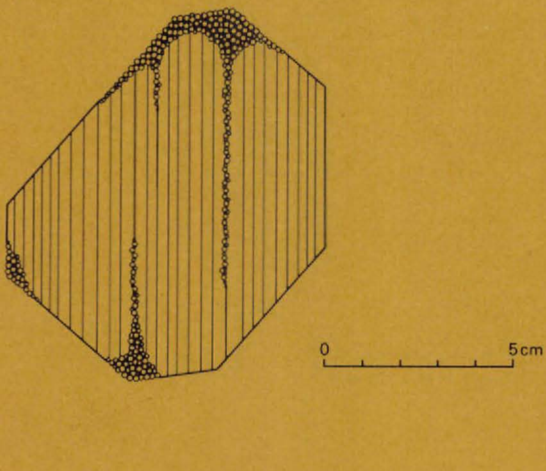
GRØNLANDS GEOLOGISKE UNDERSØGELSE
Bulletin No. 150

Stratigraphy and structure of the Fiskenæsset Complex, southern West Greenland

by

John S. Myers



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Partly recrystallised igneous plagioclase crystal from anorthosite in the Fiskenæsset Complex.

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Abstract

The Fiskenæsset Complex is a deformed and metamorphosed, sheet-like, layered basic instrusion. It is an exceptionally well preserved example of a suite of anorthositic rocks which are a widespread, but fragmentary, component of the Archaean gneiss complex of Greenland. In the Fiskenæsset region these rocks can be seen to be part of a single layered intrusion which consists of seven major lithostratigraphic units. In ascending order, these units are the Lower Gabbro (50 m), Ultramafic (40 m), Lower Leucogabbro (50 m), Middle Gabbro (40 m), Upper Leucogabbro (60 m), Anorthosite (250 m) and Upper Gabbro (50 m) Units.

The Fiskenæsset Complex crystallised from tholeiitic magmas which were probably emplaced in at least 3 stages. It was intruded into basic volcanic rocks (now amphibolites) and, together with these rocks, was fragmented by the intrusion of granitoid sheets (now gneisses) associated with thrusting 2900–2800 m.y. ago. The Fiskenæsset Complex and metavolcanic rocks occur as thin layers and trains of inclusions within these gneisses which form about 80% of the region. All these rocks were folded together into large recumbent, nappe-like, folds (F1), and then refolded into dome-and-basin interference patterns by two sets of folds with steep axial surfaces at high angles to each other (F2 and F3). Deformation was locally heterogeneous, and all stages can be seen from undeformed to very strongly deformed rocks. Recrystallisation continued in amphibolite, and locally in granulite, facies during waning tectonic activity, and most rocks have equigranular mosaic textures.

Although it is extensively recrystallised and deformed, the Fiskenæsset Complex locally preserves a variety of little deformed igneous structures and cumulate textures which demonstrate the important influence of both gravitational and current action during igneous crystallisation. Various kinds of igneous layering provide primary way-up structures which can be used to interpret the facing directions of F1 folds, and various stages of deformation of cumulate textures provide evidence of the tectonic and metamorphic processes by which banded anorthositic gneisses and amphibolites were derived from igneous rocks by progressive deformation.

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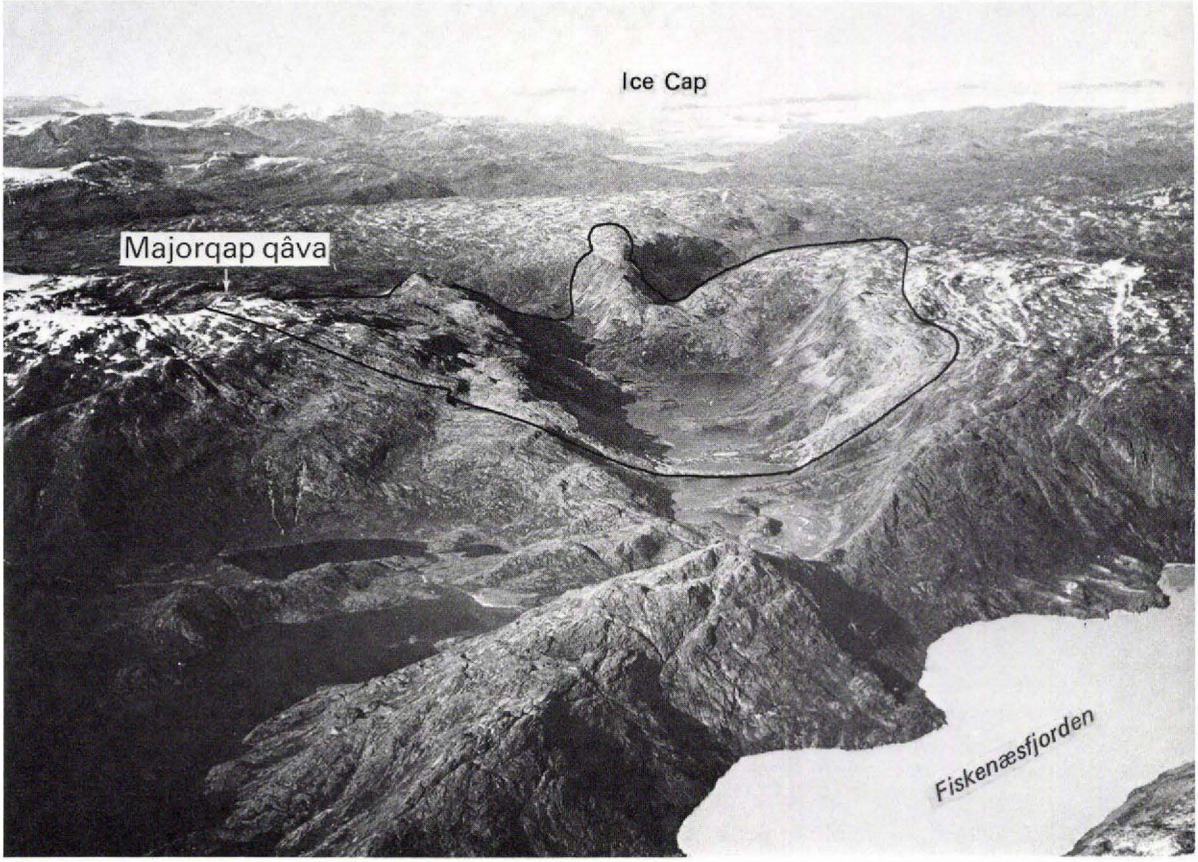


Frontispiece. The Majorqap qâva outcrop of the Fiskenæsset Complex viewed from the south-west from an altitude of 2000 m. Majorqap qâva (mountain) has an altitude of 1049 m and is about 40 km from the inland ice cap. A map of this outcrop is shown in plate 1.

INTRODUCTION

Metamorphosed calcic anorthosites and associated leucogabbroic and gabbroic rocks are widespread throughout the Archaean complex of Greenland as discontinuous layers and trains of inclusions in quartzo-feldspathic gneiss (Bridgwater *et al.*, 1976). They are derived from layered basic intrusions with very coarse grained cumulate textures, which were fragmented by the emplacement of enormous volumes of granitoid magmas associated with thrusting 2900–2800 m.y. ago. They were repeatedly deformed, folded into complex interference patterns, and recrystallised in amphibolite or granulite facies. They provide one of the best marker horizons for tracing the regional tectonic structures of the Archaean gneiss complex. These rocks are most abundant in the Fiskenæsset region of southern West Greenland (plate 2), where they also form large outcrops (frontispiece, fig. 1) in which a major igneous stratigraphy is well preserved, and small-scale igneous structures, textures and minerals are better preserved than in other parts of the Greenland Archaean complex.

The main anorthosite outcrops of the Fiskenæsset region show stratigraphic details which suggest that they form part of a single major intrusion, called the Fiske-

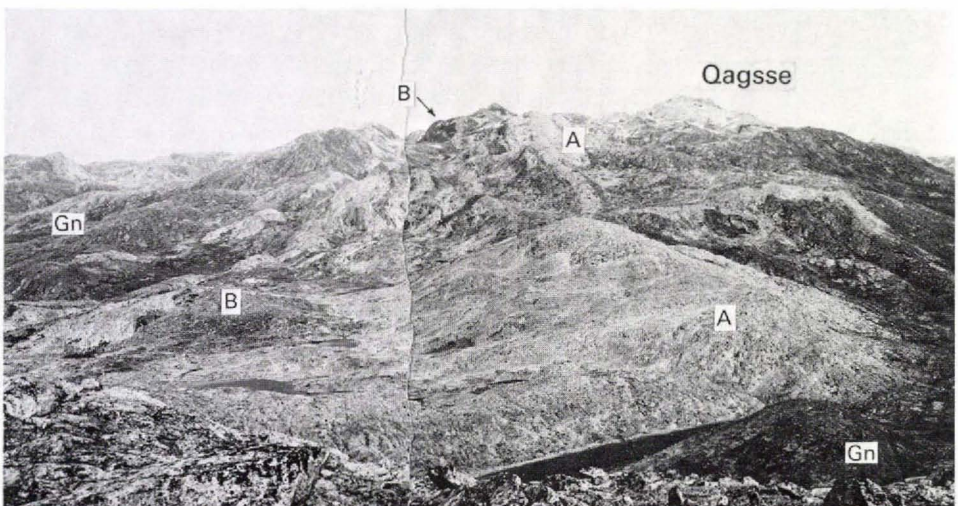


Fig. 1. T he Fiskenæsset Complex between Tasiussâ and Qagsse showing steeply dipping white layers of anorthosite (A), black layers of gabbro (B) and grey granitoid gneiss (Gn). View to the north-east from near locality 35.

næsset Complex. Because of the fragmentary nature of the complex and the lack of diagnostic features in many small outcrops, it is impossible to delineate the real boundary of the Fiskenæsset Complex. For convenience this name is applied to all the outcrops of anorthosites and associated leucogabbros, gabbros and ultramafic rocks which occur in the Fiskenæsset region between latitudes 63°30′ and 62°45′ (plate 2). Within this region most of these rocks are concentrated in a belt 50 km wide and 100 km long which extends in a north-east direction from Fiskenæsset town to the inland ice cap. The Fiskenæsset Complex occurs as sheets generally less than 500 m thick with outcrop lengths of up to 25 km, and as trains of smaller inclusions in granitoid gneisses.

This bulletin describes the igneous stratigraphy of the Fiskenæsset Complex, and uses this stratigraphy to interpret the major tectonic structure of the region.

Regional geology

Most of the region consists of quartzo-feldspathic gneiss derived by deformation and metamorphism from tonalite, granodiorite and granite (Myers, 1978a), emplaced as sub-concordant sheets into the Fiskenæsset Complex and older basic volcanic rocks (now amphibolites). The granitoid sheets were intruded about 2880 m.y. ago (Moorbath & Pankhurst, 1976; Kalsbeek & Pidgeon, 1980) in a regime of horizontal tectonic movements, and many were intruded along active thrust zones or were themselves the sites of later thrust movements (Myers, 1981).

This layered sequence was folded into large recumbent isoclinal, nappe-like, structures (F1). Sub-concordant sheets of porphyritic granite (Ilivertalik granite, Kalsbeek & Myers, 1973) were intruded about 2800 m.y. ago (Kalsbeek & Pidgeon, 1980) and, together with their host rocks, were folded into large scale tight folds (F2) with steep southward dipping axial surfaces, generally trending E–W. These F2 structures form the main tectonic grain of the region (plate 2). They were refolded by F3 folds with steep N–S trending axial surfaces which formed large scale dome-and-basin interference patterns with the F2 structures (plate 2). Metamorphism reached amphibolite facies throughout the region and locally, especially around Fiskenæsset, it reached granulite facies during and after F3 deformation about 2800 m.y. ago (Pidgeon & Kalsbeek, 1978). Subsequently (perhaps about 2600 m.y. ago, Kalsbeek & Pidgeon, 1980) mineral assemblages were extensively retrogressed to lower amphibolite facies and locally, in patches and along fault and fracture zones, they were retrogressed to low greenschist facies. These altered rocks are cut by undeformed and unaltered dolerite dykes, probably of early Proterozoic age.

Previous work

In 1809 K. L. Giesecke collected a blue mineral, later identified as sapphirine (see Bøggild, 1953), from the sheared upper contact of the Fiskenæsset Complex in

Fiskenæsset town, but neither Giesecke nor Kornerup (1879), who first mapped the outline of the regional geology, recognised the anorthosites. The latter were discovered in the middle of the next century and mapped in a number of places in the Archaean gneiss complex of West Greenland (Ellitsgaard-Rasmussen & M. Mouritzen, 1954; Noe-Nygaard & Ramberg, 1961). They were thought to result from regional 'basification' during high grade metamorphism at depth (Ramberg, 1952), or to be derived from calcareous sedimentary rocks (Sørensen, 1955; Berthelsen, 1957) or volcanic ash bands (Berthelsen, 1960).

The discovery of chromite in anorthosite near Fiskenæsset town in 1964 led to the recognition that the anorthosites are part of a, now deformed and metamorphosed, suite of layered igneous rocks (Ghisler, 1966; Windley, 1966). Molybdenite from these rocks was found to have a Re/Os age of 3080 ± 70 m.y. (Herr *et al.*, 1967).

Igneous stratification was first recognised by Ghisler (1966, 1970) and Ghisler & Windley (1967), who mapped some of the chromitite horizons in detail, and Gormsen (1971), who mapped layers of amphibolite, meta-anorthosite, metagabbro, ultrabasic rocks and chromitite. The coarse grained relic igneous textures were thought to be the result of high grade metamorphism (Windley, 1967).

Windley (1969a) proposed the name Fiskenæsset Complex for this suite of anorthositic rocks and associated amphibolites near Fiskenæsset, and he further subdivided the complex into a sequence of 10 zones which were repeated as a mirror image about the centre of a 400 m wide layer on Qeqertarssuatsiaq (north) (plate 2) (Windley, 1971). At the same time Bowden (1970) independently discovered that the chemical differentiation trend shown by whole rock analyses of a suite of samples collected across the same layer showed a similar repeated pattern. In addition, the differentiation trend from mafic to felsic rocks, and the trend towards iron-enrichment shown by the geochemical results, were taken to indicate the primary way-up of the layered intrusion. The outer, more mafic, parts of the layer were interpreted as the lower part of the intrusion, whereas the inner more leucocratic parts were interpreted as the top of the intrusion, and the whole 400 m wide layer was thus considered to represent an isoclinal syncline. Bowden (1970) also discovered that the amphibolites (zones 1 and 10 of Windley, 1971) were much richer in iron than other parts of the complex and lay off the differentiation trend, and therefore did not appear to be part of the same layered complex as the anorthosites and gabbroic rocks.

Up to 1970, only that part of the Fiskenæsset Complex in the western part of the region close to Fiskenæsset town had been investigated. Between 1970–1975 the whole region which contains the Fiskenæsset Complex was systematically mapped by GGU. Bowden's (1970) opinion that the amphibolites were not part of the anorthosite complex was substantiated as, firstly mica schists and calc-silicates of probable supracrustal origin (see Kalsbeek, 1972; Kalsbeek & Myers, 1973), and then relic pillow lava and pyroclastic structures (Escher & Myers, 1975), were found in many of the amphibolites bordering the anorthosites. It was therefore suggested that the name Fiskenæsset Complex be restricted to the anorthosites and associated layered plutonic rocks (Myers, 1975a). As the systematic mapping of the region progressed, much larger outcrops of anorthosite were discovered in eastern, inland areas than those first described from the coastal area around Fiskenæsset. The igneous textures, structures and stratigraphy of the anorthositic rocks were found to be much better preserved in many of these inland outcrops (Myers, 1973; Walton, 1973).

Aspects of the whole rock geochemistry of small parts of the Fiskenæsset Complex are described by Windley (1973), Windley *et al.* (1973), Hutt (1974), Myers (1975b), Henderson *et al.* (1976), Morgan *et al.* (1976), Myers (1976a), Page *et al.* (1980) and Weaver *et al.* (1981). The mineral chemistry of parts of the Fiskenæsset Complex is described by Hutt (1974), Windley & Smith (1974), Myers & Platt (1977), Steele *et al.* (1977) and Bishop *et al.* (1980), and aspects of isotope geochemistry are described by Gancarz (1976). Some of the chromite deposits are described in detail by Ghisler (1976).

A variety of origins have been proposed for the Fiskenæsset Complex. It has been interpreted as a tectonic slice of lower crust (Windley, 1969b) and as an intrusion similar to the Bushveld Complex (Windley, 1969a). It has been likened to lunar anorthosites, and thought to be part of the Earth's primordial crust (Windley, 1970), as well as to the Great Dyke of Zimbabwe (formerly Rhodesia) (Windley, 1971). It has been regarded as a sill intruded between metavolcanics and a gneissic basement (Windley *et al.*, 1973), and equated with modern deep ocean floor ophiolites (Sutton & Windley, 1974) and with the early components of Cordilleran batholiths (Windley & Smith, 1976).

Scope of present work

This bulletin is based on a detailed study of one of the best preserved parts of the Fiskenæsset Complex at Majorqap qâva between 1971–1974, followed by a survey of all the main outcrops of the Fiskenæsset Complex in 1975. Most of these outcrops were mapped between 1970–1974 by a number of people in various degrees of detail. The only detailed accounts of any of these outcrops are given by Windley *et al.* (1973), Hutt (1974), Myers (1975a) and Ghisler (1976).

The survey in 1975 was aimed at correlating the stratigraphy of the Fiskenæsset Complex throughout the region, and providing a consistent map of the main rock units and regional structure. This involved various amounts of remapping some areas and the measurement of 47 stratigraphic sections. A simplified map of the Fiskenæsset Complex is given on plate 2, together with the location of the measured sections. A selection of stratigraphic sections are shown on figs 7, 11, 21, 34, 62 and 63. The Fiskenæsset Complex and the regional geology are shown in more detail on the GGU 1:100 000 scale coloured map sheets of Grædefjord (63 V1 S), Sinarssuk (63 V2 S) and Bjørnesund (62 V1 N), which also contain insets naming the geologists who first mapped various parts of the complex.

STRATIGRAPHY

Introduction

The names of various stratigraphic units of the Fiskenæsset Complex were changed as the mapping progressed, and these changes are outlined below to avoid confusion that may arise from reference to previous literature.

Windley (1971) divided a small part of the complex on the island of Qeqertarssuatsiaq (north) (plate 2) into 10 zones (table 1, a, column 1) which were slightly modified by Windley (1973) (table 1, a, column 2). In this area the complex is generally very strongly deformed and the thicknesses of the units given by Windley (1971, 1973) are maximum thicknesses which are of only local significance. This subdivision (table 1, a, column 2) was followed by Walton (1973), Windley *et al.* (1973) and Windley & Smith (1974), and in essence (table 1, a, column 3) by Steele *et al.* (1977), Bishop *et al.* (1980) and Weaver *et al.* (1981). However, as most of the amphibolites, zones 1 and 9 (10), were found to be metavolcanic host rocks into which the anorthosite complex was intruded (see Kalsbeek & Myers, 1973; Bridgwater *et al.,* 1976) it was necessary to revise the zonal subdivision of the complex. In addition, in many parts of the complex chromitite layers occur throughout a large section of the upper part of the complex and the subdivision into zones 6, 7 and 8 (or zones LA, Ch and UA) made on Qeqertarssuatsiaq (north) was found to be of only local significance. The rock names of some of the zones, such as zone 4, were also found to be inappropriate.

Myers (1975a) therefore avoided using the zonal terminology of Windley (1971, 1973) and Windley *et al.* (1973) and divided the anorthosite complex at Majorqap qâva into 6 major lithostratigraphic units. The word unit was preferred to zone simply to avoid further confusion with the variously used zonal names and numbers already in the literature. During the 1975 regional study of the whole Fiskenæsset Complex (Myers, 1975a, 1976b) an additional, seventh, unit was discovered above the Anorthosite Unit (table 1, column b).

Fig. 2 shows the complete stratigraphic succession of the whole complex, constructed from a large number of outcrops where the stratigraphy is best preserved. The thicknesses are based on the general thickness of each unit where it is least deformed, but because all the sections are to a certain extent deformed or disrupted, the original thickness of the instrusion is unknown.

Mineral-graded layers occur within all the major units, but most layers are isomodal, and the composition and colour index of most rocks fall into narrow ranges: anorthosite < 5% mafics, leucogabbro 15–30% mafics, gabbro 45–50% mafics, and

Table 1. Previous stratigraphic subdivisions of the Fiskenæsset Complex

a. *Qeqertarssuatsiaq (north)*						b. *whole Fiskenæsset region*	
Windley (1971)	Windley (1973)		Steele *et al.* (1977) Weaver *et al.* (1981)			Myers (1975a)	
Zones	Zones	Maximum thickness in metres	Zones		Maximum thickness in metres	Units	General deformed thickness in metres
10 Pyroxene amphibolite	9 Pyroxene amphibolite	50					
9 Ultramafic group							
8 Garnet anorthosite	8 Garnet anorthosite	75	UA	Upper Anorthosite	75		
7 Hornblende chromitite	7 Major chromitite	20	Ch	Chromitite	20	7 Upper gabbro	50
6 Anorthosite	6 Anorthosite	130	LA	Lower Anorthosite	130	6 Anorthosite	250
5 Ophitic gabbro	5 Homogeneous leuco-gabbro	250	ULG	Upper Leucogabbro	250	5 Upper leucogabbro	60
4 Mafic gabbro	4 Dark gabbro	60	G	Gabbro	70	4 Middle gabbro	40
3 Lower layered group	3 Lower layered zone	100	LLG	Lower Leucogabbro	100	3 Lower leucogabbro	50
2 Ultramafic group	2 Magnetite-rich layered ultramafics	100	UM	Ultramafic Rocks	100	2 Ultramafic	40
1 Pyroxene amphibolite	1 Pyroxene amphibolite	200				1 Lower gabbro	50

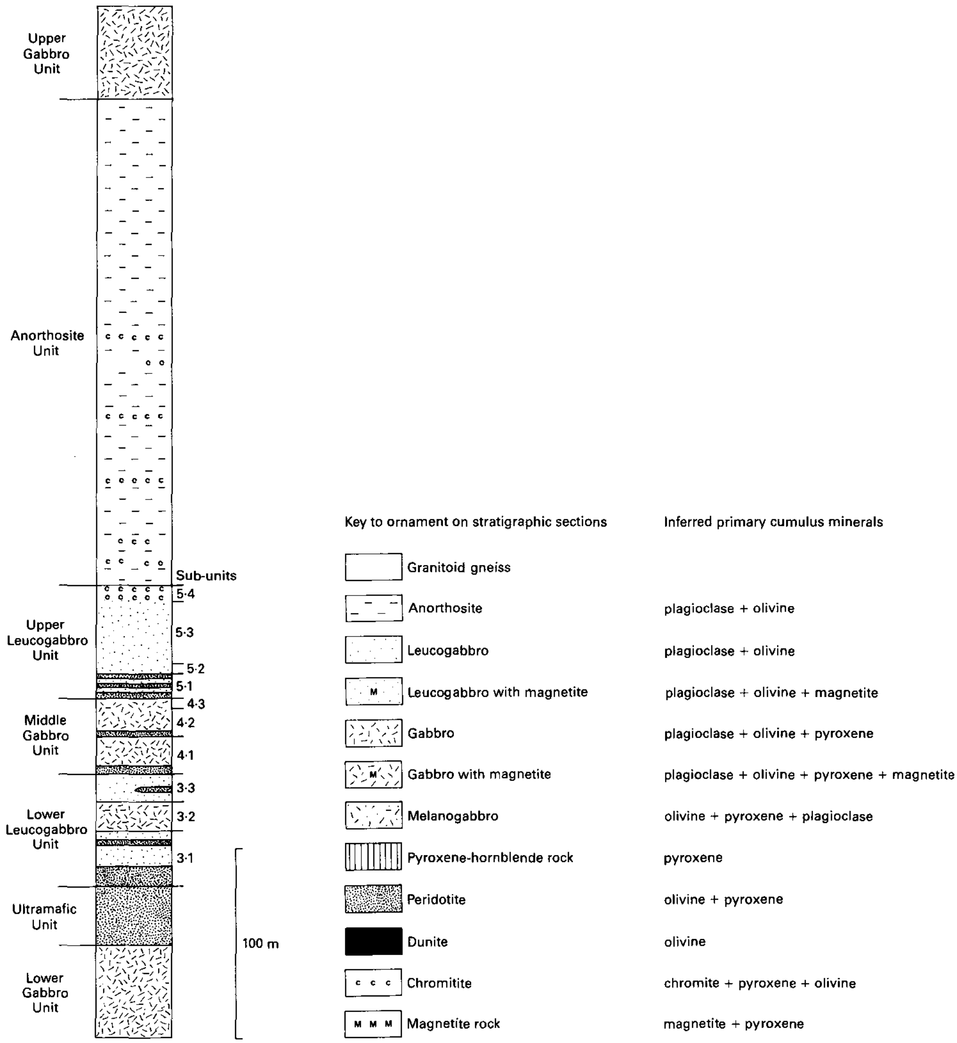


Fig. 2. Simplified stratigraphic succession of the Fiskenæsset Complex compiled from a number of outcrops.

ultramafic rock > 95% mafics. The igneous textures and kinds of igneous layering show that most rocks are cumulates, but most rocks were either partly or completely recrystallised during metamorphic and tectonic events, which followed the primary crystallisation of the igneous complex, and consist of metamorphic plagioclase and hornblende. Igneous mineral assemblages are completely preserved in only a few places but, for simplicity in description, igneous rock names are used without the prefix 'meta', regardless of the degree of metamorphic recrystallisation.

Because of extensive metamorphic recrystallisation, different kinds of cumulate textures such as adcumulate, mesocumulate and orthocumulate, cannot be distinguished. Equant plagioclase is the most widely preserved relic igneous mineral and the textures shown by aggregates of these crystals fall into two groups. In one group the igneous and relict igneous equant plagioclase crystals are separated by a matrix of metamorphic hornblende, probably derived from igneous pyroxene. These plagioclases may represent igneous primocrysts which were trapped in large fast-growing pyroxenes, either on the floor of the magma chamber or in suspension in the magma. In the latter case the plagioclase primocrysts would have settled as inclusions within giant pyroxenes and have accumulated on the floor of the magma chamber as secondary cumulates. In a second group, igneous and relict igneous equant plagioclase crystals are in contact with each other and so could either represent similar secondary cumulates (which sank as inclusions in larger pyroxenes) or primary cumulates – that is plagioclase crystals which settled individually or as aggregates onto the floor of the magma chamber.

In the following description, the words cumulus and cumulate are used in a broad sense to include crystals and crystal deposits which accumulated (on the floor of the magma chamber) either individually or as monomineralic crystal aggregates, or as inclusions within other settling crystals (such as inferred pyroxenes). Igneous pyroxenes, partly replaced by metamorphic hornblende, are only seen in a few of the felsic rocks but, as will be argued later, most hornblende is considered to be derived by metamorphism from igneous pyroxenes and olivines. Identical textures to those of undeformed Fiskenæsset leucogabbros can be seen in the Archaean Windimurra Intrusion of Western Australia (Ahmat & de Laeter, 1982) which has not suffered high grade metamorphism. In the Windimurra Intrusion the equant plagioclases can be seen to occur as inclusions within large pyroxene crystals.

This chapter concentrates on describing the succession of igneous rock types from places where the stratigraphy is best preserved, that is where the stratigraphy is least deformed and disrupted by granitoid intrusions. The effects of various amounts of progressive deformation on the appearance of the main igneous rock types are illustrated by photographs.

Volcanic host rocks

The Fiskenæsset Complex was intruded as a sub-concordant sheet into basic and ultrabasic volcanic rocks. Most of the volcanic rocks are now massive amphibolites with tholeiitic or komatiitic compositions showing geochemical affinities with ocean floor basalts (Rivalenti & Rossi, 1975; Rivalenti, 1976; Friend *et al.*, 1981), but showing little field evidence of their origin. In some places, however, they show

Fig. 3. Deformed pillow lava structure in amphibolite adjacent to the roof of the Fiskenæsset Complex at locality 41. The cores of pillows are leucocratic amphibolite and the rims of pillows and matrix between them are more melanocratic. The hammer is 28 cm long.

Fig. 4. More strongly deformed pillow lava structure in amphibolite north–east of locality 13.

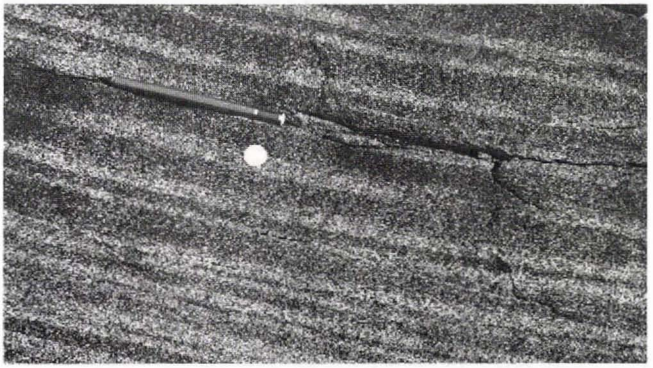


Fig. 5. Very strongly deformed pillow lava structure with leucocratic cores of pillows streaked out to form banding, in amphibolite, north-east of locality 13.

pillow lava, pillow lava breccia and pyroclastic structures (fig. 3), cut by dense swarms of basic dykes. These rocks can be traced through zones of increasing deformation (fig. 4) until they become banded or massive amphibolites (fig. 5).

Intrusive contacts of the Fiskenæsset Complex with these rocks are rarely seen. No intrusive bottom contact has been recognised, but the roof contact is well preserved at locality 41 (plate 2) (Escher & Myers, 1975), where the Anorthosite Unit is in contact with amphibolite with pillow lava structure. The Upper Gabbro Unit does not appear to have been developed at this locality. The anorthosite locally veins the metavolcanics, and rafts of these rocks occur within the anorthosite up to 20 m below the roof contact (fig. 6). The metavolcanic rafts are progressively more recrystallised the further they occur below the roof, and the associated anorthosite appears to be progressively more contaminated. Volcanic structures are well preserved in rafts up to 5 m below the roof, and garnet is absent, but below this level the volcanic structures become fainter and are obliterated as garnet becomes more abundant in both the anorthosite and amphibolite.

In some places (such as localities 4, 34, 41 and 47, plate 2) where the Upper Gabbro Unit does not appear to have been developed, the top of the Anorthosite Unit is in contact with ultramafic rocks adjacent to metavolcanic amphibolites or metasedimentary rocks (Herd, 1973). In these places a zone of metasomatic reaction occurs up to 2 m thick containing the rare minerals sapphirine and kornerupine, together with plagioclase, hornblende, enstatite, gedrite, pargasite, anthophyllite, spinel, ruby corundum and phlogopite (Herd *et al.*, 1969; Herd, 1972, 1973; Walton, 1973; Rivalenti, 1974; Friend & Hughes, 1977). These assemblages are interpreted by Herd (1973) as the result of partial assimilation of host rocks by the anorthosite during its intrusion, followed by the introduction of potash feldspar from the granitoid gneisses during amphibolite facies metamorphism. Lenses of similar rocks also occur as xenoliths within the Anorthosite Unit.

Lower Gabbro Unit (unit 1)

This unit is poorly preserved, it is generally strongly deformed and thoroughly recrystallised with a completely metamorphic texture, and is therefore difficult to recognise. The lowest part of the unit, which is clearly distinct, is in contact with younger granitoid intrusions (now gneisses). No clearly original contact with metavolcanic amphibolites was observed.

The Lower Gabbro Unit is best preserved at localities 25 and 35 (plate 2). At locality 35 near Tasiussâ a 50 m thick sequence of this unit occurs below the lowest dunite horizon of the Ultramafic Unit. Most rocks are massive amphibolite composed of plagioclase and hornblende in equal amounts with equigranular metamorphic textures. The origin of these rocks alone is not clear, but they contain layers of leucocratic and melanocratic amphibolite, they grade into less deformed layers with coarser grained, relict igneous cumulate textures, and range in composition

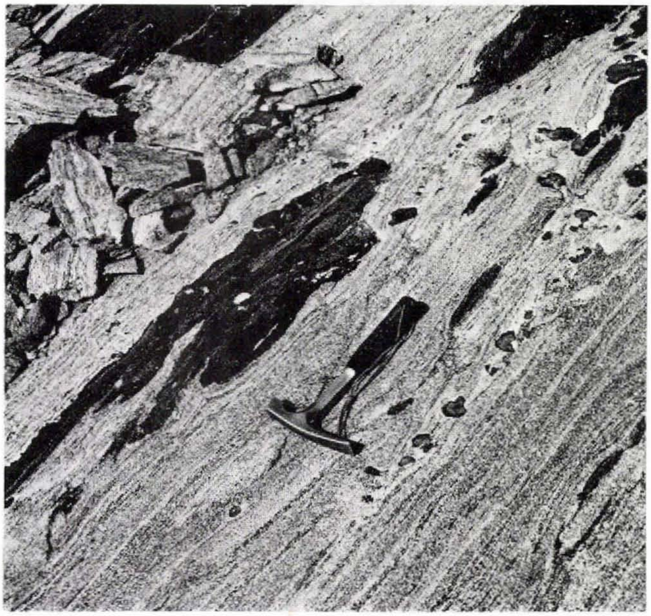


Fig. 6. Xenoliths of amphibolite with banding derived from deformed pillow lava structures in anorthosite with large garnets (colour zoned from pale pink cores to dark red rims), 10 m below the roof of the Fiskenæsset Complex at locality 41.

from melanogabbro to gabbro and leucogabbro. These distinctly plutonic layers occur intermittently throughout the recognised sequence of the Lower Gabbro Unit and suggest that most of the associated massive amphibolite is derived from gabbro.

A similar sequence of rocks, 10 m thick, occurs at locality 25 at Majorqap qâva below the basal olivine hornblendite of the Ultramafic Unit. These rocks show finer-scale layering than those near Tasiussâ with thin layers of leucocratic and melanocratic amphibolite a few centimetres or less thick. Some layers show poorly developed relict igneous mineral-grading. Like the rocks of the Lower Gabbro Unit near Tasiussâ, they are interpreted as a sequence of mainly gabbro with layers of melanogabbro and leucogabbro.

Ultramafic Unit (unit 2)

The Ultramafic Unit consists of a number of cyclic mineral-graded sub-units of dunite, peridotite and hornblende rock with a total thickness of about 40 m. Although many of these rocks are strongly deformed and have thoroughly metamorphic textures, igneous textures, minerals and relict igneous minerals have survived to a much greater degree than in the Lower Gabbro Unit. The rocks are interpreted as mainly olivine, olivine + pyroxene + spinel and pyroxene + spinel ± hornblende cumulates.

At locality 35 near Tasiussâ (plate 2), 15 cyclic sub-units occur (fig. 7c). The top and bottom contact of each sub-unit is sharp, but internally each sub-unit is different – some are massive whereas many show mineral grading upwards from dunite to olivine-hornblendite to hornblendite. Many dunite layers preserve igneous cumulus olivines which are size-graded from 5 mm in diameter upwards to < 1 mm in diameter.

At locality 25 at Majorqap qâva (plate 2) the Ultramafic Unit consists of 9 cyclic mineral-graded sub-units of dunite, olivine hornblendite and hornblendite (fig. 7b), similar to those near Tasiussâ. The Ultramafic Unit is also well preserved at locality 13 east of Uivfaup nûa (plate 2) (fig. 7a), where 9 cyclic mineral-graded sub-units occur on one side of an F1 downward-facing syncline. On the other side of the fold core these sub-units are more disrupted by granitoid intrusions and so are less distinct. Plagioclase also appears as a cumulus mineral at the top of some of these sub-units in which ultramafic layers grade upwards into gabbro or thin anorthosite.

Some layers of dunite and olivine-hornblende rock show size-graded cumulus

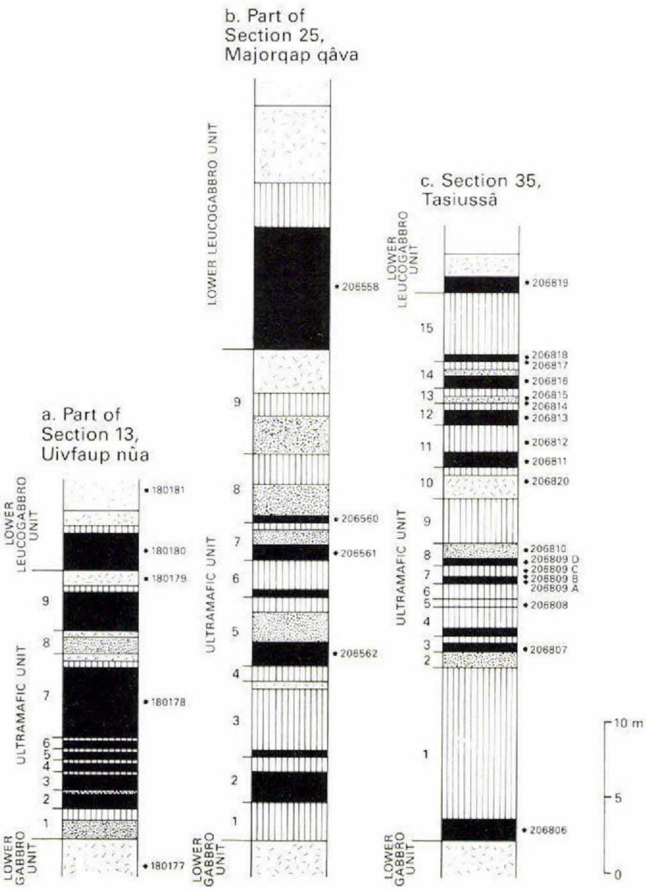


Fig. 7 *a*. Part of section 13 east of Uivfaup nûa of the Ultramafic Unit on one limb of a downward-facing F1 syncline. *b*. Part of section 25 at Majorqap qâva of the Ultramafic Unit showing 9 cyclic sub-units. *c*. Section 35 near Tasiussâ of the Ultramafic Unit showing 15 cyclic sub-units. Key to ornament in fig. 2.

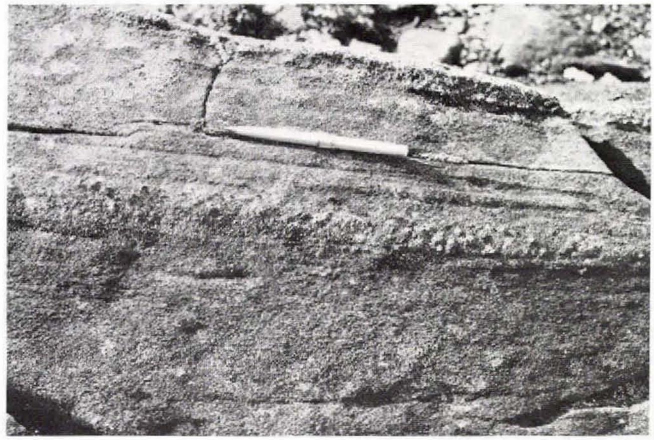


Fig. 8. Size-graded olivine cumulate dunite (paler) above peridotite (darker). Ultramafic Unit, locality 25 at Majorqap qâva. Original way up towards top of photograph. The pen is 13 cm long.

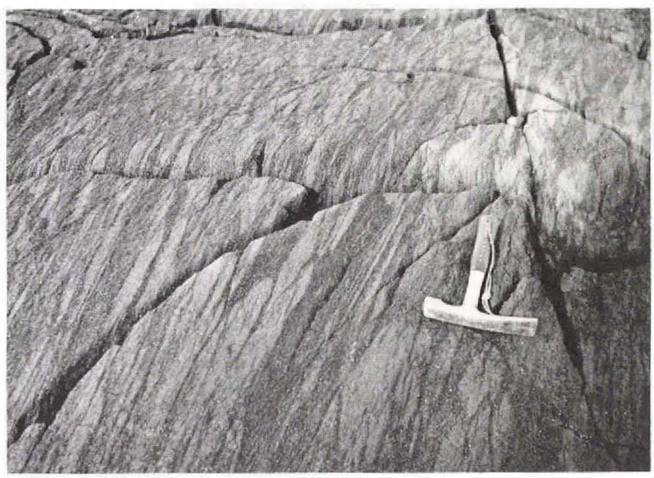


Fig. 9. Ellipsoidal aggregates of metamorphic olivine derived from deformed large cumulus olivines. Ultramafic Unit, locality 13 east of Uivfaup nûa.

Fig. 10. Trough layers of plagioclase cumulate anorthosite (below hammer) in peridotite of the Ultramafic Unit below leucogabbro (white) of the Lower Leucogabbro Unit. 3 km north-east of locality 35. Original way up towards top of photograph.

olivine (fig. 8), and in others two different cumulus grain sizes of olivine occur together. Some relict cumulus olivines are very large, and individual cumulus crystals are now deformed to ellipsoidal aggregates of metamorphic olivine with long axes of 15 × 2 × 2 cm (fig. 9). Some lenticular trough layers of gabbro occur in olivine hornblende rocks and these, together with lenticular anorthosite trough layers in dunite 3 km north-east of locality 35 (plate 2, fig. 10), are interpreted as channel deposits formed by current action, similar in structure and origin to ultramafic channel deposits in leucogabbro described later from the lower part of the Upper Leucogabbro Unit.

Lower Leucogabbro Unit (unit 3)

This unit mainly consists of leucogabbro, some gabbro and minor ultramafic layers, and has a total thickness of 50 m. Igneous structures, textures and minerals are well preserved in many places, and 3 major sub-units can be widely recognised, called 3.1, 3.2 and 3.3 (fig. 2).

Near Tasiussâ (plate 2, section 37; fig. 7c) and at Majorqap qâva (plate 2, section 25; fig. 7b) the base of the unit is marked by a layer of dunite which grades upwards through peridotite, hornblendite and gabbro to leucogabbro. East of Uivfaup nûa (plate 2, section 13; fig. 7a) it is marked by dunite which grades upwards through hornblendite and gabbro to leucogabbro.

Although the basal contact is only locally preserved at Majorqap qâva and the lowest part of the unit is generally in contact with younger granitoid intrusions (now gneisses), the rest of the unit together with the overlying Middle Gabbro and Upper Leucogabbro Units are exceptionally well preserved and have been studied in most detail at this locality. The summit cliff section at Majorqap qâva (plate 2, section 18) is therefore described as the type section of these units.

In this section (fig. 11) the lower and upper sub-units 3.1 and 3.3 mainly consist of leucogabbro with equant igneous or relict igneous plagioclase 1–5 mm in diameter set in a matrix of hornblende (fig. 12). Most layers are massive and show a uniform grain size (fig. 12), but some layers contain two or a variety of grain sizes of igneous plagioclase up to 2 cm in diameter set in a matrix of uniform leucogabbro with igneous or relict igneous plagioclase 1–5 mm in diameter (fig. 13). Some of these layers show size grading of igneous plagioclase (fig. 14).

The lower sub-unit 3.1 contains a number of lenticular ultramafic layers. These layers have sharp bases and show mineral grading from olivine-pyroxene lower portions to hornblende-rich upper portions. Some layers are rich in spinel and magnetite.

The middle sub-unit of gabbro 3.2 at Majorqap qâva consists of fairly uniform gabbro with equant relict igneous plagioclase 1–5 mm in diameter, similar in texture to the main part of sub-units 3.1 and 3.3.

Leucogabbro of the upper sub-unit 3.3 is characterised by schlieren of anortho-

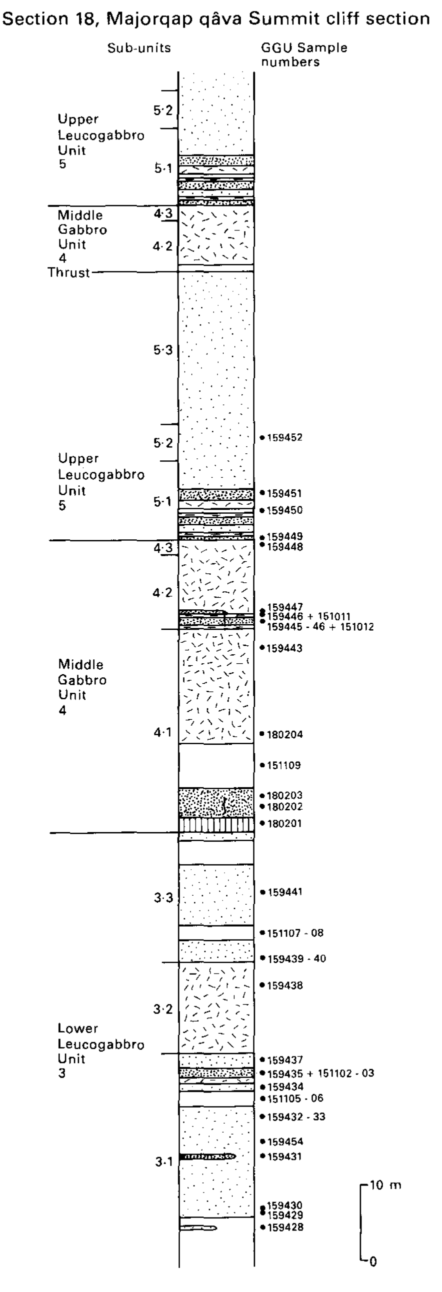
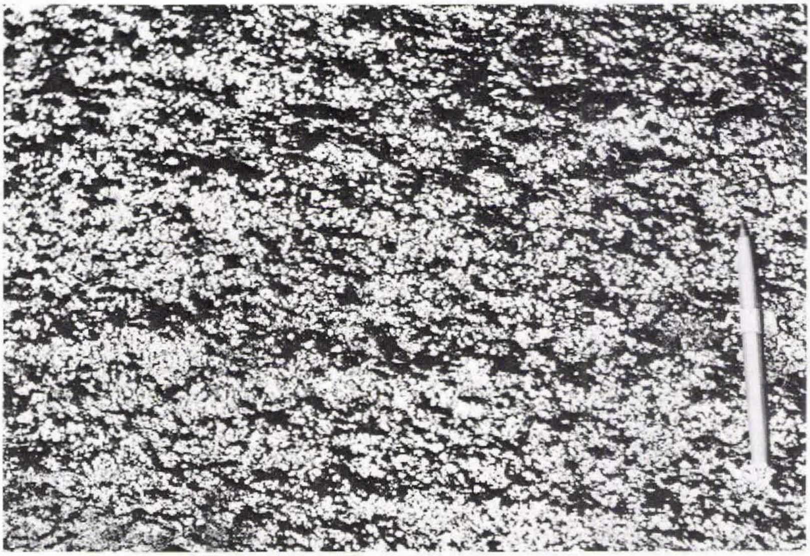


Fig. 11. Section 18 at Majorqap qâva showing the Lower Leucogabbro, Middle Gabbro and lower part of the Upper Leucogabbro Units. A thrust near the top causes repetition of part of the succession. Key to ornament in fig. 2.



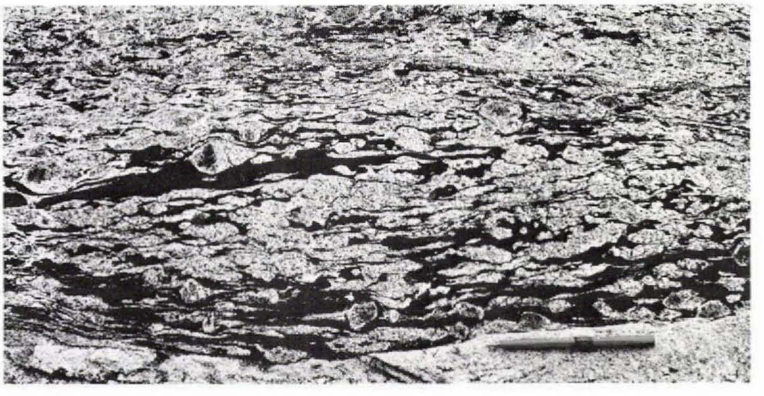
a

b

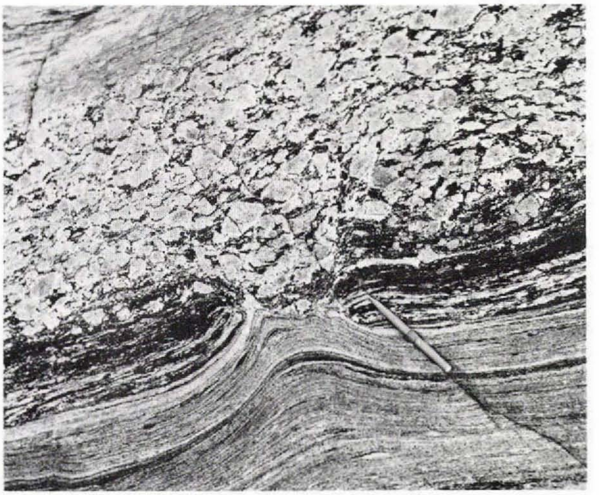
Fig. 12. Recrystallised leucogabbro with equant relict igneous plagioclase, typical of sub-units 3.1 and 3.3 of the Lower Leucogabbro Unit. Locality 18 at Majorqap qâva. *a*. Undeformed, note the clustering of plagioclase crystals. *b*. Showing the development of schistosity and boudinage of less deformed layers.

site, probably derived from sub-spherical clusters of equant igneous plagioclases (originally enclosed in large pyroxenes). These rocks are generally deformed and completely recrystallised, but they occur in a little deformed state east of Uivfaup nûa (plate 2, section 13; fig. 15).

a



b



c

Fig. 13. Partly recrystallised leucogabbro with equant igneous plagioclase of two main grain sizes. Sub-unit 3.3 of the Lower Leucogabbro Unit, locality 26 at Majorqap qâva. *a*. Undeformed. *b*. Moderately deformed with migration of plagioclase to form metamorphic 'tails' to large igneous plagioclases. *c*. Strongly deformed with tectonic banding below, boudinaged deformed mafic layer in the centre, and associated conjugate fracturing of less deformed layer above.

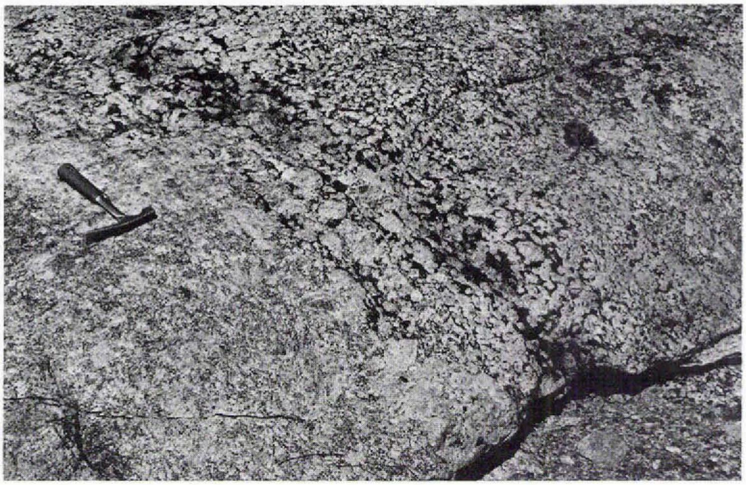


Fig. 14. Undeformed, partly recrystallised size-graded plagioclase in leucogabbro, sub-unit 3.3 of the Lower Leucogabbro Unit, locality 19 at Majorqap qâva. Original way up towards the top right.

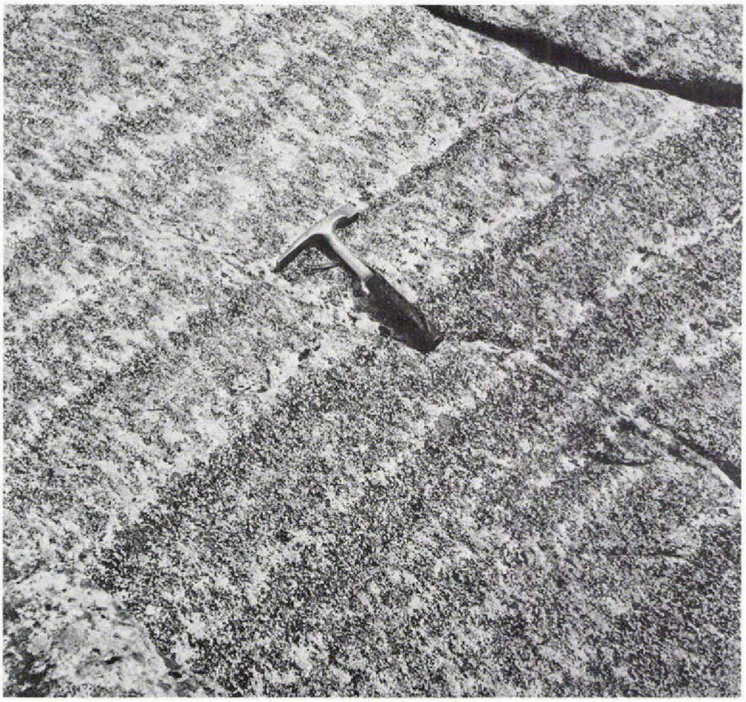


Fig. 15. Little deformed, recrystallised clusters of plagioclase in leucogabbro, typical of sub-unit 3.3 of the Lower Leucogabbro Unit, here associated with mineral-graded layers. Locality 13, east of Uivfaup nûa. Original way up towards the bottom right. The hammer shaft is parallel with poorly developed F1 schistosity.

Middle Gabbro Unit (unit 4)

The Middle Gabbro Unit mainly consists of gabbro 40 m thick which can be divided into 3 main sub-units (fig. 2). In the summit cliff section at Majorqap qâva (plate 2, section 18; fig. 11) the base of the unit is marked by 6 m of ultramafic rocks: a layer of hornblende-orthopyroxene-spinel rock overlain by peridotite. The rest of the lower sub-unit 4.1 consists of uniform gabbro with poorly defined mineral-graded layers of great lateral extent. This is overlain by a fairly massive peridotite layer 1 m thick enclosed by anorthosite, and then by another peridotite 50 cm thick (fig. 11). These form the basal layers of the middle sub-unit 4.2 which mainly consists of uniform gabbro with sporadic, well defined, mineral-graded trough layers 25 cm thick arranged in vertical columns (fig. 16). The upper sub-unit 4.3 consists of uniform gabbro with scattered tabular plagioclase crystals up to 2 cm long, rounded clusters, 20 cm in diameter, of equant or wedge-shaped plagioclase either 1–5 cm in diameter or up to 10 cm long (fig. 17), angular fragments of leucogabbro with hornblende-rich rims, and thick mineral-graded layers of melanogabbro (fig. 18).

Fig. 16. Mineral-graded layers of gabbro in uniform gabbro, sub-unit 4.2 of the Middle Gabbro Unit. Original way up towards top of photographs. *a*. Undeformed mineral-graded layers (above) cut by strongly marked tectonic schistosity (below), south-west of locality 35 near Tasiussâ.

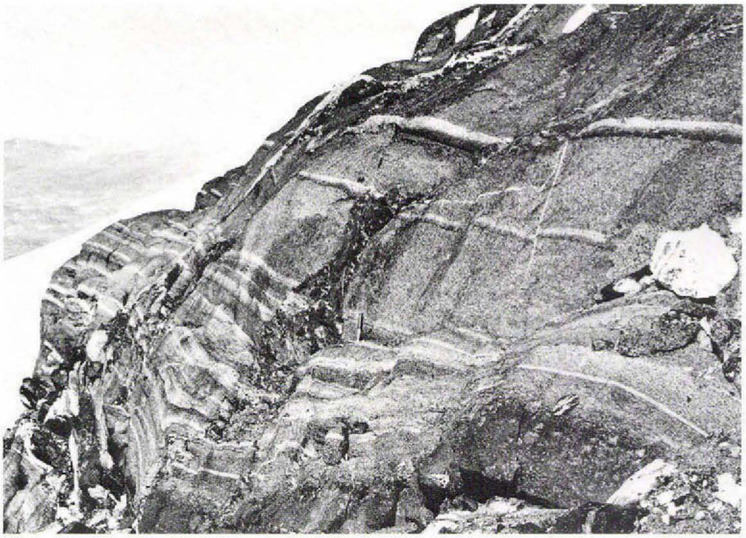


Fig. 16 *cont. b.* Little deformed vertical column of trough layers, locality 18 at Majorqap qâva.

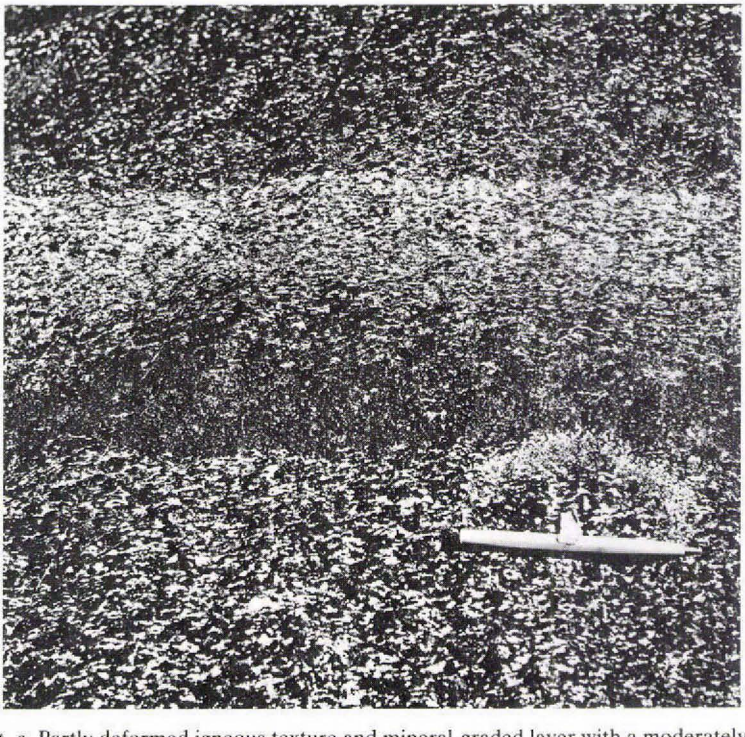


Fig. 16 *cont. c.* Partly deformed igneous texture and mineral-graded layer with a moderately developed tectonic schistosity parallel with the pen. Locality 18 at Majorqap qâva.

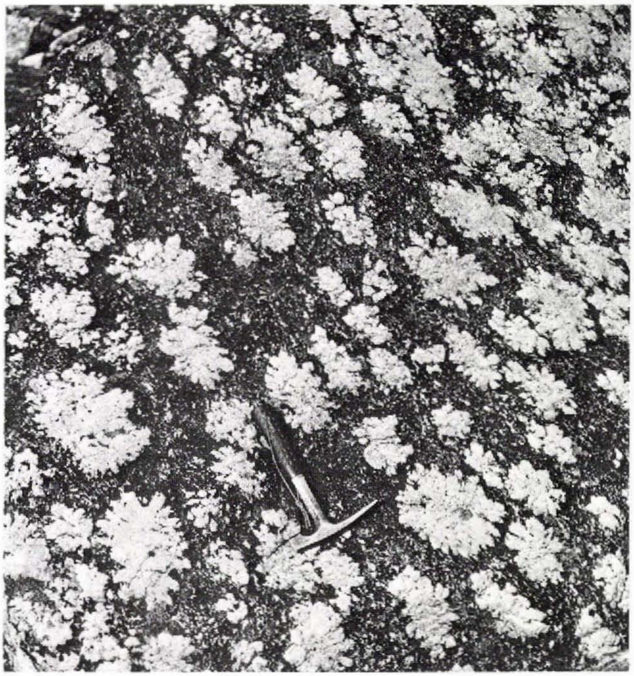


Fig. 17. 'Snow-flake' like clusters of plagioclase with radial structure in uniform gabbro, sub-unit 4.3 of the Middle Gabbro Unit. Locality 23 at Majorqap qâva, plan view of a bedding surface.

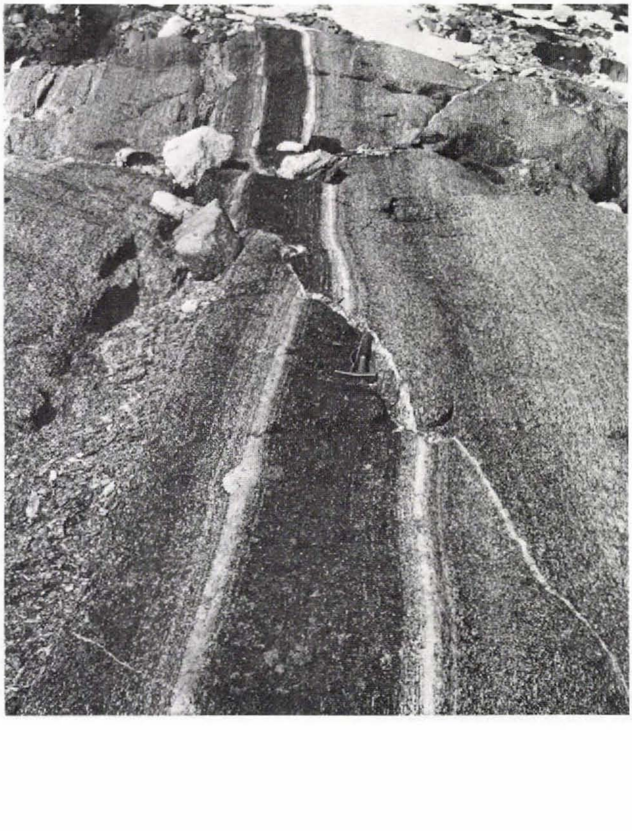


Fig. 18. Mineral-graded layers typical of sub-unit 4.3 of the Middle Gabbro Unit. Note the isolated cluster of plagioclase crystals to the lower left of the hammer. Locality 43, original way up towards the left.

Upper Leucogabbro Unit (unit 5)

This unit comprises coarse grained leucogabbro with minor layers of ultramafic rocks and chromitite. It is generally 60 m thick and can be divided into 4 sub-units (fig. 2).

At Majorqap qâva (plate 2, section 18; fig. 11) the lowest sub-unit 5.1 is 10 m thick and consists of lenticular mineral-graded ultramafic trough layers (fig. 19) interlayered with plagioclase and plagioclase-cluster cumulates. The latter consist of 20 cm diameter clusters of equant or wedge-shaped plagioclase crystals 5–20 mm in diameter or up to 5 cm long in a matrix of hornblende or hornblende and plagioclase (fig. 20). The ultramafic layers are interpreted as deposits from turbidity currents in scoured channels (Myers, 1976c). At Majorqap qâva (plate 2, section 26; fig. 21a) many consist of massive peridotite underlain by zones rich in spinel, and overlain by plagioclase-chromite cumulate and hornblende chromitite (figs 19, 22), but further east (plate 2, section 31; fig. 21b) many of these layers are mineral-graded from dunite upwards through peridotite and hornblendite to leucogabbro and anorthosite. Chromite is concentrated in thin layers in some of the dunites and peridotites (fig. 23).

At Majorqap qâva (plate 2, sections 18, 19, 26) these deposits are overlain by

Fig. 19. Peridotite trough layer (centre) in leucogabbro, sub-unit 5.1 of the Upper Leucogabbro Unit, overlain to the right by slumped blocks of hornblende chromitite (see map fig. 49). Note the 4 cm thick dark hornblende rim around the peridotite. Locality 26 at Majorqap qâva. Original way up towards the right.

a

b

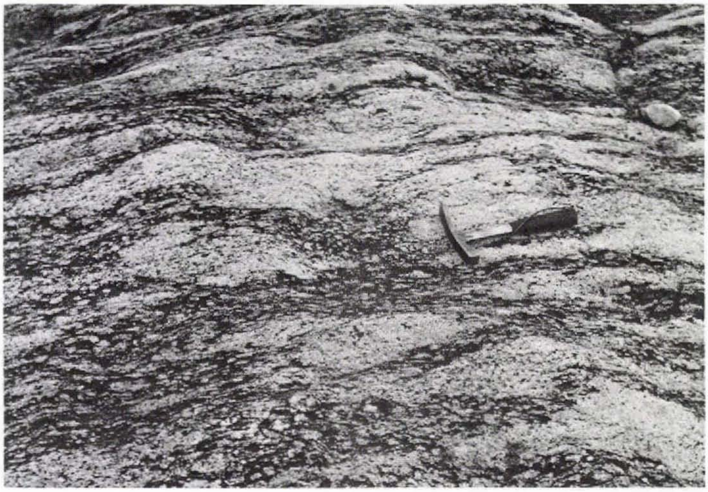


Fig. 20. Clusters of plagioclase in leucogabbro typical of sub-unit 5.1 of the Upper Leucogabbro Unit. Locality 26 at Majorqap qâva. Original way up towards top of photographs. *a*. Undeformed. *b*. Deformed.

sub-unit 5.2, 5 m thick, which consists of leucogabbro with equant igneous and relict igneous plagioclase 2–5 cm in diameter in a matrix of leucogabbro. This has a sharp wavy contact (fig. 24) with sub-unit 5.3, 35 m thick, in which igneous and relict igneous plagioclase lies in a matrix of hornblende. Within sub-unit 5.3 there is a steady upward increase in the size of equant igneous and relict igneous plagioclase from 2–5 cm to 10 cm in diameter (figs 25, 26). Above this, chromite is a prominent cumulus mineral in sub-unit 5.4, 10 m thick, with 10 cm diameter equant igneous plagioclase in a matrix of metamorphic hornblende and biotite.

This sequence of sub-units is widespread throughout the eastern part of the Fiskenæsset Complex, but in the west it is less distinct or absent, and the size of igneous and relict igneous plagioclase is generally less.

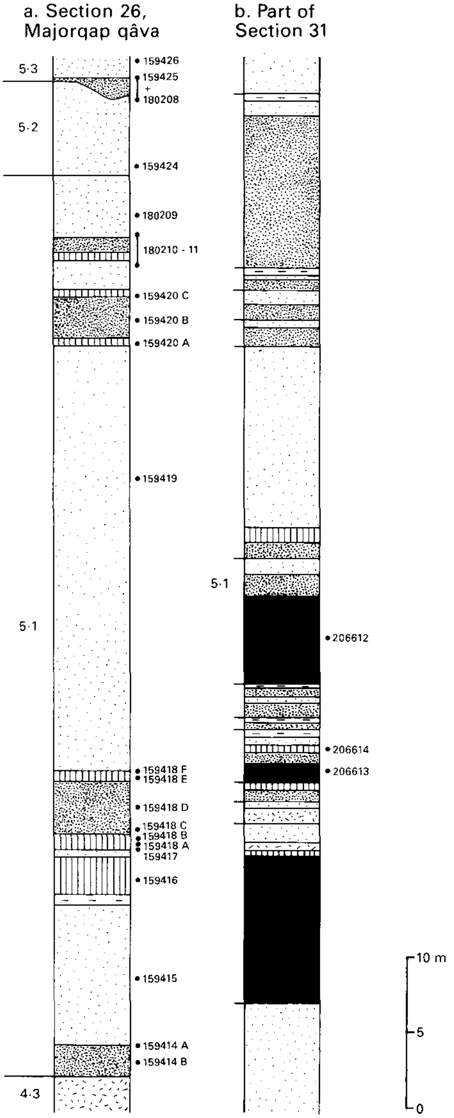


Fig. 21. Stratigraphic sections of sub-unit 5.1 of the Upper Leucogabbro Unit. *a.* Section 26 at Majorqap qâva. *b.* Western part of section 31. Key to ornament in fig. 2.

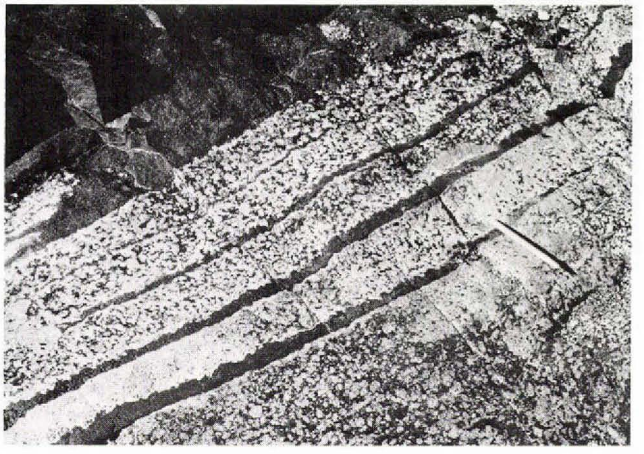
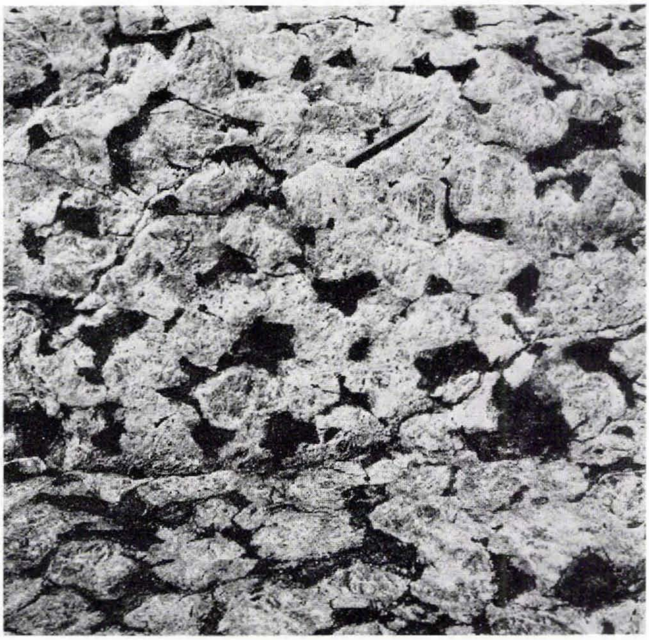


Fig. 22. Layers of chromite cumulate showing irregular bases (lode casting) and smooth tops, in leucogabbro above massive peridotite. Sub-unit 5.1, locality 26 at Majorqap qâva. Original way up towards the bottom right.

Fig. 23. Layers of chromite in a peridotite trough layer in sub-unit 5.1 of the Upper Leucogabbro Unit. Note the smooth sharp base and irregular mineral-graded top of the thickest layer. These layers contain the highest concentration of platinum yet found in the Fiskenæsset Complex (Page *et al.*, 1980). Locality 31, original way up towards top of photograph.

Fig. 24. Sharp wavy contact between sub-unit 5.2, with partly recrystallised large igneous plagioclase in a matrix of fine grained hornblende and plagioclase (right), and sub-unit 5.3, with large igneous and relict igneous plagioclase in a matrix of metamorphic hornblende (left). South-west of locality 26, original way up towards the bottom left.



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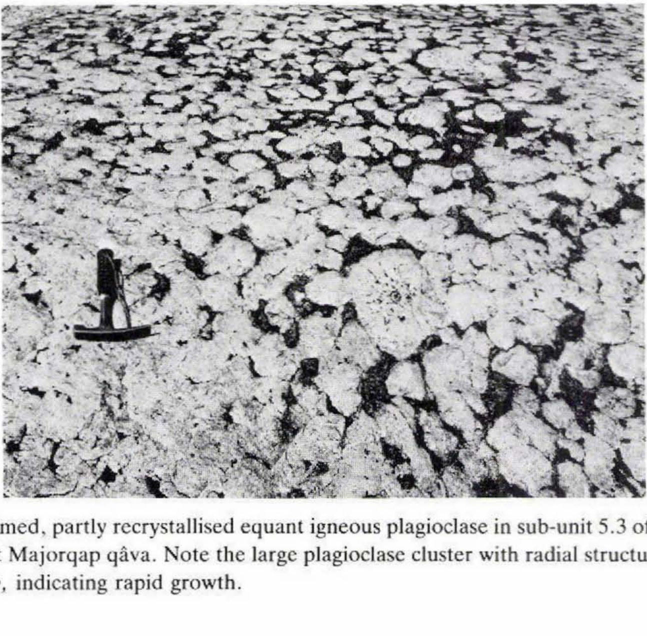


b

Fig. 25. Undeformed, partly recrystallised equant igneous plagioclase in sub-unit 5.3 of the Upper Leucogabbro Unit at Majorqap qâva. Note the large plagioclase cluster with radial structure to the right of the hammer in *b*, indicating rapid growth.

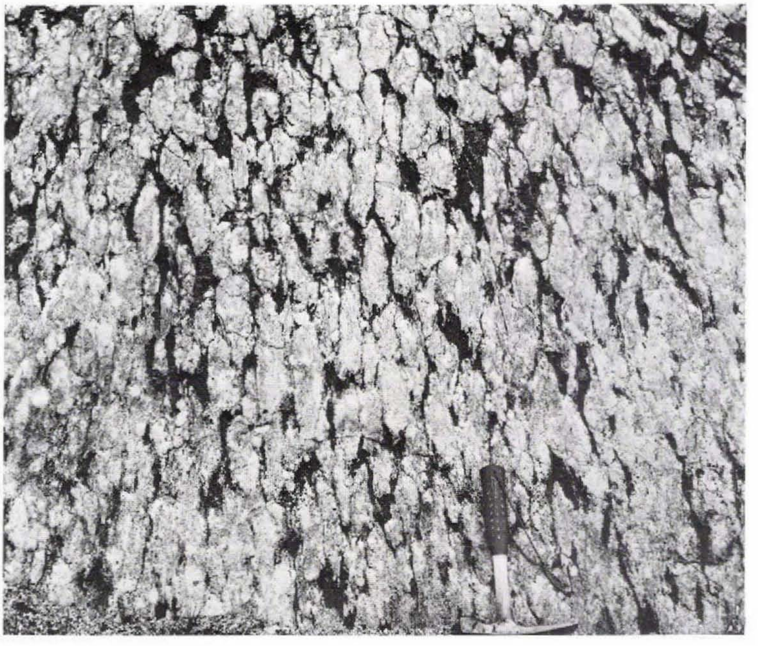
Anorthosite Unit (unit 6)

The Anorthosite Unit, 250 m thick, dominates the upper part of the intrusion. In spite of its relatively great thickness it lacks any marked internal stratigraphy and the most prominent layering within the unit is tectonic rather than igneous in origin. Overall, the Anorthosite Unit is much more strongly deformed than the Lower Leucogabbro, Middle Gabbro and Upper Leucogabbro Units. Where it is least deformed at Majorqap qâva and in the large anorthosite outcrop between localities 43 and 44 (plate 2), it consists of irregular sub-spherical patches of leucogabbro typically 30 cm in diameter in a matrix of anorthosite (fig. 27). A smaller number (locally up to 20%) of the mafic patches consist of an equigranular mosaic of plagioclase and garnet. Where the rocks are least recrystallised both the leucogabbro patches and anorthosite matrix are seen to consist of a variety of grain sizes of equant igneous plagioclase 1–10 cm in diameter dispersed between even grained metamorphic plagioclase 1–5 mm in diameter (fig. 28). With increasing deformation, the leucogabbro and garnet-plagioclase patches are seen to become ellipsoidal schlieren (fig. 29) which are then streaked out into a fine banding (fig. 30). Large igneous plagioclases are seen to be recrystallised to finer grained aggregates. Recrystallisation outlasted deformation and most anorthosite has a polygonal mosaic texture.

In the central and eastern part of the complex, chromitite occurs in discontinuous layers and lenses throughout most of the Anorthosite Unit. Locally it forms massive mafic layers or lenses of chromite-hornblende-biotite rock up to a metre thick, but chromite generally occurs evenly distributed with hornblende and biotite between igneous and relict igneous plagioclase 1–2 cm in diameter, in layers or lenses 1–2 m thick where cumulus plagioclase makes up about 80% of the rock (fig. 31). The regularity with which chromite occurs with igneous or relict igneous plagioclase, which is coarser grained than the adjacent anorthosite, suggests that this association is primary (fig 32). Some of these lenticular chromite-bearing rocks are trough shaped and are interpreted as deposits formed from currents in scoured channels (Myers, 1976c; and discussed later).

In the western part of the complex, chromite appears to be restricted to a smaller number of more continuous and thicker layers in the upper part of the Anorthosite Unit (Ghisler & Windley, 1967; Windley *et al.*, 1973; Ghisler, 1976). These layers are typically 1–2 m thick and, as elsewhere in the complex, contain igneous and relict igneous cumulus equant plagioclase 1–2 cm in diameter, larger than in the adjacent anorthosite, with equant cumulus chromite 0.1–0.5 mm in diameter. In some places, especially on Qeqertarssuatsiaq (north) and near Tasiussâ (plate 2) these layers were disrupted by magmatic slumping and, on a regional scale, were pulled apart into trains of tabular blocks or piled up into a chaotic jumble of chromitite blocks (fig. 33).

Metamorphic garnet occurs in the upper 50 m of the Anorthosite Unit (Walton,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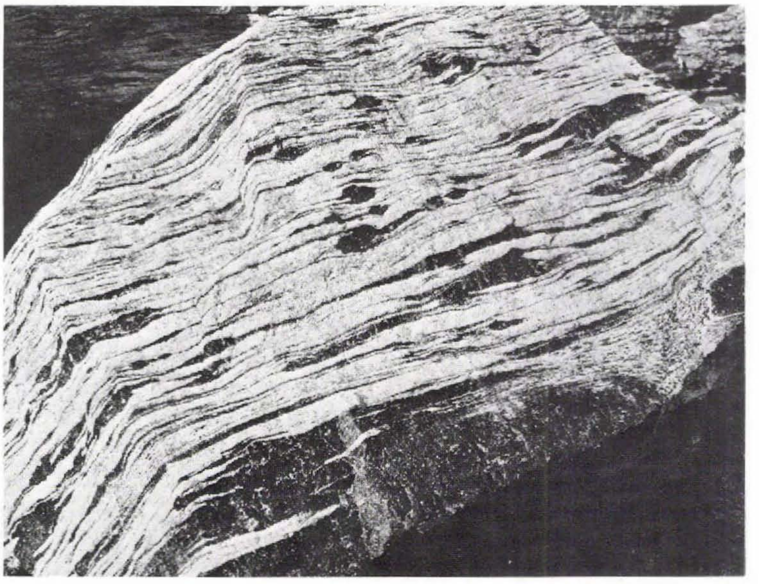
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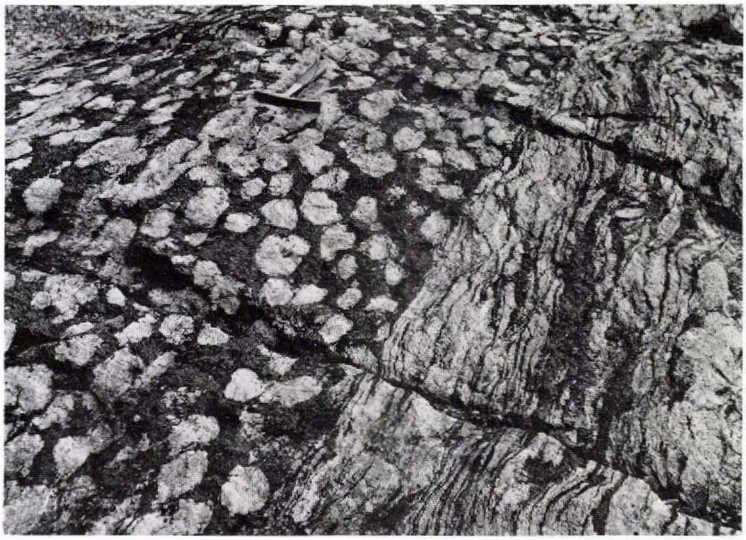

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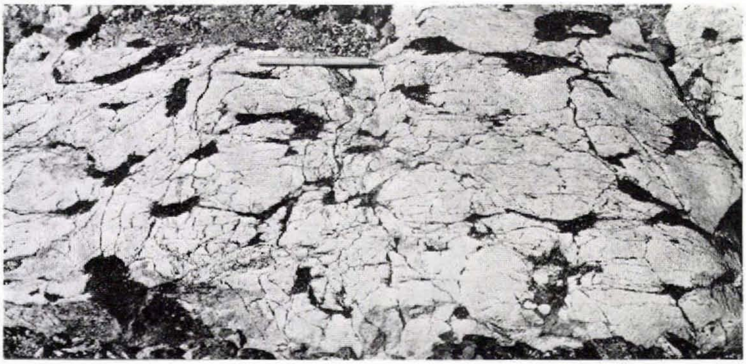


d

Fig. 26. Deformed, recrystallised equant igneous plagioclase in sub-unit 5.3 of the Upper Leucogabbro Unit at Majorqap qâva. Metamorphic aggregates after large igneous plagioclases are prolate ellipsoids in *a*, and oblate ellipsoids in *b* and in the right of *c* in a shear zone cutting undeformed leucogabbro to the left. *d*. Brittle deformation of relict igneous plagioclase during amphibolite facies metamorphism.

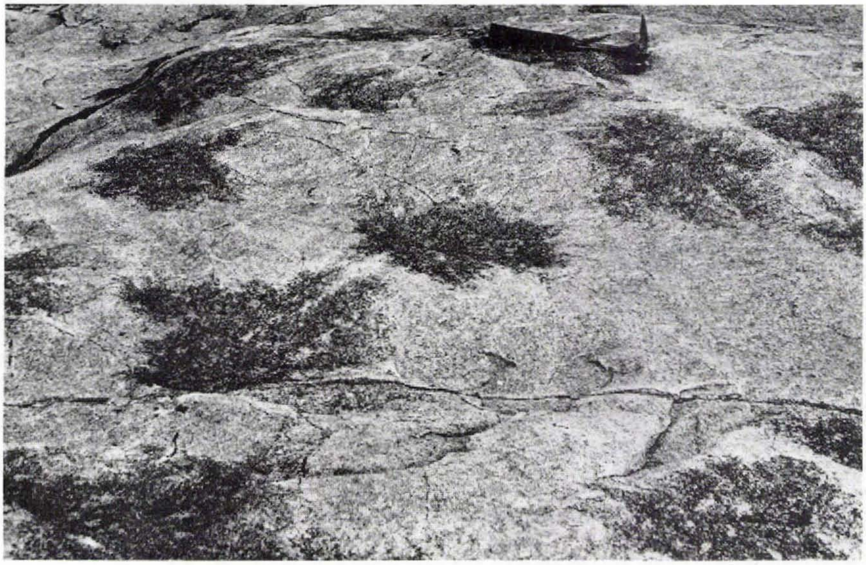


Fig. 27. Undeformed, recrystallised sub-spherical patches of leucogabbro in anorthosite, typical of the least deformed anorthosite of the Anorthosite Unit at Majorqap qâva.

Fig. 28. Undeformed, partly recrystallised patches of leucogabbro in anorthosite in the Anorthosite Unit near Qagsse summit. The least recrystallised part of the Anorthosite Unit showing a variety of grain sizes of igneous plagioclases in both the leucogabbro patches and anorthosite matrix.

Fig. 29. Recrystallised, deformed anorthosite with schlieren of leucogabbro derived from sub-spherical patches, in the Anorthosite Unit at Majorqap qâva.

a

b

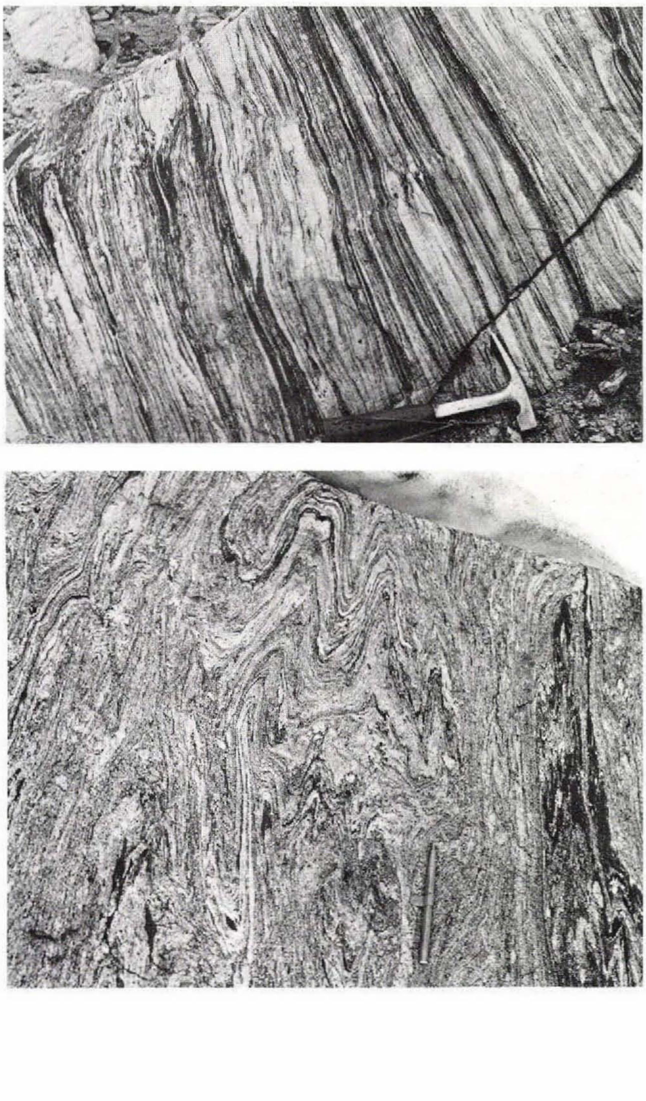


Fig. 30. Recrystallised, strongly deformed anorthosite with leucogabbro patches streaked out to form a tectonic banding in *a*, which is folded in *b*. The Anorthosite Unit at Majorqap qâva.

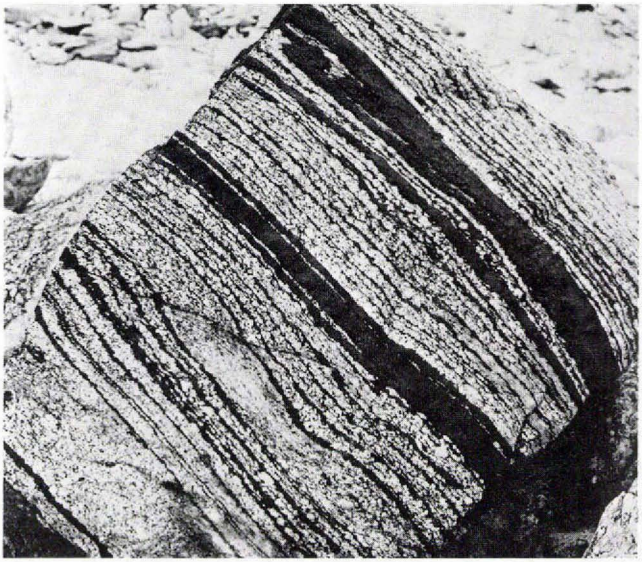


Fig. 31. Layers of chromite-hornblende-biotite and plagioclase- chromite-hornblende in a 2 m thick chromite-bearing layer in anorthosite of the Anorthosite Unit, 8 km west of locality 41. The thickest mafic layers are 6 cm thick.

1973; Windley *et al.*, 1973). At locality 41 (plate 2) the occurrence of garnet is associated with the recrystallisation and partial assimilation of rafts of basic volcanic roof rocks, and thus the presence of garnet in this part of the unit may be the result of contamination (fig. 6). At this same locality and elsewhere (plate 2, localities 32, 33, 42, 44, 46) the uppermost 5–30 m of the Anorthosite Unit is leucogabbro with chrome-green coloured hornblende.

Upper Gabbro Unit (unit 7)

This unit is best preserved in the central and eastern part of the complex (plate 2, localities 32, 33, 42, 45; fig. 35) and occurs up to 50 m thick. It consists of gabbro with layers of peridotite, dunite, melanogabbro, leucogabbro and anorthositte. Magnetite and ilmenite are prominent and locally form massive layers up to 1 m thick, or occur with equant relict igneous plagioclase in magnetite-rich leucogabbro (fig. 34). Where the rocks are strongly deformed and recrystallised, they consist of massive equigranular amphibolite or layers of melano- and leuco-amphibolite in which garnet is generally prominent. Some of the ultramafic and magnetite-rich leucogabbro layers show mineral-grading, which provides the best evidence of the primary way-up within the unit. The top of the unit is in contact with amphibolite with deformed relict pillow lava structure on the nunatak 12 km south-east of locality 29 (plate 2).

Fig. 32. *a*. Undeformed, little recrystallised igneous plagioclase with igneous chromite and metamorphic biotite and hornblende in a 2 m thick chromitite layer in anorthosite of the Anorthosite Unit, west of Qagsse summit. a

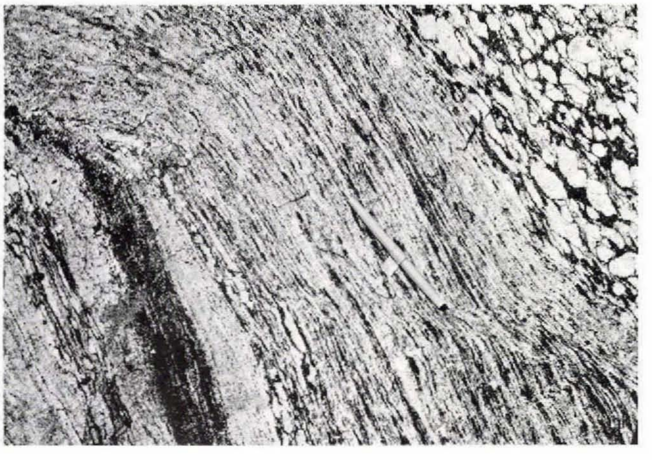


Fig. 32 *b*. Shear zone with pronounced schistosity cutting partly deformed relict igneous plagioclase with chromite and hornblende in a 1 m thick chromitite layer in anorthosite, locality 15 on Qeqertarssuatsiaq (north). b

Fig. 32. *c*. Deformed and recrystallised, relict 5–10 cm diameter igneous plagioclase and chromite-hornblende layers in a 1 m thick chromite-bearing layer with folded igneous layering and schistosity, in anorthosite of the Anorthosite Unit at Majorqap qâva. c

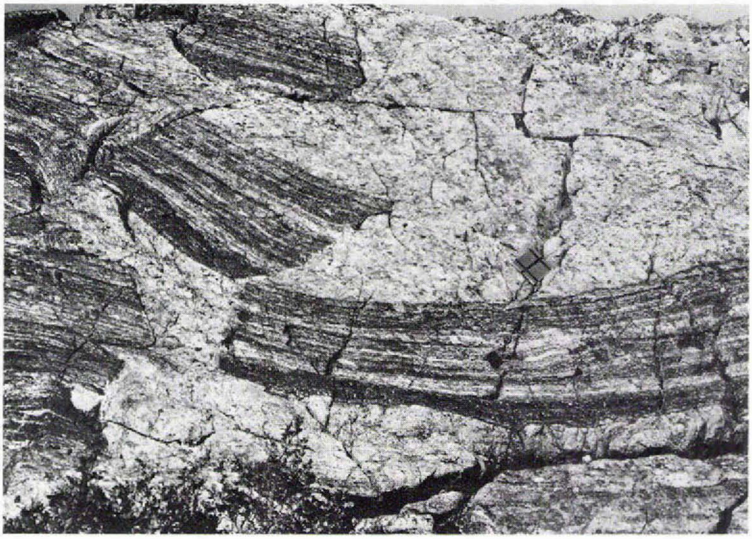
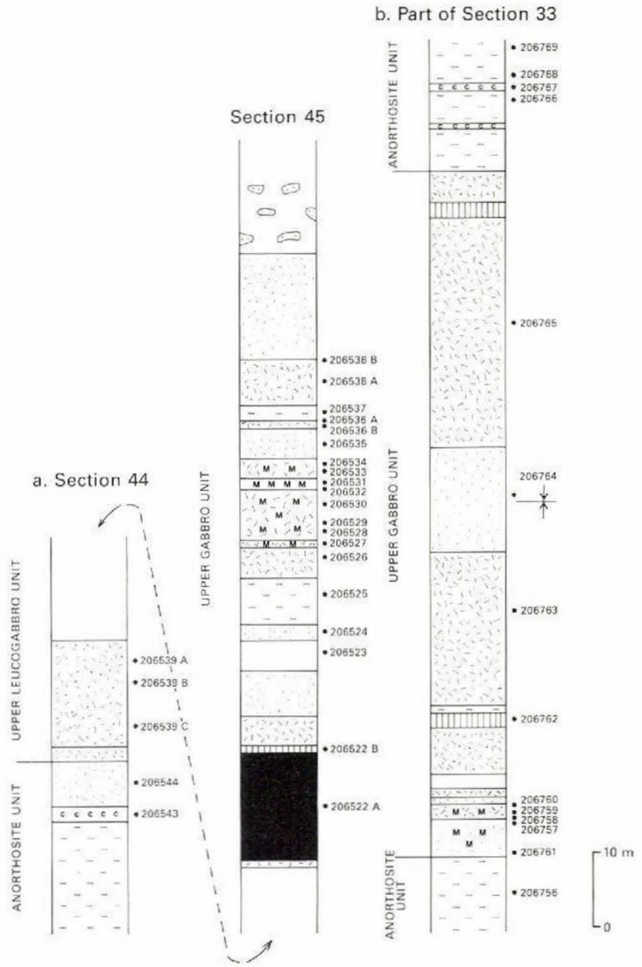


Fig. 33. Fragments of a chromitite layer disrupted by slumping in anorthosite of the Anorthosite Unit, east of locality 16 on Qeqertarssuatsiaq (north). The map case is 28 cm square.

Fig. 34. Layers of magnetite-ilmenite-hornblende and plagioclase-hornblende rock in the Upper Gabbro Unit at locality 32. Original way up towards the right.

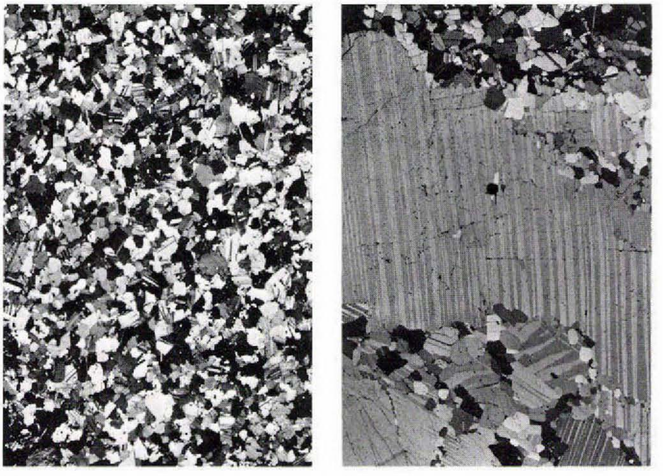
Fig. 35. *a*. Stratigraphic sections 44 and 45 of the Upper Gabbro Unit. *b*. The western part of stratigraphic section 33 of the Upper Gabbro Unit in a fold core of a downward-facing F1 syncline.



Petrography

The mineral assemblages and textures of most rocks are relatively simple, and the rocks are much more spectacular at outcrop scale than in thin section. The petrography of the rocks is, however, briefly discussed because of the importance of distinguishing between igneous and metamorphic minerals and textures in the interpretation of the Fiskenæsset Complex.

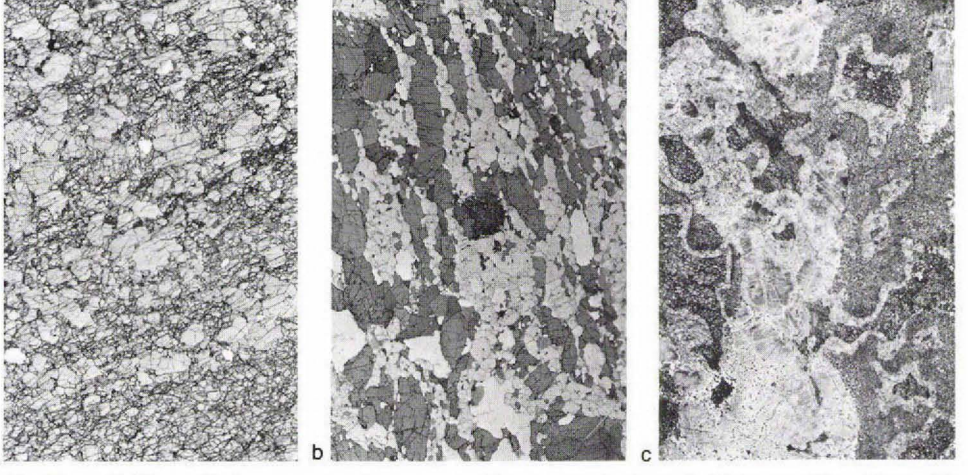
Most anorthosite and leucogabbro consists of plagioclase and hornblende with post-tectonic equigranular metamorphic textures (fig. 36a). Gabbros consist of metamorphic plagioclase + hornblende ± diopside with similar metamorphic texture, and most ultramafic layers contain largely metamorphic olivine-orthopyroxene-clinopyroxene-hornblende-spinel-magnetite assemblages. The general pe-



a b

Fig. 36. *a*. Anorthosite with post-tectonic metamorphic texture. *b*. Partly recrystallised igneous plagioclase from leucogabbro, note the inclusions of hornblende in the igneous plagioclase. Both viewed with crossed nicols. The narrow dimension of each photograph represents 2 cm.

trography of a large number of mainly metamorphic textured samples has been described from Qeqertarssuatsiaq (north) by Herd (1972) and Windley *et al.* (1973), and from the eastern part of the complex by Hutt (1974). These descriptions include most of the main metamorphic textures and mineral assemblages, and so only a few examples of the main rock types are illustrated here by photomicrographs (figs 36, 37) and are not described in detail. The petrography of chromite-bearing rocks has been described in detail by Ghisler (1976) and briefly by Steele *et al.* (1977), and Gorman (1980) has given a detailed account of deformation textures in leucogabbros and the mechanisms of formation of these textures. Modal and whole



a b c

Fig. 37. *a*. Gabbro with igneous or early metamorphic corona structure. *b*. Metamorphic gabbro with deformed relict igneous texture. *c*. Metamorphic hornblende peridotite. Scale as in fig. 36.

Table 2. Modal and whole rock chemical compositions of typical rock types

	Anorthosite 125763	Leucogabbro 159452	Gabbro 159448	Peridotite 151012
Plagioclase	98	85	50	
Hornblende	1	15	45	30
Biotite	1			
Olivine				60
Pyroxene			5	4
Magnetite				4
Spinel				2
SiO_2	48.97	47.30	48.44	39.55
TiO_2	0.12	0.11	0.20	0.14
Al_2O_3	31.41	29.11	15.82	2.68
Fe_2O_3	0.25	0.65	1.55	3.92
FeO	0.75	2.09	5.65	16.89
MnO	0.03	0.05	0.16	0.39
MgO	0.90	2.70	10.20	32.60
CaO	14.63	14.57	13.80	2.67
Na_2O	1.79	1.78	0.95	0.27
K_2O	0.46	0.19	0.24	0.16
P_2O_5	0.10	0.10	0.11	0.09
H_2O	0.44	0.64	1.32	0.56
	99.85	99.28	98.44	99.92
Li	16	44	52	20
Ba	33	19	26	22
Rb	26	4.3	2.1	14
Sr	83	82	30	0.8
Pb	5	1	1	1
Th	3	1	1	1
U	0.53	0.73	0.16	0.02
Zr	6	3	1	6
Nb	0.05	1	1	1
Mo	2	2	2	2
Sn	0.05	10.5	0.5	1
Y	2	4	5	4
La	11	3	2	10
Ce	18	2	6	6
Nd	6	2	2	2
Sc	3	10	51	26
V	28	48	171	56
Cr	232	55	1490	312
Co	13	24	51	174
Ni	25	100	273	1370
Cu	17	11	27	107
Zn	8	12	70	93
Ga	21	19	13	1
Ge	0.9	1.2	1.8	1.7
Pd				7.4 ppb
Pt				7.8 ppb
Rh				0.2 ppb
S	190	10	60	2400
Cl	25	90	180	180

Modal and major element analyses in percentages; major elements by XRF, I. Sørensen.
Trace elements in ppm by XRF, J. C. Bailey; except for U by DNA, R. Gwozdz, and Pd, Pt, Rh in ppb from Page *et al.* (1980). Analyses were selected from over 150 whole rocks analysed.

rock compositions of typical rock types are shown on table 2. In the following description and discussion emphasis is placed on the surviving igneous mineral assemblages and textures.

The most completely preserved igneous or very early metamorphic assemblages occur in gabbros of the Middle Gabbro Unit at Majorqap qâva and elsewhere throughout the eastern part of the complex. These rocks show corona structures formed by the subsolidus reaction of plagioclase and olivine (Myers & Platt, 1977; fig. 37a). The coronas consist of a core of remnant primary olivine enclosed by a zone of orthopyroxene. This in turn lies within a thin, in some places discontinuous, zone of clinopyroxene, and beyond this is a zone of pale amphibole and a spinel-amphibole symplectite. This latter intergrowth is interpreted by Myers & Platt (1977) as marking the site of primary igneous plagioclase.

A complete gradation is seen between plagioclase-bearing gabbros and plagioclase-free ultramafic rocks with corona structures. The latter appear to be derived from gabbros or melanogabbros in which all the igneous plagioclase has reacted with olivine to form pyroxene and amphibole. Therefore all hornblende-rich ultramafic layers in the Fiskenæsset Complex did not necessarily originate as ultramafic cumulates. The gabbros of the Middle Gabbro Unit contained olivine and plagioclase as the main cumulus minerals, and pyroxenes and amphiboles formed during either a post-accumulation magmatic, or early metamorphic phase.

The early microscopic textures of other rocks are not well preserved. Surviving igneous minerals, mostly plagioclase, are recognised as relatively large crystals in finer grained metamorphic mineral assemblages (fig. 36b). In rock sequences which show progressively greater amounts of deformation, it can be seen that increasing deformation was accompanied by the recrystallisation of large igneous crystals to aggregates of similar crystals with smaller grain size. Both the shapes of whole crystal aggregates derived from single igneous crystals, and individual grains within the aggregates, assumed tectonic fabrics. In general recrystallisation outlasted deformation and individual metamorphic grains are equigranular. The intensity of previous deformations is then only recorded by the shapes of the crystal aggregates derived from each igneous crystal.

Igneous plagioclase is pale green in colour because it contains numerous tiny inclusions of green hornblende (and a smaller number of inclusions of diopside and rutile). When large igneous plagioclase recrystallised to aggregates of smaller metamorphic plagioclase, it purged itself of inclusions, and metamorphic plagioclase is clear and white (fig. 38). The inclusions recrystallised forming larger crystals of hornblende and biotite.

Igneous plagioclase in the Lower and Upper Leucogabbro and Anorthosite Units shows little significant variation in composition (in the range An_{89-79}) with stratigraphic height, but igneous plagioclase in the Middle Gabbro Unit is consistently more calcic (An_{98-94}) (Myers & Platt, 1977). Some crystals show normal zoning from cores of An_{89-85} to rims of An_{84-79} and, within single large plagioclases,

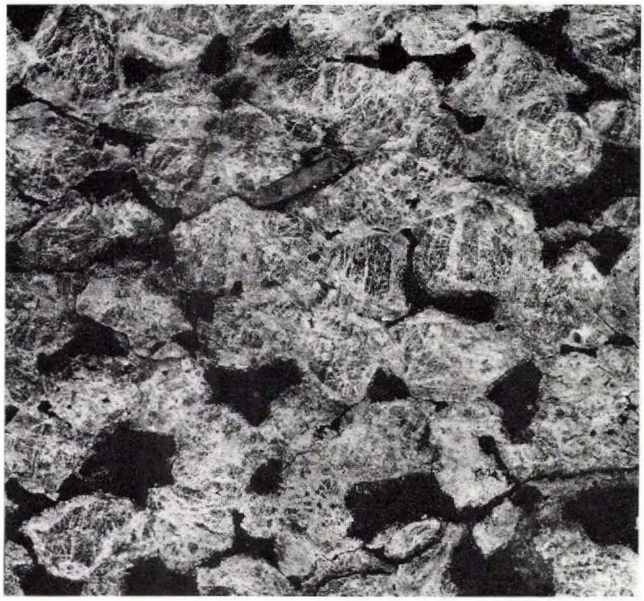


Fig. 38. Cumulus plagioclases in a matrix of metamorphic hornblende in leucogabbro, sub-unit 5.3 of the Upper Leucogabbro Unit at Majorqap qâva, showing grey remnants of igneous plagioclase and white metamorphic plagioclase around the margins of cumulus crystals and along fracture zones within them. The pen-knife is 9 cm long.

these igneous variations are generally preserved during metamorphic recrystallisation (Myers & Platt, 1977). However, in whole rock samples where igneous plagioclase does not survive there appears to be a relation between the composition of coexisting metamorphic plagioclase and hornblende with the modal composition of the rock, plagioclase being more sodic with increase in modal hornblende. The composition of metamorphic plagioclase in whole rocks also changes during dehydration with assemblages of hornblende and bytownite being replaced by clinopyroxene-anorthite-magnetite-ferropargasite (Gorman, 1980).

Green hornblende is the second main component of the Fiskenæsset Complex, but textures indicate that most is metamorphic. Igneous hornblende has only been recognised in some ultramafic rocks by Windley *et al.* (1973) and Hutt (1974), and in late magmatic gabbro pegmatite pipes (Myers, 1978b). Thus there is no direct evidence whether most hornblende was derived from igneous hornblende or pyroxene or both.

Windley *et al.* (1973) followed by Windley & Smith (1974) interpreted the abundance of metamorphic hornblende as indicative that hornblende was originally a major igneous phase, and that the magma had a high water content. Hutt (1974), however, recognised that part of the Fiskenæsset Complex further east originally crystallised under anhydrous conditions. At Majorqap qâva hornblende in some undeformed leucogabbros encloses older cores of pyroxene, and rims of hornblende occur around peridotite layers (fig. 19), suggesting that locally metamorphic hydration occurred. I consider it likely that most hornblende in the Fiskenæsset Complex is a metamorphic derivative of igneous pyroxene, but cannot disprove the contrary interpretation of Windley *et al.* (1973) and Windley & Smith (1974) that most hornblende is derived form igneous hornblende.

During amphibolite and granulite facies metamorphism recrystallisation is clearly related to deformation, with igneous minerals surviving in zones of lowest deformation. Some parts of the Fiskenæsset Complex, however, suffered post-tectonic hydrothermal alteration which affected all the rocks equally. This metamorphism in low greenschist facies, although locally associated with faulting and brecciation, is generally widespread and patchy and does not appear to be structurally controlled. It post-dates late Archaean granulite facies metamorphism and predates the intrusion of early Proterozoic dolerite dykes. All rocks affected by this alteration are reduced to assemblages of mainly quartz-chlorite-epidote-prehnite-muscovite.

MAGMATIC STRUCTURES

The Fiskenæsset Complex contains a great variety of magmatic structures and, although many of the rocks are completely recrystallised, many structures and relict igneous minerals survive in an undeformed or little deformed state. Some of these structures are well known from other layered intrusions, but some are not. The most prominent structures are summarised below.

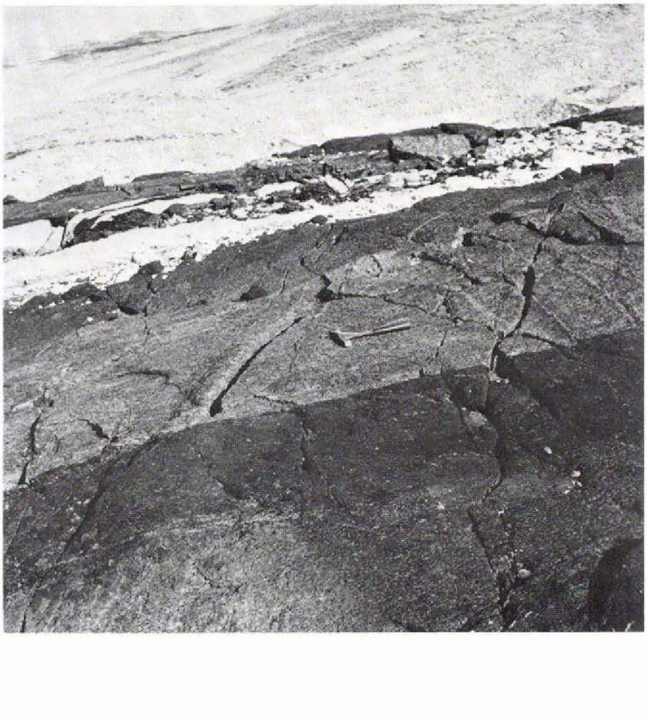


Fig. 39. Mineral-graded trough layers in sub-unit 5.1 of the Upper Leucogabbro Unit at locality 31. They range upwards across the photograph from peridotite (grey), grading up into hornblende rock (black) with a sharp top, to peridotite which grades up through melanogabbro and gabbro to anorthosite (white). Above this there is a similar graded layer followed in the distance by leucogabbro and anorthosite.

Mineral-graded layers

These are the most abundant kind of igneous layering and occur in most rock types. In many *ultramafic layers* there is an upward gradation above a sharp basal contact from mainly olivine cumulates to olivine-pyroxene cumulates and pyroxene cumulates, and some grade further into felsic cumulates to melanogabbro, gabbro, leucogabbro and anorthosite within a total thickness of 1–5 m (fig. 39). Some 1–2 m thick peridotite layers grade both upwards and downwards from cores of peridotite to pyroxene-hornblende and hornblende rocks to melanogabbro, gabbro and leucogabbro. Other peridotite layers are massive (fig. 19).

Fig. 40. Mineral-graded layers in uniform gabbro, sub-unit 4.2 of the Middle Gabbro Unit, locality 18 at Majorqap qâva. Original way up towards top of photograph.

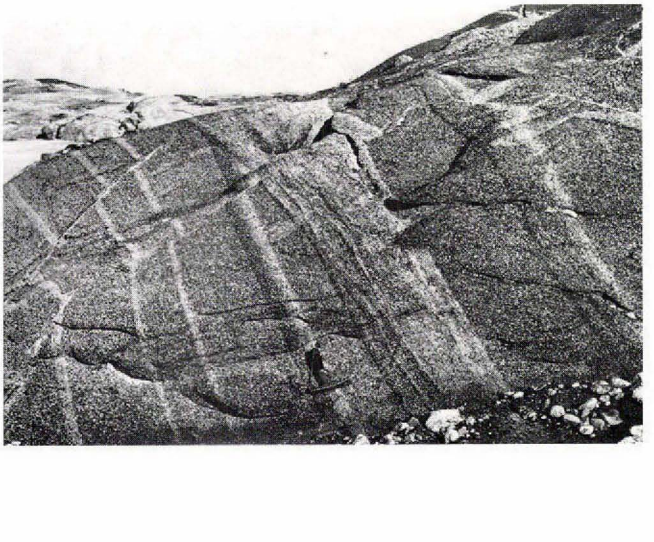


Fig. 41. Cross-bedded mineral-graded layers in gabbro, sub-unit 4.2 of the Middle Gabbro Unit, locality 26 at Majorqap qâva. Original way up towards the right.

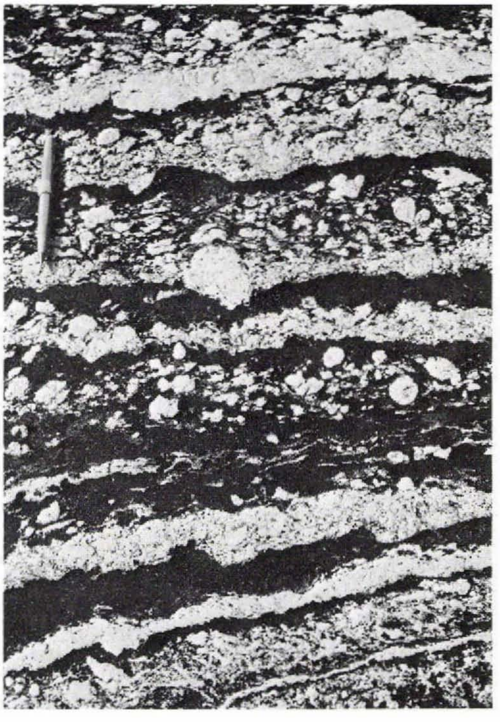


Fig. 42. Mineral-graded chromite-plagioclase-pyroxene (now hornblende) cumulate from a 50 cm thick chromitite layer in anorthosite of the Anorthosite Unit, 10 km north of locality 39. Original way up towards bottom of photograph.

Mineral-graded layering is widespread in *gabbros of the Middle Gabbro Unit.* Individual layers are typically 10–15 cm thick and are spaced at intervals of 50–100 cm in uniform gabbro. They have a sharp basal contact above which hornblende rock grades upwards through gabbro to leucogabbro or anorthosite, bounded by a sharp top contact with uniform gabbro (figs 16, 40). Some mineral-graded layers occur in continuous succession without intervening uniform gabbro, and some show cross-bedding (fig. 41). In general, mineral-graded layers in the lower sub-unit 4.1 are poorly defined (with less distinct contacts) and are of great lateral extent, whereas in sub-unit 4.2 mineral-graded layers are more clearly defined, tend to be of short lateral extent, and occur in vertical columns. Similar mineral-graded layers are well known from the Skaergaard Intrusion from which they have been described and discussed at length (Wager & Deer, 1939; Wager & Brown, 1968; McBirney & Noyes, 1979).

Mineral-graded layering also occurs in plagioclase-chromite-pyroxene (now hornblende) cumulates in the Upper Leucogabbro and Anorthosite Units (fig. 42), and in plagioclase-magnetite-pyroxene (now hornblende) cumulates in the Upper Gabbro Unit. It is also prominent in many trough layers of either peridotite, gabbro or chromitite which are described below.

Size-graded layers

These are less abundant than mineral-graded layers but occur in a variety of rocks. Size-graded olivines occur in dunite (fig. 8) and peridotite layers, especially in the Ultramafic Unit where they range in size upwards from 10 cm to 2 mm in diameter. Size-graded plagioclases occur in leucogabbro in the Lower and Upper Leucogabbro Units, and in plagioclase-chromite cumulates in the Anorthosite Unit. In different places, plagioclases either decrease (fig. 14) or increase (fig. 43) in size upwards. On a larger scale, the whole of the 35 m thick sub-unit 5.3 of the Upper Leucogabbro Unit shows size grading in which equant igneous or relict igneous plagioclase increases upwards from 2–5 cm to 10 cm in diameter.

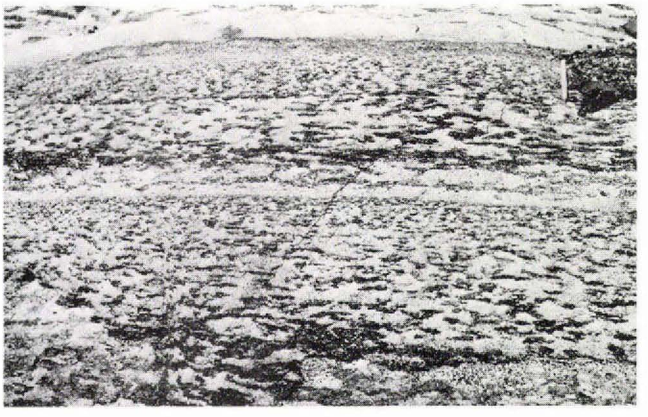


Fig. 43. Size-graded relict igneous plagioclase with size of relict igneous plagioclase increasing upwards (towards the bottom of the photograph) in sub-unit 5.1 of the Upper Leucogabbro Unit at Majorqap qâva.

Trough layers

Trough layers are widespread throughout the intrusion and occur in a variety of rocks. They are described in more detail by Myers (1976c) and are only summarised here. Most of these layers are interpreted as deposits in channels which were excavated and filled in by turbidity currents. These currents were important concentrators of olivine, pyroxene and chromite, and led to the intermittent deposition of ultramafic layers in leucogabbro and gabbro throughout the Lower Leucogabbro, Middle Gabbro and lower part of the Upper Leucogabbro Units.

Most *peridotite trough layers* in sub-unit 5.1 are 1–3 m thick, 100–500 m wide and of great but unknown length, but some are only a few centimetres thick and a few metres wide. The edges of some of these layers interfinger with plagioclase cumulates (figs 19, 44b left), and in some places the plagioclase layers are only one or two crystals thick (individual plagioclases being about 2 cm in diameter). These

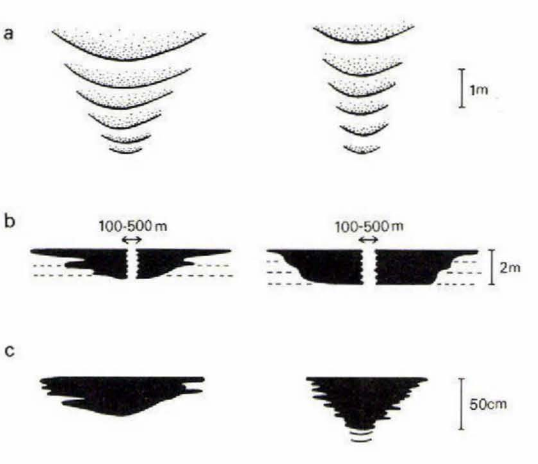


Fig. 44. Schematic profile sections of trough layers interpreted as channel deposits. *a*. Columns of mineral-graded gabbro layers in gabbro of the Middle Gabbro Unit, sub-unit 4.2. *b*. Peridotite layers in leucogabbro of the Upper Leucogabbro Unit, sub-unit 5.1. *c*. Chromitite layers in anorthosite of the Anorthosite Unit.

contacts appear to indicate local spreading of the channels from their initial sites. The edges of other ultramafic trough layers are sharp and steep (fig. 44b right). They cut across mineral-graded layering in leucogabbro, as if depressions were excavated by strong currents that immediately preceded the deposition of the ultramafic layers. They are most abundant in the lowest sub-unit 5.1 of the Upper Leucogabbro Unit, and this sub-unit marks an episode of unusual physico-chemical instability.

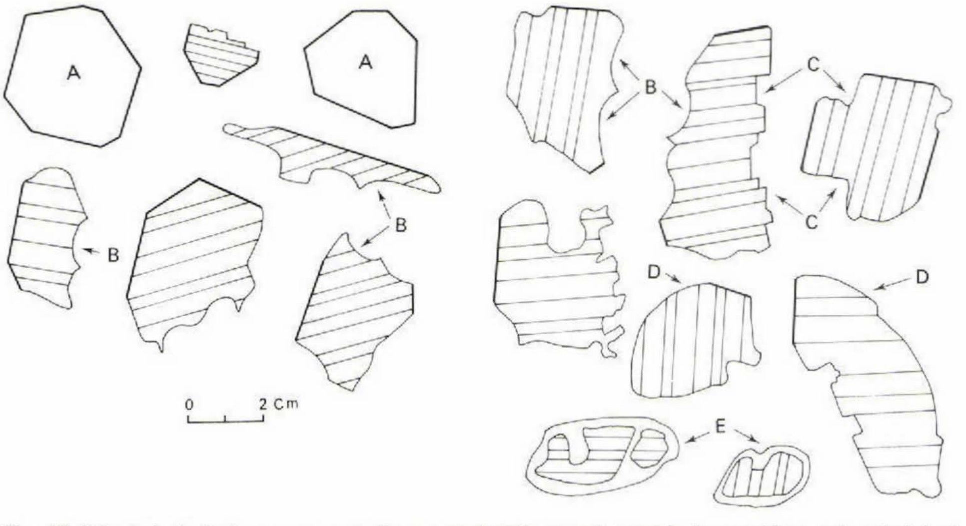


Fig. 45. Selected plagioclase xenocrysts from a peridotite trough layer in leucogabbro, sub-unit 5.1 of the Upper Leucogabbro Unit, 3 km south-east of locality 29. Thick lines mark igneous crystal faces, thin lines are twin lamellae. Note igneous crystal shapes A, conchoidal fractures B, fractures along twin planes C, rounding caused by mechanical abrasion D, and chemical overgrowth of hornblende E over corroded plagioclase crystals.

The main part of many of these trough layers consists of massive peridotite, olivine-pyroxene-spinel-chromite cumulates (fig. 37c), but some are mineral-graded from olivine-rich (dunite) lower portions to pyroxene-rich (peridotite) upper portions which grade upwards to gabbro and anorthosite (fig. 39). Most of these ultramafic layers are enclosed by a 10–20 cm thick shell of hornblende which is probably the result of metamorphic hydration (fig. 19).

Some of the massive peridotite trough layers contain irregularly distributed xenocrysts of plagioclase, generally 5 cm or less in diameter. Many xenocrysts show evidence of both mechanical abrasion and chemical reaction (fig. 45), similar to rock and crystal fragments in ignimbrites and fluidized plutonic intrusions (fig. 11 of Myers, 1975c). Many of the xenocrysts are broken along cleavage planes and conchoidal fractures. Both angular and rounded fragments occur together, and some angular fragments of plagioclase have secondary overgrowths of pyroxene and hornblende which fill irregularities in their surfaces. These features suggest that the xenocrysts were in vigorous motion prior to their deposition, and were deposited from fluidized systems with olivine and pyroxene crystals which intermittently descended as a coherent mass through the more slowly settling plagioclase cumulates.

Larger and more *complex sequences of trough layers* also occur, such as in a sub-unit of leucogabbro within the Middle Gabbro Unit at Majorqap qâva (fig. 46). This complex trough is both underlain and overlain by more extensive trough layers of peridotite, but cuts across layering within the intervening leucogabbro.

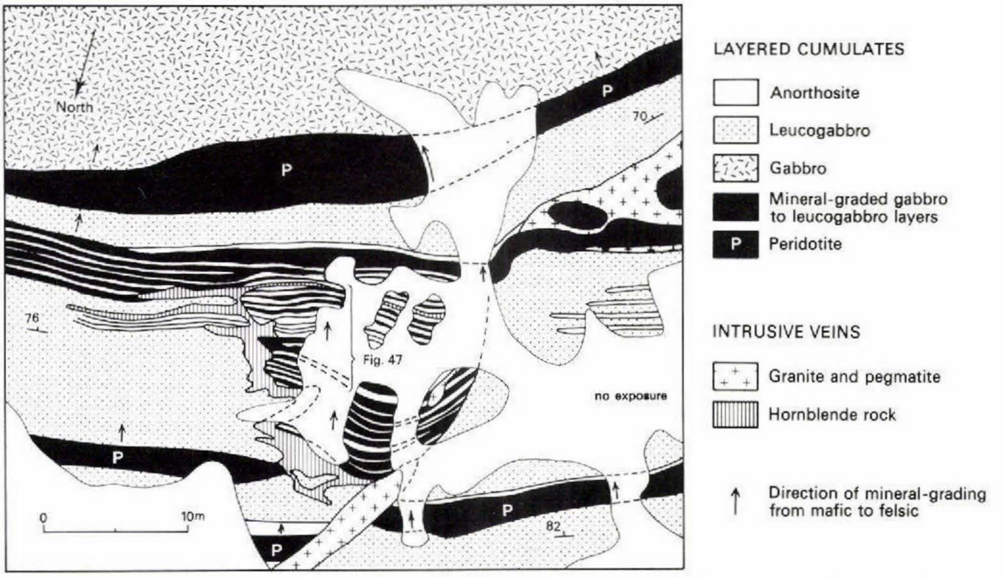


Fig. 46. Map showing a near profile section of an unusually large trough with mineral-graded layers in a local sub-unit of leucogabbro within the Middle Gabbro Unit at Majorqap qâva, locality 21.

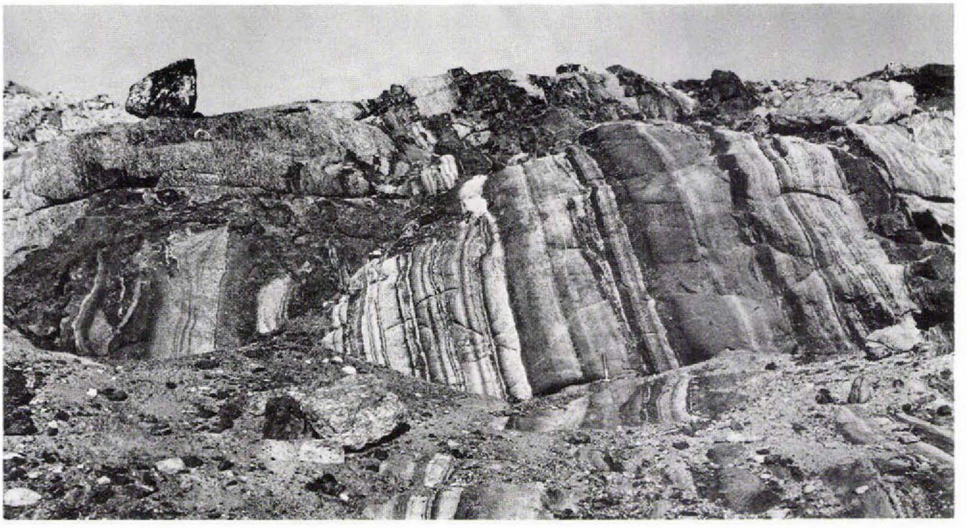


Fig. 47. Mineral-graded layers of plagioclase and hornblende in the large trough shown in fig. 46. Note the slumped layer above the hammer shaft in the centre of the photograph.

Many individual layers within the trough are mineral-graded and they range in compostion from hornblende rock upwards through melanogabbro and gabbro to leucogabbro and anorthosite, and the mineral-grading varies in detail from one layer to another (fig. 47). The uppermost layers of the trough extend eastward (fig. 46) beyond the confines of the initial trough. Some layers show slump structures which face inwards towards the centre of the trough (fig. 47).

Many *plagioclase-hornblende chromitite layers* in the Anorthosite Unit in the central and eastern part of the complex also appear to be trough layers, but they

Fig. 48. Hornblende-chromitite trough layer interfingering laterally with plagioclase cumulates in anorthosite of the Anorthosite Unit, 8 km west of locality 41. Original way up towards the bottom right.

are generally poorly defined. Most form lenticular deposits from a few centimetres to 2 m thick, and from tens of centimetres to hundreds of metres wide. The smaller layers are only a few metres long, whereas the thicker layers are of great but unknown length. Many of these layers occur in narrow vertical columns, and individual layers interfinger laterally with plagioclase cumulates (figs 44c, 48). Some trough layers cut across layering in the adjacent anorthosite, and some show internal cross-layering. Some layers show mineral- and size-grading of plagioclase and chromite (fig. 42) but most are uniform with cumulus plagioclase 2 mm to 2 cm in diameter and equant chromite 0.1–0.5 mm in diameter with interstitial hornblende and biotite.

Slump structures

Slump structures occur in a variety of stratigraphic positions in a variety of rocks, but seem to be most abundant in chromitite horizons. A chromitite layer above a peridotite trough layer at Majorqap qâva is fragmented whereas the plagioclase

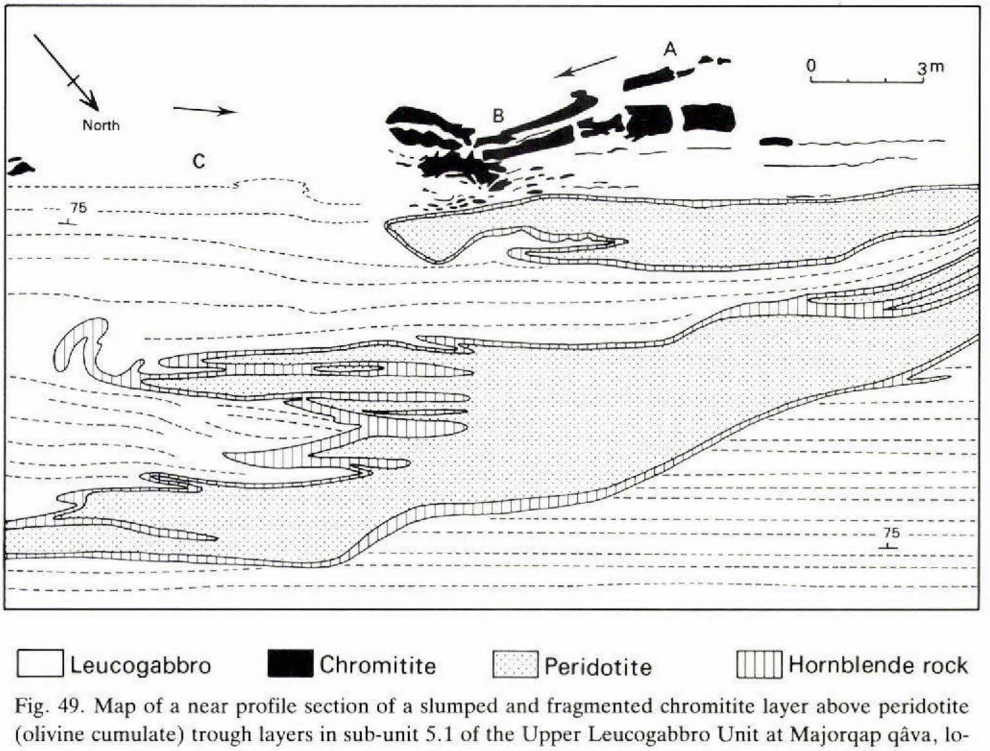


Fig. 49. Map of a near profile section of a slumped and fragmented chromitite layer above peridotite (olivine cumulate) trough layers in sub-unit 5.1 of the Upper Leucogabbro Unit at Majorqap qâva, locality 26. Original way up towards the top of the map. Dashed lines indicate compositional layering in leucogabbro. Textures indicate that, except for the slumped chromitite layer, the rocks are undeformed and the interfingering of olivine and plagioclase cumulates is a primary feature and not the result of folding. The hornblende rock is a metamorphic hydration product of the peridotite (see photograph fig. 19).

Fig. 50. Hornblende chromitite layer disrupted by slumping in leucogabbro sub-unit 5.1 of the Upper Leucogabbro Unit at Majorqap qâva, locality 26, and located near A on fig. 49. Original way up towards the right.

cumulates above and below are not (figs 49, 50). The chromitite layer therefore appears to indicate local slumping of this particular horizon before the overlying plagioclase cumulates were deposited. The chromitite layer was torn apart at A (figs 49, 50) and segments slid to the left and piled up at B (fig. 49). Chromitite is absent at C, but chromitite fragments occur again to the left of C. Fragments to the left of B therefore appear to have slid to the right from C.

Slumped chromitite layers are most abundant in the western part of the complex on Qeqertarssuatsiaq (north) and near Tasiussâ where the thickest chromitite layers occur (fig. 33).

'Snow-flake' structures

'Snow-flake' structures are sub-spherical clusters, 20 cm in diameter, of equant plagioclase crystals 0.5–5 cm in diameter, and radiating tabular wedge-shaped plagioclase crystals up to 10 cm long (figs 51, 52). They occur in the upper sub-unit 4.3 of the Middle Gabbro Unit, either in isolation (fig. 18) or as layers ranging from one cluster to 5 m thick in uniform layered gabbro (fig. 17), and in leucogabbro of the lowest sub-unit 5.1 of the Upper Leucogabbro Unit with peridotite trough layers (fig. 20). In places where they are associated with mineral-graded layers, it can be seen that they accumulated by sinking onto the floor of the intrusion.

Fig. 51. 'Snow-flake' structures, clusters of both large equant plagioclase crystals in radial chains and large radial wedge-shaped plagioclases, in gabbro sub-unit 4.3 of the Middle Gabbro Unit at Majorqap qâva, locality 23. Original way up towards top of photograph.

The wedge-shaped crystals radiate from the centre of the clusters and thicken outwards (fig. 52), showing that they grew outwards from central nucleii. Similar but smaller 'snow-flake' like clusters of plagioclase have been described from troctolite in the Hettasch intrusion of Labrador and interpreted as the products of rapid growth during supercooling (Berg, 1980). As in the Fiskenæsset Complex, these structures occur in a transition zone between medium grained and very coarse grained rocks.

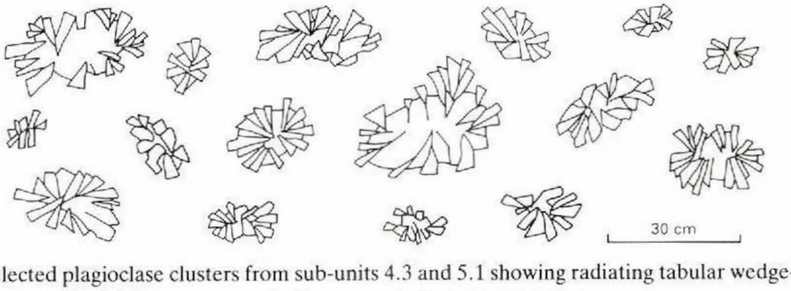


Fig. 52. Selected plagioclase clusters from sub-units 4.3 and 5.1 showing radiating tabular wedge-shaped plagioclase crystals which grew outwards from a central cluster of nucleii.

Pipes

Pipe-like bodies of mafic pegmatite cut across the igneous layering in the Lower Leucogabbro, Middle Gabbro and Upper Leucogabbro Units. They are especially abundant and well preserved at Majorqap qâva (plate 1) (Myers, 1978b). Some pipes form cylindrical bodies between 10 cm and 100 m in diameter (fig. 54), but in detail most pipes have irregular outlines and are interconnected by an irregular network of veins (fig. 53). The margins of the pipes are generally sharp and cut through igneous layering without dilation (fig. 55). Some pipes contain angular and rounded fragments of adjacent host rocks (fig. 56) or rocks transported from different stratigraphic levels, such as leucogabbro fragments in pipes cutting gabbro (figs 53, 55), but fragments are not generally abundant. None of the fragments are distinct enough rock types to indicate their direction of transport.

The pipes comprise both vertical and horizontal portions. All vertical parts of pipes contain an irregular infill of hornblende-rich pegmatite which is generally structureless (fig. 55). In contrast, horizontal parts contain mixtures of structure-

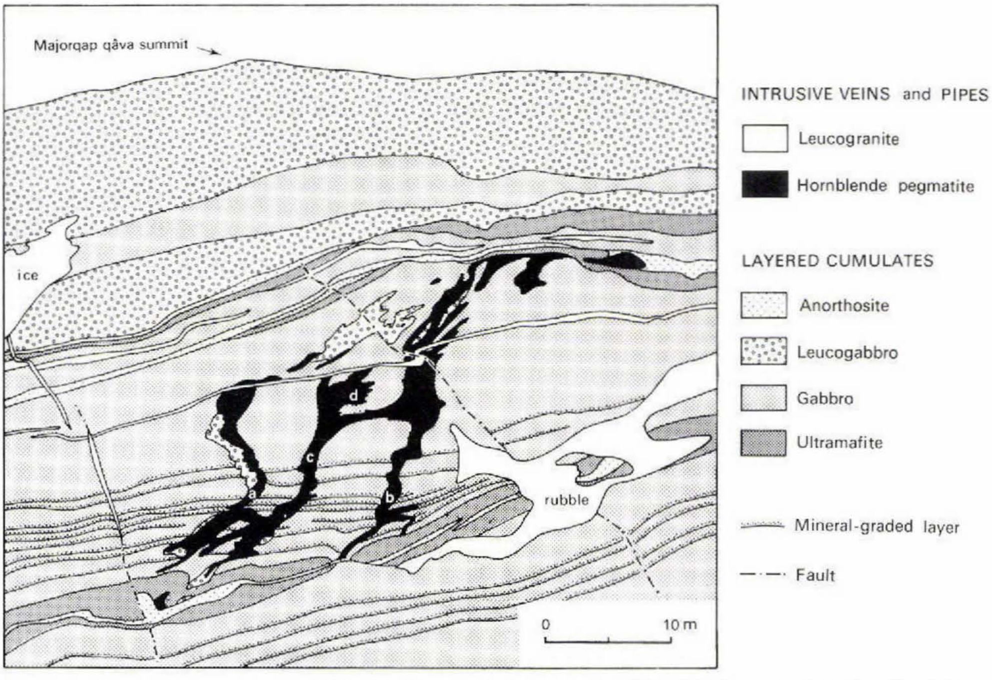


Fig. 53. Map of the upper part of the north–west facing summit cliff of Majorqap qâva (locality 18) showing hornblende pegmatite pipes in longitudinal section in the upper part of the Middle Gabbro Unit. *a*. Location of fig. 55. *b*. Location of fig. 16a in which the pipe at b is just beyond the hammer. Both photographs are views to the left (north). Note the fragmented blocks of leucogabbro in some pipes.

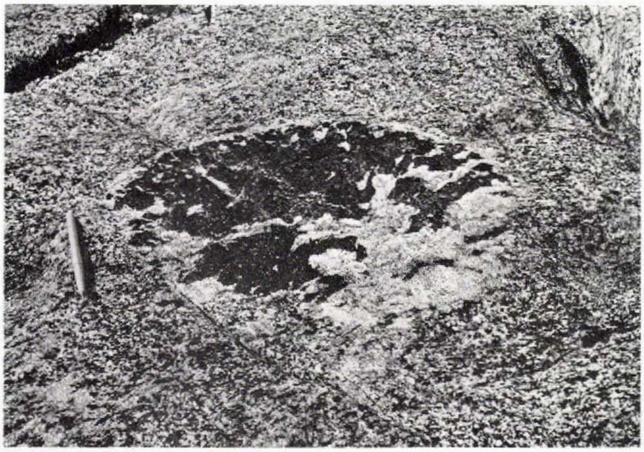


Fig. 54. Profile section of a horizontal hornblende-plagioclase pipe in gabbro of the Middle Gabbro Unit at Majorqap qâva, locality 23.

less hornblende-rich pegmatite and mineral-graded layers (fig. 57). Pegmatite forms the upper portion of most horizontal pipes, and in some cases different episodes of pegmatite growth can be distinguished in the walls (fig. 56).

Mineral-graded layers occur in the lower and central parts of horizontal pipes (fig. 57). Most of this layering grades upwards from gabbro to leucogabbro to anorthosite by increase in the ratio of plagioclase to hornblende. Each mineral-graded layer is overlain by a hornblende comb layer which consists of tabular hornblende crystals generally between 1–5 cm long, but in some places up to 25 cm long,

Fig. 55. Pipe of hornblende pegmatite (P) cutting sharply across mineral-graded layering in gabbro of the Middle Gabbro Unit, sub-unit 4.2, at Majorqap qâva, locality 18, and located at *a* on fig. 53. Note the displaced blocks of leucogabbro (L) within the pipe.

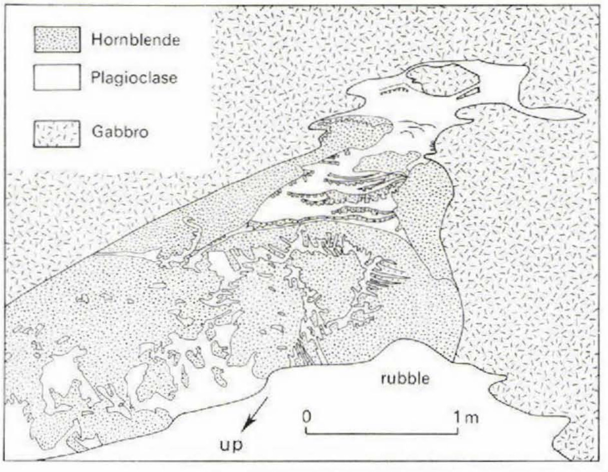


Fig. 56. Oblique cross-section of a horizontal hornblende pegmatite pipe in gabbro of the Middle Gabbro Unit, sub-unit 4.2, at Majorqap qâva, locality 25. The arrow indicates the original way up. Note the fragments of gabbro at the bottom of the pipe (top of the figure) enclosed by plagioclase cumulates, the comb layering, and upward sequence of pipe infilling from comb layering to massive pegmatite.

aligned perpendicular to the layering (figs 57, 58). All these layers are one crystal thick, and the growth pattern shown by inclusions within the hornblendes indicates that the crystals grew upwards from a number of nucleii scattered on the surface of the graded cumulate layer below.

Most of the pegmatite pipe infills are melanocratic and comprise more than 95% hornblende or hornblende-diopside, with less than 5% plagioclase. Most minerals show polygonal mosaic textures and appear to be metamorphic, but some of the

Fig. 57. Oblique view of a cross-section of a horizontal hornblende pegmatite pipe in gabbro of the Middle Gabbro Unit, sub-unit 4.1, at Majorqap qâva, locality 21. The arrow indicates the original way up. Note that the lower part is layered whereas the upper part is massive hornblende pegmatite (P). Note mineral-graded layering (M) and comb layering (C).

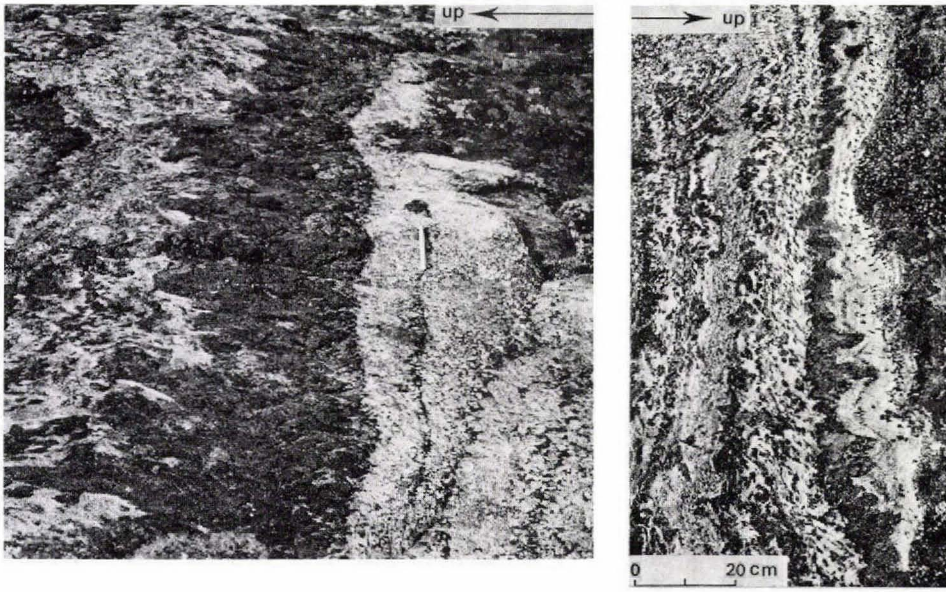


Fig. 58. Comb layering marked by large relict igneous hornblendes in horizontal hornblende pegmatite pipes at Majorqap qâva. The arrow indicates the original way up. *a*. Sub-unit 5.1 of the Upper Leucogabbro Unit. *b*. Sub-unit 4.2 of the Middle Gabbro Unit.

large relict hornblendes in both comb layers (figs 56, 57, 58) and structureless pegmatite retain cores of igneous hornblende rimmed by metamorphic aggregates of fine grained hornblende and diopside.

The shapes of the pipes, their sharp contacts and lack of dilation, suggest that the pipes formed by magmatic drilling rather than metasomatic replacement. Transported xenoliths of host rocks provide further evidence of stoping and movement of stoped fragments within the pipes. The growth of hornblende rather than pyroxene, and the large size of many crystals suggest a high water vapour pressure in the pipes. The alternations of mineral-graded and comb layers suggest that magma with plagioclase phenocrysts was flushed through the pipes in pulses, perhaps associated with erratic propagation of cracks and magmatic drilling.

The absence of pipes from the Anorthosite Unit suggests that this unit was still unconsolidated when the pipes were emplaced (no pipes were seen in the Lower Gabbro Unit, but this unit is poorly preserved and is strongly deformed).

TECTONIC STRUCTURES

Granitoid intrusions and thrusting

The quartzo-feldspathic gneisses, which now make up 80% of the region, were intruded into the Fiskenæsset Complex and metavolcanic amphibolites as sub-concordant sheets associated with thrusting. The evidence of intrusion is widespread in areas of low deformation where veins of granitoid rocks are locally discordant and net-vein layers of the Fiskenæsset Complex and amphibolites. On a larger scale the granitoid intrusions split the Fiskenæsset Complex and metavolcanic amphibolites into thin layers and trains of inclusions (fig. 59), some of which can be followed over well exposed ground for 30 km or more even when they are no more than 1 m wide. Most thin layers and inclusion trains are parts of individual stratigraphic units of the Fiskenæsset Complex rather than mixtures of different parts of the anorthosite stratigraphy. The granitoid intrusions were thus emplaced as sheets, sub-concordant with the igneous stratification of the Fiskenæsset Complex (fig. 59), rather than as large discordant plutons.

The widespread evidence of brittle fragmentation of the amphibolites and Fiskenæsset Complex during the emplacement of the gneiss (fig. 60), the lack of major chemical reaction between fragments of amphibolite and anorthosite and the enclosing gneiss (fig. 61), and the presence of gneiss sheets separating previously adjacent layers of anorthosite, leave little doubt that most of the gneiss was intrusive and was not derived by metasomatic alteration or remobilisation of older sedimentary, volcanic or plutonic layers at the level now exposed.

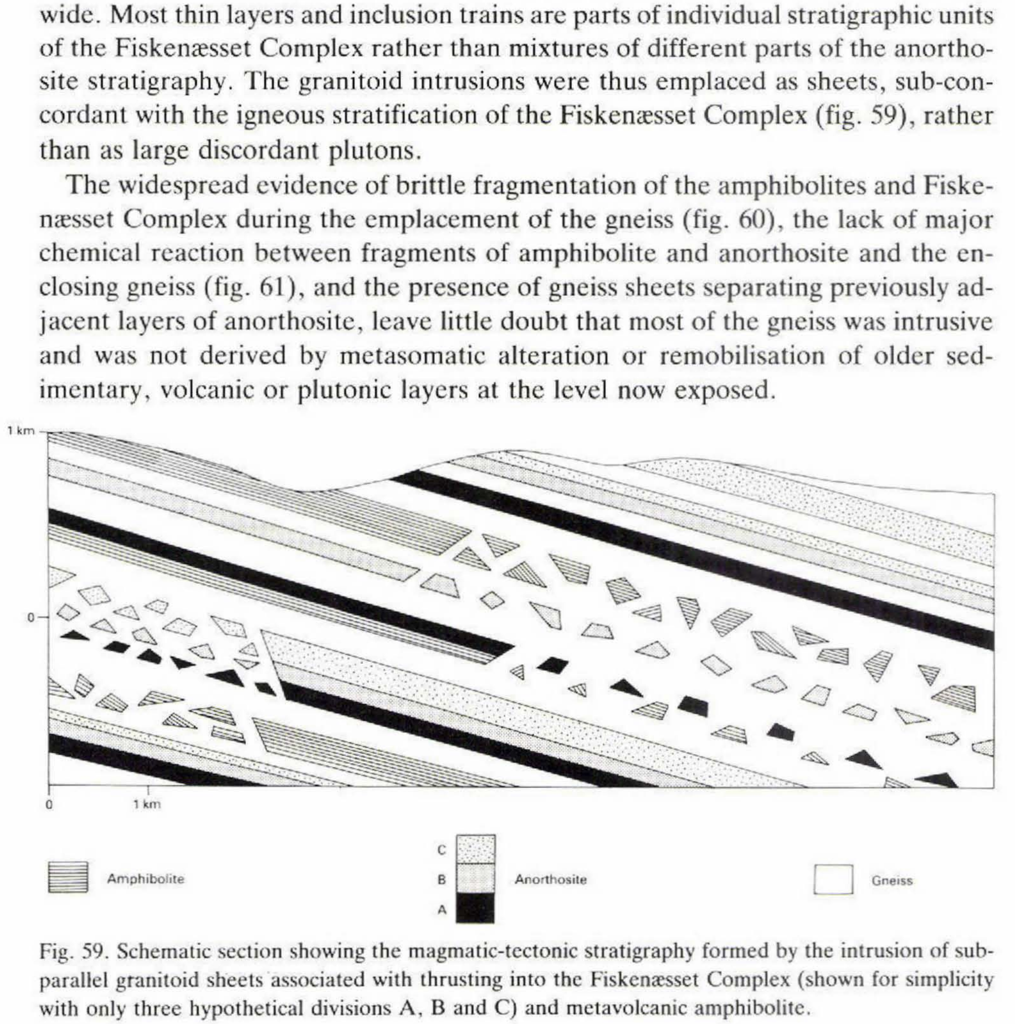


Fig. 59. Schematic section showing the magmatic-tectonic stratigraphy formed by the intrusion of subparallel granitoid sheets associated with thrusting into the Fiskenæsset Complex (shown for simplicity with only three hypothetical divisions A, B and C) and metavolcanic amphibolite.

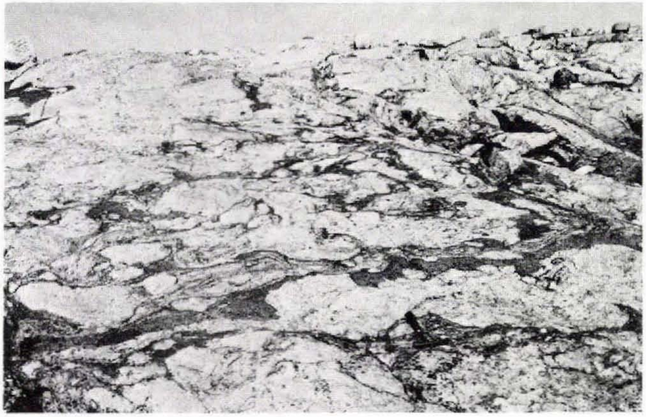


Fig. 60. Anorthosite (white) fragmented during the syntectonic intrusion of quartzo-feldspathic gneiss (grey) at locality 28.

In many places where the granitoid sheets cut at low angles across the Fiskenæsset Complex, it can be seen that the granitoid sheets were emplaced along active thrust zones which cut out or duplicated parts of the stratigraphy (figs 11, 62, 63). When these movements occurred, the granitoid sheets were more ductile than the adjacent Fiskenæsset Complex. Angular fragments of the complex were detached from both sides of the granitoid sheets and were disoriented. The granitoid sheets are strongly deformed and have a planar fabric parallel with their margins (fig. 64).

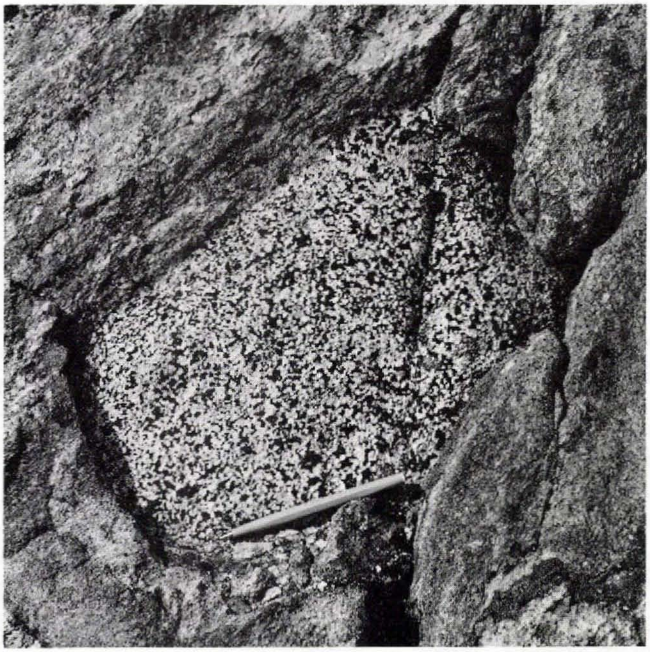


Fig. 61. Undeformed fragment of leucogabbro from the Lower Leucogabbro Unit in quartzo-feldspathic gneiss. Note the absence of any marked chemical reaction between the gneiss and leucogabbro. 5 km west of locality 18.

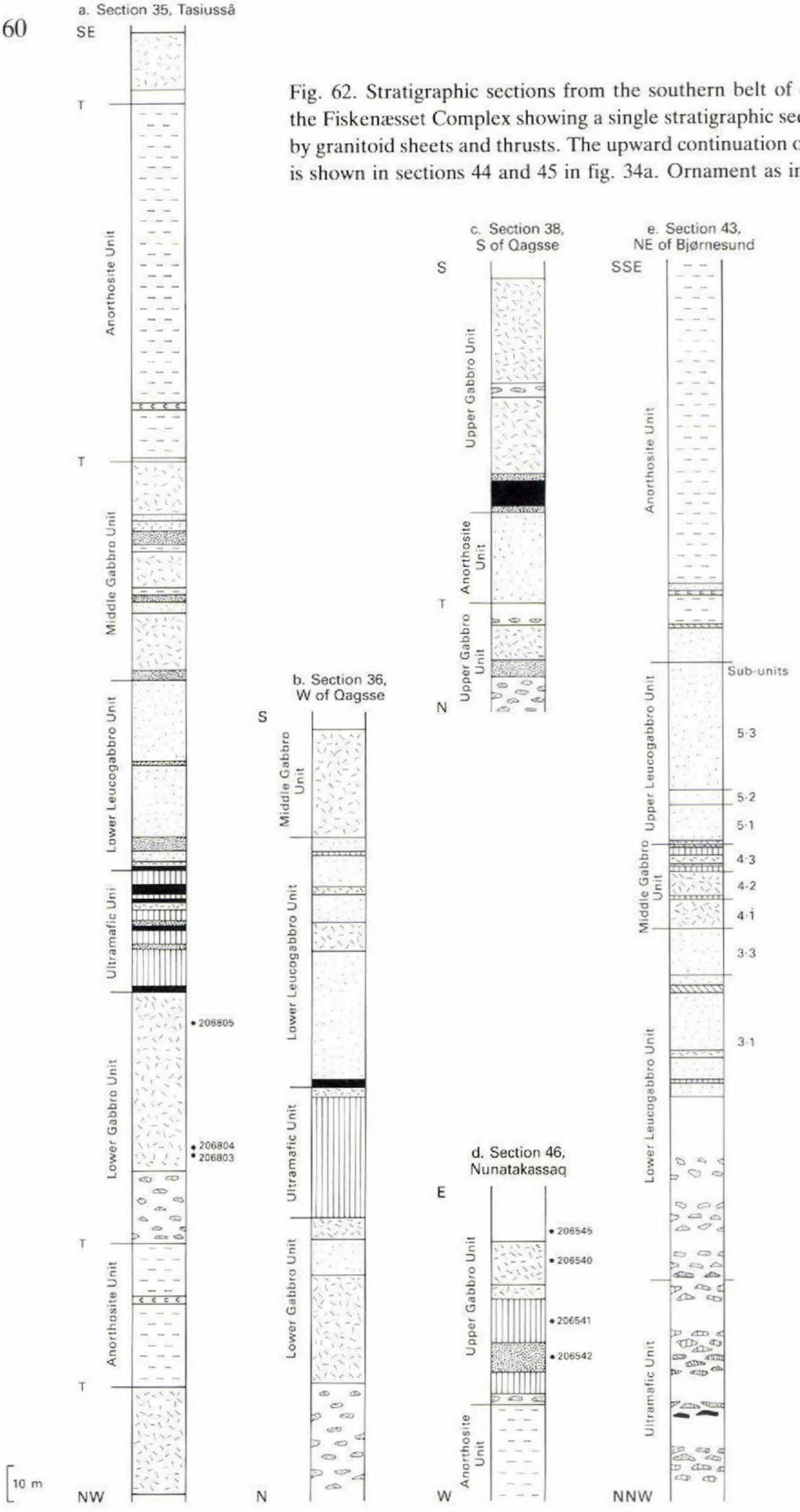
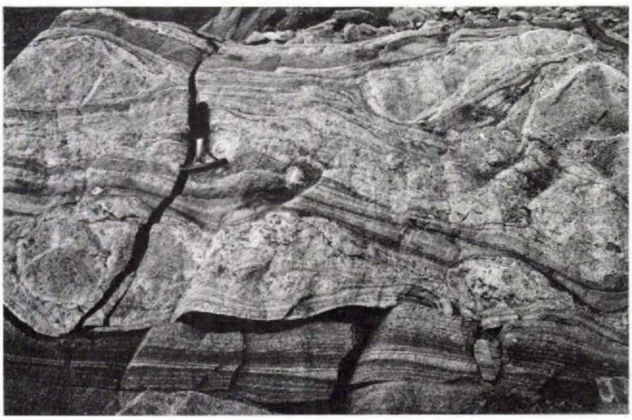


Fig. 62. Stratigraphic sections from the southern belt of outcrops of the Fiskenæsset Complex showing a single stratigraphic sequence split by granitoid sheets and thrusts. The upward continuation of section 43 is shown in sections 44 and 45 in fig. 34a. Ornament as in fig. 2.

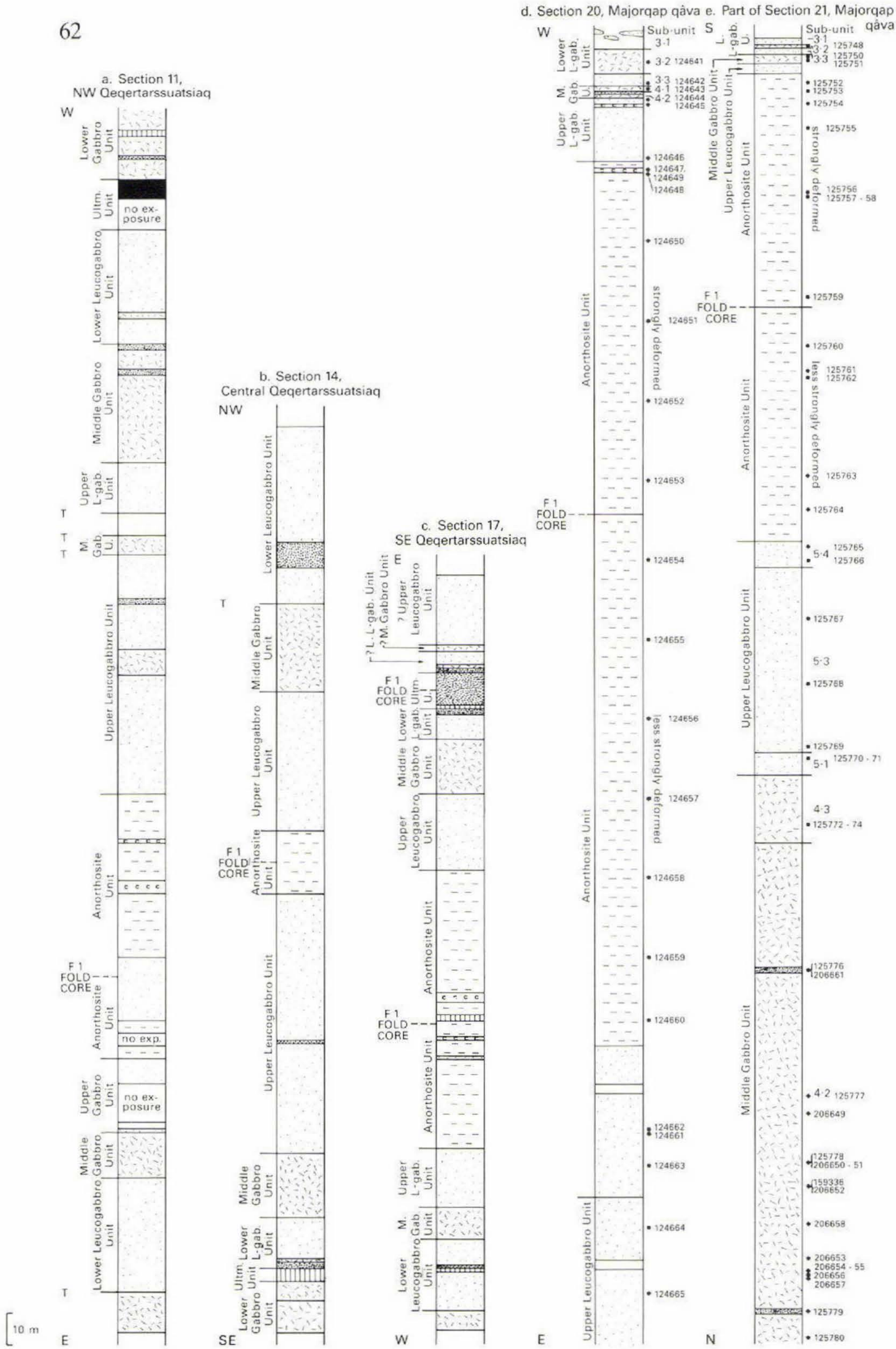
a



b

Fig. 64 *a*. Thin sheet of strongly deformed granitoid gneiss (gn) with fragments of leucogabbro, along a thrust plane between sub-units 5.3 of the Upper Leucogabbro Unit (above) and sub-unit 4.2 of the Middle Gabbro Unit (below) at Majorqap qâva, locality 18. Original way up towards bottom of photograph. The effect of this thrust in repeating part of the stratigraphic sequence can be seen in fig. 11, of which fig. 64a is part.
b. Strongly deformed granitoid gneiss with marked gneissosity and less deformed fragments of anorthosite at locality 46.

Both the Fiskenæsset Complex and metavolcanic amphibolites are seen to have been strongly deformed before they were fragmented by the intrusion of the granitoid sheets, because angular fragments of deformed anorthosite and amphibolite occur as xenoliths in the granitoid sheets. Variations in the intensity of this early deformation can locally be mapped but, because of the extensive fragmentation of the anorthosite and amphibolite and the complexity of later folding, the regional significance of these local variations in early deformation intensity are unknown.


a. Section 11, NW Qeqertarssuatsiaq
W
Lower Gabbro Unit
Ultm. Unit
no exposure
Lower Leucogabbro Unit
Middle Gabbro Unit
Upper L-gab. Unit
M. Gab. U.
Upper Leucogabbro Unit
Anorthosite Unit
F 1 FOLD CORE
Anorthosite Unit
no exp.
Upper Gabbro Unit
no exposure
Middle Gabbro Unit
Lower Leucogabbro Unit
E
10 m
b. Section 14, Central Qeqertarssuatsiaq
NW
Lower Leucogabbro Unit
Middle Gabbro Unit
Upper Leucogabbro Unit
F 1 FOLD CORE
Anorthosite Unit
Upper Leucogabbro Unit
Middle Gabbro Unit
Lower L-gab. Unit
Ultm. Unit
Lower Gabbro Unit
SE
c. Section 17, SE Qeqertarssuatsiaq
E
? Upper Leucogabbro Unit
? M. Gabbro Unit
? L. L-gab. Unit
F 1 FOLD CORE
Ultm. U.
Lower L-gab. Unit
Middle Gabbro Unit
Upper Leucogabbro Unit
Anorthosite Unit
F 1 FOLD CORE
Anorthosite Unit
Upper L-gab. Unit
M. Gab. Unit
Lower Leucogabbro Unit
W
d. Section 20, Majorqap qâva
W
Lower L-gab. Unit
M. Gab. U.
Upper L-gab. Unit
Anorthosite Unit
F 1 FOLD CORE
Anorthosite Unit
Upper Leucogabbro Unit
E
Sub-unit 3·1
3·2 124641
3·3 124642
4·1 124643
4·2 124644
124645
124646
124647
124649
124648
124650
strongly deformed
124651
124652
124653
124654
124655
less strongly deformed
124656
124657
124658
124659
124660
124662
124661
124663
124664
124665
e. Part of Section 21, Majorqap qâva
S
L. L-gab. U.
Middle Gabbro Unit
Upper Leucogabbro Unit
Anorthosite Unit
F 1 FOLD CORE
Anorthosite Unit
Upper Leucogabbro Unit
Middle Gabbro Unit
N
Sub-unit
3·1
3·2 125748
125750
3·3 125751
125752
125753
125754
125755
strongly deformed
125756
125757 - 58
125759
125760
less strongly deformed
125761
125762
125763
125764
5·4 125765
125766
125767
5·3
125768
125769
5·1 125770 - 71
4·3
125772 - 74
125776
206661
4·2 125777
206649
125778
206650 - 51
159336
206652
206658
206653
206654 - 55
206656
206657
125779
125780

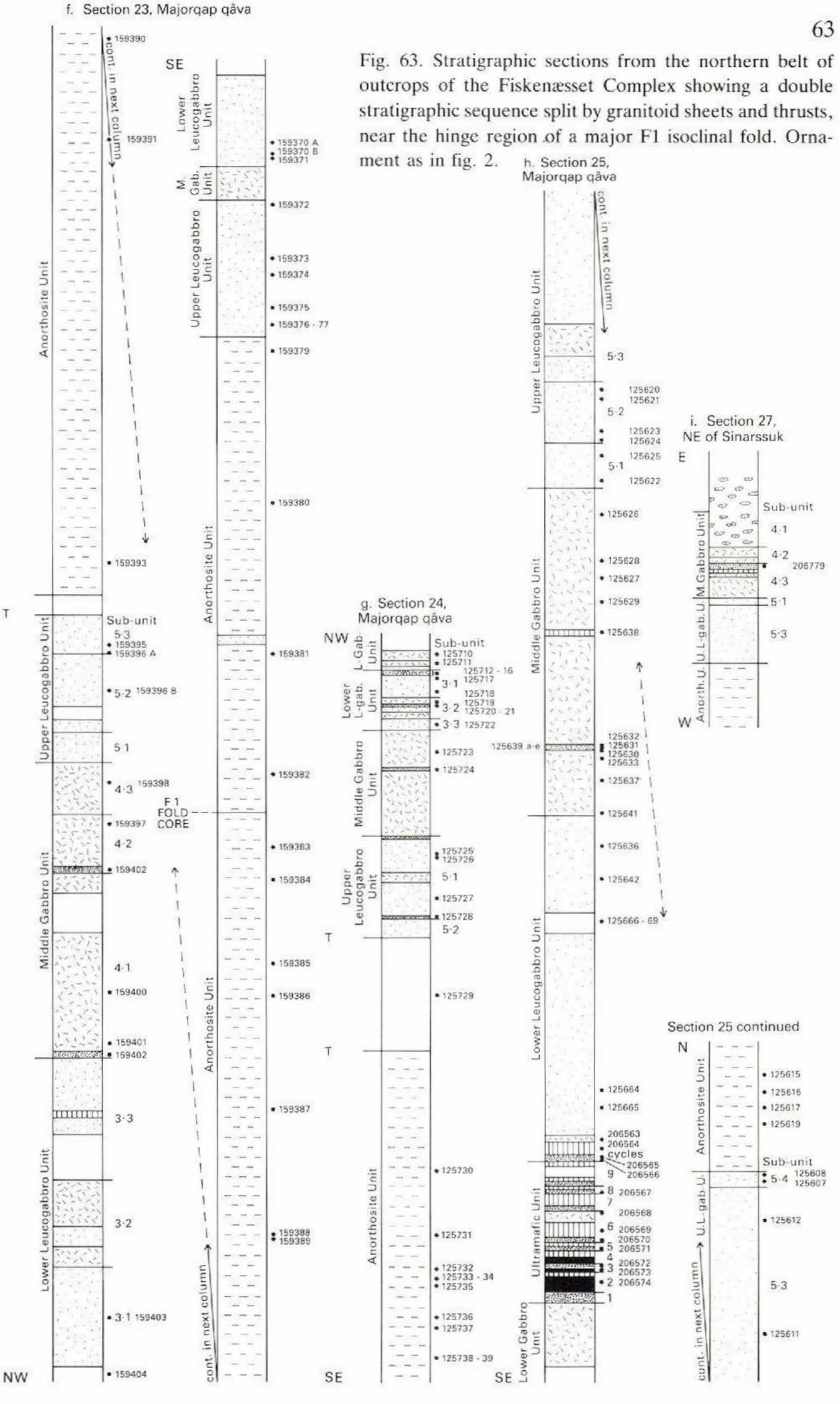


Fig. 63. Stratigraphic sections from the northern belt of outcrops of the Fiskenæsset Complex showing a double stratigraphic sequence split by granitoid sheets and thrusts, near the hinge region of a major F1 isoclinal fold. Ornament as in fig. 2.

Regional fold structures

The igneous layering of the Fiskenæsset Complex and sub-parallel layers of amphibolite and granitoid gneiss were folded during three major episodes of ductile deformation F1, F2 and F3 (Kalsbeek & Myers, 1973), post-dating the emplacement of the granitoid rocks. Each episode of deformation resulted in a complex sequence of fold and fabric development.

The first major folds, F1, are large scale recumbent isoclinal structures and, where they involve the Fiskenæsset Complex, the known stratigraphy and way-up structures enable the facing directions of these folds to be determined (figs 65–67). These structures were refolded twice into folds with steep axial surfaces at high an-

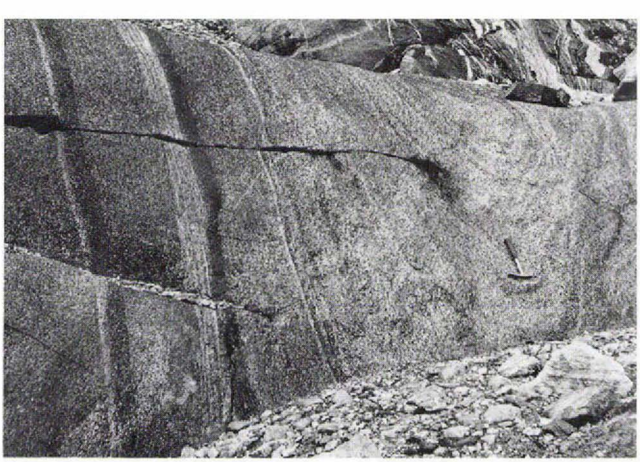


Fig. 65. Minor F1 fold, a downward-facing anticline of mineral-graded layering in gabbro sub-unit 4.2 of the Middle Gabbro Unit at Majorqap qâva, locality 26.

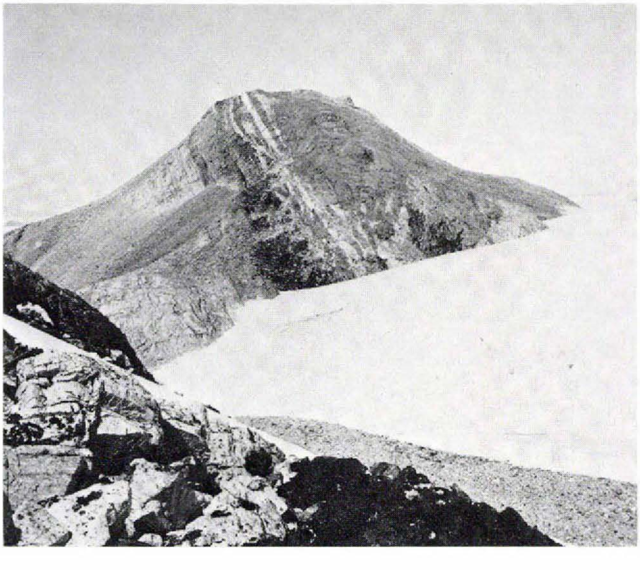


Fig. 66. Major F1 fold, a downward-facing syncline of gabbro (black) – the lower part of the Upper Gabbro Unit, leucogabbro (grey) at the top of the Anorthosite Unit, and anorthosite (white) of the Anorthosite Unit. A near profile section viewed from the SSE from locality 32.

gles to each other. F2 folds are tight or isoclinal, they impart the main (E–W) tectonic grain to the region (plate 2), and are associated with strongly marked planar fabrics.

F3 folds have vertical axial surfaces that trend NW–SE and plunge gently or moderately to the SE. It is these F2 and F3 folds which form the main dome-and-basin interference patterns of the region (plate 2). The F3 folds were contemporaneous with early stages of the last major metamorphic episode which formed amphibolite facies assemblages over most of the region. Granulite facies conditions were reached locally close to Fiskenæsset town and, as this area is approached, ductile deformation in F3 fold limbs is replaced by increasingly more brittle deformation and faulting.

The three main episodes of folding F1, F2 and F3 are well displayed at Majorqap qâva (plate 1, fig. 67) where it can be seen that the oldest structure, F1, is an isoclinal syncline which is refolded by a tight F2 synform and crossed by a number of steep F3 folds. A similar F1 synclinal structure was postulated by Windley *et al.* (1973) for the more deformed part of the Fiskenæsset Complex on Qeqertarssuatsiaq (north), based on the symmetrical repetition of the overall stratigraphy and using geochemical trends as indicators of the original way-up.

The F2 synform at Majorqap qâva extends westwards through the anorthosite outcrops on Qeqertarssuatsiaq (north) (plate 2). It similarly extends eastwards through Sinarssuk until it is folded northwards by a major F3 fold and is obscured by the inland ice cap (plate 2). Throughout this region outcrops of the Fiskenæsset Complex show a repeated stratigraphy and are parts of the hinge region of a major

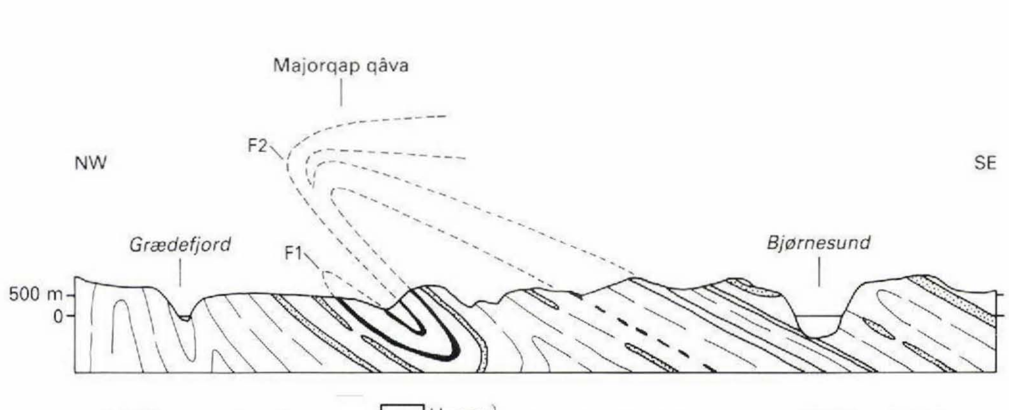


Fig. 68. Simplified cross-section of the Fiskenæsset region through Majorqap qâva, relating the northern (Qeqertarssuatsiaq – Majorqap qâva – Sinarssuk) and southern (Kangârssuk – Tasiussâ – Qagsse – Bjørnesund) belts of outcrops of the Fiskenæsset Complex as parts of a refolded major F1 recumbent syncline. For convenience of presentation, the Fiskenæsset Complex is divided into a lower part consisting of the Lower Gabbro to Middle Gabbro Units, and an upper part consisting of the Upper Leucogabbro to Upper Gabbro Units.

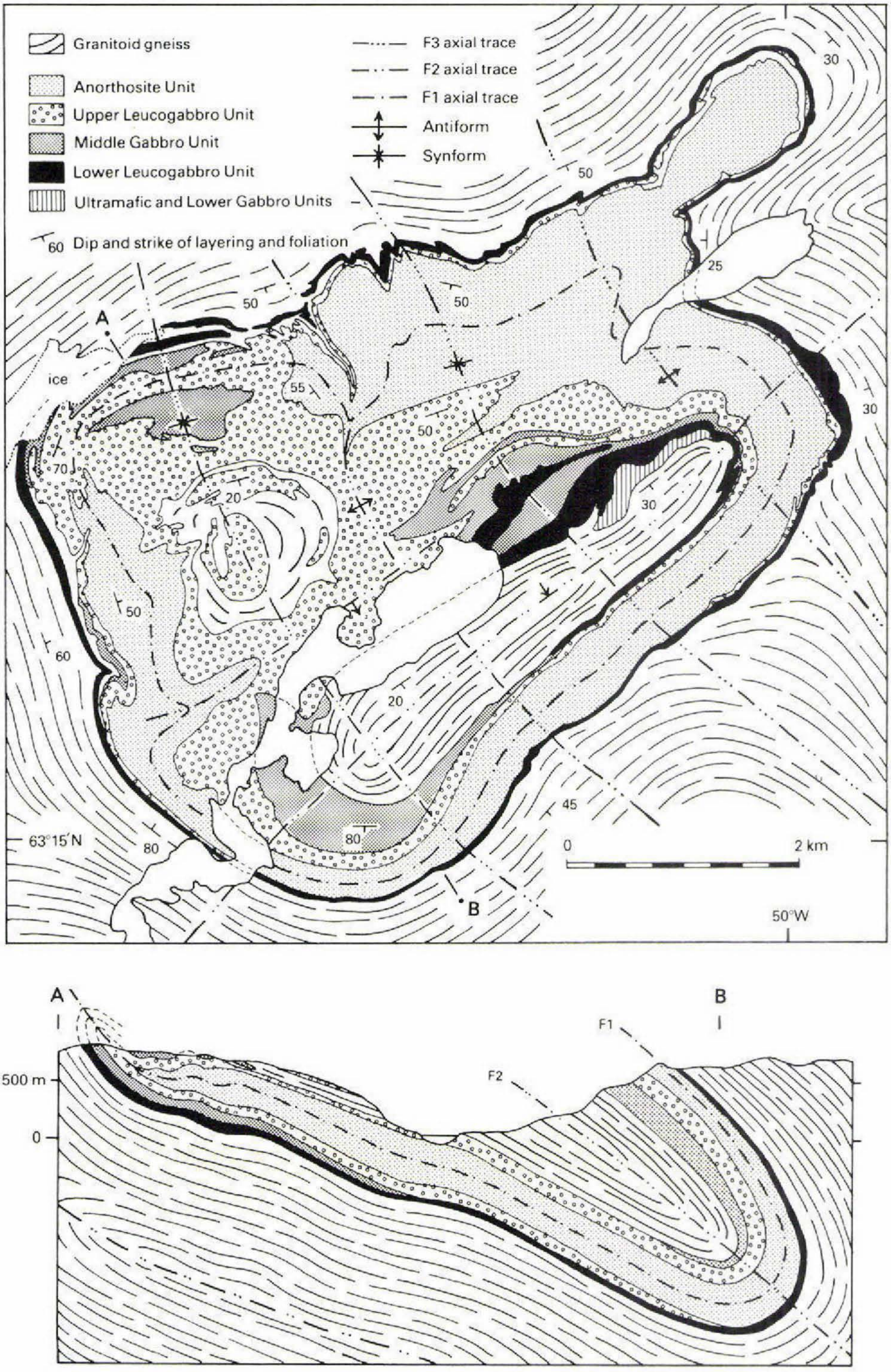


Fig. 67. Simplified geological map and cross section of the Majorqap qâva outcrop of the Fiskenæsset Complex showing the major fold structures.

F1 recumbent syncline (fig. 63). In contrast, most of the Fiskenæsset Complex outcropping in a second major belt to the south from Kangârssuk eastwards through Tasiussâ, Qagsse and north and east of Bjørnesund (plate 2) appears to be part of a single stratigraphic sequence which is inclined to the south-east and is the right way up (fig. 62). This southern belt could be interpreted as the lower limb of the major recumbent syncline at Majorqap qâva (fig. 68). In detail the structure is much more complicated because of the disruption of the Fiskenæsset Complex by granitoid intrusions and thrusting before F1, and later refolding of the recumbent F1 folds by major F2 and F3 folds.

DISCUSSION AND CONCLUSIONS

A number of conclusions can be drawn from the field evidence and petrography on the nature of the magma and its crystallisation as the Fiskenæsset Complex.

The structures and textures of the Fiskenæsset Complex indicate that this layered intrusion formed by both the deposition of crystals and crystal growth *in situ* from a magma. The initial composition of this magma cannot be determined as most rocks are cumulates which need not have been derived from a single magma in a closed chamber, and no unequivocal chilled phase is known. The Lower Gabbro Unit may largely represent a chilled phase of the initial magma, but this unit is thoroughly recrystallised and disrupted by granitoid veins and is thus poorly preserved. The typical rocks of this unit are, however, broadly similar in composition to the metavolcanic amphibolites which form the host rocks of the intrusion. These metavolcanic rocks, which range in composition from tholeiitic basalt to alkali olivine basalt and basaltic komatiite (Rivalenti & Rossi, 1975; Rivalenti, 1976; Friend *et al.*, 1981), may provide the best available indication of the composition of the magma(s) from which the Fiskenæsset Complex was derived. This interpretation is also favoured by Weaver *et al.* (1981) on the basis of the trace element geochemistry of the Fiskenæsset Complex and associated amphibolites on Qeqertarssuatsiaq (north), but contrasts with the calc-alkaline affinities suggested for the magma by Windley *et al.* (1973) and Windley & Smith (1976). A tectonic model which explains both the generation of the Fiskenæsset Complex and its subsequent incorporation as tectonic slices within granitoid gneisses is outlined in fig. 69 and is described in more detail by Myers (1981).

There is abundant evidence of crystal settling and magmatic currents in the Fiskenæsset Complex, including mineral-graded trough layers of various compositions, cross-bedding, and the depression of fine grained layers by large crystals at the base of undeformed overlying layers. The origin of mineral-graded layers in gabbro is less clear. Similar layers occur in the Skaergaard Intrusion and were first interpreted by Wager & Deer (1939) as the result of crystal settling and current ac-

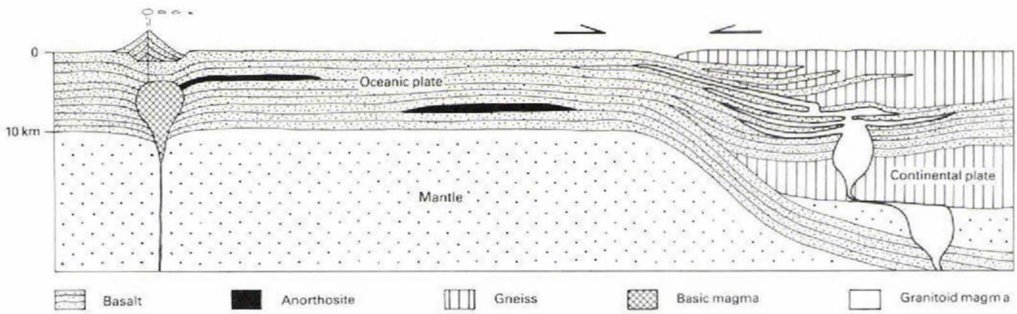


Fig. 69. Schematic section showing a tectonic model for the generation of layered anorthosite intrusions such as the Fiskenæsset Complex together with tholeiitic volcanic rocks at an oceanic spreading centre (left). On the right, the model shows how these rocks could subsequently have been interleaved by thrusting with granitoid magmas derived by subduction processes beneath the edge of a continental plate. The model shows an early stage of these processes.

tion. Recently, however, McBirney & Noyes (1979) reinterpreted these layers as formed by a mechanism of oscillatory nucleation and diffusion-controlled crystallisation *in situ*. This mechanism may explain the origin of intermittent, poorly defined mineral-graded layers of gabbro of great lateral extent in uniform gabbro, such as occur in the lower sub-unit 4.1 of the Fiskenæsset Complex. But this mechanism alone cannot explain the origin of similar, but more clearly defined, mineral-graded layers of gabbro restricted to narrow vertical columns of trough layers in sub-unit 4.2, and cannot be applied to the intermittent trough layers of peridotite in leucogabbro of the Lower and Upper Leucogabbro Units.

In discussing the Skaergaard Intrusion, McBirney & Noyes (1979) question whether minerals in most large igneous intrusions grew rapidly enough to sink or float faster than the advancing front of solidification. This generalisation may not apply to much of the Fiskenæsset Complex in which there is abundant direct evidence of crystal settling and current action. Evidence of rapid crystal growth is shown by the 'snow-flake' structures and abundant large hollow plagioclase crystals in the Upper Leucogabbro and Anorthosite Units, and many cumulus minerals throughout the complex are unusually large. The combined evidence suggests that a major part of the Fiskenæsset Complex formed by the gravitational settling of individual cumulus crystals (especially olivine, plagioclase, pyroxene and chromite), clusters of crystals (especially plagioclase), and giant poikilitic crystals (especially plagioclase enclosed by pyroxene).

The composition of minerals in the Middle Gabbro Unit are distinctly different from those in other parts of the intrusion (Myers & Platt, 1977), and this unit appears to represent the influx of a second batch of magma with higher MgO/FeO ratio (Myers, 1975a, 1975b) than the magma which formed the underlying Lower Gabbro, Ultramafic and Lower Leucogabbro Units (fig. 2). The leucogabbros of the Lower and Upper Leucogabbro Units are broadly similar to each other. This could indicate that they crystallised from a single (first) magma, intruded by a second magma which crystallised as the Middle Gabbro Unit between the already consolidated Lower Leucogabbro Unit and overlying remaining first magma (which subsequently crystallised as the Upper Leucogabbro and Anorthosite Units). This interpretation is not considered likely as the second magma, based on the composition of the minerals it crystallised, is likely to have been hotter than any remaining first magma. An alternative explanation is that the Upper Leucogabbro and Anorthosite Units are differentiation products of the same, second, magma which formed the Middle Gabbro Unit. The structures and rock types at the bottom of the Upper Leucogabbro Unit indicate a major episode of physico-chemical instability and supercooling. This suggests a third alternative, that the Upper Leucogabbro and overlying units were derived from a third major influx of magma, similar to the first, which was emplaced on top of the second magma towards the end of consolidation of the Middle Gabbro Unit.

The occurrence of the relatively thin (540 m) Fiskenæsset Complex over the whole Fiskenæsset region (plate 2) in folded layers with first fold amplitudes in excess of 20 km confirms the interpretation that the first major folds were recumbent. It also indicates that the region as a whole only represents a relatively thin layer of the Earth's crust which was individually thickened by tectonic distortion, thrusting and folding, detached from underlying and any overlying layers. The absence of major ultramafic bodies in the region suggests that, if the Fiskenæsset Complex was derived by crystal fractionation from a tholeiitic magma, then an associated ultramafic residue was at sufficient depth not to have been brought to the presently exposed crustal level during recumbent folding and thrusting. Therefore such an ultramafic component probably lay in a lower crustal layer or sank back to the mantle, perhaps during the intrusion of the enormous volume of granitoid material which engulfs the Fiskenæsset Complex. Any alkali-rich portion which would have formed during fractionation of the Fiskenæsset Complex from a tholeiitic magma may be part of, and not easily distinguished from, the flood of granitoid intrusions which make up most of the Fiskenæsset region.

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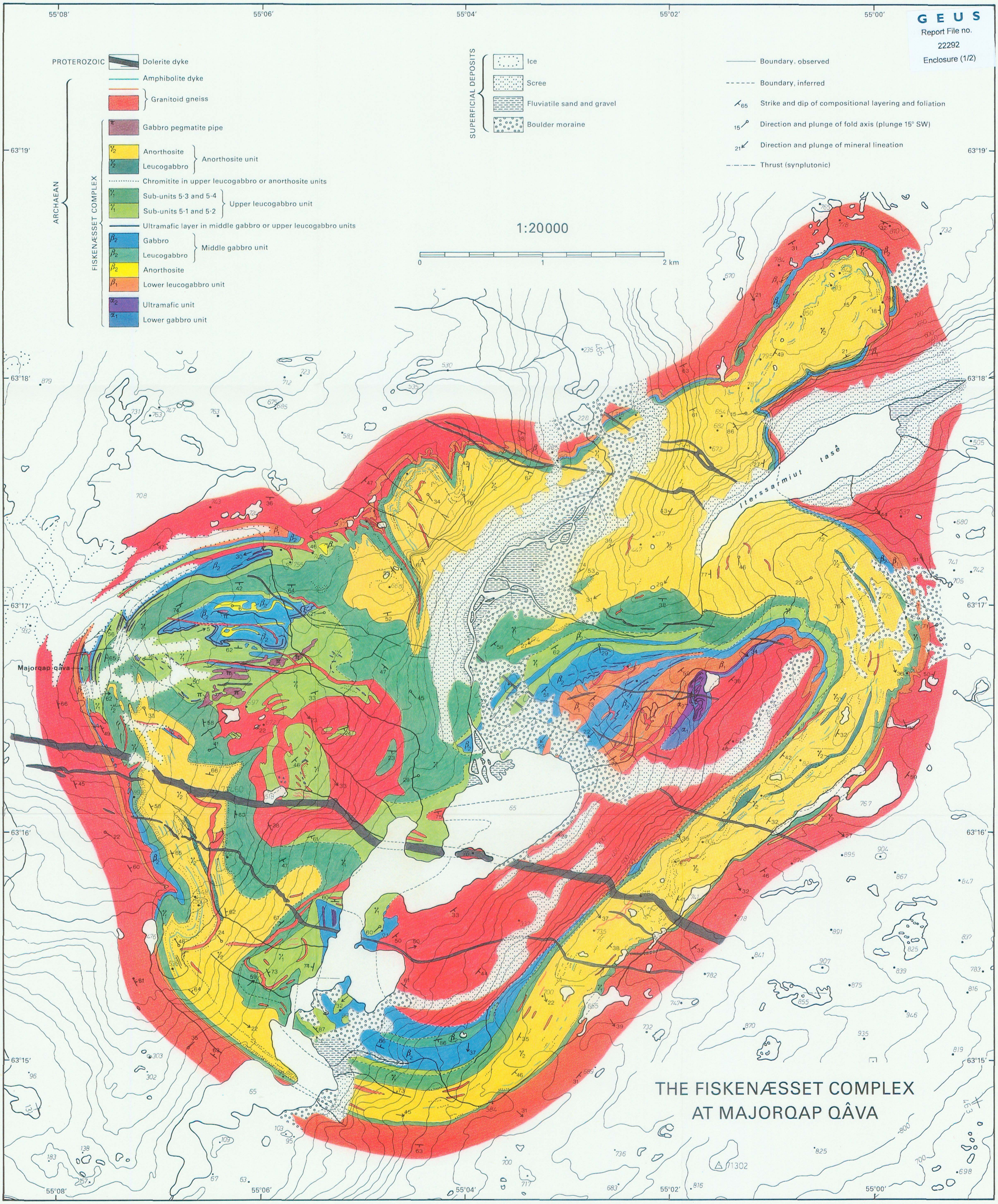
Plate 1. Geological map of the Majorqap qâva outcrop of the Fiskenæsset Complex.

Plate 2. Simplified geological map of the Fiskenæsset region showing the outcrop and structure of the Fiskenæsset Complex and the location of measured sections 1–47. These numbers are also used as locality numbers to locate observations or photographs in the vicinity of the measured sections.


GEUS
Report File no.
22292
Enclosure (1/2)
PROTEROZOIC
Dolerite dyke
Amphibolite dyke
Granitoid gneiss
ARCHAEAN
FISKENÆSSET COMPLEX
Gabbro pegmatite pipe
Anorthosite
Leucogabbro
Anorthosite unit
Chromitite in upper leucogabbro or anorthosite units
Sub-units 5·3 and 5·4
Sub-units 5·1 and 5·2
Upper leucogabbro unit
Ultramafic layer in middle gabbro or upper leucogabbro units
Gabbro
Leucogabbro
Anorthosite
Middle gabbro unit
Lower leucogabbro unit
Ultramafic unit
Lower gabbro unit
SUPERFICIAL DEPOSITS
Ice
Scree
Fluviatile sand and gravel
Boulder moraine
Boundary, observed
Boundary, inferred
Strike and dip of compositional layering and foliation
Direction and plunge of fold axis (plunge 15° SW)
Direction and plunge of mineral lineation
Thrust (synplutonic)
1:20000
0
1
2 km
Majorqap qâva
Iterssarmiut tasê
THE FISKENÆSSET COMPLEX
AT MAJORQAP QÂVA
55°08'
55°06'
55°04'
55°02'
55°00'
63°19'
63°18'
63°17'
63°16'
63°15'


GEUS
Report File no. 22292
Enclosure (2/2)
Grædefjord
ice
Majorqap qâva
Midgård
Uivfaup nûa
Puilassut qâva
Qeqertarssuatsiaq (north)
Amigfik
Fiskenæsfjorden
Manisât
Sinarssuk
Sungâsa nûat
Qeqertarssuatsiaq (south)
Taseq
Qagsse
Itivdlik Fault
Tasiussâ
Angnertussoq
Kangârssuk
Kigssaviarssup qâva
Naujarssuit
Nakaissorssuaq
Tasiussarssuaq
Tâterât qâva
Bjørnesund
Perserjôrssuaq
51° W
50° 30′ W
50° W
49° 30′ W
63° 30′ N
63° 15′ N
63° N
0
5
10 km
LEGEND
Granitoid gneiss, intruded as sheets into the Amphibolite and Anorthosite Complex
Granitoid gneiss, with fragments of Upper Gabbro Unit
Granitoid gneiss, with fragments of Anorthosite and Upper Leucogabbro Units
Granitoid gneiss, with fragments of Middle Gabbro and Lower Leucogabbro Units
Granitoid gneiss, with fragments of Ultramafic and Lower Gabbro Units
Granitoid gneiss, with fragments of Amphibolite of mainly volcanic origin
Fiskenæsset Anorthosite Complex
Upper Gabbro Unit
Anorthosite and Upper Leucogabbro Units
Middle Gabbro and Lower Leucogabbro Units
Ultramafic and Lower Gabbro Units
Amphibolite of mainly volcanic origin
Boundary, exposed
Boundary, inferred
Strike and dip of compositional layering and foliation
Direction and plunge of fold axis
Axial trace of anticline and syncline, F1, 1u upward facing, 1d downward facing
Axial trace of antiform and synform, F2
Axial trace of antiform and synform, F3
Way-up of mineral-graded layering
Fault
Measured section