minerals)

1.	Gnt		Hyp		Hb					Plg					
2.	Gnt		Hyp		Hb	Cumm				Plg					
3.			Hyp		Hb					Plg					
4.			Hyp			Anth	Bi			Plg					
5.	Gnt			Di				Ep		Plg		(K-sp)	Cb		
6.	Gnt			Di				Ep	Qtz	Plg	Scap	Micr	Cb		Sphene
7.	Gnt			Di	Hb					Plg					
8.	Gnt	(Sp)		Di	Hb					Plg					
9.	Gnt			Di	Hb			Ep	Qtz	Plg			Cb		
10.				Di	Hb					Plg					
11.	Gnt				Hb	Cumm	Bi		Qtz	Plg					
12.					Hb					Plg					
13.		Sp			Hb					Plg					
14.	Gnt	Sp			Hb		(Bi)			(Plg)				(Mag)	
15.	Gnt				Hb		Bi		Qtz					(Mag)	
16.	Gnt	Sp					Bi			(Plg)					
17.	Gnt					(Cumm)	Bi			(Plg)					

Accessory minerals and alterations: corundum, chlorite, carbonate, magnetite, pyrite, Fe stain

Plagioclase pegmatites (0–35 % mafic minerals)

1.		Hyp	Di	Hb		(Ep)	Qtz	Plg	
2.		Hyp		Hb		(Ep)	Qtz	Plg	
3.	Gnt	Hyp		Hb	Cumm			Plg	
4.			Di	Hb		Ep	Qtz	Plg	Scap

Plagioclase pegmatites (cont.)

5.		Musc	(Ep)		Plg		(Cb)	
6.		(Musc)		(Qtz)	Plg	(K-sp)		
7.	Bi			(Qtz)	Plg			(Mag)

Accessory minerals and alterations: chlorite, quartz, scapolite

Mafic pegmatites (35–90 $^0/_0$ mafic minerals)

1.			Hb	Cumm	Bi	Qtz		(Pyr)	
2.	Hyp	Di	Hb				(Plg)		(Zircon)

Alteration: Fe stain

Ultramafic schlieren (90–100 $^0/_0$ mafic minerals)

1.	Ol	Opx	Di	Hb

Carbonate-rich rocks (*"marbles"*)

1.	Gnt		Di	Hb	Ep	Scap	Cb	(Mag)	(Ilm)		(Sphene)	
2.		Ol	Di	Hb			Cb			(Pyrrh)		(Ap)

Sillimanite rocks

1.	Gnt	Sill		Bi	Qtz	Plg	K-sp		
2.		Sill	Cn	Bi		Plg		Hematite	Rutile

Mica schists

1.	Gnt	Bi	Plg
2.		Bi	Plg

Notes on the preceding lists of mineral assemblages:

1) Abbreviations used:

Anth	Anthophyllite	Gnt	Garnet	Phlg	Phlogopite
Ap	Apatite	Hb	Hornblende (see note 6 a)	Plg	Plagioclase
Bi	Biotite	Hyp	Hypersthene	Pyr	Pyrite
Cb	Carbonate	Ilm	Ilmenite	Pyrrh	Pyrrhotite
Chl	Chlorite	K-sp	Potash feldspar	Qtz	Quartz
Cn	Corundum	Mag	Magnetite	Scap	Scapolite
Cumm	Cummingtonite	Micr	Microcline	Serp	Serpentine
Di	Diopside	Musc	Muscovite	Sill	Sillimanite
Ep	Epidote	Ol	Olivine	Sp	Spinel (see note 6 b)
Fe	Iron	Opx	Orthopyroxene	Trem	Tremolite
Ged	Gedrite	Orth	Orthoclase		

2) Bracketed abbreviations, as (Plg), indicate that the mineral is in relatively minor abundance (approximately 5 $^0/_0$ or less).

3) Mineral percentages have, in the majority of cases, been estimated using the visual charts of Terry & Chilingar (1955); percentage estimates by this method are probably accurate to within 5 $^0/_0$.

4) Minerals within the assemblages are listed in a specific order, i.e. with mafic minerals on the left and felsic minerals to the right, with accessory minerals on the far right. This allows each assemblage to be given a proper lithological name—e.g. anorthosite assemblage 1 is either hypersthene–hornblende anorthosite or hornblende–hypersthene anorthosite, depending on the relative abundance of the 10 $^0/_0$ contained mafics.

5) Mineral assemblages within each rock type are listed in order of decreasing metamorphic grade. By this device, the petrogenesis of the rocks is graphically portrayed to some degree (see further under 'Petrogenesis').

6a) "Hornblende" is used for several different amphiboles, ranging in colour from dark olive brown to bluish green in thin section. In general, hornblendes in the higher grade assemblages (see note 5) are olive-brown to dark green.

6b) "Spinel" is used for several different spinellid phases, i.e. the translucent varieties ranging in colour from grey to pale or dark green, and from dark brown to olive green. The term spinellid is used to denote isometric minerals of the AB_2O_4 type, where A = Mg, Fe^{2+}, Zn, Ni etc. and B = Al, Fe^{3+}, Cr^{3+} etc.

7) These assemblage lists are thought to be least complete for the mafic gabbros, marbles, sillimanite rocks, mica schists, mafic pegmatites and ultramafic schlieren.

STRATIGRAPHY

Part I

Rock units

The well-preserved primary stratigraphy of the complex can be followed persistently along the strike for at least 60 km on Qeqertarssuatsiaq (Plate 1), west and east of the lake Taseq, and along the southern coast of Fiskenæsfjorden. However, there has been considerable tectonic thinning of many layers especially along fold limbs, and there are primary lateral variations of certain layers, particularly the dark gabbros. These interruptions of the layered succession mean that any one traverse across the complex may display an incomplete stratigraphic section. By taking these variations into account the following maximum, composite stratigraphy has been found:

	Zone	*Major Units*	*Minor Units* (may or may not occur)
Top	9	Pyroxene Amphibolite	
			----- Sulphide-bearing amphibolite
			----- Marbles and minor ultramafics
			----- Sapphirine rocks
	8	Garnet Anorthosite	----- Rare 1 m ultramafics
	7	Hornblende Chromitite	
	6	Anorthosite	----- Dark gabbro
			----- Rare 1 m ultramafics
			----- Dark gabbro
Fiskenæsset intrusion	5	Homogeneous Leuco-gabbro	----- Dark gabbro
			----- Rare 1 m ultramafics
	4	Dark Gabbro	----- Rare 1 m ultramafics
			----- Chromite-layered gabbros and bronzitites
	3	Layered leuco-gabbro with anorthosite laminae	----- Rare 1 m ultramafics
	2	Ultramafic Group	
			----- Rare gneiss layers
	1	Pyroxene Amphibolite	----- Sulphide-bearing amphibolite
Bottom			

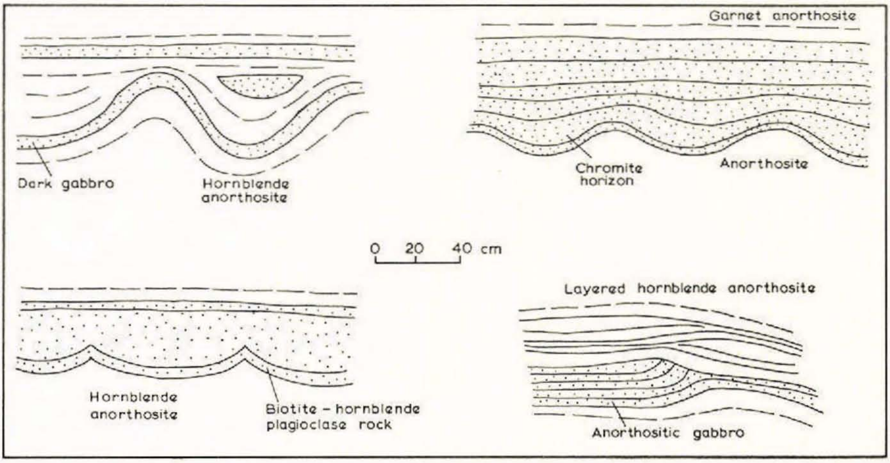

Fig. 3. Four examples of bottom structures which all give a consistent and correct way-up for the stratigraphy.

The term Fiskenæsset 'complex' is used to embrace the complete group of intercalated mappable rock units given above. Cryptic layering, petrography (especially the recognition of relict cumulate textures, cf. MYERS, 1973), and contrasts in bulk composition, show that part (zones 2 to 8) of the 'complex' belongs to a stratiform igneous 'intrusion'; the remaining part (bordering amphibolites—zones 1 and 9) belongs to the cover rocks into which the 'intrusion' was emplaced.

The stratigraphy was recognised in 1970 but the names formerly given to zones 3, 4, and 5 (WINDLEY, 1971) were altered after discussion with members of the Fiskenæsset team in the summer of 1972. The changes are: zone 3, lower layered group (mostly gabbros) to layered leuco-gabbro with anorthosite laminae; zone 4, mafic gabbro to dark gabbro; and zone 5, ophitic gabbro to homogeneous leuco-gabbro.

Petrographic descriptions of the rocks within these major stratigraphic units now follow. A detailed assessment of the double-reversed nature of the present stratigraphic sequence is then given, followed by a discussion of the chemistry of the rocks as shown in a single traverse across the complex. The way-up of the stratigraphy as indicated in the preceding table is based on chemical evidence. Primary bottom structures (fig. 3) suggest the same facing direction for the stratigraphy as is indicated by the chemical analyses.

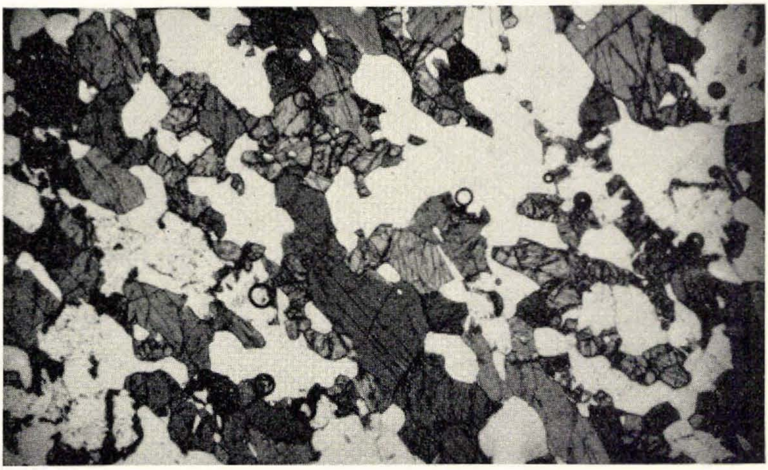

Fig. 4. Photomicrograph of a lineated amphibolite with brown hornblende and diopside. GGU 68699, plain light, ×17.

Petrographic descriptions

Zone 1

The Lower Pyroxene Amphibolites are foliated, lineated rocks, locally showing layering of units 10 cm to 1 m thick. Discrete layers are defined by the relative concentrations of hornblende, pyroxene, garnet, and plagioclase. The mineral orientation (hornblende prisms, other mafic clots) lies in the plane of the layering.

The mafic minerals usually form a dense lineated network homogeneously with the felsic minerals. Hypersthene and/or diopside form rounded xenoblastic granules up to 1 mm across in layers in the plagioclase matrix or as ragged groups of granules within mafic layers. Hornblende is the most abundant mafic mineral, although small patches of the rock may be hornblende-free. There is some evidence that the hornblende is derived in part from the pyroxenes, in that larger hornblende crystals may enclose other mafic granules. Hypersthene and diopside, and especially the orthopyroxene, may form spongy or sieve-like intergrowths with plagioclase. This texture is taken to indicate partial instability of the pyroxene. Hypersthene may alter in part to talc, carbonate, serpentine and ore granules along rims and cleavages. Most of the ore must be derived in this way. Hornblende forms grains of a similar to much larger size than the pyroxenes, and may exhibit sub-idioblastic forms with curved grain boundaries against the matrix plagioclase (fig. 4). It is deep green, greenish brown, or brown in colour. Garnet is sporadically distributed as small rounded grains with the other mafic minerals, or shows a strong correlation with diopside where hornblende is scarce. Quartz forms rare xenoblastic grains in the plagioclase subfabric, or siliceous layers and veinlets (possibly introduced silica).

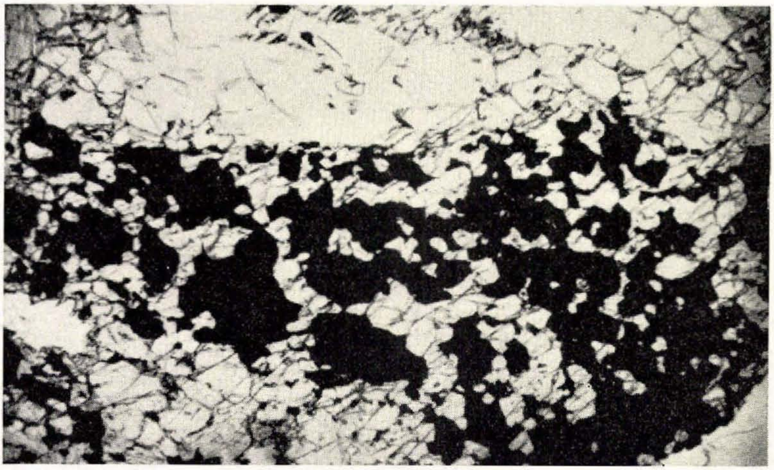

Fig. 5. Photomicrograph of spinel layer in hornblende peridotite of zone 2. GGU 86935, plain light, ×17.

Mineral Assemblages:

Amphibolites:	1.	Garnet–hypersthene-diopside-biotite 'granulite'
	3, 4.	Two pyroxene amphibolites
	8, 11.	Diopside amphibolites
	10.	Garnet–diopside amphibolite
	14.	Amphibolite
Mafic Lens:	1.	Garnet–hypersthene–hornblende–plagioclase

Zone 2

The Ultramafic Group contains spinel-bearing dunites, pyroxenites, and hornblende peridotites rich in magnetite–ilmenite, magnetite-bearing hornblende pyroxenites, and hornblendites. Chromian spinel is known from this zone.

Most of the rocks are weakly oriented, although some hornblendites are intensely lineated. Grain size ranges up to about 8 mm, commonly with great variation due to megacrysts of pyroxene or olivine. A granular texture of mostly equidimensional grains is prevalent, large megacrysts being strained with the formation of subgrain boundaries and mosaics. Dark green or olive spinel, magnetite, and chromite, may form layers or be scattered throughout the matrix (fig. 5). The spinellids are generally associated with olivine, or olivine and orthopyroxene, as inclusions or along contacts of these minerals with hornblende. Spinel is found as inclusions in hornblende only in hornblendites. The oxide phases may make up 50 % of the rock, entirely intergrown with, and including, the silicates. Spinel grains reach 4 mm in these rocks. Larger spinel grains associated with hornblende while smaller spinels are included within olivines, indicate the metamorphic nature of the bulk of the spinel. Hornblende is green, pale grey green, or bluish green, and reaches maximum grain size in equilibrium textures in the hornblendites. In such cases spinel, orthopyroxene, and olivine are scattered as a net of smaller grains in the hornblende matrix. Olivine in all rocks may show some serpentinisation.

Mineral Assemblages:

Dunites:	1. Spinel-magnetite dunite
Pyroxenites:	1. Spinel-magnetite bronzitite
Hornblende dunites:	2. Olivine-hornblende
Hornblende peridotites:	1. Two pyroxene-olivine-magnetite
	4. Hypersthene-olivine-spinel
	6. Hypersthene-olivine-spinel-magnetite
Hornblende pyroxenites:	2. Two pyroxene-magnetite
Hornblendites:	2. Spinel-hypersthene
	6. Hornblendite (bluish green amphibole)

Zone 3

This zone displays a complicated stratigraphy of rock types caused by variation in the relative amounts of pyroxene, hornblende and plagioclase. The rocks are well-layered, purely plagioclase layers rapidly alternating with more mafic layers in places; these layers may be up to 30 cm thick. There is usually a hornblende orientation lying in the plane of the layering.

The rocks have a marked granular texture, grain size not exceeding 1.5 mm. Plagioclase, unzoned, but intensely twinned, and with wavy extinction, has an equilibrium texture with triple junctions and planar to curviplanar grain boundaries. There may be three distinct sizes of plagioclase grains: large grains in plagioclase-rich segments of the matrix, medium size polygonal grains making up the bulk of the matrix, and smaller polygonal grains in association with the mafic minerals. This grain-size variation produces a rough augen or web texture, and may be founded upon an igneous cumulate texture. Hornblende is almost always present, as green to pale bluish green sub-idioblastic prisms and aggregates, scattered separately in the plagioclase matrix, or intimately associated with pyroxene in the mafic layers. These rocks show the most superbly developed hypersthene–plagioclase sponge symplectites (fig. 6). The hypersthene granules within the symplectites are rounded but of constant optical orientation. The symplectites are stretched and oriented parallel to hornblende prisms in the plane of the layering. Elsewhere it is clear that the amphibole is developing in part from the pyroxene, confirming the metastable nature of the symplectites.

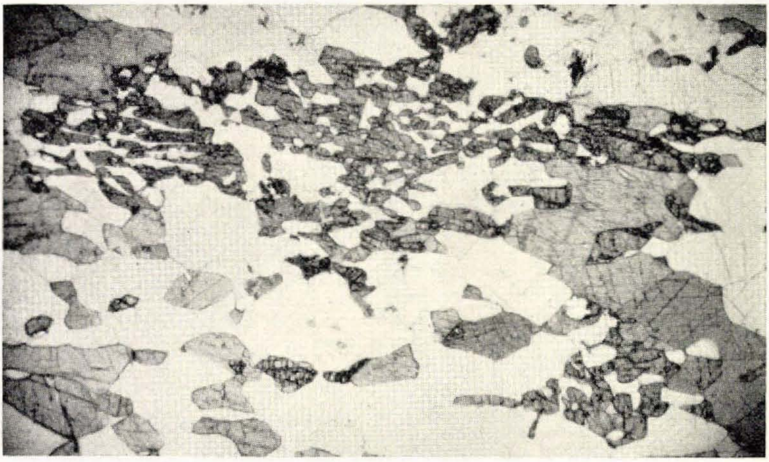

Fig. 6. Photomicrograph of spongy hypersthene–plagioclase intergrowth in hornblende norite of zone 3. The hypersthene granules are all in optical continuity. GGU 74497, plain light, ×17.

3 Enstatite + 1 Diopside + 1 Anorthite + 1 Water = 1 Common Hornblende. Some amphibole, however, may be derived from igneous amphibole. Distinct layers of pyroxene and plagioclase in equilibrium, without hornblende, may alternate with layers containing dark green hornblende and plagioclase, also in equilibrium. Biotite is found where these rocks are quartz-bearing. This is obviously the effect of local metasomatism, as the quartz replaces plagioclase, and mutual sutured contacts are found.

Mineral Assemblages:

Gabbroic anorthosites:	2.	Hypersthene-hornblende
	8.	Hornblende
	14.	Biotite-quartz
Anorthositic gabbros:	2.	Two pyroxene-hornblende
Gabbros:	1.	Two pyroxene-hornblende
	4.	Hypersthene-hornblende

Zone 4

The Dark Gabbro zone is composed of interlayered gabbros, anorthositic gabbros, and melagabbros. Textures in these rocks are similar to those described in the zone 3 rocks, although the higher mafic content allows further conclusions about the inter-relationship between amphibole and pyroxene to be drawn. Grain size ranges up to 2 mm. Plagioclase may be present as large zoned crystals, or more commonly as equilibrium granular polygons among the mafic concentrations. The mafic minerals are usually present as equidimensional grains as well, aggregated or scattered in the plagioclase matrix. Grains of pyroxene occur as mosaics, or small individuals, invariably altering to hornblende. Relics of diopside may be found where hornblende development is extensive. Note that the equation quoted above with its 3:1 ratio of enstatite to diopside explains the predominance of diopside relics, rather than hypersthene relics, in hornblende. Orthopyroxene is preferentially used up in the hornblende-forming reaction. The hornblende ranges in colour from dark olive green to pale bluish green. Some hypersthene-plagioclase sponge symplectites are present. Where the rock as a whole approaches melagabbro in composition, or within mafic streaks, plagioclase occurs as interstitial grains with concave boundaries. Some augen or net textures of the mafics with respect to the plagioclase are recognised.

Mineral Assemblages:

Anorthositic gabbros:	7. Diopside-hornblende
	10. Hornblende
	11. Hornblende-magnetite
Gabbros:	1. Two pyroxene-hornblende
	4. Hypersthene-hornblende
	5. Diopside-hornblende-magnetite
	8. Hornblende
Melagabbros:	1. Two pyroxene-hornblende

Zone 5

The Homogeneous Leuco-gabbro zone contains rocks grading in composition from anorthosite to gabbro, linked by the occurrence of distinctive cumulate textures within them. What appear to be euhedral or subhedral plagioclase crystals (up to 1 cm or more across) in hand sample, can be seen in thin section to consist of a regular mosaic of equidimensional plagioclase grains up to 1.5 mm across, replacing larger relict plagioclase megacrysts. The latter show breakdown into subgrain mosaics, and re-entrant angles against the equilibrium mosaic of the secondary plagioclase. All plagioclase grains may show twinning and zoning (sometimes pronounced), and wavy extinction. The outlines of the plagioclase-rich areas are marked by hornblende and pyroxene aggregates. Hornblende is dark green to pale bluish green, and the paler amphibole is derived from pyroxene, as diopside relics in optical continuity with one another occur in it (fig. 7). Breakdown of the larger plagioclase grains may be accompanied by the development of hornblende microcrystals within the body of the larger grains (fig. 8); this explains their greenish colour often observed in hand sample. Hypersthene, as spongy relics, is rare within this zone, although a spinel gabbro is found to have hypersthene.

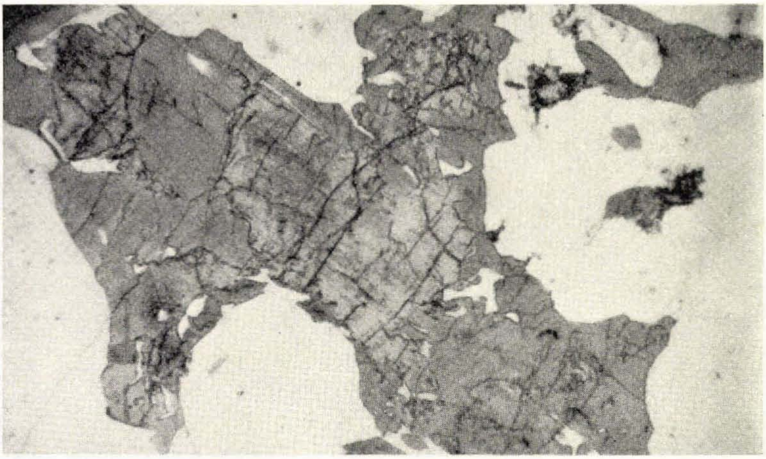

Fig. 7. Photomicrograph of diopside relics in secondary pale green hornblende, zone 5. GGU 68696, plain light, ×34.

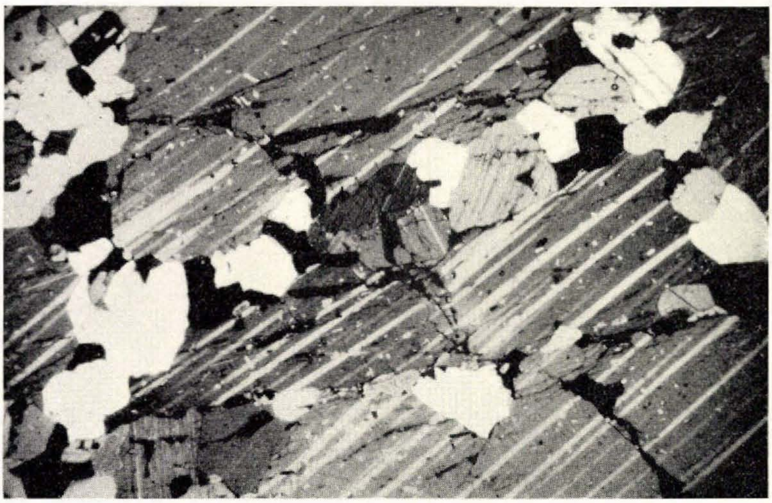

Fig. 8. Photomicrograph of polygons of secondary plagioclase replacing twinned plagioclase megacryst in a cumulate leuco-gabbro. Note the hornblende microcrystals in the body of the megacryst. Zone 5. GGU 92578, crossed nicols, ×17.

Mineral Assemblages:

Anorthosites:	3. Hornblende
Gabbroic anorthosites:	4. Diopside-hornblende
	8. Hornblende
Anorthositic gabbros:	7. Diopside-hornblende
	10. Hornblende
	15. Hornblende-biotite-chlorite
Gabbros:	3. Spinel-two pyroxene-hornblende
	4. Hypersthene-hornblende

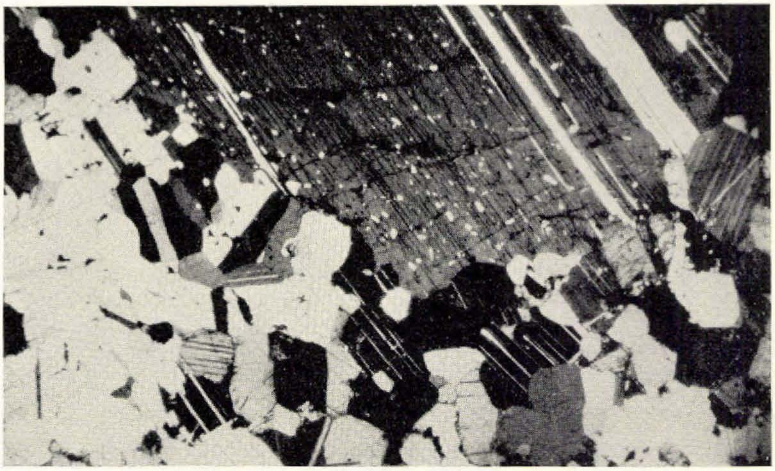

Fig. 9. Photomicrograph of large relict plagioclase megacryst being replaced by polygonal secondary plagioclase. Hornblende microcrystals are developed within the megacryst. Zone 6. GGU 86913, crossed nicols, × 17.

Zone 6

This major Anorthosite zone of the complex contains both anorthosites and gabbroic anorthosites which share many of the textural and mineralogical features described for the zone 3, 4, and 5 rocks. Coarsely foliated types are present as well as rocks with relict augen or cumulate textures. Grain size does not exceed 3 mm generally but some megacrysts reach 8 mm in size. Plagioclase is twinned and zoned, and shows strain effects through wavy extinction, especially if altered to late epidote, scapolite, etc. Hornblendes and pyroxenes occur as individual granules, or aggregates of granules along plagioclase–plagioclase grain boundaries. The rocks are typically dominated by a well-formed equilibrium fabric with planar boundaries and triple junctions. However, hornblende aggregates especially show re-entrant angles against the plagioclase matrix. These are dark green hornblendes with a slightly bluish rim. Bluish hornblendes with diopside and hypersthene relics are also seen, and it is obvious that they have developed from original pyroxene through amphibolite facies metamorphism. The green hornblendes with the bluish rims must belong to higher grade assemblages. Hornblende microcrystals are also observed in these rocks, accompanying the recrystallisation of plagioclase megacrysts as in the cumulate leucogabbros (fig. 9). More significantly, trains of these microcrystals can be seen cutting the grain boundaries of secondary plagioclase polygons (fig. 10); this indicates that large plagioclase crystals once occupied such sites.

Fig. 10. Photomicrograph of secondary plagioclase polygons with aligned hornblende microcrystals indicating the prior existence of a megacryst. Zone 6. GGU 86930, crossed nicols, × 34.

Mineral Assemblages:

Anorthosites:	1. Hypersthene-hornblende
	3. Hornblende
	8. Hornblende-biotite-epidote-quartz
	9. Hornblende-epidote
Gabbroic anorthosites:	4. Diopside-hornblende
	5. Diopside-hornblende-epidote
	8. Hornblende
	10. Hornblende-biotite
	12. Hornblende-biotite-epidote
Anorthositic gabbros:	10. Hornblende

Zone 7

For a more detailed account of the chromite-bearing rocks, see Ghisler & Windley (1967). There are two types: a) Layered type in which the chromite is confined to the hornblende-rich layers which alternate with others rich in plagioclase. Layers are usually up to about 3 cm thick, chromite grains up to 0.7 mm across, and aggregates of chromite up to 0.5 cm across. The chromites form euhedral to subidioblastic grains set in a hornblende or hornblende–biotite matrix (fig. 11). Hornblendes (green, maximum length 1 mm) occur as short interlocking prisms, biotites are rare (2 mm maximum) and plagioclase forms an equilibrium mosaic of xenoblastic granules. Rutile builds rare grains between chromites. b) Augen type (fig. 12) is characterised by

Fig. 11. Photomicrograph of hornblende chromitite showing chain structure and some euhedral chromite crystals. Zone 7. GGU 53282, plain light, ×17.

plagioclase augen (5 cm × 2 cm maximum) consisting of a mosaic of plagioclase grains (up to 2.5 mm across) with an equilibrium texture. The outlines of the augen are defined by seams of chromite-rich hornblende-biotite 'matrix'.

Mostly the augen lie with their longest axes in the plane of the layering, but they locally occur oblique to it, forming an augen lineation. Recent field study has shown that there were at least two generations of augen orientation, during the F_2 and F_3 fold phases, the augen being aligned along the axial planes of the two sets of minor folds.

Relict plagioclase megacrysts are present especially in the augen type rocks (fig. 13), and therefore it may now be concluded that megacrysts of plagioclase are characteristic of zones 3, 5, 6 and 7 of the complex. The augen in these rocks are deformed and recrystallised megacrysts.

GHISLER (1970) has distinguished two types of chromite in these chromite horizons: large poikilitic crystals with silicate inclusions, and inclusion-free pure chromite grains. He has also noted the occurrence of chain structure in the chromite aggregates, which may or may not be indicative of igneous aggregation of the grains (cf. JACKSON, 1961, p. 34).

Biotite is present in all but a few chromite-bearing assemblages. It may be implied that it formed under amphibolite facies conditions due to introduction of potash and water, but the possibility exists that it is an expression of significant potash concentration in the original igneous chromite zones.

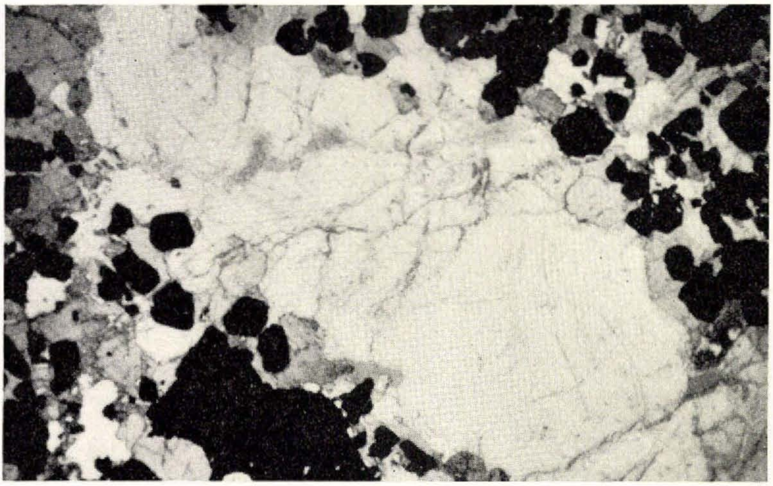

Fig. 12. Photomicrograph of plagioclase augen in hornblende chromitite matrix. The amphibole is very pale green, and the chromite is semi-translucent and dark brown. Zone 7. GGU 86881, plain light, × 17.

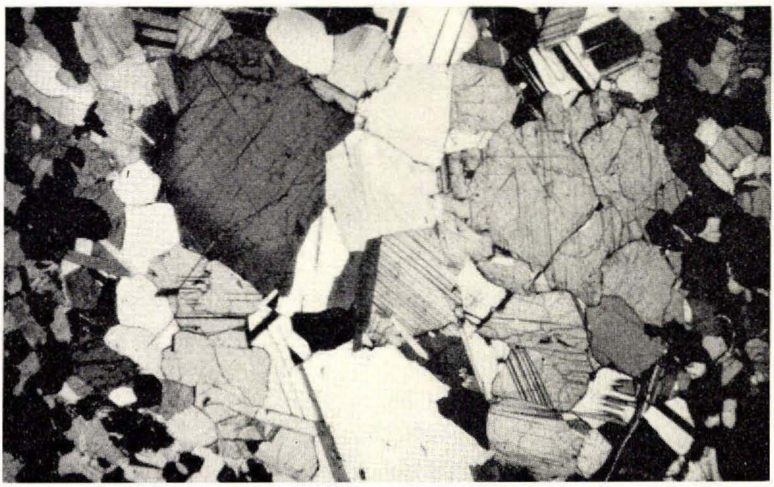

Fig. 13. Photomicrograph of zoned relict plagioclase within an augen in a hornblende–biotite chromitite matrix. Zone 7. GGU 86917, crossed nicols, × 17.

All hornblendes in these rocks have a characteristic pale green colour, and are probably chromiferous.

Mineral Assemblages:

Chromiferous gabbroic anorthosites:	4. Muscovite-epidote-rutile
Chromiferous anorthositic gabbros:	1. Hornblende-rutile
	2. Hornblende-biotite-rutile
	3. Hornblende-biotite
Chromiferous gabbros:	2. Hornblende-biotite-rutile

Zone 8

Garnet Anorthosite consists largely of plagioclase with a predominant equilibrium fabric and subsidiary garnet, biotite and hornblende. Garnet and biotite are intimately associated. Garnets have grown up to 2 cm across as rounded xenoblastic grains with inclusions of plagioclase but they are scattered widely and sporadically throughout the rock. The largest garnets have leucocratic haloes consisting solely of plagioclase with a similar texture to that of the matrix plagioclase. Thin coarse-grained mafic layers in the garnet anorthosites consist of garnet–spinel–hornblende–biotite. In some cases it can be seen that garnets have nucleated on magnetite or another spinellid, as in the mafic layers just mentioned (fig. 14). Elsewhere they contain olive green hornblende while the matrix hornblende is bluish green, or they are associated with diopside grains in hornblende-poor layers. Obviously several different kinds of garnet are present. Recent analyses of two garnets by FRISCH (1971) indicate that garnet within the anorthositic rocks is magnesian almandine in composition, while garnet in a clinopyroxene garnet vein in anorthosite proves to be grossular. Late stage shearing has been noted where potash feldspar and quartz have been introduced into the plagioclase matrix, destroying the otherwise superb equilibrium textures.

Mineral Assemblages:

Anorthosites:	6.	Garnet-hornblende-biotite
	10.	Garnet-biotite
	11.	Garnet-biotite-muscovite
	12.	Garnet-biotite-muscovite-K-feldspar
Gabbroic anorthosites:	6.	Garnet-diopside-hornblende
	9.	Garnet-hornblende-cummingtonite-biotite
	15.	Garnet-biotite
Anorthositic gabbros:	12.	Garnet-hornblende-biotite
	13.	Garnet-hornblende-biotite-quartz
	14.	Garnet-hornblende-biotite-chlorite-ore
	17.	Garnet-biotite-quartz
Gabbros:	10.	Garnet-hornblende-quartz
Mafic layer:	14.	Garnet-spinel-hornblende-biotite-ore

Zone 9

The Upper Pyroxene Amphibolites have a granular texture, and a granoblastic to nematoblastic fabric, depending on the degree of hornblende lineation, which may be marked. Grain size is up to about 2 mm, and mafic and felsic minerals are present in approximately equal proportions. Hornblende is generally dark green or olive brown, less commonly

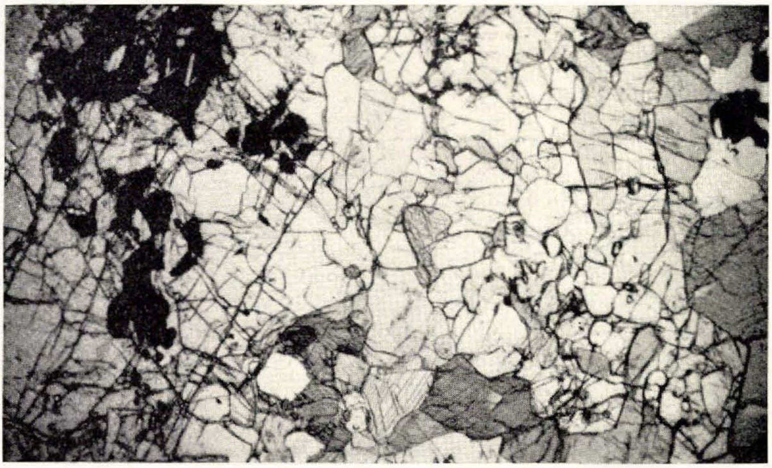

Fig. 14. Photomicrograph of spinel inclusions in garnet in a garnet hornblendite layer from within zone 8. Note the inclusions of mica in the spinel. GGU 73124, plain light, × 17.

bluish green. Layers lacking hornblende may have diopside garnet and plagioclase as their sole constituents; these layers grade into others having brown hornblende in addition to the anhydrous phases. Quartz is a characteristic accessory in the matrix as lenticular to sutured grains. The pyroxene may form a net with respect to the hornblende plagioclase matrix.

Mineral Assemblages:

Amphibolites:	4.	Two pyroxene amphibolite
	6, 7.	Hypersthene amphibolites
	8, 11.	Diopside amphibolites
	12.	Garnet-diopside amphibolite
	16.	Amphibolite (with quartz)
Mafic layer:	7.	Garnet-diopside-hornblende-plagioclase

Petrological summary

Assessment of mineral assemblages

On the basis of the mineral assemblages listed for the rocks of the Fiskenæsset complex it is evident that at least two metamorphic events have been responsible for their variety. No further comments need be made on the assemblages in general. They are strong evidence that granulite and amphibolite facies conditions have occurred in the Fiskenæsset area. This evidence is corroborated by studies made on the gneisses, but a discussion of the latter rocks is beyond the scope of this paper.

Nothing is shown, however, in the lists of mineral assemblages about the nature of the hornblende. In the above petrographic descriptions comments have been made on the nature of the amphiboles in each rock type and within each unit, as the variation in hornblende colour has proved a useful guide in assessing the metamorphic grade of the hornblende-bearing assemblages. There are at least four major types of hornblende: 1) brown to olive-brown; 2) dark olive-green to dark green; 3) pale green to bluish green; 4) pale green and pale brown (pargasites). These colour variations are thought to represent real variations in the chemistry of these amphiboles in response to both metamorphic conditions and host rock chemistry. Their content of minor elements such as chromium may also affect their colour.

Brown and olive brown hornblendes are limited in their occurrence to the border amphibolites of zones 1 and 9, and to a few layers within the dark gabbros of zone 4. Their deep brown colour is taken to be a result of both hornblende granulite facies conditions, and the relatively high iron and/or titanium content of their host rocks. Deep green to olive-green hornblendes are found within the layered rocks of zones 3 to 8. They are interpreted as being hornblende granulite facies hornblendes formed in rocks of lower iron tenure—the anorthosites and gabbros of the intrusion. Bluish green hornblendes are of widespread occurrence within zones 2 to 8, where they can be seen developing from pyroxene and from darker green hornblendes, often occurring only as narrow rims on the latter. These observations indicate that such bluish green amphiboles are of amphibolite facies origin. The pale green and pale brown pargasitic amphiboles are found within the ultramafic rocks at the 8–9 boundary almost exclusively. They are interpreted as representing amphibolite facies metamorphism of rocks rich in magnesium and aluminium and relatively low in iron.

Thus it is obvious that an assessment of the nature of the hornblendes in the Fiskenæsset rocks can provide further evidence of the interplay of the granulite and amphibolite facies metamorphisms in the production of the various mineral assemblages. Using evidence from the hornblendes, another petrographic observation can be assessed—that of the occurrence of hornblende-free layers with the mineralogical composition garnet–diopside–plagioclase. As these layers may be observed grading into others containing brown or olive-green amphiboles, in some cases with an intervening zone of vermicular diopside-plagioclase symplectites, these anhydrous layers must be interpreted in one of two ways. Either they represent relics of an earlier higher grade clinopyroxene granulite facies metamorphism, or they are rocks that remained dry under hornblende granulite facies conditions. Significantly, they occur mostly within the border amphibolites of zones 1 and 9 or as lenses and layers of isolated extent within zones 2 to 8. It is possible that they represent dry areas within the metavolcanic(?) amphibolites of the border zones, and inclusions of the same rocks or other dry sediments within the intrusion of zones 2 to 8. Analyses of the clinopyroxenes of these layers may solve the problem of their origin.

It has been noted that good reaction rims and coronas of amphibole on pyroxene are generally lacking in the rocks of the complex, and this is essentially the reason that amphibole colour has been invoked as a guide to the metamorphic grade of hornblende-bearing assemblages. Hence it is gratifying to note that in the few instances where such coronas have been found, the rimming amphiboles show colour zonation confirming that two metamorphisms have occurred. Some of the plagioclase pegmatites found as schlieren within the other rocks of the complex contain hypersthene or diopside rimmed successively by dark olive green hornblende and bluish hornblende. Epidote may occur as a further rim. As olive green amphibole rims the pyroxene, there is the possibility that this pyroxene is igneous in origin, and such coronas although rare may record the whole history of the complex: igneous pyroxene, hornblende granulite facies hornblende, amphibolite facies hornblende, and epidote amphibolite facies epidote. It would be interesting to investigate the oxygen isotope ratios in the pyroxene and in the successive rims.

Although amphiboles are the most abundant hydrous silicates in the Fiskenæsset rocks, micas do occur. Biotite is present in rock types ranging from felsic to ultramafic. Some red-brown biotites within the ultramafics of zone 2 may be derived from primary igneous biotites. They exhibit little evidence of being derived from either hornblende or bronzite in a hornblende pyroxenite. Phlogopite takes the place of biotite in magnesium-rich ultramafics, e.g. those along the contact of zones 8 and 9. Biotite is more common in zones 7 and 8 than in any other major zones of the complex. Muscovite is of uncommon occurrence being present only where assemblages are being downgraded to epidote amphibolite facies in felsic rocks. It is usually found with biotite or epidote. Biotite is also found where there has been local metasomatism, shearing, and introduction of material from the surrounding gneisses.

In summary it may be said that the mineral assemblages within the Fiskenæsset rocks reflect both the nature of the metamorphisms within the area and the original bulk chemistry of the rocks. It should especially be noted that the border amphibolites of zones 1 and 9 contain quartz, brown amphiboles, and show more prominent mineral orientation in the form of a penetrative foliation and/or lineation. The equivalent basic rocks of zones 3 to 8 are quartz-free cumulate gabbros, and the rocks clearly belonging to the stratiform igneous intrusion possess a typical granular fabric of equidimensional grains, and an equilibrium granoblastic-polygonal texture with planar grain boundaries and triple junctions shown particularly by plagioclase. A discussion of the textures in these rocks now follows.

Assessment of textures

It is clear from the foregoing descriptions that the rocks of the Fiskenæsset complex possess textural evidence that can assist in understanding their petrogenesis. The first and most obvious point is that the amphibolites of zones 1 and 9 are generally more foliated and lineated than any of the rocks of comparable mineralogy from within zones 2 to 8. This suggests immediately that the amphibolites may have had a different origin and history than the other metagabbros, anorthosites, etc.; mineralogical and chemical evidence bear out this viewpoint (see later).

Among the plagioclase-bearing rocks of zones 3 to 8, it is obvious that another textural observation is of importance in unravelling the history of the complex: the anorthosites, leucogabbros, and gabbros possess textures that can be related to plutonic igneous crystallisation followed by two metamorphic events.

In general, textures within these rocks vary from non-lineated, non-foliated aggregates of hornblende, plagioclase, and pyroxene, through cumulate, sub-ophitic and augen textures, to slightly lineated 'stretched' equilibrium textures. Rocks with pronounced foliation or lineation have been found to possess brown amphiboles and as such can be classed as amphibolite inclusions within zones 2 to 8. Careful examination of thin sections makes it increasingly clear how important and widespread the cumulate, augen, web, or mesh textures are in these rocks. Most anorthositic and gabbroic rocks can be inferred to have possessed such textures at one point in their history. Within the homogeneous and layered cumulate leuco-gabbro marker horizons of the complex, these textures are most noticeably preserved. It is probable that these textures stem from an original igneous texture in the rocks, modified by metamorphic events which variably recrystallised igneous megacrysts of plagioclase. The assumption is that the megacrysts of plagioclase (and megacrysts of orthopyroxene, clinopyroxene and olivine in ultramafic rocks) are relict igneous crystals. Original planar grain boundaries have been partially destroyed by high-temperature polygonisation and equilibration; the initial period of equilibration and partial megacryst recrystallisation is correlated with the granulite facies metamorphism.

It should be noted that JACKSON (1961) figures cumulate textures for olivine chromitites of the Stillwater complex that are very reminiscent of those described here. It thus seems possible that the texture of the cumulate leuco-gabbros and of the 'augen-type' chromite horizons are founded on igneous cumulate textures. That is, in those zones of the intrusion, plagioclase was the cumulus phase, and ferromagnesian silicates and chromite were the intercumulus phases, respectively.

Subsequent to the granulite facies metamorphism, the rocks were subjected to a further period of equilibration under amphibolite facies conditions. Textures characteristic of partial or complete amphibolite facies equilibration are abundant, especially within the homogeneous cumulate leuco-gabbro marker horizon. Smaller plagioclase polygons with mutual curved interfaces and 120° triple junctions are seen to replace megacrysts. This process proceeds by the megacrysts first developing sub-grain mosaics, and then becoming progressively incorporated into the granular secondary plagioclase matrix. Grain boundaries of the large crystals may show re-entrant angles where partially replaced by the secondary mosaic (fig. 9). As yet no chemical evidence is available on the changing An values of the plagioclase that must accompany this recrystallisation, but the phenomenon of different An value plagioclases in one rock is known from many other areas (cf. KALSBEEK, 1970). Both the primary megacrysts and secondary polygonal crystals may be twinned and zoned. Another phenomenon accompanying this plagioclase recrystallisation is in some instances the development of hornblende microcrystals within the body of the megacrysts (figs 8, 9). This phenomenon probably means that the resulting secondary plagioclase will have a higher An value than the primary megacryst, as alkalis will preferentially enter the hornblende (J. V. SMITH, pers. comm.). These hornblende microcrystals eventually are crowded to grain-boundary areas of the secondary plagioclase, but they may persist cutting across the secondary equilibrium texture, as evidence that a large megacryst once occupied the site (fig. 10).

Evidence of two metamorphic episodes is also evident in the relation of the plagioclase to the mafic minerals. While the plagioclase megacrysts are being replaced by smaller polygons, the mafic minerals interstitial to the plagioclase change as well. Initially large pyroxene and green hornblende crystals break down into sub-grain mosaics, and then these smaller grains are separated along grain boundaries by plagioclase invading the mafic mass from the matrix. Hypersthene characteristically forms rather grotesque sieve or sponge textures with respect to the plagioclase (fig. 6). The grain size of the mafic minerals is usually smaller than that of the plagioclase, and this indicates that the plagioclase recrystallisation proceeds more quickly—i. e. the matrix recrystallisation was in an advanced state before the mafic minerals began to recrystallise. Plagioclase that invades the mafic areas is generally of a smaller grain size as well, so that the resulting texture in many rocks is of augen of plagioclase polygons surrounded by rims composed of finer-grained plagioclase and rounded, small grains of the mafic minerals. This texture is modified by the development of bluish green hornblendes which replace and include relics of the higher grade mafic minerals—green to brown hornblende, hypersthene and diopside. In some cases it is clear that this bluish green hornblende is oriented in response to stress conditions coincident with the amphibolite facies conditions, and thus there exists a foliated hornblende–plagioclase rock, with an overall nematoblastic texture, which contains a relict texture of plagioclase augen surrounded by mafics and fine-grained plagioclase as described above.

Under lower *P-T* conditions, epidote will replace the amphibolite facies hornblende and plagioclase in part, the boundaries of the hornblende with the plagioclase will begin to show vermicular growth of untwinned plagioclase (albite?) or quartz, and plagioclase polygons will show a rim of lower An content plagioclase as well. Further late stage alteration will produce extensive epidote development throughout the whole rock, accompanied by chlorite, and by scapolite in less advanced cases.

It should be noted that green hornblende rimmed by slightly more bluish hornblende is characteristic of cumulate rocks in which the plagioclase megacrysts are being replaced by the polygons, but in which the mafic minerals have not yet been separated into smaller grains. The shape of interstitial mafics in these rocks is also important. Masses of green hornblende, for example, have curved boundaries, concave inwards toward the hornblende mass. This is obviously due to the effect of the plagioclase matrix recrystallising before the mafic clots. Only when the ratio of mafic minerals to plagioclase becomes high (as in the mafic gabbros), is the reverse relationship observed, i.e. a matrix of recrystallised mafic minerals containing sutured interstitial plagioclase grains.

Web or augen textures are observed in ultramafic rocks as well. In their case, hornblende recrystallises first with respect to the pyroxenes and olivine, and the latter minerals end up as clots and rims of finer grain size in a hornblende matrix.

It should be noted that the textural evidence discussed above argues against ideas of plagioclase homogenisation under metamorphic conditions, as postulated by BOWDEN (1970). It is thus probable that a relict systematic change in An content of the plagioclase with position in the layered complex may be found by analysis of plagioclase in the layered rocks, especially if megacrysts are chosen for analysis.

STRATIGRAPHY
PART II

This section deals with the present stratigraphy that has been considerably complicated by folding and tectonic thinning.

Fourteen measured sections were made in traverses across the complex on Qeqertarssuatsiaq. Sections 5–10 and 11–14 occur on opposite limbs of an isoclinal fold, the closure of which lies to the south-west of the area (fig. 2). The traverses are shown diagrammatically and intercorrelated in Plate 2, which illustrates the following stratigraphic features.

1) There are considerable variations in thickness of the layers along strike; this is largely caused by attenuation along limbs and thickening in fold cores. Usually most major zones are equally or correspondingly thinned in the fold limbs with the result that the same succession is seen there as in the cores. Besides the thickening of all zones in the fold cores, however, there are commonly additional layers present, in particular layers of dark gabbro and cumulate leuco-gabbro.
2) The complex has a double or repeated stratigraphy arranged symmetrically in reverse order. One passes from zone 1 at the 'outside' through zones 2 to 8 to zone 9 at the 'centre' and then symmetrically again through zones 9 to 1. In spite of the fact that some zones are locally missing due to tectonic thinning, the symmetry is easily recognised in most areas.

3) The higher zones 6–9 close internally so that the lower mafic zones 1–5 continue along the strike (e.g. to the right of section 8 and to the left of section 11 in Plate 2).

4) Zone 7 chromitites are confined to the upper anorthosites and are antipathetic with all other zones. Therefore, where the complex consists only of the lower mafic zones beyond the internal closures, e.g. traverses 1, 2, 3, 9, 10 and 14 of Plate 2, it is clear that the chromitites cannot be present and for prospecting purposes need not be looked for.

The pyroxene amphibolites of zones 1 and 9 (figs 15, 16, 17) occur along the borders and centre of the complex. They may be up to 200 m

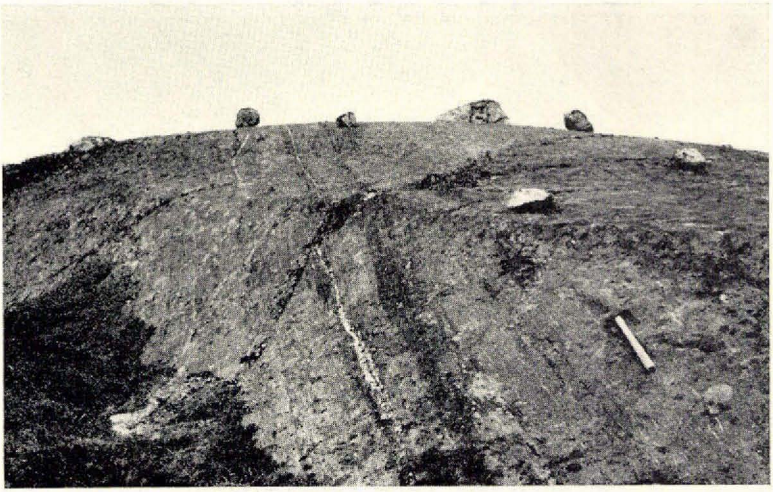

Fig. 15. Zone 1: General view of pyroxene amphibolite showing layering. Hammer shaft is 40 cm in length. Southern central Qeqertarssuatsiaq.

Fig. 16. Zone 9: Pyroxene amphibolite showing narrow banding and foliation, with pegmatitic schlieren. West side of peninsula, west of Qeqertaq.

thick (section 13) and are often attenuated in fold limbs. They commonly have marked centimetre-scale layering reflecting variable concentrations of hypersthene, diopside, garnet, hornblende and plagioclase. The presence of garnet in these rocks distinguishes them from many garnet-free amphibolite layers in the gneisses. However, garnet appears to be more common in some areas such as west Qeqertarssuatsiaq than others. They locally contain ultramafic layers up to a few metres thick and scattered small hypersthene pegmatites. Both amphibolites often have

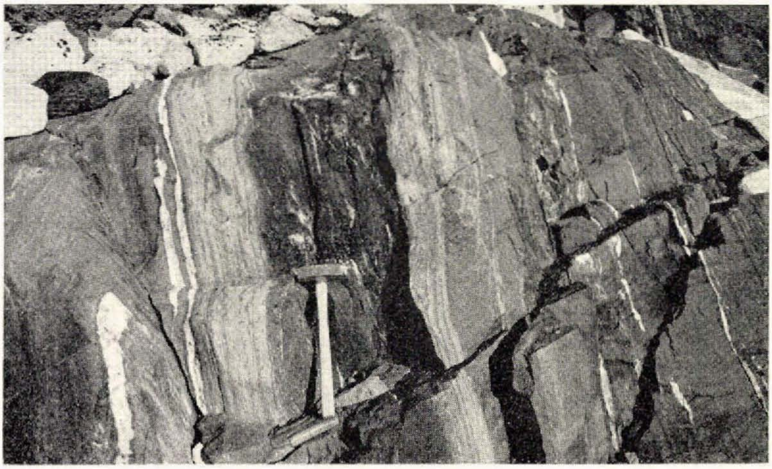

Fig. 17. Zone 9: Pyroxene amphibolite with broad bands of mafic and felsic material, the latter garnetiferous. Centre of the complex (cf. fig. 16). West side of peninsula, west of Qeqertaq.

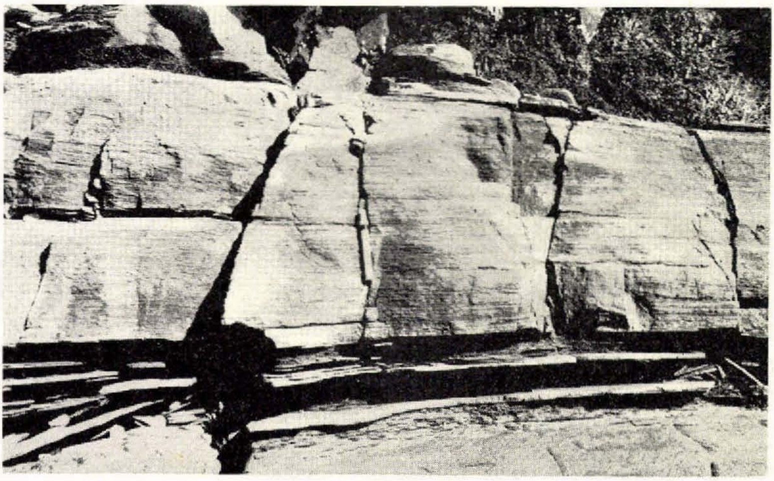

Fig. 18. Zone 2: A magnetite-layered serpentinised hornblende dunite. Pen is approx. 12 cm long. West side of peninsula, west of Qeqertaq.

a rusty sulphide-bearing zone up to about 10 m thick against the adjacent ultramafics.

Some field evidence suggests that the upper amphibolites may be metavolcanics: between zones 8 and 9 there are minor marble layers up to 2 m thick and the amphibolites themselves contain rare layers of mica schists up to 1 m thick and rare sillimanite-plagioclase-corundum biotite nodules. The high aluminium content of these nodules would suggest they are derived from supracrustal bauxitic or argillaceous material. On the other hand there is little unequivocal field evidence as to the origin of the lower amphibolites. Chemically, however, they show dissimilarities to other rocks definitely belonging to the intrusion, and are probably metavolcanic on the basis of this evidence (see 'Geochemistry').

Throughout the complex there are conformable layers of amphibolite as inclusions within the other zones, as well as coarser-grained mafic rocks as layers and lenses, pods and lenses of carbonate-rich material etc., which may be interpreted as rafted supracrustal material from the borders of the intrusion.

The ultramafic zone 2 is up to 100 m thick but its presence is highly variable due to extensive thinning and boudinage, and thus the rocks occur in isolated tectonic pods and lenses. They are commonly spinel dunites (fig. 18), peridotites and pyroxenites, and locally there are hornblendites. The best stratigraphic sequence found to date within zone 2 is on the south side of Fiskenæsfjorden, as follows:

	Thickness	*Rock Type*
Top	40 cm	Biotite–hornblende bronzitite
	30 cm	Biotite–hornblende pyroxenite
	1 m	Hornblende peridotite
	2 m	Hornblende peridotite (weakly serpentinised)
	2 m	Hornblende peridotite
	1 m	Magnetite-rich serpentinised hornblende dunite
	15 cm	Hornblende dunite (sheared, weakly serpentinised)
	30 cm	Hornblende dunite
	15 cm	Hornblende dunite (with secondary talc and anthophyllite)
	2 m	Diopside hornblendite

These layers have sharp boundaries and continue along the strike for several metres. It should be noted that serpentinisation is confined to certain narrow layers.

Zone 3 has a complex stratigraphy of intercalated streaks and thin layers of anorthositic gabbro, gabbroic anorthosite, layered cumulate gabbro, dark gabbro and minor anorthosite (fig. 19). This zone reaches a maximum thickness of 100 m and it has been possible to divide it further into sub-units recognisable along the strike. There are locally 1 m thick ultramafic layers in the lower part of the zone which are probably an extended part of zone 2.

Dark gabbros (fig. 20) are characterised by their dark appearance both in the field and on the aerial photographs. They are listed as zone 4 as they occupy this stratigraphic position throughout the most exten-

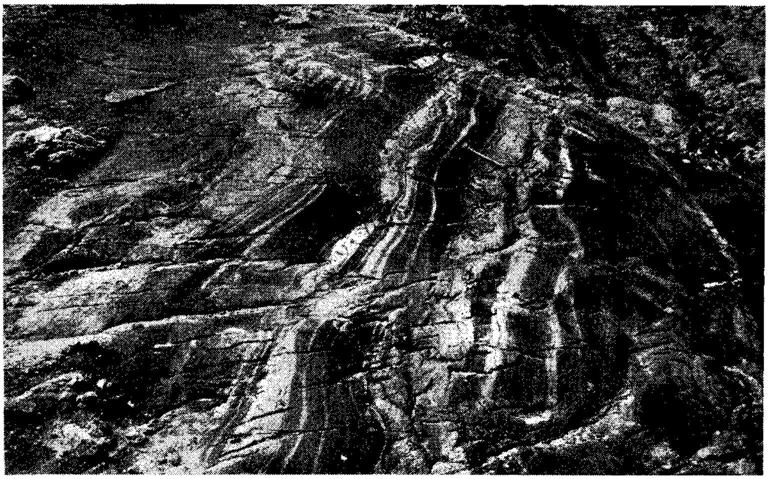

Fig. 19. Zone 3 showing the first (lowest) anorthositic layers in a largely mafic sequence. Area north-east of Fiskenæshumplen, Qeqertarssuatsiaq.

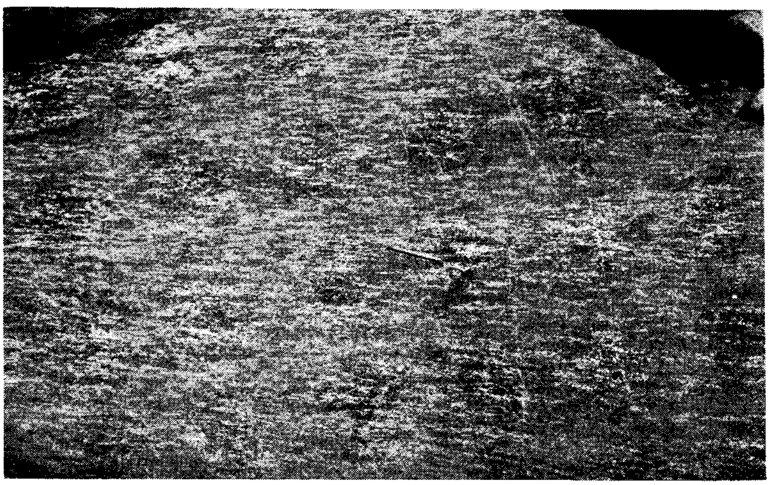

Fig. 20. Zone 4: Dark gabbro showing well layered hornblende and pyroxene. Scale is given by the match. West side of peninsula, west of Qeqertaq.

sive areas of central and east Qeqertarssuatsiaq. However, east of Taseq similar rocks occur between zones 5 and 6, and west of Taseq within both zones 5 and 6. This change in stratigraphic position should not be seen as zone 4 occurring locally within zones 5 and 6, but rather as zones 5 and 6 having their own dark gabbro layers. They probably vary in composition accordingly (analysed dark gabbros have a lower Fe/Mg ratio than the cumulate leuco-gabbro of zone 5, and a higher value than rocks of zone 3). The dark gabbros are up to 40 m thick and have a

Fig. 21. Zones 3–4: Chromite-layered hornblende bronzitite. A 30 cm thick layer. Southern central Qeqertarssuatsiaq.

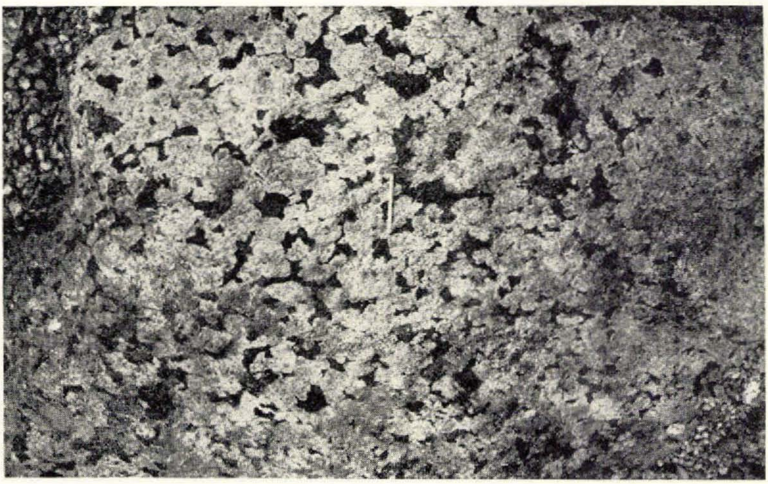

Fig. 22. Zone 5: Leuco-gabbro showing typical cumulate texture. Scale is given by the match. East of Fiskenæshumplen.

distinct affinity for the cumulate leuco-gabbros with which they are often associated. Locally they also contain 1 m thick ultramafic layers.

Three layers of chromite-layered bronzitite and one of chromite gabbro about 30–60 cm thick are so far known throughout the complex (fig. 21). They definitely occur in the lower mafic part of the complex, but their precise position in the succession is uncertain. In traverse 13 they appear to lie along the contact of zones 3 and 5 where the dark gabbro zone is missing.

Fig. 23. Zone 6: Anorthosite with thin hornblende laminae. Pen is approx. 12 cm long. Southern central Qeqertarssuatsiaq.

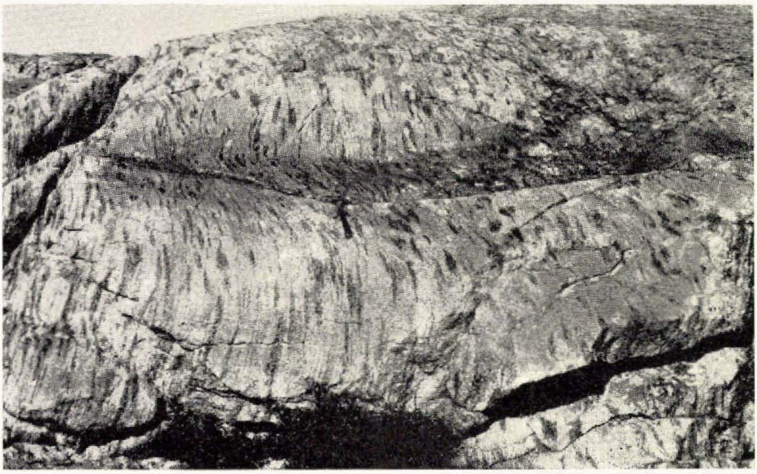

Fig. 24. Zone 6: Anorthosite containing elongate hornblendic schlieren. Hammer shaft is 40 cm long. Eastern Qeqertarssuatsiaq.

The cumulate leuco-gabbro of zone 5 is one of the most characteristic and easily recognisable rocks of the complex, on account of its distinctive cumulate texture and its lack of layering and mineral orientation (fig. 22). It is usually present as a marker horizon throughout most of the complex reaching a maximum thickness of 150 m in the largest fold cores. It keeps its stratigraphic position below the main anorthosite zone and can usually be found symmetrically on both sides of the complex. It locally contains 1 m thick ultramafic layers. On north-east Qeqertarssuatsiaq beyond the internal closure of the inner zones 6–9,

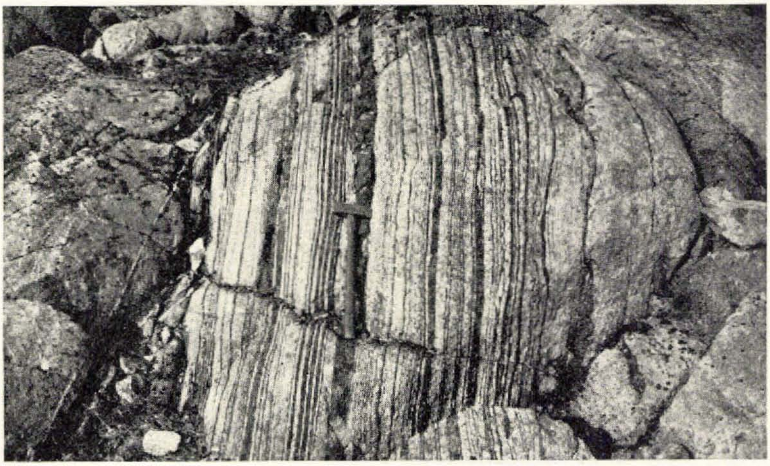

Fig. 25. Zone 7: Chromite horizon showing extremely well-layered alternating chromitite and plagioclase. Eastern Qeqertarssuatsiaq.

Fig. 26. Zone 7: Chromite horizon showing broad chromitite layers. Chisel is 22 cm long. Eastern Qeqertarssuatsiaq.

the whole width of the complex is occupied by cumulate leuco-gabbro (250 m stratigraphic thickness, section 14, Plate 2).

The anorthosite of zone 6 varies to gabbroic anorthosite where it has more than 10 % hornblende. It contains hornblende-rich streaks and lenses of anorthositic gabbro, particularly in a layer against the zone 7 chromite horizons (figs 23, 24). This anorthosite layer bordering the chromitites is often garnet-bearing and so has affinities with zone 8. Alternatively, it could be considered that the chromite horizons occur in the lower part of zone 8; in other words they are often bordered by garnet-bearing anorthosite on both sides. Zone 6 reaches a maximum

Fig. 27. Zone 7: Chromitite rich in plagioclase megacrysts. Scale is given by pencil. East of Taseq.

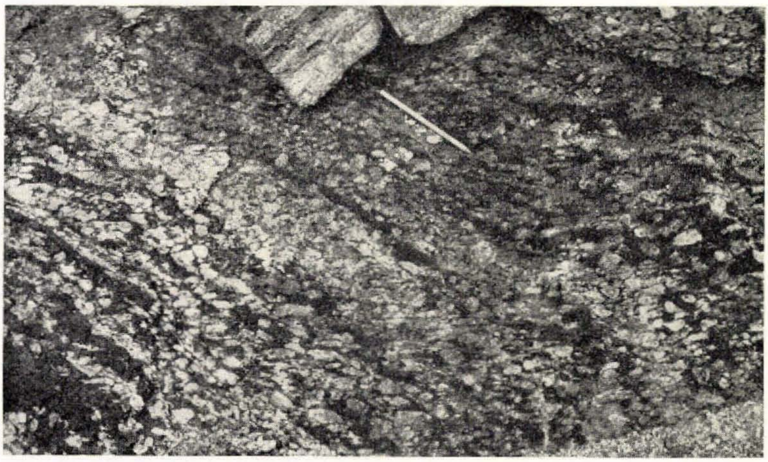

Fig. 28. Zone 7: Chromitite with plagioclase megacrysts deformed as augen. Same outcrop as figure 27. Note the lineation of the augen oblique to the layering. East of Taseq.

thickness of 130 m, but where it is thin, it may consist entirely of garnet-bearing anorthosite (see traverses 7 and 8 of Plate 2).

The zone 7 chromitites (figs 25–28) are grouped together with alternating anorthositic laminae into chromite horizons which occupy the same stratigraphic position throughout the whole length of the complex that has been studied. Although the chromite horizons reach a maximum thickness of 10 m in two fold cores on Qeqertarssuatsiaq, they are usually between 30 cm and 3 m thick. Because of their competency relative to that of the surrounding anorthosite they have been much attenuated

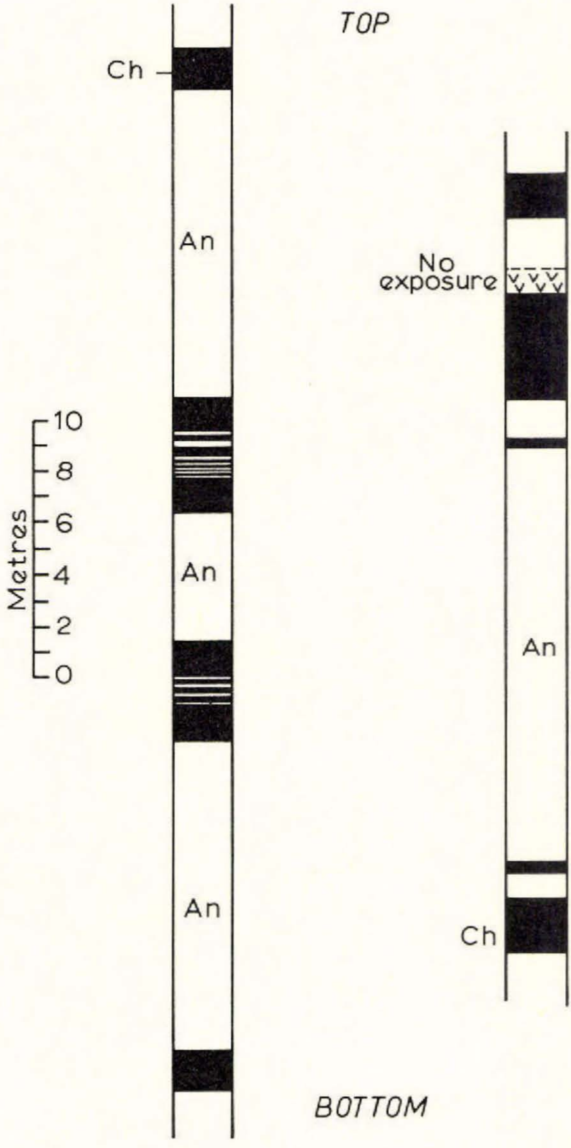

Fig. 29. Two measured sections of chromite horizons of zone 7, each with an aggregate thickness of about 10 m, made up of smaller chromitite units (black) interlayered with anorthositic units (white).

and disrupted by boudinage in fold limbs, so that they occur as discontinuous layers and lenses as thin as 2 to 10 cm. These can still be recognised within the white anorthosite where they keep to their stratigraphic position. The presence of garnets in the anorthosite has been found to be a valuable marker for locating the relics of the chromite horizons. Thus by positioning oneself in the correct position in the

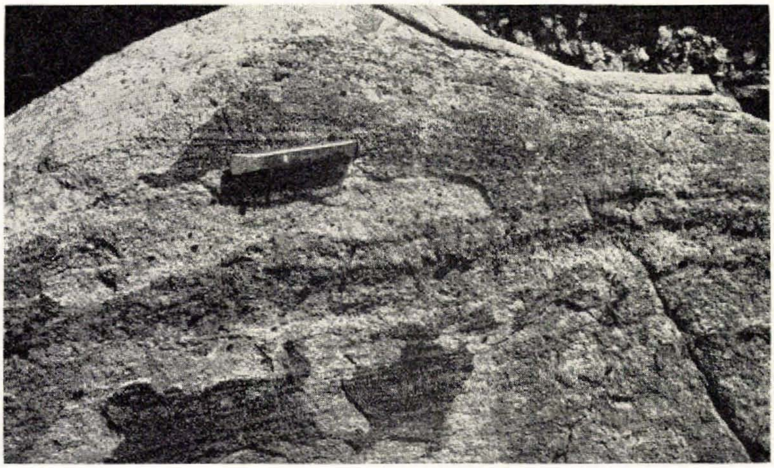

Fig. 30. Zone 8: Garnet anorthosite with garnets confined to certain layers. Southern central Qeqertarssuatsiaq.

stratigraphy, i.e. in the lower part or on the cumulate leuco-gabbro side of the garnet-bearing anorthosite zone, the chromite horizons can be readily located. By doing so, it has been found that there is far more chromite in the complex than was reported by GHISLER & WINDLEY (1967).

There is usually one chromite horizon in zone 7, but in places there are two separated by up to 30 m of garnet anorthosite (section 13, Plate 2). Where they are 10 m thick, they are in the form of several smaller horizons about 1 to 1.5 m thick interbedded with anorthositic layers throughout stratigraphic thicknesses of up to 40 m. Two measured sections of chromite horizons are given in fig. 29. The section on the left gives the impression of a cyclic succession of anorthosites and minor chromite seams. The layers marked in black consist of chromitite seams only a few centimetres thick alternating with anorthositic laminae of similar size.

Garnet anorthosite constitutes a marker zone of the complex reaching a maximum thickness of about 70 m. The hornblende content may increase giving rise to gabbroic anorthosite and even locally to anorthositic gabbro. The presence of garnets is the diagnostic feature of this zone; they are often concentrated in discrete layers up to about 1 cm thick (fig. 30) or are scattered sporadically throughout the anorthosite. They are commonly only a few millimetres across and thus require close inspection for identification. Their development is irregular, but throughout the area studied they are almost always present. In a few localities they are absent, but one has to pass not far along the strike before they appear. Within this zone there are rare but distinctive coarse-grained layers up to about 30 cm thick with garnet-biotite or garnet-biotite-plagioclase, and also rare 1 m thick ultramafic layers.

Between zone 8 and the amphibolites of zone 9 ultramafic rocks may be developed sporadically. They occur as isolated short layers and lenses and are extensively thinned by boudinage. Some are magnetite-rich, while others are magnesian spinel-bearing peridotites and pyroxenites. As the zone 9 amphibolites may be supracrustal in origin, rocks occurring on the zone 8/zone 9 contact may be genetically connected to the plutonic igneous rocks or to supracrustal volcanic rocks and their associates. On the basis of what is now known of the bulk chemistry of the amphibolites (see 'Geochemistry') and of the differentiation of the igneous rocks, the iron-rich (i.e. magnetite-bearing) ultramafics in this contact zone may be related to the amphibolites. It is noticeable that where the upper amphibolite of zone 9 is absent from the stratigraphy of the complex, so too are ultramafic rocks next to zone 8.

The magnesian ultramafics along the zone 8/zone 9 contact are more difficult to interpret. They may occur in association with sapphirine- and corundum-bearing rocks which have also been localised along this particular anorthosite/amphibolite contact in a number of cases. A preliminary study of the sapphirine localities (HERD, WINDLEY & GHISLER, 1969) indicated that such ultramafic rocks were the parental material from which the sapphirine rocks developed. It was postulated that the ultramafics were igneous in origin, and genetically connected to the layered igneous rocks of the complex. This view of the connection of the sapphirine rocks and the magnesian ultramafics not only to each other, but also to the chromite-layered anorthosite complex, was supported by the presence of spinel layering in the ultramafics, spinel acting as the parent mineral for sapphirine formation, and by unpublished geochemical data showing that minerals in both ultramafics and the sapphirine rocks contained traces of Cr, Ni, V etc.

Careful field mapping by the second writer (RKH), and discoveries of new sapphirine localities by many GGU geologists and personnel of Platinomino A/S, indicate that previous views on the origin of the sapphirine rocks must be modified. It was found that spinel-bearing magnesium-rich ultramafic rocks were not characteristic at all sapphirine localities. Further, sapphirine-bearing rocks were found to be interlayered with anorthosite (i.e. to merge with zone 8) and gradually to pass into plagioclase-poor material nearer to the zone 9 contact at several localities. Amphibolite was also found within areas of sapphirine rock. In addition layers and lenses of material were found at the sapphirine localities which are almost certainly supracrustal in origin, such as sulphide-bearing carbonate rocks, sillimanite-rich rocks, and ultramafic(?) rocks in which microscopic study has revealed rounded zircons and monazite? included in silicate and oxide phases.

No account is given in the present work of the chemistry of the complicated suite of rocks occurring along the zone 8–zone 9 contact. However, it may be noted that whole rock analyses are now available and indicate that the magnesian ultramafic rocks may still tentatively be related to the layered igneous rocks. On the other hand the chemical data on the sapphirine rocks support a new hypothesis, inferred from the field data, that the sapphirine rocks originated by partial assimilation of aluminous supracrustal material by the igneous anorthositic rocks. Such aluminous material would have been hornfelsed and metasomatised by the magma, acquiring not only major elements such as Ca, Mg, and Fe, but also traces of Cr, Ni, V etc. as observed. Loss of Si and alkalis from the supracrustal material would have resulted in spinel-rich mineral assemblages. Further metasomatism and metamorphism would then give rise to sapphirine-bearing rocks as previously postulated. Minor amounts of sapphirine occurring in the magnesian spinel-layered ultramafics probably formed contemporaneously with sapphirine developed from spinel-rich assimilated supracrustal material. Supracrustal material less rich in aluminium, as well as blocks of amphibolite and ultramafic rocks of zone 9, also partially assimilated by the anorthositic magma, must have given rise to the carbonate rocks, hornblendites, and sillimanite rocks observed in association with the sapphirine rocks, or indeed found elsewhere within the upper part of zone 8.

It should be noted that sapphirine-bearing aluminous xenoliths occurring within the Critical Zone of the Bushveld complex are strikingly similar both mineralogically and texturally to some of the sapphirine-bearing and associated rocks from Fiskenæsset (cf. Willemse & Viljoen, 1970). Textures characteristic of both thermal and regional metamorphism have now been found in the Fiskenæsset sapphirine- and corundum-bearing rocks.

GEOCHEMISTRY

The aim of this section is not to present a detailed discussion of the geochemical characteristics of the individual rock types and zones, or of the differentiation trend of the complex as a whole, but rather to give preliminary geochemical evidence that indicates the way-up of the complex and that corroborates the symmetrically reversed stratigraphy obtained from the field. The geochemical analyses were made by the third writer (BOWDEN, 1970) at Leeds University during the summer of 1970 when the other two writers were in Greenland. Thus his conclusions that the complex has a double reversed stratigraphy were made from geochemical evidence alone, totally independent of similar field conclusions made by the first writer (BFW) at the same time in the Fiskenæsset area.

Most major elements were determined with the Philips 1212 automatic X-ray fluorescence spectrometer, while ferrous iron oxide, bonded and unbonded water and the oxides of sodium and potassium were determined by wet chemical methods. The silica and chromate contents of three samples were also determined by chemical methods. Trace elements were analysed with the 1212 XRF spectroscope and by ultraviolet spectrography with the Hilger-Watts spectrograph. The precision of the analyses is discussed in detail by BOWDEN (1970). In general the major element analyses performed by XRF are accurate to $\pm 1\ ^0/_0$, while most trace element values are accurate to $\pm 12\ ^0/_0$.

The stratigraphic position of 18 analysed samples (Table 1) with respect to the zones of the complex is given below (for locality see fig. 1).

Analyses in Table 1 show that these rocks are basic to ultrabasic in composition, while the C.I.P.W. norm calculations show that they are also mafic to ultramafic. Analysis 6 has not been plotted on variation diagrams, because it contains $6.14\ ^0/_0$ Cr_2O_3 and FeO has not been determined.

To illustrate the course of differentiation of the intrusion the $Na_2O + K_2O - Fe_2O_3 + FeO - MgO$ triangular diagram has been used. Figs 31 A and 31 B demonstrate that the southern and northern halves of the intrusion differentiate 'inwards' towards the 'centre', thus re-

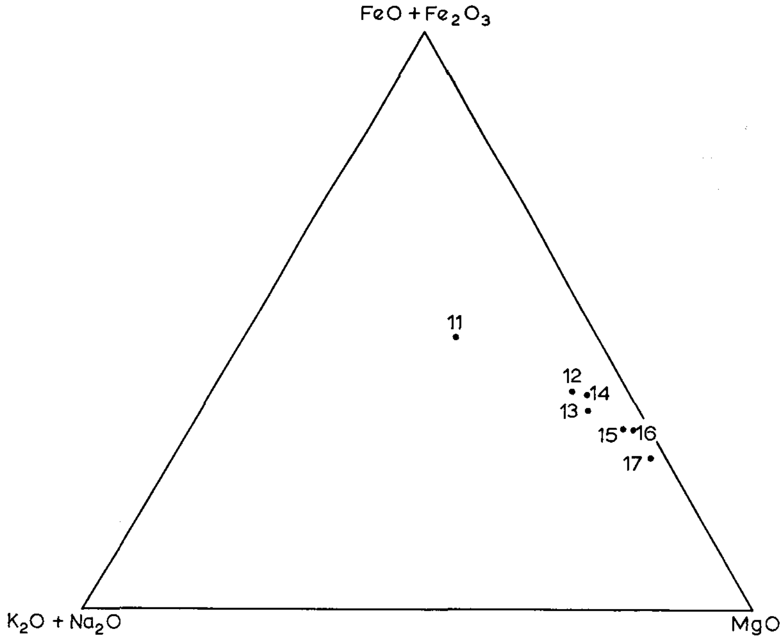

Fig. 31 A.

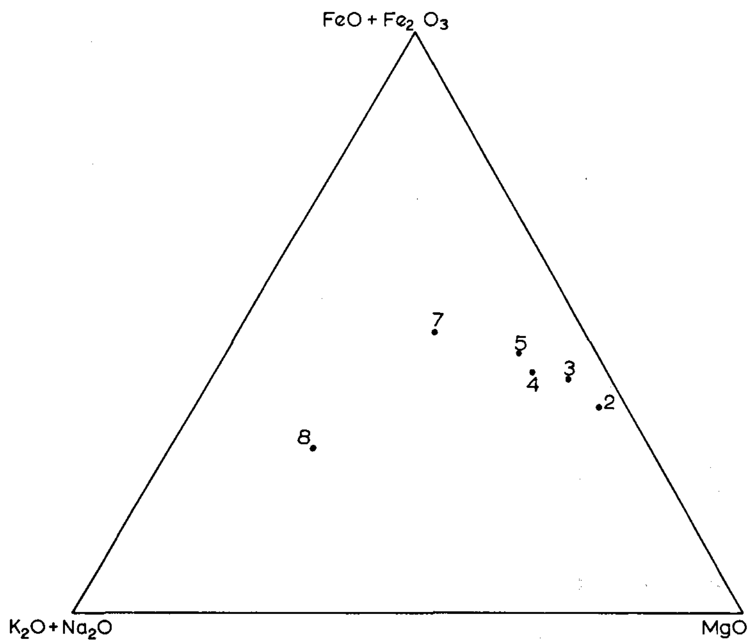

Fig. 31 B.

Fig. 31. AFM (wt. %) diagrams for the northern (A) and southern (B) halves of the intrusion. The analysed samples are numbered according to their place in the traverse.

Table 1. *Major element analyses of the Fiskenæsset complex*

Sample No.	1	2	3	4	5
Zone	*1*	*2*	*4*	*4*	*5*
GGU No.	86921	74482	86923	86922	86924
SiO_2	49.85	42.40	45.15	47.30	46.45
TiO_2	0.94	0.12	0.08	0.17	0.37
Al_2O_3	12.20	13.80	15.50	21.40	23.15
Cr_2O_3	n.d.	n.d.	n.d.	n.d.	n.d.
Fe_2O_3	3.23	4.57	2.85	1.57	1.32
FeO	13.31	7.23	8.29	4.98	5.12
MnO	0.22	0.16	0.15	0.11	0.09
MgO	6.61	20.00	14.82	7.53	6.31
CaO	8.08	8.50	9.79	13.95	14.15
Na_2O	3.45	1.21	1.43	1.56	1.58
K_2O	0.24	0.15	0.37	0.22	0.16
P_2O_5	0.05	0.03	0.02	0.01	0.05
H_2O+	0.59	1.40	1.28	1.13	0.92
H_2O-	0.09	0.06	0.20	0.11	0.08
	98.9	99.6	99.9	100.0	99.8
		C.I.P.W. Norms			
Q	—	—	—	—	—
co	—	—	—	—	—
or	1.4	0.9	2.2	1.3	0.9
ab	29.2	10.2	12.1	13.2	13.4
an	17.1	31.8	34.8	50.7	55.6
ne	—	—	—	—	—
di	18.9	8.1	11.0	14.8	11.6
hy	17.4	2.3	7.0	7.6	5.5
ol	7.6	38.0	27.1	8.5	9.1
mt	4.7	6.6	4.1	2.3	1.9
il	1.8	0.2	0.2	0.3	0.7
ap	0.1	0.1	—	—	0.1
H_2O	0.7	1.5	1.5	1.2	1.0
	98.9	99.7	100.0	99.9	99.8

n.d. = not determined

(continued)

Table 1 (cont.)

Sample No.	6*	7**	8	9	10
Zone	*7*	*7*	*8*	*9*	*9*
GGU No.	86925	74489	92579	86928	74484
SiO_2	29.00	45.10	47.10	47.10	47.30
TiO_2	0.57	0.23	0.13	0.82	0.81
Al_2O_3	27.10	27.60	30.60	15.70	16.60
Cr_2O_3	6.14	n.d.	n.d.	n.d.	n.d.
Fe_2O_3	16.00	1.19	0.57	2.75	2.31
FeO	—	3.21	1.36	9.23	8.10
MnO	0.25	0.06	0.03	0.22	0.22
MgO	6.98	2.70	1.45	6.95	8.24
CaO	6.70	14.90	13.70	13.85	12.40
Na_2O	0.60	1.91	2.52	1.59	2.16
K_2O	1.83	0.15	0.96	0.27	0.44
P_2O_5	0.02	0.03	0.02	0.08	0.04
H_2O+	0.92	0.32	0.45	0.88	0.52
H_2O-	0.12	0.04	0.07	0.12	0.04
	96.2	97.4	99.0	99.6	99.2

C.I.P.W. Norms

	6*	7**	8	9	10
Q		—	—	—	—
co		—	0.6	—	—
or		0.9	5.6	1.6	2.6
ab		14.6	16.9	13.5	18.3
an		66.3	67.8	34.9	34.3
ne		0.8	2.4	—	—
di		6.1	—	27.3	21.9
hy		—	—	9.9	2.3
ol		6.1	4.0	5.6	14.3
mt		1.7	0.8	4.0	3.3
il		0.4	0.2	1.6	1.5
ap		0.1	—	0.2	0.1
H_2O		0.3	0.5	1.0	0.6
		97.3	98.8	99.6	99.2

* chromite plus matrix
** matrix to chromitite

(continued)

Table 1 (cont.)

Sample No.	11	12	13	14
Zone	*6*	*6*	*5*	*5*
GGU No.	86929	86930	86931	86932
SiO_2	46.20	44.30	43.80	44.50
TiO_2	0.47	0.06	0.09	0.09
Al_2O_3	27.90	27.25	21.60	21.05
Cr_2O_3	n.d.	n.d.	n.d.	n.d.
Fe_2O_3	1.22	1.37	1.96	1.81
FeO	3.45	3.14	4.83	5.88
MnO	0.08	0.06	0.10	0.12
MgO	3.12	6.31	11.20	11.60
CaO	14.80	14.80	12.80	11.65
Na_2O	1.94	0.93	1.38	1.27
K_2O	0.16	0.06	0.18	0.07
P_2O_5	0.04	—	0.04	0.03
H_2O+	0.44	0.55	1.22	1.37
H_2O-	0.03	0.05	0.09	0.06
	99.9	98.9	99.3	99.5

C.I.P.W. Norms

Q	—	—	—	—
co	—	—	—	—
or	0.9	0.4	1.1	0.4
ab	16.3	7.9	9.4	10.7
an	66.9	70.0	52.2	51.5
ne	—	—	1.3	—
di	5.1	2.7	8.8	4.9
hy	—	5.5	—	6.5
ol	7.3	9.8	22.2	21.1
mt	1.8	2.0	2.8	2.6
il	0.9	0.1	0.2	0.2
ap	0.1	—	0.1	0.1
H_2O	0.5	0.6	1.3	1.4
	99.8	99.0	99.4	99.4

(continued)

Table 1 (cont.)

Sample No.	15	16	17	18
Zone	*3*	*2*	*2*	*1*
GGU No.	74487	86935	86927	86934
SiO_2	48.30	43.80	47.70	45.50
TiO_2	0.14	0.22	0.15	1.44
Al_2O_3	15.70	9.62	6.86	13.55
Cr_2O_3	n.d.	1.67*	0.23*	n.d.
Fe_2O_3	0.90	3.52	3.99	3.23
FeO	4.74	7.23	5.55	14.21
MnO	0.14	0.17	0.19	0.24
MgO	11.60	22.70	25.60	5.74
CaO	15.60	6.99	6.35	10.60
Na_2O	0.75	0.77	0.72	2.30
K_2O	0.01	0.17	0.17	1.18
P_2O_5	0.03	0.03	0.03	0.09
H_2O+	0.36	1.40	0.83	1.54
H_2O-	0.08	0.07	0.25	0.03
	98.4	98.4	98.6	99.7
C.I.P.W. Norms				
Q	—	—	—	—
co	—	—	—	—
or	0.1	1.0	1.0	7.0
ab	6.3	6.5	6.1	17.5
an	39.4	22.3	15.0	23.2
ne	—	—	—	1.1
di	30.2	9.6	12.9	24.2
hy	14.6	20.3	35.7	—
ol	5.7	29.9	20.6	17.6
mt	1.3	5.1	5.8	4.7
il	0.3	0.4	0.3	2.7
ap	0.1	0.1	0.1	0.2
H_2O	0.4	1.5	1.1	1.6
	98.4	96.7	98.6	99.8

* values not included during norm calculation
Analyst A. A. Bowden

Sample No.	*GGU No.*	*Zone*	*Description (mineral assemblage no. in brackets)*
18	86934	*1*	Foliated, lineated cg. amphibolite (11)
17	86927	*2*	Chrome-spinel-layered hornblende peridotite (4)
16	86935	*2*	Chrome-spinel-layered hornblende peridotite (4)
15	74487	*3*	Noritic cumulate gabbro (1)
14	86932	*5*	2–10 cm thick layer of gabbro (4)
13	86931	*5*	Layered gabbro (3)
12	86930	*6*	Medium to fine-grained anorthosite (1)
11	86929	*6*	Patchy gabbroic anorthosite (8)
10	74484	*9*	Weakly foliated amphibolite (8)
9	86928	*9*	Foliated, lineated, fine-grained amphibolite (11)
8	92579	*8*	Garnetiferous gabbroic anorthosite (15)
7	74489	*7*	Garnetiferous anorthositic gabbro (13)-matrix to chromitite
6	86925	*7*	Chromite-layered anorthositic gabbro (2)
5	86924	*5*	Cumulate anorthositic gabbro (10)
4	86922	*4*	Layered anorthositic gabbro (10)
3	86923	*4*	Coarse gabbro (4); 3 m inclusion in 86922
2	74482	*2*	Lineated hornblende peridotite (6)
1	86921	*1*	Foliated medium-grained amphibolite (3)

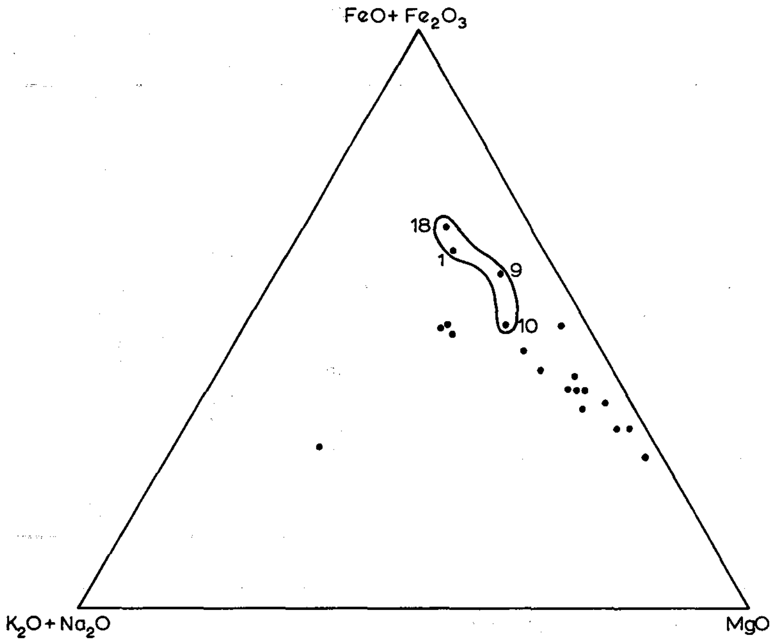

Fig. 32. An AFM diagram for all analyses of the complex compiled from those in figs 31 A and 31 B together with 3 analyses from WINDLEY (1969 a). The four analyses of the border amphibolites fall off the differentiation trend of the intrusion and are accordingly grouped together.

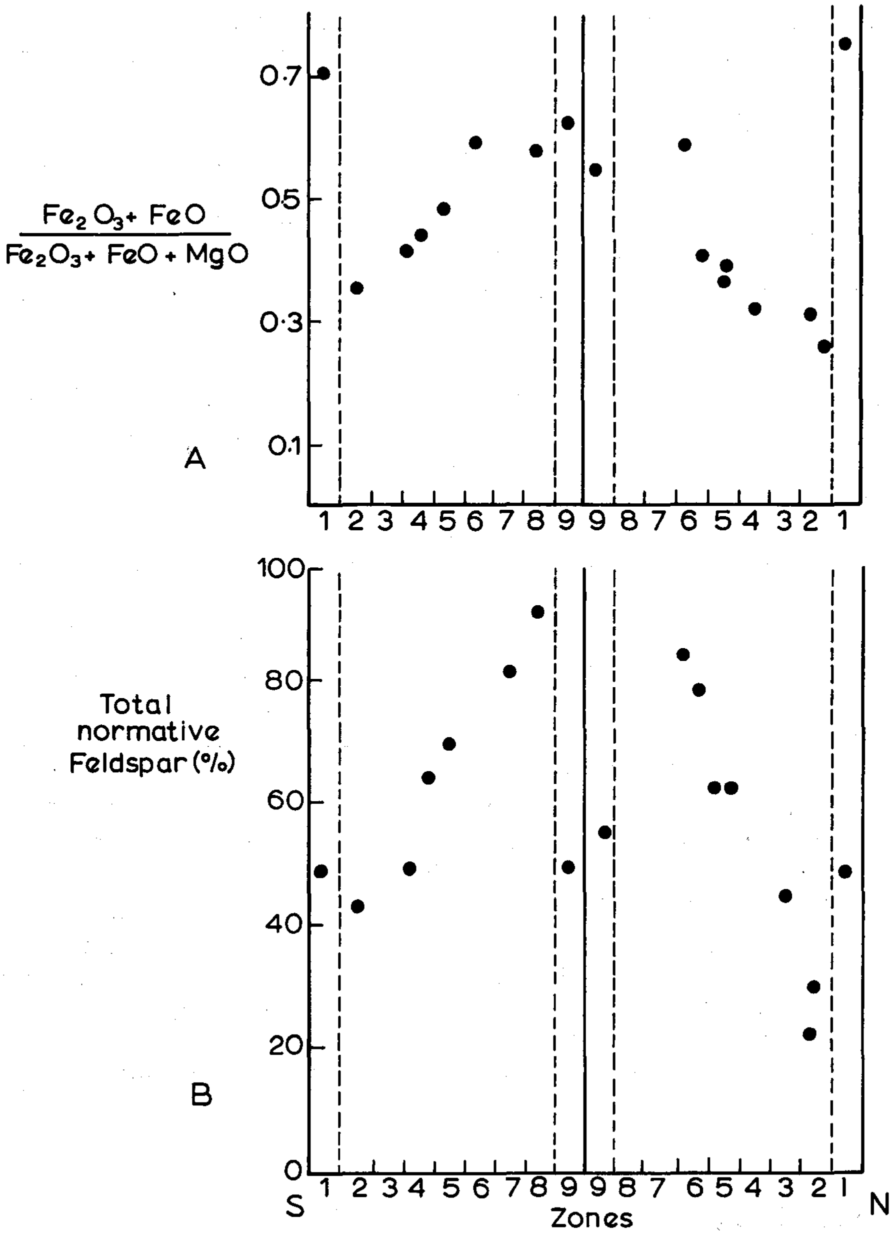

Fig. 33 A. The variation of Fe/Mg ratios with respect to zones across the complex.
B. The variation in total normative feldspar across the complex.

flecting the symmetrical reversed order of the stratigraphic zones. Fig. 31 shows that both halves exhibit very similar compositions and that their differentiation trends mirror each other. The systematic increases in Fe/Mg ratio suggest a relative way-up of the layered body, which differentiated from what is now the 'outside' to the 'inside'. This means that the mafic zones 2 to 5 underlie the anorthositic zones 6 to 8 and that the orthopyroxene chromitites (or chromite-layered bronzitites) formed earlier than the hornblende chromitites (or chromite-layered anorthosites). Analysed rocks from the Fiskenæsset complex are compiled on fig. 32.

The increase in total Fe relative to Mg is demonstrated in fig. 33 A showing that there is a systematic increase in this ratio on both sides of the intrusion. The differentiation towards an anorthositic centre is illustrated in fig. 33 B in which the total feldspar content in the norms is plotted against the sample position in the complex. The concentration of ultramafic and mafic layers in the lower part and of plagioclase in anorthosite in the upper part of the complex compares with the similar situation in the Critical Zone of the Bushveld complex (CAMERON, 1963) and in the Stillwater complex (HESS, 1960). As a result of this differentiation sequence total alkalis, alumina and calcium increase, and iron and magnesium decrease towards the top of the intrusion (see figs 34 A, 34 B, 34 C).

With regard to fractionation of the trace elements (Table 2) the Co/Ni ratio increases towards the 'centre' of the complex and is comparable with the Mg/Fe path. The cobalt content decreases upwards (fig. 34 D) as this element follows the concentration of mafic minerals. Likewise the Ni/Mg ratio decreases and the Zr content increases slightly toward the centre. Chromium appeared as an appreciable trace element early in the history of the intrusion when it entered the lattices of olivine, pyroxene, and spinel in the lower ultramafic and mafic layers. As differentiation continued and the temperature dropped, less ferromagnesian minerals were formed and chromate became supersaturated in the magma, being precipitated as chromite in minor amounts along the zone 3–4 contact and extensively within the anorthosites as zone 7. This was iron-rich chromite (GHISLER & WINDLEY, 1967) and thus the Fe/Mg ratio of zone 8 was appreciably reduced by the chromite precipitation (see fig. 33 A).

Geochemically the zone 1 amphibolite (sample numbers 1 and 18) is totally different from any of the rocks belonging to the layered intrusion. It is extremely rich in iron and so lies far from the differentiation trend of the layered igneous rocks (fig. 32). All other major and

Fig. 34 A. Weight % MgO + FeO + MnO with respect to the zones of the complex.
B. Weight % Al_2O_3 with respect to the zones of the complex.
C. Total alkalis (weight %) with respect to the zones of the complex.
D. The variation of cobalt across the complex.

All analyses for both halves of the complex have been grouped together for fig. 34.

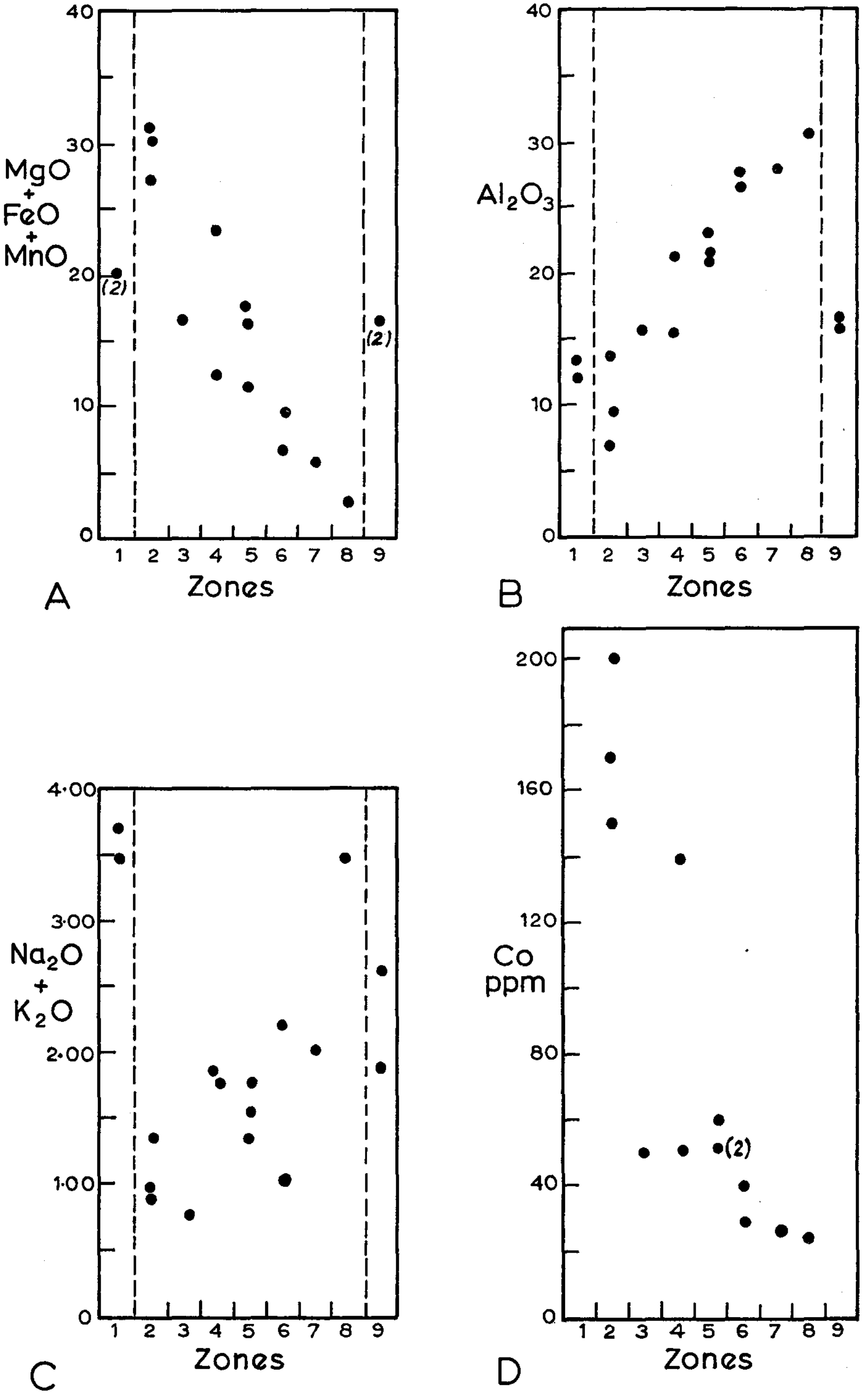

Fig. 34 A–D.

Table 2. *Trace element analyses of the Fiskenæsset complex*

Sample No.	1	2	3	4	5
Zone	*1*	*2*	*4*	*4*	*5*
GGU No.	86921	74482	86923	86922	86924
XRF trace element analyses					
Ni	200	1100	570	170	225
Zn	110	65	75	50	40
Sr	40	13	17	65	60
Rb	6	4	10	8	6
Zr	110	13	10	17	30
S	70	450	25	25	150
U. V. Spectrography trace element analyses					
Ba	70	←	3	12	7
Co	100	150	140	50	50
Cr	130	500	450	75	450
Cu	300*	300*	9	30*	90*
Ga	30	9.5	12	10	17
Li	14	20	34	19	26
Ni	130	760	400	130	140
Pb	←	2	5	7	5
Sc	10	←	8	30	6
Sr	65	←	←	75	60
V	130	30	60	70	100
Y	←	←	←	←	←
Yb	1	3	1	←	←
Zr	60	←	12	←	←

* visual estimate
← too little to detect
→ too much to measure
n.d. not determined
< less than

(continued)

Table 2 (cont.)

Sample No.	6	7	8	9	10
Zone	*7*	*7*	*8*	*9*	*9*
GGU No.	86925	74489	92579	86928	74484
XRF trace element analyses					
Ni	800	50	60	160	230
Zn	210	40	40	100	90
Sr	50	90	80	85	80
Rb	80	3	125	7	4
Zr	40	30	25	80	55
S	12	20	50	360	60
U. V. Spectrography trace element analyses					
Ba	90	16	180	50	28
Co	n.d.	25	14	70	60
Cr	→	85	22	350	←
Cu	4	8	40*	100*	80*
Ga	26	24	22	20	16
Li	25	16	14	28	30
Ni	450	50	45	130	190
Pb	←	7	4	←	←
Sc	13	7	←	35	8
Sr	40	100	100	80	60
V	400 +	80	15	170	75
Y	←	←	←	30	←
Yb	←	←	←	4	5
Zr	< 10	←	←	36	40

(continued)

Table 2 (cont.)

Sample No.	11	12	13	14
Zone	*6*	*6*	*5*	*5*
GGU No.	86929	86930	86931	86932
XRF trace element analyses				
Ni	60	275	540	500
Zn	35	40	40	60
Sr	90	60	45	45
Rb	5	6	4	4
Zr	35	5	11	7
S	16	60	35	30
U. V. Spectrography trace element analyses				
Ba	26	7	3	7
Co	30	40	50	60
Cr	380	←	500	140
Cu	20*	30*	25*	30*
Ga	24	24	16	15
Li	19	15	25	16
Ni	60	200	400	430
Pb	1	8	3	4
Sc	40	←	←	6
Sr	120	80	34	40
V	90	10	50	45
Y	←	←	←	←
Yb	←	←	←	←
Zr	13	< 10	12	< 10

(continued)

Table 2 (cont.)

Sample No.	15	16	17	18
Zone	*3*	*2*	*2*	*1*
GGU No.	74487	86935	86927	86934
XRF trace element analyses				
Ni	220	1700	1100	90
Zn	45	70	50	140
Sr	40	7	7	140
Rb	3	5	8	10
Zr	12	20	7	90
S	20	1300	200	65
U. V. Spectrography trace element analyses				
Ba	4	3	←	65
Co	54	170	200	70
Cr	→	→	→	←
Cu	16	300*	60*	20*
Ga	15	11	8	20
Li	17	90	13	19
Ni	175	1200	700	70
Pb	15	7	7	5
Sc	132	11	13	6
Sr	60	←	30	130
V	120	160	85	180
Y	←	←	←	25
Yb	←	←	2	3
Zr	←	←	11	55

Analyst A. A. Bowden

trace elements confirm this marked difference. On the basis of this chemical disparity it is suggested that the amphibolite was a metavolcanic.

Although the zone 9 amphibolite (sample numbers 9 and 10) contains thin mica schists, rare sillimanite nodules and is locally underlain by marbles, it is not chemically similar to the zone 1 amphibolite and so not directly related to it. Fig. 33 A shows that its Fe/Mg ratio is much less than that of the lower amphibolite. Other elements confirm this relationship. However, it is also more iron rich than the rocks of zones 2 to 8, and so likewise is removed from the differentiation trend of the intrusion on the AFM diagram (fig. 32). On the basis of present evidence it can be suggested that the zone 1 and zone 9 amphibolites are metavolcanic rocks, not genetically connected. The balance of field and chemical evidence favours this opinion. In some areas such as east Qeqertarssuatsiaq zones 1 and 9 are totally absent over several kilometres, and in others (north-east Qeqertarssuatsiaq) zone 1 continues for several kilometres in the gneisses beyond the termination of the main layers of the complex—these features would be in accord with a volcanic interpretation of these layers.

KALSBEEK (1970) and KALSBEEK & LEAKE (1970) have studied amphibolites in the gneisses between Frederikshåb and Ivigtut, south of the Fiskenæsset area. There is some similarity between the petrography of the Frederikshåb amphibolites and those at Fiskenæsset (presence of garnet and diopside, some quartz) but only the zone 9 amphibolites are chemically similar to KALSBEEK's rocks. The amphibolites of zone 1 are more iron rich than all but one of 54 analysed amphibolites from Frederikshåb (KALSBEEK & LEAKE, 1970, pp. 27–32).

It is possible that the marginal amphibolites might be derived from an early volcanic phase of the magma. Loss of iron enriched in this fraction might account therefore for the lower Fe/Mg ratio of the intrusion compared with those of the Bushveld and Stillwater complexes (BOWES & SKINNER, 1969). This factor might work together with the high oxidation state caused by an elevated water pressure in accounting for the unique differentiation trend of the Fiskenæsset magma.

A postscript may be added to this section: WINDLEY & SMITH (in press) have been able to show by electron probe microanalysis of minerals within the Fiskenæsset complex that the cryptic layering of the igneous rocks is still preserved in both the felsic and ferromagnesian constituents.

PETROGENESIS

It has been shown that the mappable rock units of the present metamorphosed layered complex in part were originally stratiform horizons of igneous rock formed by gravity differentiation of a basic magma and in part were cover rocks along both upper and lower margins. Although the complex has been affected by at least two metamorphic episodes, the original igneous stratigraphy is well preserved. There is little evidence of widespread introduction of material to change the bulk chemistry of the intrusion, nor is there evidence of interchange of material amongst the layers of the complex during metamorphism. That is, metamorphism has been isochemical both on the scale of the complex and also on the scale of individual layers within it. Taking these circumstances into account, comments can be made on the nature of the original igneous magma, and on the mineral assemblages present in the parent igneous rocks.

Nature of the parent magma

The bulk chemical analyses listed in Table 1 have been averaged to give a general indication of the chemistry of the parent magma. Individual analyses have been weighted in proportion to their contribution to the total width of the complex along the traverse line. The resulting average analysis corresponds almost exactly to the analysis 5 of the cumulate anorthositic gabbro 86924. The fact that this rock shows normative olivine, and no normative quartz or nepheline suggests that the parental liquid had the composition of a high-alumina basalt.

Note that allanite pegmatites confined to the complex are the only rocks within it to contain abundant potash feldspar. Their presence may be explained by the fact that the accumulation of plagioclase as anorthosite would leave a residual liquid enriched in potassium (YODER, 1969).

The norms of the analysed rocks as listed in Table 1 give no indication of the water content of the original magma, but evidence for water content can be deduced from both the relict mineral assemblages and the differentiation trends.

Original mineral assemblages

Examination of the highest grade mineral assemblages within each rock type shows that the following rocks probably made up the igneous suite:

Anorthosites:	Plagioclase + orthopyroxene ± clinopyroxene ± hornblende ± spinel
Leuco-gabbros:	Plagioclase + orthopyroxene ± clinopyroxene ± hornblende ± spinel
Gabbros and Norites:	Plagioclase + orthopyroxene ± clinopyroxene ± hornblende ± spinel ± ore
Chromitites:	Chromite ± orthopyroxene ± hornblende ± rutile
Dunites:	Olivine ± hornblende ± spinel ± ore
Pyroxenites:	Orthopyroxene ± clinopyroxene ± hornblende ± spinel ± ore
Peridotites:	Olivine + orthopyroxene ± clinopyroxene ± hornblende ± spinel ± ore

A number of conclusions can be drawn from this list of assemblages. First of all, the lack of olivine in any of the plagioclase-bearing rocks is significant. The norm values of Table 1 show the felsic and mafic rocks to be olivine normative, but not a single specimen has been found to contain relict olivine. No armoured relics, so common in metagabbros elsewhere, have been found with olivine centres. A single example of a hypersthene armoured relic in a gabbro is known, and this information, along with observations of rare igneous-type intergrowths between ortho- and clinopyroxene in pyroxene aggregates, suggests that if olivine had been present in the original plagioclase-bearing rocks, then some evidence of its presence would have survived.

On the other hand it may be argued that no olivine survives in the felsic and mafic rocks because of introduction of water under metamorphic conditions. This process would favour destruction of olivine, while pyroxene would survive. It seems unlikely that there has been much H_2O metasomatism throughout the complex in general, and most of the olivine in the ultramafic rocks has escaped amphibolisation and/or serpentinisation. Olivine, pyroxene, and hornblende are interlayered on a millimetre-scale in some ultramafics (see below).

Second, the occurrence of hornblende in all assemblages is noteworthy. Certainly there is evidence of the development of hornblende from pyroxene and plagioclase during metamorphism, but several lines of reasoning all lead to the conclusion that some primary amphibole existed in the original igneous rocks. An intimate interlayering of horn-

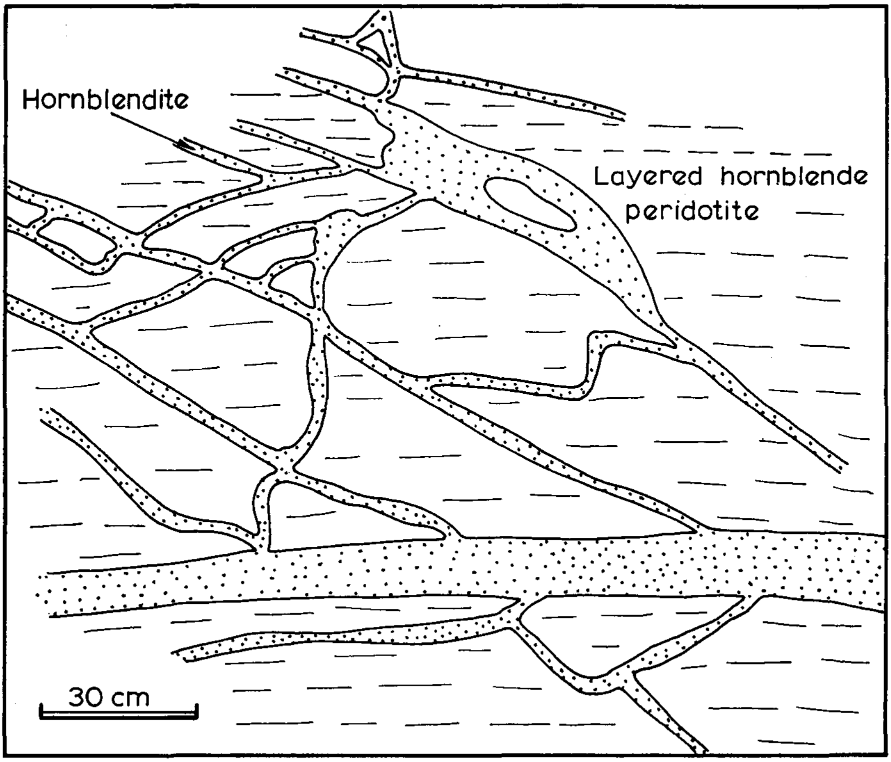

Fig. 35. Veining of a layered hornblende peridotite by metasomatic amphibolite facies hornblendite. Qeqertarssuatsiaq.

blende-free and hornblende-bearing assemblages is found within felsic and mafic rock types; if the hornblende is all metamorphic in origin, what mechanism has confined it to certain laminae only? The hornblende in such cases does not contain relict pyroxene and furthermore it is present in apparent equilibrium intergrowth with olivines and pyroxenes in ultramafic rocks. More significantly, it may form single millimetre-scale laminae in the ultramafic rocks, alternating with laminae of anhydrous silicates.

It may be argued that the hornblende in the ultramafics was derived by metasomatism and that the laminae of hornblende were formed by deformation accompanying this introduction of material. Field evidence that this cannot be so is shown in fig. 35, in which a hornblende peridotite can be seen to be cut by a network of metasomatic hornblende veins. The hornblende veins must have developed under amphibolite facies conditions, so the hornblende in the peridotite pre-dates the amphibolite facies events. No evidence is known of introduction of alkalis and silica under granulite facies conditions elsewhere in the complex, so the horn-

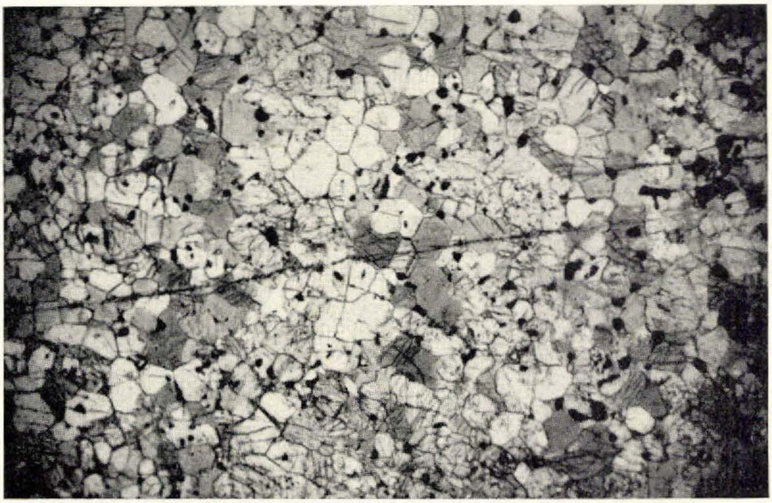

Fig. 36. Photomicrograph of a layered hornblende peridotite with olivine, pyroxene, and brown hornblende in granoblastic-polygonal equilibrium texture. Note that the hornblende is confined to certain laminae. Opaque minerals are ilmenite and magnetite. GGU 89812, plain light, ×17.

blende of the peridotite is probably recrystallised igneous amphibole. The equilibrium texture of a hornblende-bearing layered peridotite is shown in fig. 36. There is no reason to assume intense deformation to produce the fine laminae in such a rock when rhythmic and cryptic layering are so well preserved throughout all the rocks of the complex. Furthermore, recent work on magnetite-layered, hornblende peridotites from zone 2 has shown that cryptic layering is found in the magnetites and other minerals of successive layers, proving that the lamination is primary (WINDLEY & SMITH, in press).

It could be postulated that the ultramafic rocks of the intrusion are derived for the most part from plagioclase-bearing olivine melagabbros and troctolites, that have been metamorphosed and annealed to such an extent that no evidence of their original plagioclase content survives. Plagioclase is a normative mineral in the ultramafics, as is olivine. However, no textures even vaguely reminiscent of such a process are found.

The occurrence of spinel in the ultramafic rocks again throws light upon the original mineral assemblages. Certainly the amount of spinel present has been augmented by the metamorphisms. The largest spinels are found in hornblendites, and also concentrated in the axial planes and cores of minor folds in some of the pyroxenites and hornblende peridotites. Spinel forms distinct layers in the ultramafic rocks, as chromite does in the anorthositic rocks of zone 7. Sometimes it is difficult to understand how the spinel has been increased in amount during metamorphism, as the host rocks appear to be composed largely of the aluminium-free phases olivine and pyroxene. Closer examination reveals the presence of a pale amphibole in such rocks, accompanied in some cases by spinel-pyroxene symplectites. It can be postulated that the spinel and intergrown spinel and pyroxene have been derived in part by metamorphism of a primary aluminium-rich igneous amphibole. Significantly, where this process is thought to have operated, the present amphibole is pale green and may be poor in aluminium, i.e. tremolitic.

Spinel is only of rare occurrence in the anorthosites, leuco-gabbros, and gabbros of the complex. As has been noted under the petrographic description of the garnetiferous rocks of zone 8, spinel may have acted as a nucleus for the formation of garnet in these plagioclase-bearing rock types and thus spinel must be considered as a possible phase in the felsic and mafic rocks.

Note that the presence of water would facilitate the development of equilibrium textures and the thinning and thickening of the complex during deformation.

Thus there is petrographic and field evidence for a certain tenure of water in the parent igneous magma of the Fiskenæsset rocks, through the lack of olivine and the possible presence of igneous amphibole and spinel in the felsic and mafic rocks, and through the probable presence of igneous amphibole in the ultramafic rocks. The normative anorthite, albite, and orthoclase shown for the ultramafic rocks in Table 1, arise from the presence of hornblende, as do the normative olivine and alkali feldspar in the felsic and mafic rocks.

The presence of an elevated water pressure in the crystallising magma has the following mineralogical and chemical implications, which explain satisfactorily several features of the Fiskenæsset complex:

1) The eutectic of the diopside-anorthite system would be shifted from gabbro to anorthosite, increasing the amount of precipitated plagioclase (Yoder, 1969).

2) The anorthite content of plagioclase would be increased, allowing very calcic plagioclase to form (Yoder, 1969).

3) The temperature of formation of plagioclase would be lowered more than that of olivine and pyroxene, thus delaying its appearance and enabling it to concentrate as late anorthosite (Yoder, 1969; Osborn, 1969).

4) Yoder & Tilley (1962) showed that a high water pressure would increase the stability field of amphibole at the expense of pyroxene. Primary hornblende could thus form with olivine in the hornblende dunites of zone 2.

5) Chromian amphibole could form as a primary mineral from the melt. It is a typical product of many alpine ultramafic complexes in their final magmatic stages, characterised by the enrichment of volatile constituents in high level chromitites (BORCHERT, 1963). The paucity of pyroxene in the zone 7 chromitites (GHISLER & WINDLEY, 1967) could therefore be due, not to preferential water metasomatism and retrogression of these layers, but to formation of primary chromian amphibole.

6) A result of (3) above would be that the liquid would move to a higher alumina content than would be the case if no water were present, as portrayed by movement of the anorthite phase in the tetrahedron $Mg_2SiO_4 - Fe_3O_4 - CaAl_2Si_2O_8 - SiO_2$ (OSBORN, 1969).

7) The Fo-An system would be replaced in the presence of water by the following three systems (YODER, 1969):

Amph + An + Cpx + Sp 1
Cpx + Opx + Sp 2
An + Cpx + Opx + Sp 3

Assemblages 2 and 3 are known with hornblende from the ultramafics and gabbros, respectively. Although the amount of spinel has probably been augmented during metamorphism, a high water pressure would

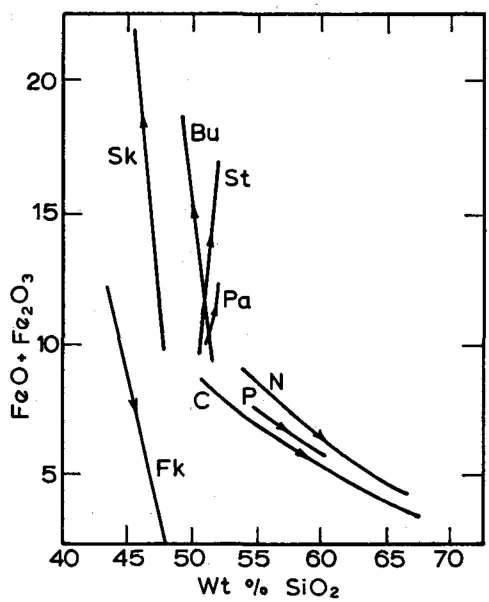

Fig. 37. A silica–total iron diagram showing liquid composition trends of various rock series including that of the Fiskenæsset complex (Fk). Curves for the layered intrusions: Bu = Bushveld; St = Stillwater; Sk = Skaergaard; Pa = Palisades diabase. Curves for calc-alkaline extrusive rock series: N = NOCKOLDS' (1954) averages; P = Paricutin lavas; C = Cascades volcanic rocks (after OSBORN, 1969, fig. 6).

have given rise to spinel-bearing igneous assemblages, thus facilitating the nucleation of metamorphic spinels.

8) Dissociation of H_2O would give rise to a high oxidation state in the melt, enabling magnetite to crystallise early in zone 2. This would give rise to a differentiation trend that had certain affinities with that of the calc-alkali series as expressed by the low *relative* increase in Fe compared with Mg seen on the AFM diagrams.

Whereas the Skaergaard, Bushveld and Stillwater complexes formed from anhydrous magmas and show *absolute* enrichment of iron with crystallisation, the Fiskenæsset intrusion must have formed from a hydrous melt similar to the calc-alkali extrusive rock series and thus shows a decrease in total iron with differentiation (see fig. 37).

Mechanism of formation of the anorthosites

The actual mechanism of formation of anorthosite in layered igneous bodies has been much discussed in the literature, in particular with regard to processes of sinking or flotation of the plagioclase crystals in the magma. The Fiskenæsset intrusion displays features which favour the plagioclase flotation model. The fact that the anorthosites occur at the top (zones 6 and 8), underlain by higher density mafic and ultramafic material, should mean that on the scale of the intrusion the density of the plagioclase-rich material was less than that of the magma. However, in zones 5, 6, and 7 the presence of plagioclase megacrysts in cumulus textures suggests that there was some degree of plagioclase sinking in certain upper layers.

Estimates of physical conditions during emplacement and metamorphism

The pressure and temperature conditions existing during emplacement of the magma of the Fiskenæsset complex are probably best given by the equation

An + Fo + Gas = An + Cpx + Opx + Sp + Gas (Yoder, 1969).

This reaction takes place (in the presence of excess H_2O) at 900–950° C and 7–7.5 kilobars pressure.

According to estimates by O'Hara (1967, pp. 14–15) on assemblages in ultramafic rocks in general, the granulite facies conditions at Fiskenæsset probably were some 800–900° C, and 7–17 kilobars pressure.

F. Seifert, Bochum, (pers. comm.) has stated that sapphirine can form with extremely low reaction rates in the system $MgO - Al_2O_3 - SiO_2 - H_2O$ at approximately 650–700° C and 5–6 kilobars pressure, and these estimates are taken as characteristic of the amphibolite facies conditions at Fiskenæsset.

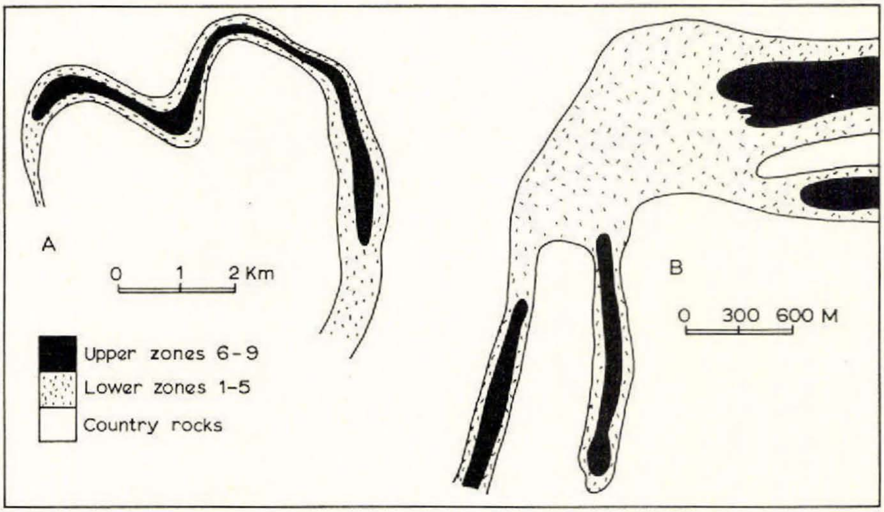

Fig. 38. Map showing the internal closures of the upper zones of the complex. The duplication in B is caused by subsequent F_2 folding (see fig. 40). A is on central to east Qeqertarssuatsiaq; B extends from west of Taseq eastwards along the south coast of Fiskenæsfjorden.

STRUCTURE OF THE COMPLEX

The Fiskenæsset complex is clearly in part the remains of a layered intrusion. The presence of chromitites indicates that the layering must have formed in a near horizontal position, but it now has an isoclinal synclinal shape as shown by the internal synclinal closures of the upper zones (fig. 38) and by the fact that the upper zones are symmetrically innermost (as indicated by the geochemistry). Originally the body (as distinct from the layering) could have been entirely flat-lying, had the form of a basin or have been funnel-shaped. The fact that the body was consistently folded into a syncline with precisely the same axis of symmetry along its whole length suggests that it most likely was originally basin or funnel-shaped. Flattening of such an initial syncline could most easily give rise to the remarkable symmetry of the present body. A suggested mode of evolution of the complex is illustrated diagrammatically in fig. 39. The repeated stratigraphy arranged in reverse order was derived by flattening symmetrically the two halves or limbs of the original intrusion.

Anorthositic rocks occur in the Archaean gneisses to the north and to the south of Fiskenæsset (Berthelsen, 1960 b; Kalsbeek, 1970; Sørensen, 1955). Windley (1969 a) has pointed out that there are at least 400 km of such anorthositic layers as measured along their strike

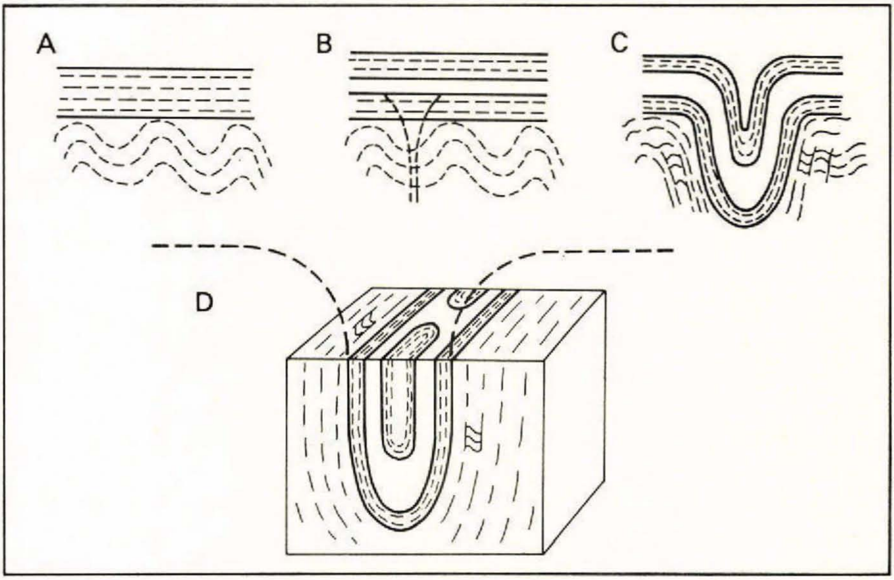

Fig. 39. Diagrammatic model for the evolution of the Fiskenæsset complex.

A. Formation of supracrustal rocks (largely basic volcanics) overlying a gneissic basement.

B. Emplacement of Fiskenæsset intrusion within the supracrustals.

C. Downfolding of the complex into the underlying basement. Obliteration of the unconformity and partial refoliation of adjacent gneisses.

D. Present mode of occurrence of the complex infolded as an isoclinal syncline in conformable gneissic basement.

length. Although they are petrologically similar to the Fiskenæsset anorthosites (especially with their calcic plagioclases) and locally are associated with cumulate leuco-gabbros, they differ in their lack of associated ultramafic rocks and chromitites and in their paucity of mafic material. Whereas the Fiskenæsset complex is remarkably well preserved as a continuous layer, the other anorthositic rocks occur in comparatively short discontinuous layers and tectonic lenses.

These similarities and differences might be explained by supposing that the 'outlying' anorthositic layers were derived from higher (i.e. gabbro anorthositic and anorthositic) stratigraphic levels of a series of anorthosite-bearing intrusions. The remarkable degree of preservation of the Fiskenæsset complex must be due in part to the fact that synclinal flattening *increased* its thickness during deformation. The 'outlying' anorthositic layers are usually thinner and more broken up than the Fiskenæsset complex and they may not have been deformed in this way. The ideas presented in this paragraph are naturally speculative but should be entertained as possibilities.

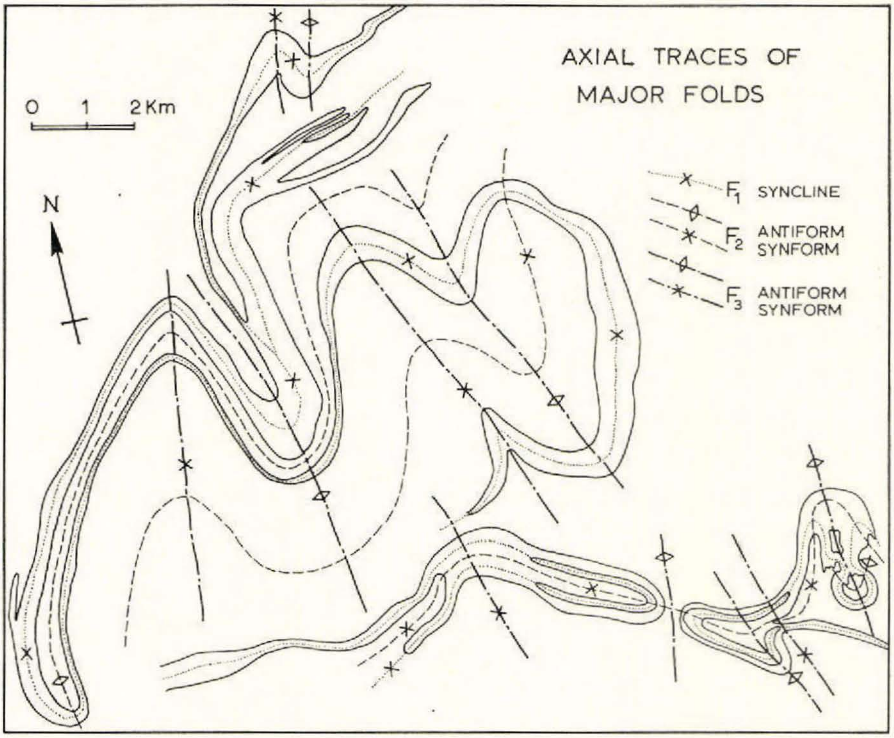

Fig. 40. Map giving the axial traces of the major folds in the Fiskenæsfjorden region.

Petrographic, chemical and field evidence have been presented which suggest that the magma was intruded into volcanic rocks. The question may be readily asked—how do the intrusion and the bordering metavolcanic amphibolites come to form a concordant layer in high grade gneisses? Consideration of the present structure of the complex very much concerns this question. The Fiskenæsset intrusion and its bordering amphibolites are almost entirely unmigmatised and have undergone little partial melting, whereas the adjacent monotonous quartzo-feldspathic gneisses, which contain extremely little metasedimentary material, contain many migmatised and agmatised amphibolite inclusions. It must be considered unlikely that the gneisses are the product of extensive granitisation and bulk homogenisation of a pile of supracrustal rocks which did not affect the intrusion and its bordering amphibolites.

An explanation for the conformability of the complex lies in a group of amphibolite dykes that are present in the gneisses but absent within the complex and its adjacent amphibolites. McGREGOR (1968) used this amphibolite dyke method in Godthåbsfjord, West Greenland, to demonstrate that a series of metavolcanic amphibolite layers and associated metasediments together with an anorthosite layer were younger than the adjacent conformable quartzo-feldspathic gneisses which must accordingly have been the crystalline basement upon which the supracrustal rocks were deposited. There is similar though less substantial amphibolite-dyke evidence in the present region which suggests that the magma was intruded into volcanic rocks overlying a crystalline gneissic basement. Early deformation of the complex obliterated the cover–basement relationship and gave rise to the conformability of all rock groups.

The first flattening deformation (termed F_1) was responsible for the isoclinal structure of the complex and probably for its conformability with the basement gneisses. The body was subsequently folded isoclinally by an F_2 series of folds the axial traces of which strike north-east–south-west, and by an open to tight F_3 fold phase with north–south to north-north-west striking axial planes. The axial traces of these three fold phases are shown in fig. 40.

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The aerial photograph (fig. 1) is reproduced with permission (A. 235/72) of the Geodetic Institute, Copenhagen.

Department of Geology,
Bennett Building,
The University,
Leicester LE1 7RH, England.

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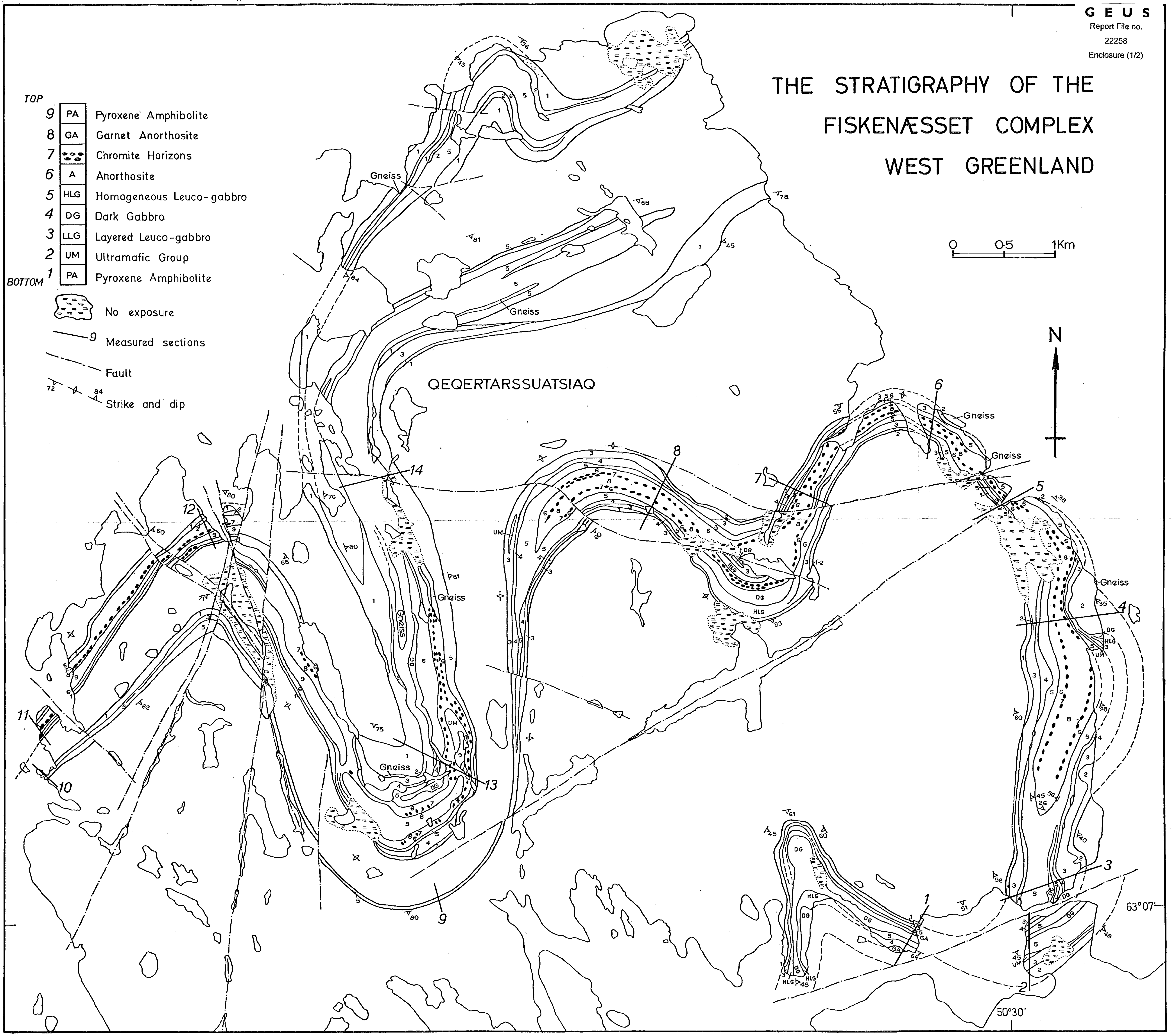
Enclosure (1/2)

THE STRATIGRAPHY OF THE FISKENÆSSET COMPLEX WEST GREENLAND



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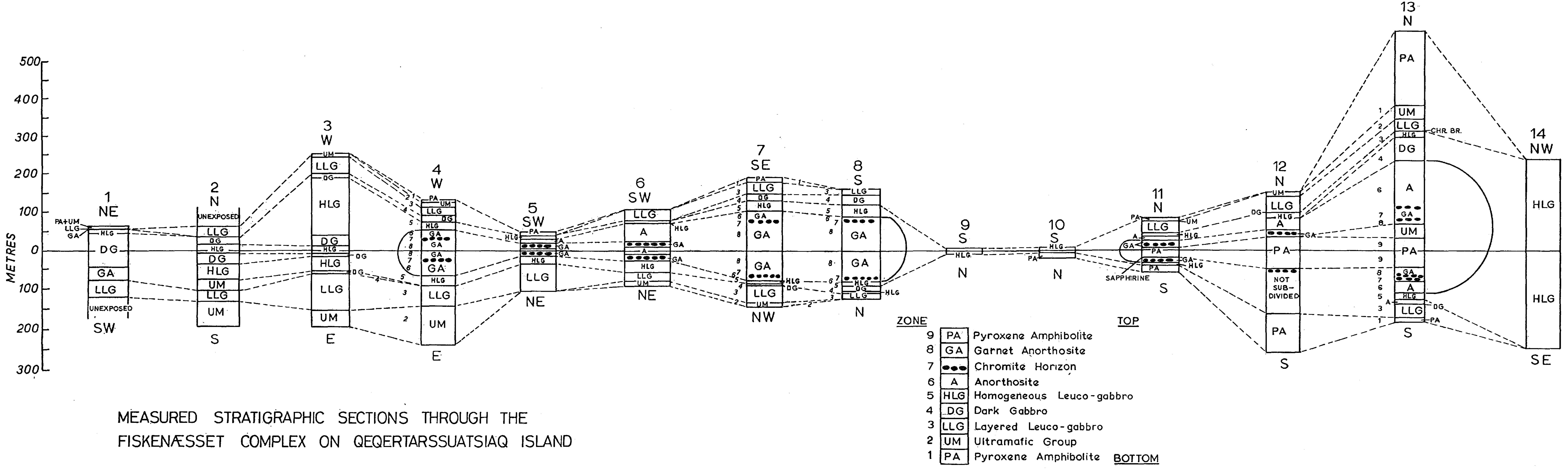

MEASURED STRATIGRAPHIC SECTIONS THROUGH THE FISKENÆSSET COMPLEX ON QEQERTARSSUATSIAQ ISLAND