GRØNLANDS GEOLOGISKE UNDERSØGELSE
BULLETIN No. 88

THE PLUTONIC HISTORY OF THE TASIUSSAQ AREA, SOUTH GREENLAND, WITH SPECIAL REFERENCE TO A HIGH-GRADE GNEISS COMPLEX

BY

PETER R. DAWES

WITH 53 FIGURES AND 1 TABLE IN THE TEXT,
8 PLATES AND 1 MAP

KØBENHAVN
BIANCO LUNOS BOGTRYKKERI A/S
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Abstract

Two Precambrian rock complexes, a high-grade gneiss complex and a younger granitic complex, can be recognised in the Tasiussaq area of South Greenland. Both complexes are cut by late basic dykes and faults.

The high-grade gneiss complex is composed of a variety of gneisses and granites, together with small areas of metasediments and metavolcanic rocks displaying well-preserved primary structures. Basic and ultramafic rocks also occur. Basic granulite layers in the gneisses represent early basic intrusions while pyroxene-metadolerite dykes represent later post-gneissic intrusions.

The complex is charnockitic in nature, the high-grade mineralogy being the result of granulite facies regional metamorphism. Hypersthene and garnet are common minerals but amphibole, mainly a brown hornblende, has a sparse distribution occurring in rocks containing no free quartz. Biotite is common and it occurs in equilibrium with the anhydrous pyroxene in the presence of H_2O.

Three phases of migmatisation post-dating an early granulite facies metamorphism can be recognised in the complex. This migmatisation was not connected to a regional downgrading of the complex and retrogressive mineral changes are local. Two structural units can be recognised, a lower unit composed of folded and migmatised gneisses and an upper unit composed of relatively undeformed and unmigmatised meta-arkoses and metaconglomerates. Evidence for four phases of deformation exists in the rocks of the lower unit, the most important structures being large-scale recumbent isoclines and nappe-like folds.

Discussion on the age of the Tasiussaq complex is given in the light of the geology of the whole of the Tasermiut fjord region. The idea is favoured that the granitic gneisses, gneisses, schists, metasediments and metavolcanics of the region have not all been derived from a single supracrustal pile. The possibilities that the meta-arkoses and metaconglomerates might represent cover rocks to older basement gneisses, or that such metasediments may represent syn-orogenic flysch-type deposits laid down on rocks of the same geological cycle, are considered.

The granitic complex is composed of granites of four ages. Hornblende- and biotite-bearing rapakivi granites, which are associated with the intrusion of a suite of dolerite-norite rocks, post-date an autochthonous microcline granite. A late microgranite was emplaced following the potash metasomatism of the rapakivi granites. Diorite, dolerite and ultramafic rocks were intruded during the late stages of the granitic complex.

Comments on the nature of the Precambrian evolution conclude the paper. It is suggested that the plutonism which resulted in the formation of the granitic complex is complementary to that forming the high-grade gneiss complex, indicating a tendency for acidic magma generated at depth during the early stages of a developing orogen to rise to higher levels. The late dolerite dykes and faults which cut both complexes are considered as post-plutonic or epeirogenic events.

CONTENTS

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INTRODUCTION

Scope of the paper

This paper presents the main features of the Precambrian geology of the Tasiussaq area, South Greenland, but places special emphasis on the development of the high-grade gneiss complex. The Tasiussaq area forms the basis of an unpublished Ph. D. thesis by the author which is however mainly concerned with the genesis of rapakivi granite. The rapakivi granites and associated basic rocks of the Tasiussaq area, which are members of the granitic complex (*i.e.* post-high-grade gneiss complex), are to be treated in detail in later publications and thus they are only briefly mentioned here to give a complete picture of the chronology.

Rock types of the high-grade gneiss complex and their relationships to each other are described in some detail but petrographical descriptions are kept to a minimum. Aspects of the metamorphism and structural development of the complex are dealt with at some length. For a full appreciation of the age and significance of the high-grade gneiss complex, comparisons and correlations to other areas in the immediate vicinity of the Tasiussaq area are considered and comments are made on the geology of the whole of the Tasermiut fjord region of which the Tasiussaq area is a part.

Wegmann (1939, 1948) drew attention to the importance of the Tasermiut fjord region in the understanding of the regional geology of this part of South Greenland. Unfortunately the region formed the south-eastern boundary of the detailed geological team mapping carried out by the Geological Survey of Greenland (GGU) and which finished in 1963 with the result that parts of the region have not been mapped in detail, while other parts, especially in the north towards the Inland Ice, have not been visited at all. Until mapping or at least geological reconnaissance has been carried out in the whole of the Tasermiut fjord region, statements dealing with the interpretation of the regional geology should be regarded purely as appraisals of the knowledge available at the time and it is with this in mind that the author gives the ideas set down in this paper.

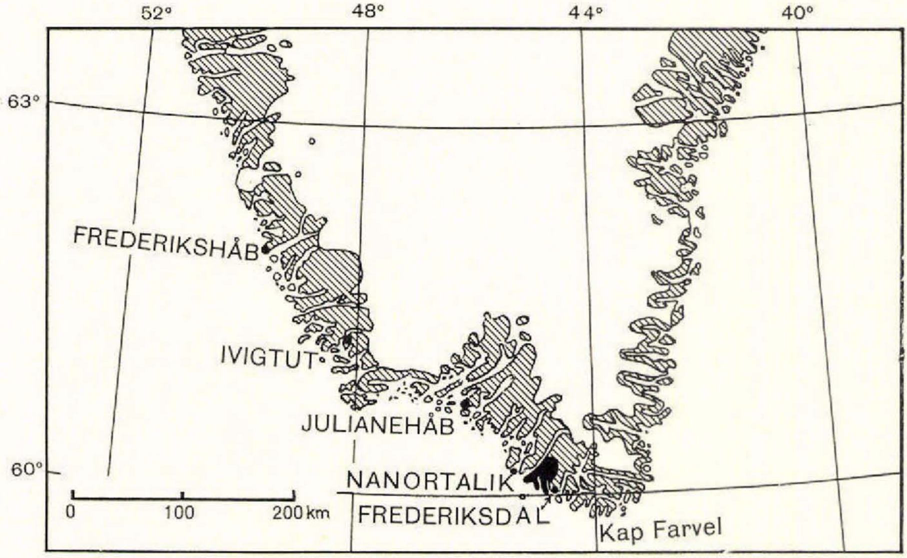

Fig. 1. Location of the Tasiussaq area (black) in South Greenland.

The Tasiussaq area takes its name from the small settlement named Tasiussaq situated in the north of the area at the mouth of what is referred to in the text as "Tasiussaq bay" (bugt). Reference to the settlement in the text is by "Tasiussaq village". A third form "Tasiussaq" is also used in the text for the square-ended inlet in the west of the area. All three names are given on Map 1.

Location and field work

The Tasiussaq area is situated on the south-west coast of Greenland between 60°3′N and 60°18′ N (Fig. 1, Plate 1). Tasermiut fjord, the southernmost part of which borders the Tasiussaq area on the north-west, lies approximately 75 km to the north-west of Kap Farvel, the southern point of Greenland.

The field work leading to the present account was carried out by the author in the summers of 1962 and 1963 as part of a systematic mapping project in South Greenland by GGU. In the years 1956 to 1963, GGU mapped on the scale of 1:20,000 the whole coast from north of Ivigtut to Frederiksdal (Fig. 1). The Tasiussaq area is the most southeasterly area investigated during the team mapping which finished in 1963. W. S. WATT investigated the Frederiksdal rapakivi massif which borders the Tasiussaq area on the south while WALLIS (1966) working independently from GGU has studied the area immediately to the north. ESCHER (1966) has mapped the south-western part of the Nanortalik peninsula to the west of Tasermiut fjord, the north-eastern part towards the Inland Ice having been investigated by J. WATTERSON. In 1965 and 1967 reconnaissance investigations were carried out by GGU to the east and south-east of the Tasiussaq area in the Kap Farvel region (BRIDGWATER, SUTTON and WATTERSON, 1966; SUTTON and WATTERSON, 1968).

The main mapping of the Tasiussaq area (approximately 450 km²) was carried out from 19 helicopter and light-weight camps (see Map 1) but the motor-cutter "J. P. Koch" was used as a base to map the coastal areas of Tasermiut fjord, Tasiussaq, Narssap sarqâ, the eastern side of Igdlukasîp tunua and also for a cursory examination of the rocks from Torssukátak fjord to Augpilagtoq, to the east of the Tasiussaq area. A small outboard motor-boat was used to visit localities on both sides of the lake Taserssuaq.

The majority of the mapping was done on 1:20,000 maps supplemented by aerial photographs, all supplied by the Geodetic Institute, Copenhagen.

Physical features of the Tasiussaq area

The area ranges in altitude from sea-level to the high peaks of the eastern area, the highest summit reaching 1763 m (see Map 1). The topography varies from the flat-lying coastal districts around Tasiussaq bay to the lichen-covered low plateau (around 700 m) of the western part of the area (west of Tasiussaq), to the high eastern and northern area with its rugged and alpine mountains with sharp peaks, in places needle-like, rising above ice-filled corries (see Plate 1). About twenty-five corrie glaciers exist and in the eastern part of the region permanent snow-fields are common.

Excellent exposures exist at the water-smoothed coasts although locally both moraine and scree hindered observations. Glacially polished surfaces in the high eastern area provide locally perfect exposures although here scree and talus slopes from the steep, jagged peaks obscure much. The upper parts of the mountains, however, are cleanly exposed but are only accessible in places. The land around Taserssuaq supports many willow and birch trees and Qínguadalen, locally famous for its sizeable trees, is situated at the eastern end of the lake.

Regional geological setting

In South Greenland remnants of two Precambrian fold-belts are preserved: the pre-Ketilidian (ca. 2700–2500 m.y.) and the Ketilidian (ca. 1800–1500 m.y.). In the Ivigtut area, where the relationship between these belts has been determined, the pre-Ketilidian is represented by folded crystalline gneisses (the "Ivigtut gneisses") and a metamorphosed supracrustal series known as the Tartoq Group (HIGGINS and BONDESEN, 1966). These rocks form an old basement which in the area north of Ivigtut is overlain unconformably by Ketilidian sediments and lavas. These Ketilidian supracrustal rocks can be traced into higher grade metamorphosed and migmatised rocks to the south and south-east.

WEGMANN (1938a) recognised the existence of these two fold-belts referring to them as "cycles" and he also recognised the later Gardar events of sedimentation and cratogenic magmatism, events which took place between 1300–1000 m.y. ago. Unmetamorphosed Gardar supracrustal rocks are restricted to the Tunugdliarfik trough region, north-north-east of Julianehåb, but Gardar igneous rocks have a much wider distribution. Rocks of WEGMANN's Ketilidian exist between the Ivigtut area and Kap Farvel but, as WEGMANN (1938b, 1939, 1948) pointed out, the presence of reworked areas of pre-Ketilidian rocks is to be expected. Comments on the age of the rocks of the Tasiussaq area and on correlation with rocks elsewhere in South Greenland, particularly those of Tasermiut fjord, are given in this paper.

Following later investigations in the Ivigtut area, however, BERTHELSEN (1961) erroneously interpreted WEGMANN's pre-Ketilidian gneisses around Ivigtut as "transformed Ketilidian supracrustal rocks" and this has led to some confusion in chronological terminology, since WEGMANN's chronology was subsequently revised and new terms introduced (see ALLAART 1964; BRIDGWATER, 1965). For a discussion of the problems arising from this and for a more detailed account of the geology of the Precambrian basement of South Greenland, the reader is referred to the papers quoted above and to the accounts of BRIDGWATER and WALTON (1964), WATTERSON (1965), WALTON (1965), ALLAART, BRIDGWATER and HENRIKSEN (1969) and HENRIKSEN (1969).

Previous investigations in the Tasiussaq area

Few references exist in the literature mentioning geological aspects of the Tasiussaq area and although the coasts were certainly passed by early investigators, little was written down. STEENSTRUP, KORNERUP and HOLM all visited Tasermiut fjord but they were more concerned with the glaciers, alpine mountains and Norse ruins towards the head of the fjord (STEENSTRUP and KORNERUP, 1881; HOLM, 1883). Also, the Frederiksdal rapakivi massif, bordering the Tasiussaq area on the south, held the attention of the earlier workers (SYLOW, 1883; TÖRNEBOHN, 1886; KNUTSEN and EBERLIN, 1889). However EBERLIN (KNUTSEN and EBERLIN, 1889, p. 252) describes rocks from his southern granite belt to the east of the Tasiussaq area as coarse-grained, striped garnet granites having transitional relationships to the gneiss. These rocks are most probably equivalent to the garnet-bearing charnockitic gneisses of the Tasiussaq area and their granitic variations (see p. 23).

A. Jessen was the first to visit the Tasiussaq area for geological purposes during his trip into Tasermiut fjord in 1894 and he remarks about the region which is the northern part of the Tasiussaq area (Jessen, 1896, p. 140): "On the southern coast of the fjord the gneisses reach the big lake Tasersuak, the south coast of which is composed of a high jagged gneiss field, partly composed of iron gneiss covered by vast screes. A little way south-west of Tasersuak the lower side of the fjord consists of younger, coarse-grained granite, and north of the lake one finds only granite, partly the same younger, although completely subordinate except around glaciers in the bottom of the fjord, and partly older, grey, fine-grained granite" (translation). "Tasersuak" is the old name for Taserssuaq. Jessen's younger, coarse-grained granite is the biotite rapakivi granite of this report (see p. 89) and the iron gneiss is the sulphide-bearing gneiss which forms layers in the hypersthene-quartzo-feldspathic gneiss south of the lake (see p. 33). Jessen's fine-grained older granite however represents in fact grey metasedimentary psammites which in many places look like homogeneous gneiss (see p. 29). This interpretation is clear since Jessen (1896, p. 140) goes on to describe a sharp and clear contact to the west of Tasermiut fjord between dark-coloured gneisses and the fine-grained granite. This contact is in fact between dark metavolcanic rocks and underlying lighter-coloured psammites.

Jessen was the first to map the contact of the biotite rapakivi massif although his boundary has been revised (see fig. 2). Wegmann, although not visiting the rapakivi in 1936, mentions its presence as erratics (Wegmann, 1938a, p. 115). In a small-scale map in a later publication, Wegmann (1939) includes the western part of the Tasiussaq area, apart from the "New granite", *i.e.* the biotite rapakivi granite, within the grey gneiss of the "Upper Sermilik group", with the gneisses of the eastern part of the area as higher grade gneisses being part of the granulite complex of the Kap Farvel region.

The only other observer prior to the mapping in 1962 and 1963 was the late G. H. Francis who, as a member of the British Museum Cape Farewell Expedition, visited the northern part of the Tasiussaq area for approximately two weeks in the summer of 1957, to make collections of the basement gneisses and younger granites. Unfortunately Francis died before publishing on the geology although a manuscript (Francis, 1957), rock samples, diaries and a rough field map are available in the British Museum archives (Natural History Section), together with a semi-popular article (Francis, 1958) in which some comments on the geology are given.

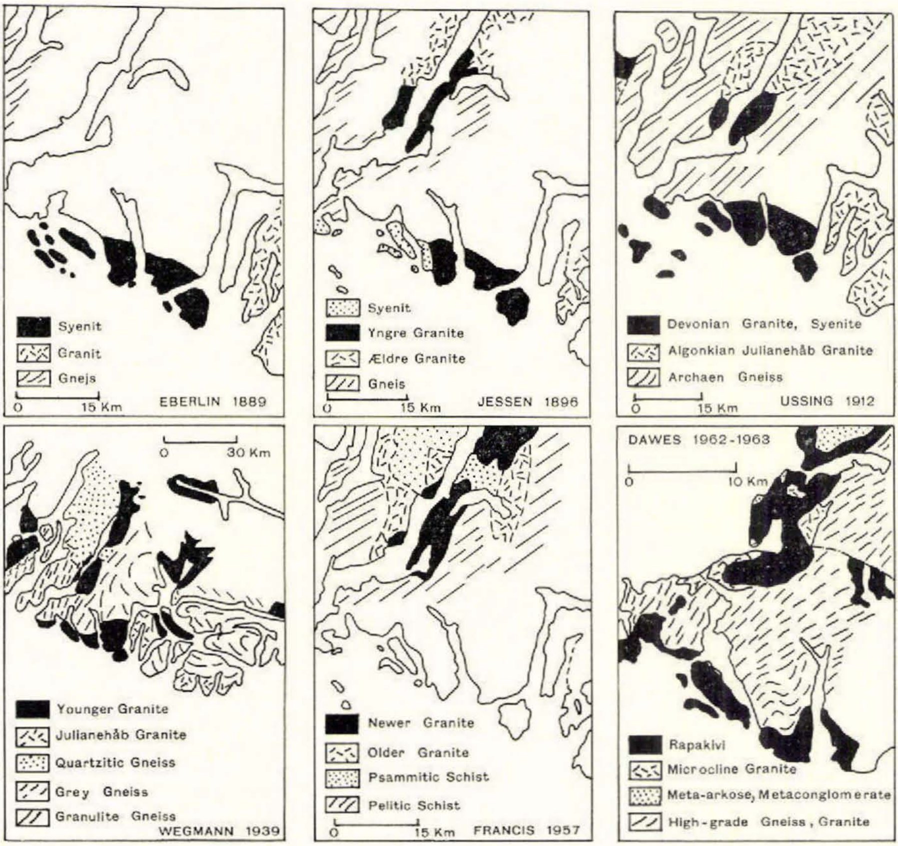

Fig. 2. Compilation diagram showing the previous geological maps and sketches pertaining to the Tasiussaq area and environs.

All the small-scale maps in the past literature and from other sources pertaining to the Tasiussaq area and its surroundings have been compiled in figure 2.

General geology of the Tasiussaq area

Two rock complexes exist in the Tasiussaq area (Fig. 3) both of which are cut by late basic dykes. A high-grade gneiss complex is composed of a variety of gneisses and granites of charnockitic affinities together with smaller areas of basic and ultramafic rocks. Small isolated areas of mainly unmigmatitic meta-arkoses, metaconglomerates and metavolcanics also exist. All the rocks of this complex have been metamorphosed under granulite facies conditions. Some high-grade pyroxene-

Fig. 3. View taken from the north of Tasiussaq village looking south-eastwards over Tasiussaq bay. The low foreground is composed of rocks of the granitic complex while the high snow- and ice-covered mountains in the background are composed of gneisses and granites of the high-grade gneiss complex. The boundary between the two complexes is situated at the base of the mountains. The highest peak seen is 1400 m. June 10th, 1963.

metadolerite dykes cut the gneisses and granites of the complex. The complex may contain rock material of both basement and a younger cover although no conclusive evidence is forthcoming.

The high-grade gneiss complex is cut by a granitic complex which has a smaller extent but which is composed of granites of four ages with associated pegmatites and aplites. Autochthonous and allochthonous granite types exist. Rapakivi granites, which have a complicated history and which display both magmatic and metasomatic characters, form the most important rock type of the complex and they are associated with a suite of norite-gabbro-dolerite-diorite rocks which occurs as dykes and small bodies. Further basic intrusions are represented by post-rapakivi metadolerite dykes and diorite sheets.

Numerous late events can be recognised in the Tasiussaq area, divisible into basic intrusion, represented by three generations of dolerite dykes which cut sharply through rocks of both complexes, and tectonic events, including faulting and mylonitisation.

The chronology of the Precambrian of the Tasiussaq area is diagrammatically represented in Table 1 and, in Map 1, the Precambrian geology of the area is shown on the scale 1:40,000. Data on the island Igdlukasik and the position of the boundary between gabbro and granite to the west of Frederiksdal have been supplied by W. STUART WATT while some of the geological boundaries in the extreme north-east of the map to the north of the eastern end of Taserssuaq have been modified from J. H. ALLAART's field work.

HIGH-GRADE GNEISS COMPLEX

Extent and nature of the complex

Rocks of the high-grade gneiss complex form approximately 300 km^2 of the Tasiussaq area. The complex is charnockitic in nature and is composed of a variety of gneisses, granites and schistose rocks together with metasediments and metavolcanics displaying well-preserved primary structures. Basic and ultrabasic rocks also exist. Migmatitic features vary in size from small streaks, flecks and veins of quartzo-feldspathic material to larger sheets of pegmatite which in places give the gneisses a conspicuous banded appearance.

The metasediments (meta-arkoses and metaconglomerates) and the metavolcanics have only a small occurrence in the complex and they are not found intercalated with the gneisses. They differ from the gneisses in that they are not generally migmatitic and they are relatively undeformed but the exact relationship between the two rock groups is not clear. In the north of the area, meta-arkoses and metaconglomerates are not in contact with the gneisses but only with the younger biotite rapakivi granite while to the west of Tasiussaq a small outcrop of metavolcanics and meta-arkoses appears to be of tectonic origin but even here relations to the underlying gneisses are obscure. Thus the question arises as to the age relationship between the metasediments and the migmatitic gneisses and whether the high-grade gneiss complex is composed of material of more than one age.

Comments on charnockites and granulites

WEGMANN (1939) referred to the hypersthene- and garnet-bearing gneisses in this part of South Greenland, including the gneisses in the northern and eastern part of the Tasiussaq area, as "granulites" and also as "gneisses of the charnockite series" comparing them with the charnockitic rocks described by WAGER (1934) from the Angmagssalik area of south East Greenland. Thus it is of some interest to see how far the terms "charnockite" and "granulite" can be used to describe the rocks of the Tasiussaq complex.

From the earliest days of geological research the occurrence in close association of charnockites and granulites has been noted. ESKOLA (1952) outlined his ideas on the connection between the two groups of rocks with reference to the rocks of Lapland and remarks (p. 142) that "all the rocks of the granulite facies are granulites" and that (p. 133) "the granulite facies appear to be identical with that facies represented by the charnockites, the specific characteristics of the foliated granulites being due to their tectonic history". Hence, according to ESKOLA, some charnockites are granulites. Clearly the petrogenesis of the charnockites and granulites are linked through the metamorphic facies concept but the use of both terms or derivatives of each term must be qualified before usage, a practice made necessary through the expanded use of both terms by geological workers. Perhaps a simplified but practical way of looking at the problem in general terms is to consider that the charnockites and charnockitic rocks bear the same relationship to the granulites as granites and granitic rocks do to gneisses.

The controversy stemming from the extended use of HOLLAND's (1900) terms "charnockite" and "charnockite series" is well known. His wishes that the terms should not be used outside Peninsular India and that "charnockite" should not be used for "any hypersthene granite" have not been met. Workers outside India have freely adopted HOLLAND's terminology to describe all types of granulitic pyroxene-bearing rocks and this has directly led to the many discussions on the nature and genesis of the hypersthene-bearing rocks of India which have developed into a controversy embracing actual terminology and the petrogenesis of hypersthene-bearing rocks in general. The discussions appear to have reached a stage where it is accepted that both igneous and metamorphic charnockites occur in nature and PICHAMUTHU (1953) has stated that "there are charnockites and charnockites" while PARRAS (1958) has listed charnockites as occurring in the six continents of the world and in Greenland and Antarctica. His Greenland examples are from both the west and east coasts (WAGER, 1934; KRANCK, 1935; RAMBERG, 1951). The hypersthene-bearing gneisses and granites of the Tasiussaq complex clearly belong to the charnockitic realm of rocks providing an example of a charnockitic complex derived through granulite facies regional metamorphism. The majority of the gneisses are metamorphosed sedimentary and igneous rocks and "igneous charnockites" are restricted to the few charnockite pegmatite veins cutting the gneisses. Some areas of the G_1 granites and the hypersthene-quartzo-feldspathic gneisses, where fusion, melting and recrystallisation of felsic material may have occurred, form a genetical link between the "metamorphic" and "igneous" charnockitic rocks.

Granulites in the original German and Scandinavian sense are gneisses in which the mineral paragenesis is strictly anhydrous but due to the connection of granulites with granulite facies metamorphism

(Eskola, 1952), the term may now cover a variety of rocks. Thus there are granulites in the facies sense and granulites in the classical sense. In the former sense, the rocks of the Tasiussaq area can be referred to as granulites but they are not comparable in character to the original and classical granulites of Saxony or to granulite *sensu stricto*. Furthermore the rocks are not comparable to granulites in the common British usage of the term where granulite is a fine-grained quartzo-feldspathic rock without conspicuous schistosity, *e.g.* Moine granulite, nor are they equivalent to granulites in the French sense of a fine-grained muscovite granite. However some of the rocks, especially the more homogeneous types, are comparable to rocks described elsewhere as granulites or granulitic gneisses, *e.g.* Hsu, 1955; Riley, 1960; Hepworth, 1964; Evans, 1965.

Nomenclature

Hypersthene is a very common mineral in the gneisses, granites and basic rocks of the high-grade gneiss complex and rock types mineralogically similar to members of the "acid" and "basic" divisions in Holland's (*op. cit.*) "charnockite series", and to Subramaniam's (1959) restricted "charnockite suite" exist. However to avoid any confusion with Subramaniam's strict redefinition of "charnockite" *sensu stricto*, the broader term hypersthene granite is preferred in this paper. Some rocks described here as hypersthene granite have close similarities to charnockite *s. str.* but the majority of hypersthene granites from the same suite vary on one or more counts from Subramaniam's redefinition (*op. cit.*, p. 328). Tilley's (1936) term "enderbite" and Spurr's (1900) term "alaskite", other members of Subramaniam's "charnockite suite", are used.

The gneisses are referred to by their characteristic mineral content (*e.g.* biotite-garnet gneiss, hypersthene-quartzo-feldspathic gneiss), although many are mineralogically similar to rocks which elsewhere have been described as paracharnockite (Parras, 1958); charnockitic granulite (Morel, 1958); charnockitic gneiss (Evans, 1965). The gneisses vary from charnockitic to enderbitic and quartz dioritic in composition. Charnockitic gneiss is reserved for the hypersthene-quartzo-feldspathic gneisses which are the dominant gneiss type of the complex and which are characterised by the charnockite assemblage: hypersthene-quartz-perthite-plagioclase with or without garnet and biotite (see p. 25).

Basic granulite is used to describe the rocks having pyriclasite and biotite-pyriclasite compositions corresponding to the basic granulites of Singh (1966) and to rocks described elsewhere as basic charnockite (Parras, 1958; Cooray, 1962), pyroxene granulite (Subramaniam, 1959) and basic charnockitic granulite (Evans, 1965).

In this paper "granulite" is used in the mineral facies sense of ESKOLA (1952) for rocks which display metamorphic mineral assemblages typical of the granulite facies while "granulitic texture" is used for a metamorphic fabric in which the quartz and/or plagioclase grains show a tendency to be flattened into lenticles either being present with or without intercalated equigranular quartz-feldspar fabrics.

Division of the complex

It is convenient to describe the rocks of the complex under three headings: paragneisses, granitic rocks, and basic and ultramafic rocks. The majority of the rocks of the complex are paragneisses formed by regional metamorphism of pelitic, psammitic and psephitic rocks containing subsidiary calcareous material. Granitic gneisses and local granites forming transitions from the veined and streaky gneisses are included under "paragneisses" but mappable non-gneissic rocks varying in composition from charnockite and granite to enderbite and granodiorite are dealt with under "granitic rocks". Metavolcanic rocks and basic igneous intrusive rocks are grouped together with the small occurrences of ultramafic rocks under "basic and ultramafic rocks".

Paragneisses

The term "paragneiss" as used here includes the metamorphic rocks, excluding the granites, known to have been derived by metamorphism of original sedimentary rocks. The paragneisses are subdivided into three groups based on the nature of the pre-metamorphic rock types. Rocks bearing sulphides are considered separately.

Pelitic rocks

Gneisses and schistose gneisses derived from pelitic rocks outcrop in the west of the area bordering Tasermiut fjord (Map 1). A gradation exists from biotite schistose gneiss to biotite gneiss with increased migmatisation, the biotite gneisses varying from veined to banded in character. An area of biotite-garnet gneiss where garnets become notably conspicuous can be mapped separately. Development of graphite and sulphides in some rocks produce rusty gneisses. Porphyroblastesis has produced local areas of augen gneiss (Fig. 6).

The boundaries indicated on Map 1 separating biotite gneiss, biotite schistose gneiss and schist, and biotite-garnet gneiss are arbitrary since gradations exist between them. Thus areas of schistose gneiss and schist occur within the biotite gneisses and garnets may occur in the biotite gneiss producing local areas of biotite-garnet gneiss.

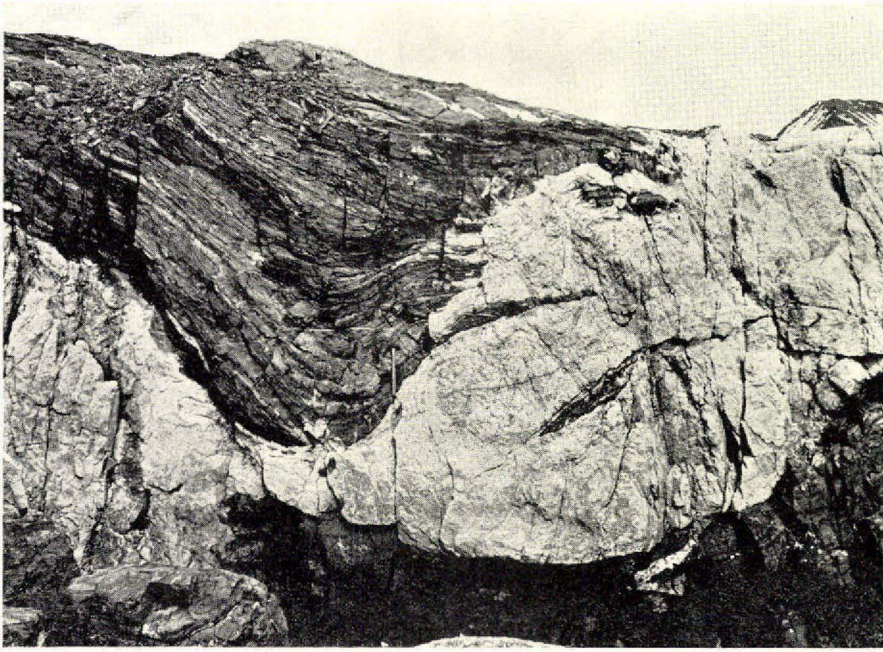

Fig. 4. Biotite schists in small F_3 synform replaced by M_3 pegmatite. Relics of schist occur in the pegmatite. Transitions from such schists to biotite schistose gneiss and biotite gneiss occur where finer-grained quartzo-feldspathic material has developed as small streaks and veins parallel to schistosity. Hammer (centre) as scale. East of Pingoq.

Biotite schistose gneisses and schists

Biotite schistose gneisses and schists are the youngest pelitic rocks of the area and they form a belt approximately 2 km wide to the south-east of Usuk. The majority of the rocks are schistose gneisses but areas of darker unmigmatitic schists do occur (Fig. 4). The rocks have a conformable relationship to the underlying biotite gneisses being situated in the centre of a broad synform.

The rocks vary from grey to dark grey in colour but occasionally rusty layers occur. The more schistose types vary from fine- to medium-grained and have a lepidoblastic texture. The more gneissic types containing more quartz have a less pronounced biotite foliation and have a grano-lepidoblastic texture. A xenomorphic-granoblastic mesostasis of quartz and feldspar is common (Plate 2 a).

Mineral assemblages from the following: biotite, orthopyroxene, plagioclase, potash feldspar, quartz, garnet, (myrmekite).

Accessory minerals: apatite, zircon, graphite, pyrite, pyrrhotite.

Alteration minerals: sericite, penninite.

Fig. 5. Succession of banded gneiss, north of Pingoq. Banding produced by concordant and sub-concordant pegmatites (M_2 and M_3) in biotite gneiss. The two melanocratic layers (right centre) are basic granulite sills which are truncated by the pegmatites. The succession isc ut by dolerite dykes (see p. 113) the largest of which fails to reach the coast. Height of the cliff is 210 m.

Biotite gneisses

Biotite gneisses underlie the biotite schistose gneisses and schists and they outcrop to the east and west of those rocks forming the lower level of the synform mentioned above. The gneisses are fine to medium in grain size and vary from dark grey to brown in colour. The dominant rock type is a veined gneiss but on the west coast of Pingoq layers of pegmatite have produced a succession of banded gneiss (Fig. 5). North of Pingoq areas of augen gneiss exist (Fig. 6). South of Pingoq the gneisses are more quartz-rich and they weather readily into a brown sand.

The rocks are characterised by a lepidoblastic to granoblastic-xenomorphic texture produced by a mesostasis of quartz and feldspar in which biotite forms a foliation. Hypersthene is rather common and garnet exists in certain horizons.

Mineral assemblages from the following: biotite, orthopyroxene, garnet, potash feldspar, plagioclase, quartz, myrmekite, muscovite.

Accessory minerals: zircon, ore.

Alteration minerals: green hornblende, chlorite, biotite, saussurite, sericite.

Fig. 6. Augen gneiss. Development of felsic material (M_2) in porphyroblastic aggregates within a pelitic rock. Flexures of the schistosity of the host rock are common. Coast, south of Pingoq.

Biotite-garnet gneisses

Biotite-garnet gneisses outcrop to the west of Tasiussaq. The gneisses dip to the north-west and they in part structurally underlie the biotite gneisses. The dominant rock type is a veined gneiss which occasionally has a spotted appearance due to the presence of garnets. The garnets occur either singly or in clusters and they exist both in the felsic microcline-bearing migmatitic veins and in the more basic component of the gneisses. Transitions to granitic gneiss are common (Fig. 7).

The gneisses are medium-grained and vary from grey, brown to reddish brown in colour. Texture varies from lepidoblastic to eugranoblastic-xenomorphic and a mesostasis of quartz and feldspar is typical. Frequently both garnet and hypersthene are poikiloblastic.

Mineral assemblages from the following: biotite, garnet, orthopyroxene, plagioclase, potash feldspar, quartz, (clinopyroxene).

Accessory minerals: graphite, ore, apatite, rutile.

Alteration minerals: chlorite, biotite, green hornblende, saussurite, sericite.

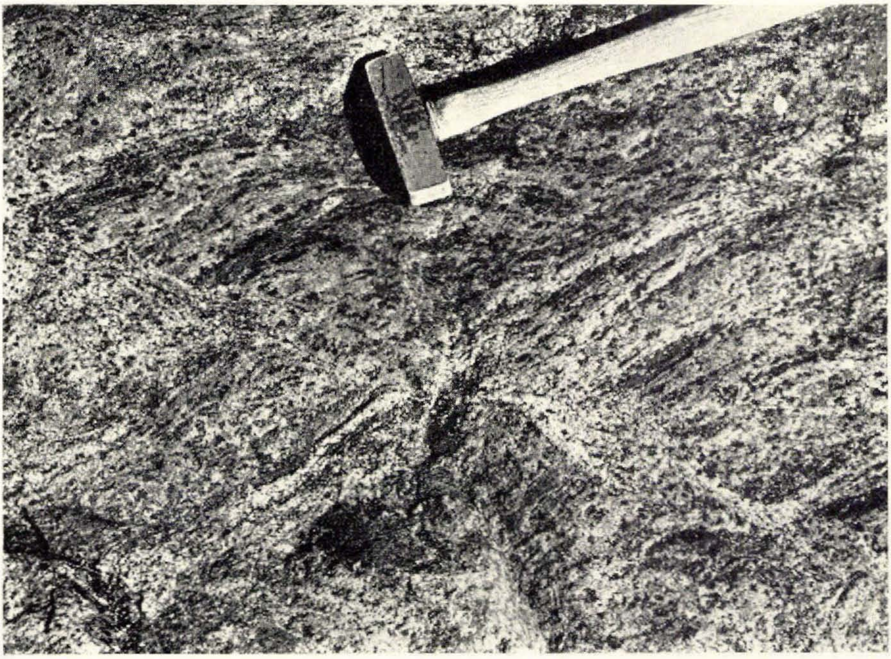

Fig. 7. Granitic biotite-garnet gneiss showing biotite-rich strips within a leucocratic granitic material. Dark spots are garnets. Transitions from biotite-garnet gneiss where the amount of quartzo-feldspathic material is relatively small to this granitised state are locally common. West of Tasiussaq.

Graphitic rocks

A few occurrences of graphite-bearing gneisses and schists exist as small horizons especially in the biotite gneisses and biotite-garnet gneisses. On the Tasermiut fjord coast to the east of Usuk, pods of pure graphite can be sampled. The rocks are fine-grained with the graphite commonly forming black streaks within the schistosity of the rock. In thin section graphite occurs in scales and thin flakes which may form discontinuous network.

Mineral assemblages from the following: graphite, biotite, plagioclase, quartz, garnet.

Accessory minerals: apatite, zircon, ore, calcite.

Alteration minerals: sericite, epidote.

Calcareous rocks

Calcareous rocks are present only locally in the gneiss complex. No pure marbles were found and the few occurrences can all be classified as:

Calc-silicate granulites

A few bands of calc-silicate rocks up to 1 m in thickness occur in the biotite gneisses and the biotite-garnet gneisses in the west of the area. A larger band, over 10 m in thickness, was noted in the hypersthene-quartzo-feldspathic gneiss in the north of the area. The bands commonly have been boudinaged and are now impersistent. The rocks vary in colour from grey to bluish grey and they commonly display a grooved or pitted surface due to weathering. Some rocks show a small-scale banding of lighter- and darker-coloured units and in places white- and cream-coloured layers exist.

The calc-silicate granulites vary from fine to medium in grain size and they have hypautomorphic to xenomorphic-granoblastic textures. Biotite and potash feldspar are notably absent.

Mineral assemblages from the following: tremolite, calcite, clinopyroxene, orthopyroxene, garnet, wollastonite, epidote, plagioclase, quartz, scapolite, olivine.

Accessory minerals: sphene, spinel, apatite, ore.

Alteration minerals: sericite, saussurite.

Psammitic and psephitic rocks

Psammitic and psephitic rocks are represented by gneisses, schistose gneisses, and granitic gneisses, by migmatised psammites and by unmigmatitic meta-arkoses, and metaconglomerates. Macroscopic sedimentary features are excellently preserved in the unmigmatised rocks and even locally in the migmatised psammites. A transition exists from migmatised psammites to the hypersthene-quartzo-feldspathic gneisses.

Hypersthene-quartzo-feldspathic gneisses form the dominant rock type of the complex but the unmigmatised meta-arkoses and the meta-conglomerates have a restricted distribution, only occurring in two areas: in a small outlier to the west of Tasiussaq and as a roof-zone to, and as inclusions in, the biotite rapakivi granite in the north of the area (see Map 1).

Hypersthene-quartzo-feldspathic gneisses (charnockitic gneisses)

Hypersthene-quartzo-feldspathic gneisses (including hypersthene-biotite-garnet gneiss, hypersthene-garnet gneiss, hypersthene-biotite gneiss) are the dominant rocks of the high-grade gneiss complex and they outcrop to the east of the pelitic rocks as a broad belt stretching from Taserssuaq to the southern part of the Tasiussaq area around Igdlukasîp tunua. The gneisses vary from homogeneous to veined and streaky in character (Fig. 8) with local areas of flecky gneiss. There is a common transition to granites through granitic gneiss.

On fresh surfaces the gneisses vary in colour from bluish grey and greenish grey to shades of light and dark grey but on weathered surfaces the rocks display shades of brown. In places the gneisses contain greenish feldspars and/or grey quartz, and in these places especially the gneisses

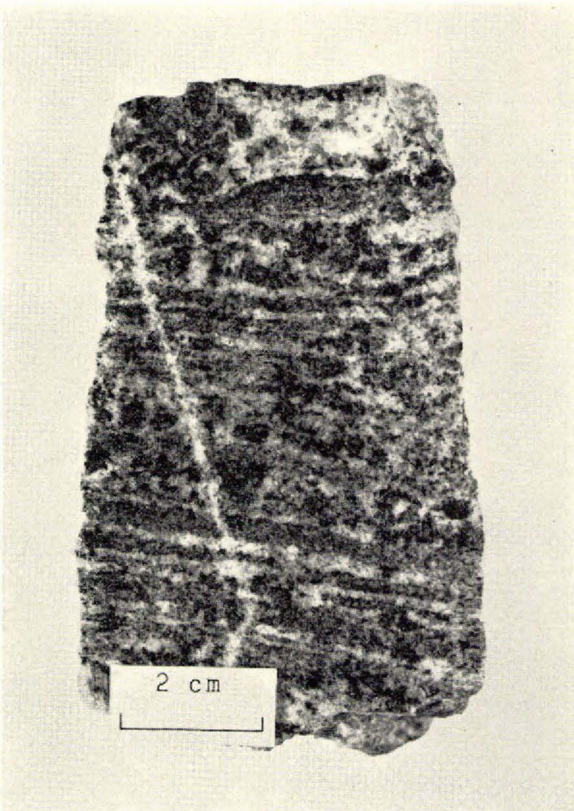

Fig. 8. Sample of charnockitic gneiss. The dark rounded areas are garnets which commonly exist in the leucocratic (greenish cream in colour) quartzo-feldspathic part of the rock which forms veins in the darker matrix. This matrix is bluish grey in colour containing abundant hypersthene. A later felsic vein cuts the rock. Sample GGU 67181.

have a waxy lustre. The gneisses are fine- to medium-grained and have granoblastic-xenomorphic textures which vary from eu- to hemigranoblastic (Plate 2b). Lepidoblastic textures characterise the fabrics rich in biotite. Blastopsammitic textures exist occasionally especially in types grading toward migmatised psammites. Some fabrics show a tendency towards granulitic texture.

Hypersthene is the main mafic mineral of the gneisses, being ubiquitous. It often occurs intimately associated with garnet (Plate 3a) and both minerals may display poikiloblastic texture (Plate 3b). Garnet frequently occurs in the migmatitic quartz-feldspar veins and streaks. Biotite is also a common mineral but it exists in much smaller amounts than in the pelitic rocks. Diopside occurs in some places, commonly along with hypersthene. Quartz, potash feldspar, perthite and plagioclase are the main felsic minerals.

The dominant mineral assemblage of the gneisses is the normal assemblage of charnockites: hypersthene-quartz-potash feldspar-plagioclase with or without garnet and biotite. Tracts in the gneisses are relatively deficient in potash feldspar and mineral assemblages characteristic of enderbites exist.

Mineral assemblages from the following: orthopyroxene, biotite, garnet, plagioclase, potash feldspar, perthite, quartz, clinopyroxene, myrmekite.

Accessory minerals: apatite, zircon, rutile, ore.

Alteration minerals: chlorite, saussurite, hornblende, actinolite-tremolite, clinozoisite, pistacite, sericite, antigorite.

Quartz schistose gneisses

Small occurrences of quartz schistose gneiss are found intercalated with other rock types, notably with the hypersthene-quartzo-feldspatic gneisses in the east and south of the area. The rocks form concordant layers usually having gradational relationships to the gneisses. To the west of Tasiussaq some quartz schistose gneisses exist which have a migmatitic character being penetrated by veins and streaks of felsic material while some schistose gneisses in the same locality seem to have formed by shearing of the unmigmatitic meta-arkoses.

The rocks vary from grey and light grey to cream coloured and from fine to medium in grain size. Textures vary from granoblastic in the more gneissose types to more grano-lepidoblastic in the schistose rocks. Quartz is a dominant mineral.

Mineral assemblages from the following: clinopyroxene, orthopyroxene, garnet, biotite, quartz, plagioclase, potash feldspar.

Accessory minerals: muscovite, zircon, apatite, ore.

Alteration minerals: chlorite, saussurite, sericite.

Migmatised psammites

Migmatised psammites occur as small intercalations in the hypersthene-quartzo-feldspathic gneisses particularly on the peninsula to the north-west of Frederiksdal. The rocks are composed of two components; a fine- to medium-grained rock matrix varying from grey to light grey in colour, and later migmatitic material (Fig. 9). The matrix often displays a millimetre-scale layering which is parallel to the foliation of the surrounding gneisses. This is interpreted as a palimpsest sedimentary banding. The migmatitic component is a fine-grained quartz-feldspar rock usually containing garnet and this forms veins, streaks and com-

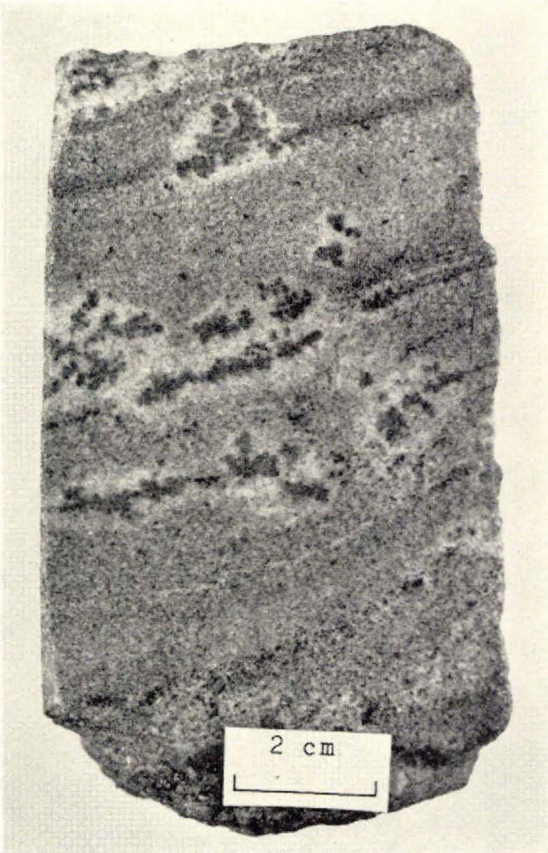

Fig. 9. Migmatised psammite. Flecks of felsic material (M_2) have developed in a grey psammitic matrix. The flecks truncate a layering which is interpreted as a relic sedimentary banding. Garnets occur in the flecks and they show some alignment parallel to the banding in the matrix. Sample GGU 54473.

monly flecks in the grey matrix. The felsic veins and streaks are subconcordant to the millimetre-scale layering but the felsic flecks and the garnets clearly truncate the layering.

Textures of the rocks are granoblastic-xenomorphic but the garnets are frequently idioblastic.

Mineral assemblages from the following: orthopyroxene, biotite, garnet, plagioclase, perthite, potash feldspar, myrmekite.

Accessory minerals: apatite, ore.

Alteration minerals: antigorite, chlorite.

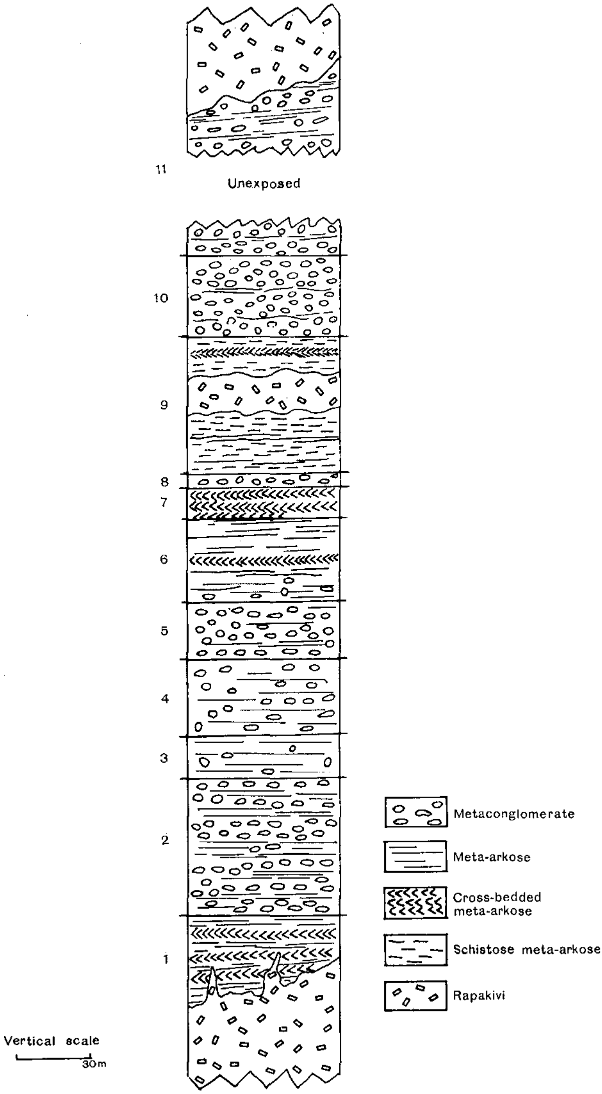

Fig. 10. General stratigraphy of a 350 m succession of metaconglomerate with intercalated meta-arkose. The succession forms part of a roof-zone inclusion in the biotite rapakivi granite of the granitic complex but cross-bedding in the meta-arkoses indicates that no repetitions exist in the succession. The following units occur: 1. Grey, banded, cross-bedded meta-arkose, 2. Dark metaconglomerate with some orange and grey meta-arkose bands, 3. Grey to light-grey meta-arkose with some metaconglomerate horizons, 4. Light-coloured metaconglomerate with intercalated meta-arkose layers, 5. Dark metaconglomerate with a few intercalated grey meta-arkose layers, 6. Grey meta-arkose with some bluish and orange intercalated layers. Some cross-bedding, 7. Cross-bedded grey meta-arkose, 8. Metaconglomerate, 9. Schistose meta-arkose with some grey cross-bedded meta-arkose, 10. Dark metaconglomerate with some dark grey meta-arkose horizons, 11. Light-coloured metaconglomerate with some grey, blue, green and orange meta-arkose layers. Tasermiut fjord coast, east of Qeqertaq.

Fig. 11. Cross-bedding in grey, banded meta-arkose. Bedding truncations of the units indicate that the youngest rocks occur to the left. Such well-preserved sedimentary features are characteristic of the psammitic and psephitic rocks in the northern part of the Tasiussaq area. Such rocks display a high-grade granulite facies mineralogy. Diameter of the lens-cap is 4.5 cm. Tasermiut fjord coast, south-southwest of Qeqertaq.

Meta-arkoses

The term "meta-arkose" is used here for metamorphic equivalents of PETTIJOHN's (1957) "arkosic sandstones" which include "arkose" and "subarkose or feldspathic sandstone" (*op. cit.*, p. 291). In all the unmigmatised psammites in the Tasiussaq area feldspar is an important mineral and although the quartz percentage of the rocks may be high, no true quartzites ("orthoquartzites" in the terminology of PETTIJOHN) exist. Some of the feldspar may be metamorphic in origin but even allowing for this, enough feldspar occurs with quartz forming the mesostasis of the rocks to conclude that the original sandstones were not pure.

Meta-arkoses occur in two areas, one to the west of Tasiussaq in a small outcrop and the other in the north of the area, where the rocks exist as inclusions in, and as a roof-zone to, the biotite rapakivi granite (see Map 1). To the west of Tasiussaq the meta-arkoses are associated with quartz schistose gneisses and in places with metavolcanic rocks while in the north they are intimately associated with metaconglomerates. North of the lake Taserssuaq a mass of intercalated meta-arkoses and metaconglomerates form a succession approximately 700 m thick. A section of over 350 m of similar rocks is given in figure 10 to illustrate the intercalated nature of meta-arkoses and metaconglomerates.

Fig. 12. Well-preserved cross-bedding in banded meta-arkose. The meta-arkose at this locality contains conspicuous garnets which are aligned parallel to the bedding. The angle of "cut-off" of the two units in the meta-arkose is clearly determinable. The diameter of the lens-cap is 4 cm. Tasermiut fjord coast, south-east of Qeqertaq.

The meta-arkoses are not migmatitic but in places they are reminiscent of "homogeneous gneisses". Sedimentary structures are represented by large- and small-scale cross-bedding which is exceptionally well preserved in places (Figs 11, 12), by occasional ripple marks, by slump structures and by disruptions in lithological bedding due to rock fragments. The cross-bedding provides evidence on the "way-up" of the meta-arkoses and this helps in the interpretation of the stratigraphy and local structure.

The meta-arkoses vary in colour from light grey and grey to pale shades of pink and green and commonly there is a colour banding in these colours. Some are banded by mafic layers and streaks, such layers being commonly rich in garnet. The meta-arkoses vary from fine to medium in grain size but in some places pebbles of quartzite up to a few centimetres in diameter may exist, usually disrupting the bedding. In places there occurs a transition from meta-arkose into metaconglomerate (especially the oligomictic metaconglomerates, see below) by increase in the abundance of pebbles or rock fragments; otherwise contacts between the two rock types are relatively sharp.

The macroscopic rock fragments in the meta-arkoses are seen microscopically as patches having differing grain sizes and occasionally composition, depicting the original form of the sandstone grains. The texture of the rocks is blastopsammitic varying from homeoblastic (eugranoblastic) to hemigranoblastic. A characteristic feature is the presence of a mesostasis of quartz and plagioclase in which clinopyroxene crystals are so arranged that the original outlines of the felsic grains are accentuated (Plate 4a). Diopside and garnet with some orthopyroxene are the main mafic minerals but biotite is sometimes present usually forming a foliation.

Mineral assemblages from the following: quartz, plagioclase, potash feldspar, perthite, orthopyroxene, garnet, clinopyroxene, biotite, epidote.

Accessory minerals: apatite, sphene, spinel, calcite, pyrite, haematite, ottrelite, zircon, ore.

Alteration minerals: hornblende, green biotite, sericite, clinochlore, penninite, pistacite.

Polymictic metaconglomerates

Polymictic metaconglomerates exist only in the northern part of the Tasiussaq area in the same localities as mentioned under meta-arkoses. The metaconglomerates are everywhere associated with intercalated meta-arkoses (Fig. 10) but single metaconglomerate units may reach a thickness of 200 m. The metaconglomerates are unmigmatitic but where occurring as inclusions and rafts in the biotite rapakivi granite they may be agmatised and veined by the granite.

The pebbles of the metaconglomerates vary in size from less than a centimetre to over 70 cm in diameter and they are characteristically rounded and smooth (Fig. 13). In places however, for example on the

Fig. 13. Polymictic metaconglomerate composed of elongated to rounded pebbles which vary both in size and composition. The conglomerates are extraformational in origin being composed of granitic, gneissic, meta-arkosic, quartzitic and basic pebbles (see p. 84). The chisel is 22 cm in length. Tasermiut fjord coast, east of the island Qeqertaq.

island Qeqertâraq in Tasermiut fjord, metaconglomerate composed of angular and fragmental pebbles occurs, the pebbles rarely reaching more than 10 cm in diameter (Fig. 14 and Plate 4b). Single beds within a metaconglomerate unit vary from less than 1 metre up to the largest recorded of approximately 75 m. In some beds a grading in pebble size is recognisable from base to top. Some beds are characterised by pebbles which have been severely stretched and elongated due to later deformation.

The overall colour of the metaconglomerates varies from grey, bluish grey and dark grey to light brown while the pebbles are composed of red, grey and leucocratic granites some of which are porphyritic,

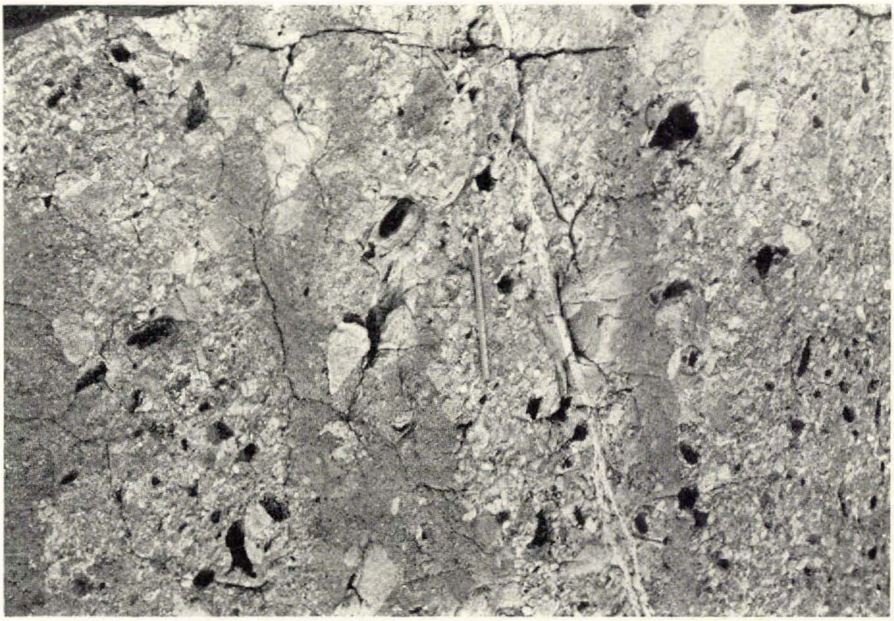

Fig. 14. Metaconglomerate composed mainly of meta-arkose pebbles, most of which are angular. Some granitic pebbles occur. Intercalated layers of meta-arkose exist. The metaconglomerate is cut by a late acid vein associated to the biotite rapakivi of the granitic complex. The island Qeqertâraq in Tasermiut fjord.

foliated grey and leucocratic granite and gneiss, light-coloured quartzite and meta-arkose, and different basic rocks. The matrix of the metaconglomerates is composed of quartz and feldspar with varying amounts of metamorphic pyroxene. Recrystallisation and production of metamorphic minerals have taken place and the original pre-metamorphic nature of the conglomerate has been superimposed by metamorphic pyroxene (hypersthene and diopside) without conspicuous alteration of the pre-metamorphic fabric. The textures are blastopsephitic (Plate 5a) and heteroblastic-xenomorphic.

The pebbles have the following main mineral assemblages: Granitic and gneissic: hornblende-microcline-plagioclase-quartz(-biotite); perthite-quartz(-hornblende-biotite); microcline-plagioclase-quartz. Quartzitic and meta-arkosic: quartz; plagioclase-quartz (-potash feldspar). Basic rocks: pyroxene-plagioclase(-quartz-biotite).

Mineral assemblages of the matrix from the following: quartz, plagioclase, potash feldspar, perthite, orthopyroxene, clinopyroxene, biotite, myrmekite, garnet.

Accessory minerals: apatite, spinel, calcite, zircon, sphene, hornblende, ore.

Alteration minerals: chlorite, hornblende, sericite, epidote.

Oligomictic metaconglomerates

A few metaconglomerates have a comparatively simple composition being composed only of quartzitic and meta-arkosic pebbles. Pebbles in these oligomictic metaconglomerates are typically rounded and smooth and they rarely reach more than 3 cm in size. The metaconglomerates form bands within the meta-arkoses in the north of the area and there is a gradation from metaconglomerate to meta-arkose with decrease in grain size. The rocks are well bedded and cross-bedding has been observed in places.

The rocks vary from grey and light grey to cream coloured. Textures are blastopsephitic and heteroblastic-xenomorphic.

Mineral assemblages from the following: quartz, plagioclase, potash feldspar, orthopyroxene, clinopyroxene.

Accessory minerals: calcite, biotite, apatite, sphene, ore, zircon.

Alteration minerals: sericite, hornblende.

Sulphide-bearing rocks

Sulphides occur as accessory minerals in the gneisses and schists derived from pelitic and psammitic rocks but in some places in both the pelitic rocks of the west (especially seen on the Tasermiut fjord coast) and in the hypersthene-quartzo-feldspathic gneisses of the east and north (*e.g.* south of the lake Taserssuaq, a locality mentioned earlier by JESSEN (1896), see page 11) sulphides are concentrated in certain layers. The layers characteristically have a rusty appearance due to the strong weathering. The actual layers containing sulphides vary in width from less than a metre up to tens of metres but due to the ochre colour of the weathered sections the layers appear thicker where screes and loose rock cover adjacent gneiss.

The main sulphides are pyrrhotite with lesser amounts of chalcopyrite and pyrite, the latter mineral being more common in some of the layers in the pelitic rocks of the west. Pyrrhotite in places has been replaced by pyrite and marcasite. Biotite is a common associate and garnet also occurs (Plate 5b). The rocks are fine- to medium-grained and textures vary from lepidoblastic in the biotite-rich rocks to granoblastic-xenomorphic in the layers in the hypersthene-quartzo-feldspathic gneisses.

Mineral assemblages from the following: pyrrhotite, chalcopyrite, pyrite, biotite, plagioclase, garnet, orthopyroxene, quartz.

Alteration minerals: sericite, saussurite, chlorite, epidote, marcasite.

Basic and ultramafic rocks

Two main groups of rocks can be separated here: those derived from volcanic rocks and those derived from igneous intrusive rocks.

Fig. 15. Metavolcanics in the outcrop to the west of Tasiussaq. Cavities and vesicles occur (centre left) and the lavas have an irregular appearance.

Some rocks of both groups are gneissic in character having been transformed and gneissified along with the paragneisses; other rocks, for example the cross-cutting pyroxene-metadolerite dykes, are later than the main gneissification.

Volcanic rocks

Volcanic rocks are represented by a restricted occurrence of unmigmatised lavas and pyroclastics with preserved primary structures and by migmatised basic schists which are interbedded with the pelitic paragneisses.

Metalavas and pyroclastics

One area of unquestionable metavolcanic rocks exists to the west of Tasiussaq. A thickness of at least 100 m occurs in the small outlier which also contains meta-arkoses and quartz schistose gneisses. In part the metavolcanic rocks contain some small units of meta-arkosic material. The metavolcanic rocks are characterised by a sequence of metalavas, pyroclastics and schists. The overall colour of the rocks is dark, varying from dark grey to black but locally lighter-coloured areas occur varying from grey to greyish green. The rocks are fine- to medium-grained.

Fig. 16. Rounded masses of lava simulating pillow structure. Some zoning is recognisable in the uppermost "pillow" (top centre). West of Tasiussaq.

The whole section of rocks has an irregular appearance with metalavas passing along the strike into agglomerate and schists. Vesicles and weathered-out cavities, together with rounded masses of lava which simulate pillow structure, help to produce the irregular character of the rocks (Figs 15, 16). Such pillow structures vary from grey and greenish grey to black but in some "pillows" the outer parts are lighter in colour than the central parts. They are commonly separated from each other by a darker-coloured matrix which weathers easily and forms cavities. The "pillows" range in size up to 1 m in length and they are relatively undeformed. Layers in the metalava vary in thickness from less than 20 cm to 5 m and these commonly display lighter-coloured borders (Fig. 17). Agglomerates are composed of rounded fragments of rock mostly above 15 cm in length, commonly within a lighter-coloured matrix.

Textures of the metalavas and pyroclastics vary from granoblastic-xenomorphic (Plate 6a) to lepidoblastic-hypautomorphic. The majority of the rocks have mineral assemblages composed from the following: brown hornblende, hypersthene, clinopyroxene, calcite, scapolite and plagioclase, but elsewhere, especially in some of the lighter-coloured pillow lava, secondary assemblages from the following minerals exist: green hornblende, pistacite, clinozoisite, calcite, tremolite, cummingtonite and plagioclase. Hornblende and epidote in some places can be

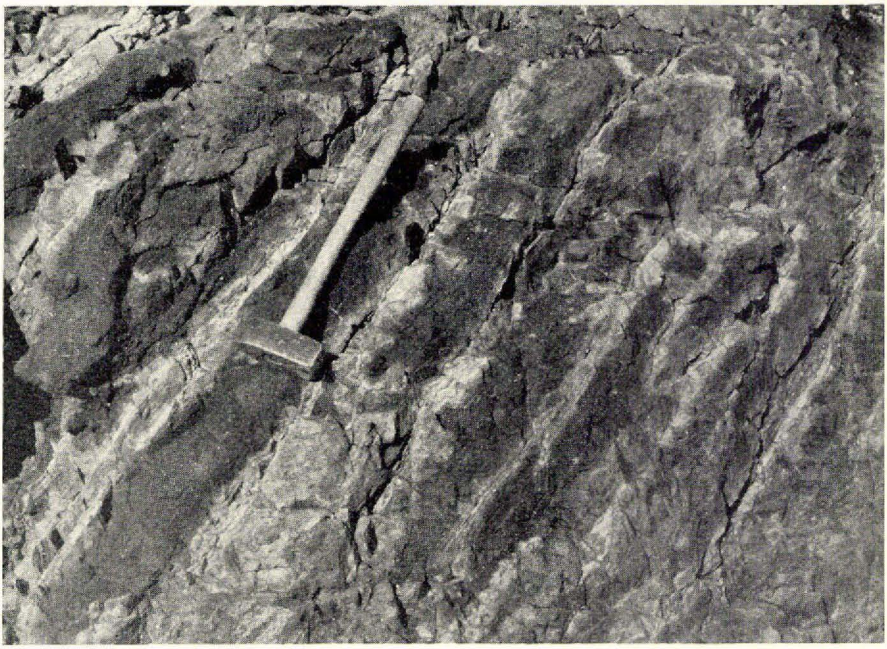

Fig. 17. Metalava showing lighter-coloured borders to the layers. West of Tasiussaq.

seen to be replacing pyroxene. Scapolite is locally common forming a matrix of large crystals in which pyroxene and/or hornblende are situated (Plate 6b). Accessory minerals are apatite, quartz and ore, and the main alteration products are saussurite, sericite, chlorite and hornblende.

Basic schists

Three main bands of basic schists occur within the pelitic paragneisses of the western part of the area. The bands have been deformed with the gneisses and they are now impersistent and broken up. The largest band reaches a thickness of 100 m. The schists have been migmatised in part, otherwise they appear very similar to the schists intercalated with the metavolcanic rocks to the west of Tasiussaq. No palimpsest volcanic structures have been seen in the schists. Schistosity is everywhere parallel to that of the enclosing gneisses.

The schists are dark grey to black in colour often with a dark green tint, and they are fine- to medium-grained. Textures are typically lepidoblastic-hypautomorphic but some rocks locally have granolepidoblastic to granoblastic-xenomorphic textures. The schists are characterised by the presence of brown and green hornblende, clinopyroxene and/or orthopyroxene and plagioclase; in composition the rocks are pyribolites.

Fig. 18. Two basic granulite sills within charnockitic gneiss. The sills converge and they are slightly discordant to the foliation of the gneiss. Migmatitic veins (immediately to the left of the hammer) parallel to the foliation of the gneiss pass through the sills. West of the islands in Igdlukasîp tunua.

Mineral assemblages from the following: hornblende, clinopyroxene, orthopyroxene, plagioclase, biotite, scapolite.

Accessory minerals: apatite, sphene, ore.

Alteration minerals: penninite, epidote, actinolitic hornblende, sericite.

Intrusive rocks

Two groups of rocks, one pyroxene-bearing the other dominantly amphibole-bearing, are considered to have been derived from intrusive igneous rocks. The pyroxene-bearing rocks are basic in composition and are described here as "basic granulites". The amphibole-bearing rocks contain little plagioclase and they are thought to represent small ultramafic intrusions.

Basic granulites

Basic granulites include the rocks already described elsewhere as pyriclasites and biotite-pyriclasites (Dawes, 1968), plus basic gneisses in which quartz percentage falls below 10 %.

The majority of basic granulites occur as sills in the paragneisses (Fig. 18) but some exist in the metavolcanic rocks to the west of Tasiu-

Fig. 19. Boudinage in basic granulite layer in charnockitic gneiss. East of Tasiussaq bay.

ssaq. Dykes, sharply discordant to the host gneisses, exist but are less common. The sills vary in thickness from a few centimetres to 4 m but two horizons are somewhat larger and reach up to 20 m. The smaller sills are commonly boudinaged and in places the sill-form can only be detected by the presence of isolated boudins in a particular horizon of gneiss (Fig. 19). Contacts between the gneiss and basic granulites are usually sharp and distinct. The sills have been folded and refolded with the gneisses (Fig. 27, and DAWES, 1968, figs 7, 8, 10) and they have been migmatised.

The quartz-poor gneisses, which are here classified as basic granulites, occur as transitions from the paragneisses. They are very minor in amount and only differ from the rock type of the sills in that they are darker in colour, are more gneissose and that they bear more migmatitic veins and streaks of felsic material. They are considered here for completeness although it is arguable whether they have been derived from basic igneous material.

The basic granulite sills vary in colour from dark grey to brown and pale green and they commonly contain a lustrous brown biotite which produces a foliation parallel to that of the host rock gneisses. The rocks are fine- to medium-grained and textures vary from eugranoblastic to hemigranoblastic, all being xenomorphic. Commonly the textures are directional and both plagioclase and pyroxene are elongated (Plate 7a). Biotite, where present, is aligned parallel to this elongation and it may form a foliation even within large pyroxene crystals (Plate 7b). Garnet and hypersthene are commonly poikiloblastic.

Mineral assemblages from the following: orthopyroxene, biotite, plagioclase (An_{60-75}), garnet, clinopyroxene, quartz.

Accessory minerals: sphene, apatite, ore.

Alteration minerals: green hornblende, sericite, penninite, pistacite.

Ultramafic rocks

Ten small ultramafic bodies occur within the pelitic gneisses of the west of the area and in the hypersthene quartzo-feldspathic gneisses of the east. The rocks form concordant to sub-concordant lenses in the gneisses, the largest of which reaches about 50 m in length. One occurrence at Naujat nûat consists of small agmatitic blocks in a granitic matrix. The rocks have a crude schistosity which is parallel to the foliation of the enclosing gneisses. Migmatitic features vary from small lenses, streaks and veins of felsic material to the more agmatitic veining as displayed at Naujat nûat.

The rocks are medium- to coarse-grained, green in colour and commonly have a pitted, fluted surface. Textures vary from eugranoblastic-hypautomorphic to grano-lepidoblastic-hypautomorphic. Hornblende is the dominant mineral of the ultramafics but it is joined by hypersthene in some rocks (Plate 8a) and in two samples hornblende is subsidiary to hypersthene. Plagioclase rarely reaches more than 5 %, mostly it is absent or replaced by sericite. Biotite occurs and it commonly replaces hornblende. Green hornblende replaces hypersthene.

Mineral assemblages from the following: hornblende, orthopyroxene, plagioclase, biotite.

Accessory minerals: apatite, ore.

Alteration minerals: sericite, chlorite, biotite.

Pyroxene-metadolerite dykes

These dykes have been described in detail elsewhere (DAWES, 1968) and a brief summary only is given here.

Approximately fifteen dykes exist varying in width from 20 cm to 5 m. Both NNW and NNE dykes have been recorded. The metadolerite varies in colour from dark brown to dark grey and in grain size from fine to medium. The form of the dykes varies according to the type and age of the host rocks. The dykes sharply cut the rocks of the high-grade gneiss complex and they appear as straight, sharp-bordered dykes, sometimes with chilled margins. In rocks affected by later reactivation and in rocks of the granitic complex, the dykes are not persistent but have been attacked and agmatised by acidic material. In the microcline granite of the granitic complex (G_4) folded pyroxene-metadolerites were noted. This deformation is probably connected to the development of the granite through reactivation of the gneisses of the high-grade complex.

The metadolerites are composed of three principal minerals: pyroxene (hypersthene and diopsidic augite), plagioclase (An_{45-60}) and biotite with minor amounts of quartz and hornblende in some dykes. Textures vary from sub-ophitic to relic sub-ophitic (blastophitic) and allotriomorphic-equigranular (granoblastic) with increased recrystallisation. The dykes are characterised by pyroxene in all stages of recrystallisation. Hornblende is only found in two dykes in small amounts replacing pyroxene. Biotite commonly replaces pyroxene and the mineral has a tendency to form in vein concentrations, suggestive of late crystallisation from potash-bearing fluid penetrations. Plagioclase is present as long laths in the rocks with sub-ophitic texture but has recrystallised to equigranular grains to produce the granoblastic types.

The metadolerites have a high-grade mineralogy and are considered to be genetically related to the high-grade gneiss complex. The dykes post-date the early granulite facies metamorphism of the complex and it has been argued that the dykes could have recrystallised under dipsenic conditions at thermal conditions below those of the early granulite facies metamorphism.

Granitic rocks

Included here are all the granitic rocks of the high-grade gneiss complex which can be distinguished in the field from the paragneisses and their granitic variants. During the field mapping emphasis was placed on the age relationships of the granitic rocks to the gneisses and to each other, rather than on rock composition, since even over short distances in one area of granite, composition can vary greatly. It also proved difficult (and later, often misleading) to apply strict names in the field to the charnockitic rocks since inconsistent colouration of feldspar and quartz often produces contrasted macroscopic characters to rocks with similar compositions. Furthermore, the quartz percentage of the charnockitic rocks proved difficult to estimate and a rock appearing rather basic in the field might well turn out to be within the granite range.

The oldest granites (G_1) and the younger G_2 granites vary in composition due to the variable ratio of plagioclase to potash feldspar and a compositional range from true granite to granodiorite exists. The youngest granitic rocks of the complex (G_3) do not vary markedly in mineral composition.

These three ages of granitic rocks can be recognised clearly on the peninsula to the north-west of Frederiksdal where they occur together although it is sometimes difficult to distinguish between G_1 and G_2 granites in the field especially where both are pyroxene-bearing. Further field work and systematic sampling would be needed to adequately portray the detailed compositional variations of the G_1 and G_2 granites.

1st generation granites — G_1 (enderbite, hypersthene granite)

The oldest granitic rocks of the complex are charnockitic in nature being characterised by hypersthene as the main mafic mineral. They correspond to rocks of the "acid" division in HOLLAND's (1900) "charnockite series" and to SUBRAMANIAM's (1959) "charnockite suite". Variation in the potash feldspar:plagioclase and perthite:antiperthite ratios produces a transition from hypersthene granite to hypersthene adamellite and to enderbite without conspicuous variation in megascopic characters, while locally, where the quartz percentage falls, hypersthene-quartz syenite, hypersthene monzonite and hypersthene-quartz diorite rocks occur.

The rocks form bands and irregular layers in the high-grade gneisses (Fig. 20). Some layers of a few metres in width have been noted in the gneisses but mainly the rocks form wider bands, a major one occurring on the peninsula to the north-west of Frederiksdal (see Map 1). This thins out rapidly to the north. Other main layers occur to the north of Kangikitsoq and to the east of Tasiussaq bay. Similar charnockitic rocks were sampled to the east of the Tasiussaq area on the west side of Torssukátak fjord and were noted around Augpilagtoq.

The bands are darker than the enclosing gneisses and they tend to form marker horizons. The rocks vary from grey, greyish green and bluish grey on fresh surfaces to shades of brown and fawn on weathered outcrops. Rock types are medium-grained and equigranular but some are foliated by clusters of pyroxene and/or biotite crystals. Textures are eugranoblastic and vary from hypautomorphic-granular to xenomorphic-granular.

Fig. 20. Layers and irregular patches of G_1 granite within a uniformly dipping sequence of charnockitic gneiss. Both gneiss and granite are cut by younger basic dykes (top left) and sills (centre and right). Vertical height from observer to the peak in centre background is approximately 1000 m. Eastern border of the Tasiussaq area, north-east of Kangikitsoq.

The following minerals exist in varying proportions: orthopyroxene, potash feldspar, plagioclase, perthite, antiperthite, myrmekite, quartz and biotite. Garnet occurs in some rocks. Accessories are apatite, zircon, and ore while sericite, chlorite and antigorite form the main alteration products.

The granites have concordant to sub-concordant relationships to the enclosing gneisses and they appear to have been folded with the gneisses. The rocks have been locally migmatised and bear in places small streaks and veins of felsic material, millimetres in width, parallel to any foliation in the rocks and usually parallel to the foliation of the gneisses. It is possible, although not readily demonstrable, that the bands of charnockitic rocks represent early igneous rocks having been affected later by folding, metamorphism and migmatisation. The rocks have been replaced by the G_2 and G_3 granites and are cut by the pyroxene-metadolerite dykes.

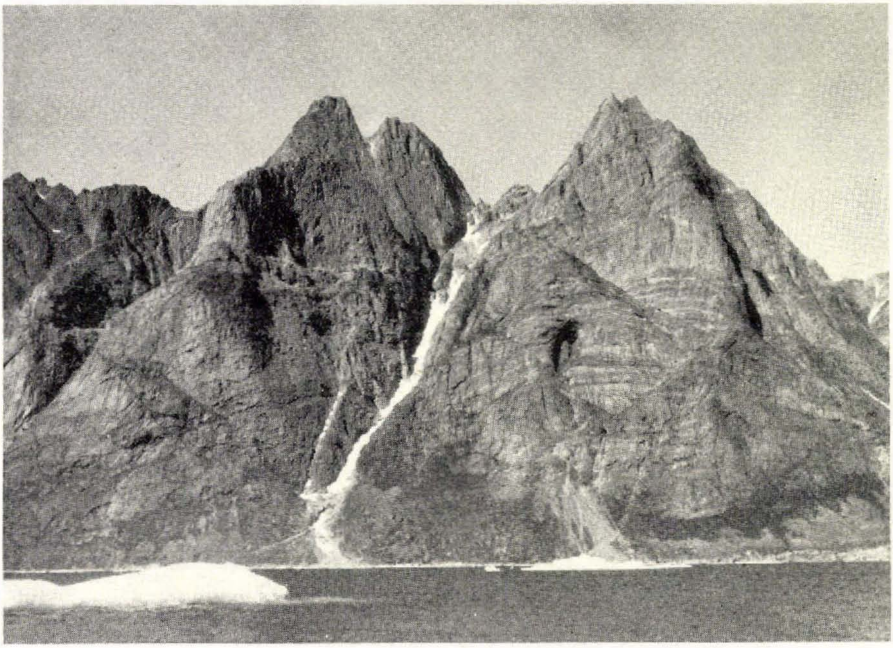

Fig. 21. G_2 granite conformably overlying charnockitic gneiss. The contact between the two dips gently to the left. The granite also veins the gneiss below the contact, and in the higher part of the mountain (top right) relics of gneiss occur in the granite. West side of Narssap sarqâ immediately north of the contact of the Frederiksdal rapakivi massif. Vertical height of exposure is 1000 m.

2nd generation granites — G_2 (pyroxene-, biotite-, garnet-bearing granites and granodiorites)

In composition G_2 rocks vary from granite to adamellite and granodiorite with fluctuating ratio of potash feldspar to plagioclase. Quartz percentage is always above 30 %. Biotite (often replaced by chlorite) and garnet are the main mafic minerals in the rocks although one or the other may be absent. Hypersthene also occurs in some of the rocks, in places to the exclusion of biotite or garnet so that some hypersthene granites of this age exist.

G_2 rocks occur mainly on the peninsula to the north-west of Frederiksdal (Fig. 21) and to the east and south-east of Tasiussaq bay, forming large masses and layers replacing hypersthene-quartzo-feldspathic gneisses. Rocks of possibly similar age occur in the core of the broad dome structure at Pingoq while porphyroblastesis of the pelitic gneisses in the higher parts of the dome is correlated with this granite phase. Commonly the granites contain *in situ* relic gneiss inclusions and in many areas on the peninsula to the north-west of Frederiksdal the amount of included material is high and the granite forms tracts within the

Fig. 22. Contact of G_2 granite and charnockitic gneiss. The gneisses conformably overlie the granite and both are veined by M_3 pegmatites. Height of vertical exposure in the centre of the picture is approximately 350 m. East of Igdlukasîp tunua.

gneiss. The contacts of the granites to the gneisses are concordant to sub-concordant and they vary from transitional to diffuse. The passage between gneiss and granite can take place over as much as 15 m; elsewhere the transition zone is only a few centimetres wide.

The granites are lighter in colour than the G_1 rocks and they vary from grey, bluish grey and buff coloured on fresh surfaces to light brown on weathered surfaces. The majority of the rocks are medium-grained and granulose but some coarse-grained porphyroblastic types occur. A megascopic foliation produced by flattened quartz grains, or clusters of quartz grains, within a felsic matrix (granulitic texture) is common and this is usually parallel to any preferred orientation of the mafic minerals. Textures are typically xenomorphic-granular. Quartz is commonly blue or bluish grey and feldspars vary in colour from cream to grey.

The following minerals occur in varying proportions: perthite, plagioclase, quartz, myrmekite, potash feldspar, biotite, garnet, orthopyroxene. Accessories are zircon, apatite, clinopyroxene, spinel and ore while sericite, chlorite, green biotite and epidote form the main alteration products.

The G_2 granitic rocks characteristically replace the high-grade gneisses and the G_1 layers in the gneisses, but they are cut by the G_3

Fig. 23. Relic strips of gneiss within garnet alaskite (G_3). The tracts have remained *in situ* during the replacement of the gneiss and contacts between the gneiss and the granitic rock are concordant to the foliation of the gneiss. The dark spots concentrated in rows parallel to the relic strips are garnets. Amitsuarssuk.

granites and associated pegmatite sheets (Fig. 22). Orthopyroxene and biotite occur in places as relic minerals from the gneisses in a state of replacement by plagioclase and quartz while elsewhere garnet and hypersthene appear to have recrystallised. (At the head of the valley overlooked by the cliff shown in figure 22 the hypersthene-quartzo-feldspathic gneisses overlie the granite with a rather sharp, flat-lying contact and no visible evidence exists of the granite replacing the gneisses. Although the contact has only been seen from the distance it raises a suspicion that parts of the G_2 granitic rocks might be older than the gneisses).

3rd generation granites — G_3 (alaskite and garnet alaskite)

G_3 granites have constant composition and are all leucocratic being composed essentially of quartz and feldspar. They are alaskites (Spurr, 1900) and garnet alaskites.

The granite forms elongated bodies within the high-grade gneisses on the peninsula to the north-west of Frederiksdal and also occurs in the core of an anticline. These bodies merge into layers of pegmatite which "vein" the gneisses and in places produce a banded appearance to the gneisses. The alaskites have no foliation apart from parallel inclusions of gneiss which in places are abundant and which occur as relic strips and ribbons which have an orientation parallel to the foliation of the gneisses. Many appear in a state of transformation and the alaskite in such cases is commonly rich in garnets (Fig. 23). Contacts of the alaskites to the gneisses are dominantly concordant to sub-concordant but some discordant contacts exist. In detail they vary from sharp to transitional.

The alaskites are medium to coarse in grain size and white in colour. The garnet alaskites frequently have a spotted appearance due to the garnets which reach up to 1.5 cm in diameter. Quartz varies from colourless to grey and bluish grey and the feldspars are white varying to cream coloured. Textures are xenomorphic-granular but in places perthitic megacrysts produce hemigranoblastic to porphyroblastic textures. Main minerals are quartz, potash feldspar and perthite with lesser amounts of myrmekite and plagioclase. Garnet is common, in places in abundance. Biotite exists but never in large amounts. Accessories are apatite, zircon, spinel, pyroxene and ore but tourmaline and beryl have been noted in pegmatites of the same age. Alteration products include sericite, epidote and chlorite.

The alaskites replace both the high-grade gneisses and the G_1 and G_2 rocks being the result of a "microclinisation" of the older rocks. In part this process has been controlled by the structure of the replaced rocks.

Migmatisation of the complex

Extent of migmatisation

Migmatisation has been widespread in the complex and the majority of the paragneisses are migmatised to some degree. However the meta-arkoses and metaconglomerates in the northern part of the area occurring as inclusions in, and as a roof-zone to, the biotite rapakivi granite, and some of the metalavas to the west of Tasiussaq, have escaped migmatisation and two units can be recognised: a migmatised unit composed of the hypersthene-quartzo-feldspathic gneisses and pelitic rocks with associated basic granulites, basic schists, calc-silicate granulites and ultramafic rocks, and a non-migmatised unit composed of the meta-arkoses, metaconglomerates and the metavolcanics.

These two units can only be distinguished in general terms since some variations in severity of the migmatisation occur in the migmatised unit while some migmatitic effects are seen in some rocks of the non-migmatised unit. For example slightly migmatised psammites occur in the hypersthene-quartzo-feldspathic gneiss terrain while parts of the

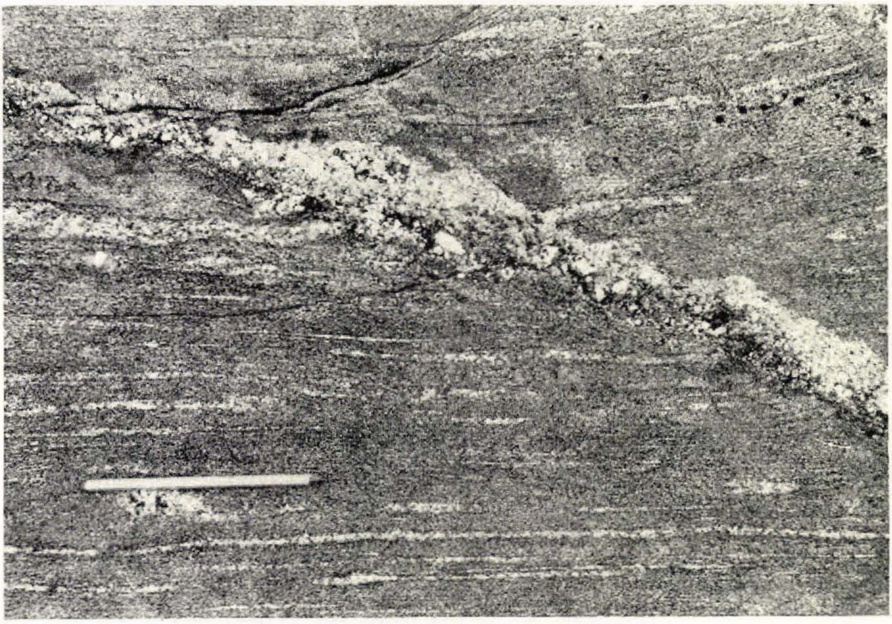

Fig. 24. Migmatisation of pelitic rock. M_1 quartzo-feldspathic veins and streaks occur parallel to the biotite foliation of the rock while a later discordant M_2 vein truncates the foliation and the M_1 material. Coast, south-east of Pingoq.

biotite schists are unmigmatitic. Conversely, some of the metalavas and meta-arkoses in the outlier to the west of Tasiussaq are migmatised in places.

The occurrence of the meta-arkoses, metaconglomerates and metalavas is too restricted to allow a detailed statement on the relationship of the migmatised to unmigmatised rocks. However the unmigmatised meta-arkoses and metaconglomerates do represent a higher stratigraphical unit than the majority of the migmatised gneisses and schists suggesting a lower migmatised unit and an upper non-migmatised unit. The meta-arkoses and metavolcanics occurring to the west of Tasiussaq are considered tentatively as members of the upper unit by comparison with the stratigraphical relationships demonstrated by Escher (1966) from the area to the north-west of Tasermiut fjord. These two units recognised in the Tasiussaq complex correspond to the two structural units distinguished (p. 51), the possible significance of which is discussed later (pp. 59,78).

Phases of migmatisation

In the pelitic rocks of the west three phases of migmatisation can be distinguished, the last two of which can be correlated with the forma-

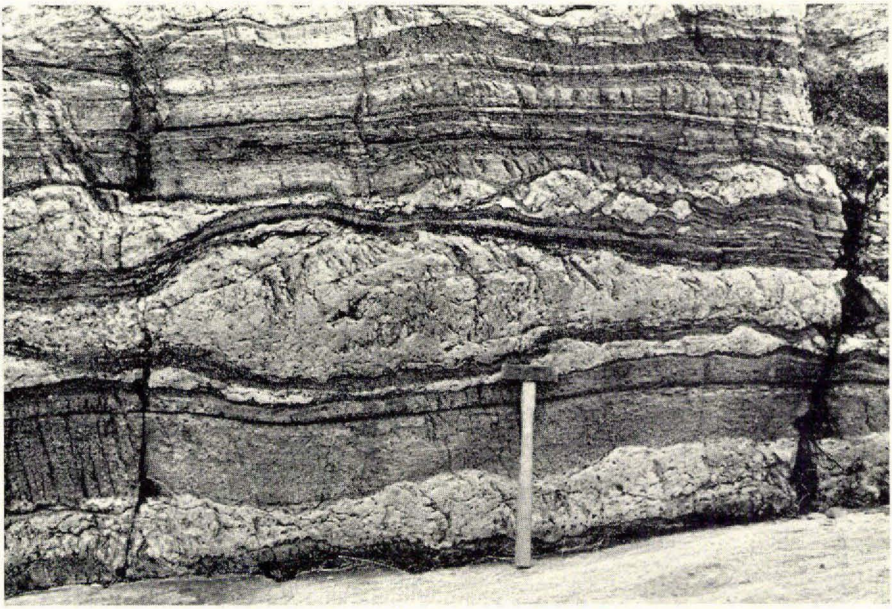

Fig. 25. Biotite gneiss displaying pinch-and-swell structure in M_2 quartzo-feldspathic material. East of Pingoq.

tion of the G_2 and G_3 granites (see p. 70). Elsewhere only one or two phases of migmatisation are clearly recognisable.

The earliest signs of migmatisation are small streaks, veins and flecks of quartzo-feldspathic material which are characteristically concordant to the foliation of the host rocks (Fig. 24). The veins and streaks are impersistent and rarely reach more than 2 cm in width. The material is fine-grained and is composed of plagioclase and quartz with smaller amounts of potash feldspar. All the veins, streaks and flecks are considered to have been formed during the first phase of migmatisation (M_1) although the time when this type of migmatisation affected the rocks may have varied in different parts of the complex.

A second phase of migmatisation (M_2) resulted in larger layers or veins of aplitic to pegmatitic material varying from a few centimetres to 1 m in width. These M_2 veins truncate the M_1 veins and streaks. They vary from concordant to sub-concordant to the foliation of the host rocks but some follow fracture-zones and are discordant (Fig. 24). Some flecks of aplitic material occur in the hypersthene-quartzo-feldspathic gneisses of the south and produce local areas of flecky gneisses, in appearance like those described by LOBERG (1963). Garnet may occur in the core of the flecks (Fig. 9). Pinch-and-swell structure (Fig. 25) is a common feature of the M_2 veins and many cases exist where the presence of feldspar has affected the foliation of the host rocks either

Fig. 26. Irregular development of M_2 quartzo-feldspathic material producing agmatitic gneiss. Garnet occurs in the felsic material. Tasermiut fjord coast, east of Usuk.

through actual growth of the feldspar or by subsequent deformation. Isolated areas of augen gneiss (Fig. 6) and agmatitic gneiss (Fig. 26) occur. The main minerals of the material are quartz, albitic plagioclase and potash feldspar with or without garnet and biotite.

A third phase of migmatisation (M_3) produced layers of pegmatite which vary from less than 1 m to 75 m in thickness together with irregular patches and these pegmatites are genetically related to the formation of the alaskites (G_3). Where concentrated, the layers produce areas of banded gneiss such as at Pingoq (Fig. 5). The layers are commonly concordant to the foliation of the host rocks and some display pinch-and-swell structure. Others are distinctly discordant. The layers cut the F_2 and F_3 folds and also G_2 (Fig. 22) and M_2 veins and layers. The pegmatite varies from medium-coarse- to very coarse-grained, with occasional feldspars reaching 30 cm in length. Main minerals are quartz, potash feldspar with smaller amounts of albitic plagioclase and garnet. Accessories are beryl, tourmaline, magnetite, muscovite and biotite.

A few orthopyroxene grains have been noted in some of the M_2 and M_3 migmatitic veins and hypersthene exists in the G_2 granite. However, true hypersthene-bearing acid veins (charnockite veins) occur in places in the gneisses and have especially been seen cutting the metalavas to the west of Tasiussaq. The veins are small, varying up to 5 cm in width, and they are generally discordant to the foliation of the host rocks. The veins are grey to greenish in colour and have a waxy appearance. The hypersthene occurs in crystals up to 5 mm in length and is situated in a white to cream-coloured matrix of quartz and feldspar (Plate 8b). The age of such charnockite veins is not clear but one vein has been seen to cut M_1 felsic streaks. The charnockite is composed essentially of hypersthene, which is pink and pleochroic, potash feldspar (microcline, which is in places perthitic), plagioclase, which has been sericitised, and quartz. Some myrmekite and biotite exist along with some ore. Some hypersthene crystals have been altered in thin peripheral zones to chlorite and uralite. Such charnockite veins in places are cut by chlorite veins up to 1 cm in width.

Nature of migmatisation

The main process of the migmatisation appears to have been migration of potash, silica and sodium from the host rocks into zones either parallel to the internal structure of the rocks (foliation) or into zones controlled by tectonic dislocations. How much fusion, melting and recrystallisation has occurred in material having a lower melting point than the host rocks is not clear but it seems likely to have occurred on a small scale. It is also possible that the charnockite veins may have originated in this way and that they were intrusive into the host rocks.

The M_1 veins and streaks are interpreted as "secretion aplites" formed by the diffusion of felsic material into tracts parallel to the foliation during regional metamorphism. RAMBERG (1949) refers to such a process as "chemical squeezing". The M_2 veins differ in their genesis from the M_1 veins and they are interpreted as "concretion pegmatites" (RAMBERG, 1949) having grown in the solid state by pushing aside the host rocks. This can be readily observed in the development of augen or porpyroblasts in the gneisses (Fig. 6). The majority of the larger M_3 pegmatites are regarded as "replacement pegmatites" and are considered to have replaced the host rocks *in situ.* Evidence for this is provided by the relic strips and ribbons of gneissic material remaining within the layers and the concentration of garnets in strips in places depicting the original rock composition and structure of the replaced rocks. Some of the more regularly shaped layers in the gneisses might have been formed by some fluid intrusion into the gneisses.

The relationship between the migmatisation, metamorphism and deformation is treated later (p. 70).

Structure of the complex

General structure

The trend pattern of the regional foliation of the complex is relatively simple and directions vary from N–S to E–W but are restricted to the NE quadrant. Characteristic of the complex, especially evident in the east, is the occurrence of large tracts of shallow dipping gneisses resulting from large-scale recumbent isoclinal folding. Few linear structures occur. Tectonic structures vary in style and intensity according to different rock types and two structural units can be recognised. No palimpsest sedimentary discordances have been recognised in the gneisses of the complex but an isolated outcrop of meta-arkoses and metavolcanics of an upper unit lies with possible structural discordance on pelitic gneisses of a lower unit. The possible significance of this outcrop is discussed later (pp. 60, 76).

The rocks

The rocks most visibly affected by deformation are the pelitic and hypersthene-quartzo-feldspathic gneisses together with the included calc-silicate granulites, basic schists, ultramafic rocks, and basic granulites; these rocks form a supposed lower structural unit than the meta-arkoses, metaconglomerates and metavolcanics which are relatively undeformed and which represent an upper unit. Small-scale folds are very scarce in the upper unit occurring only in the psammitic rocks in the lower parts of the outlier mentioned above and, although no evidence has been seen for them, large-scale folds may exist. The pelitic and hypersthene-quartzo-feldspathic gneisses have been severely folded and the rocks display both small- and large-scale folds. Concentration of folds in some areas produce areas of "small-folded gneiss".

The relationship of the pelitic and hypersthene-quartzo-feldspathic gneisses to the rocks of the upper unit (meta-arkoses, metaconglomerates and metavolcanics) is obscure. In the north of the area where the largest area of meta-arkoses and metaconglomerates exists the contact with the gneisses has been destroyed by the intrusion of the younger biotite rapakivi granite. The only other area of rocks of the upper unit is in the small outlier to the west of Tasiussaq. Here there is a thickness of about 120 m of meta-arkoses (including banded meta-arkose and quartz schistose gneiss) in places mixed with metavolcanic material. About 100 m of metavolcanic rocks occur overlying the meta-arkoses. Field relationships between the meta-arkoses and the underlying pelitic gneisses are not particularly well exposed but crushed, disrupted and sheared meta-arkoses exist in many places. On the southern side of the outlier, the meta-arkoses are in places slightly migmatitic and there is a transition

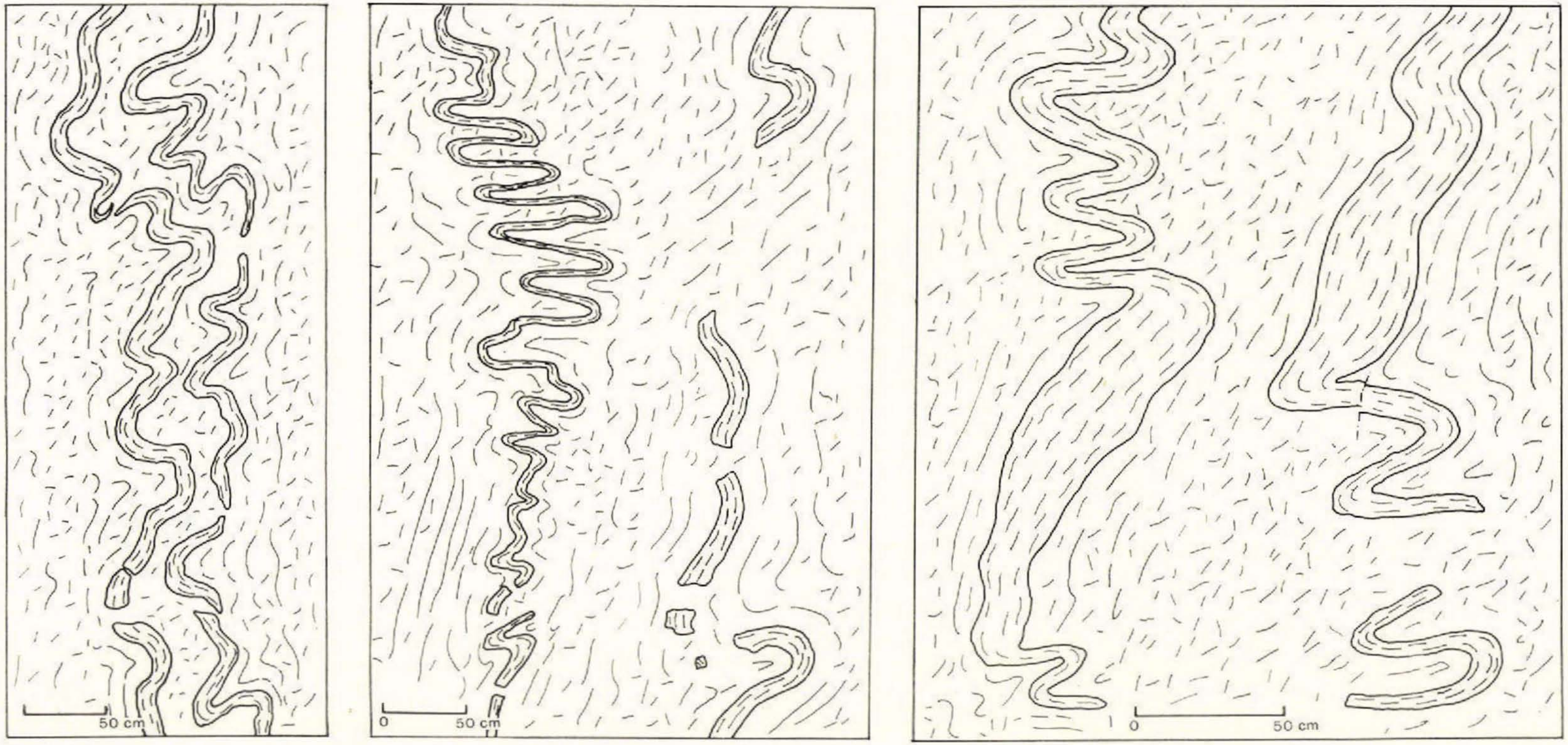

Fig. 27. Composite diagram of basic granulite layers in charnockitic gneiss illustrating the relationship between thickness and style of folding. The three parts of the diagram represent the same basic granulite layers at different localities along the strike. The two layers in fact represent the two limbs of an F_2 isoclinal fold, the visible small-scale folds being F_3I structures. The charnockitic gneisses are commonly granitic in nature, and folds, and even foliation, may be difficult to discern. In such cases the basic granulite sills are indispensable in determining the local structure, although fold style in the sills, which varies with thickness from open to close, may not correspond to that in the gneisses. In this case the gneiss is steeply inclined to vertical in attitude and the F_3I folds have flat-lying to shallow-dipping axial planes. Such folds appear to be the result of a flattening deformation. East of Igdlukasîp tunua.

to quartz schistose gneiss and gneiss in which zone some folding occurs. The surrounding gneisses however show little structural sympathy with the outcrop which is tentatively interpreted as a small tectonic outlier.

Many of the hypersthene-quartzo-feldspathic gneisses forming the main rock type of the complex are granitic in character and in many places little or no foliation exists in the rocks. Furthermore, large areas of the gneisses, especially on the peninsula to the north-west of Frederiksdal, have been replaced by the G_2 and G_3 granites and only ghosts of gneissic rocks remain. The gneisses are deficient in good marker horizons. Rust-zones are thin and impersistent and no stratigraphical horizons can be accurately traced.

The most valuable "markers" are the basic granulite sills in the gneisses and these are invaluable in deciphering local structure especially in granitic gneiss terrain where they appear distinct. However, the fold style depicted by them does not necessarily correspond to the style of folding in the gneisses, and sills of different thicknesses have developed different fold styles (Fig. 27).

The obvious difficulties of measuring fold trends and axes, and of interpreting structures, cannot be stressed enough in high-grade granitic gneiss terrain devoid of marker horizons.

Structural elements

Foliation

The regional foliation of the gneisses is marked by a set of planar surfaces (S_1) due to the preferred orientation of minerals. This fabric coincides with the original sedimentary and volcanic layering. Compositional variations in the gneisses are parallel to this foliation and migmatitic streaks and veins (M_1) also follow S_1. The regional foliation has a general NE–SW strike varying to NNE–SSW and N–S in the west of the area around Tasermiut fjord to ENE–WSE and locally to E–W in certain parts of the eastern area. The NE–SW direction of S_1 is visibly altered by later folding and this is well seen on the peninsula to the north-west of Frederiksdal.

Primary structures

Primary structures occur in the meta-arkoses, metaconglomerates and metalavas. While theoretically such sedimentary and volcanic structures have both qualitative and quantitative uses in structural analysis, they only occur in rocks which form small isolated units which are not intercalated with the gneisses of the complex. Thus their use as indicators of amount and type of stress affecting the complex generally is limited.

However, the primary structures are of considerable use locally in estimating the type of stress affecting the rocks of the upper unit and also in determining details of the pre-metamorphic stratigraphy of the rocks. Cross-bedding in the meta-arkoses and metaconglomerates provides a means of determining the "way up" of the rocks, proving exceptionally useful since much of the meta-arkosic material occurs as isolated areas within the biotite rapakivi granite.

Pebbles in the metaconglomerates and primary structures in the metalavas reveal that these rocks are relatively undeformed compared to the gneisses and they appear to form part of a different structural level. However, two factors complicate the use of pebbles for quantitative analysis. Firstly, there is clear variation in the original shape of the pebbles in the metaconglomerates; some are angular and irregular while others, elsewhere, are rounded and smooth. Secondly, the majority of the metaconglomerates are polymictic. Any consideration of stress conditions based on the deformed shape of pebbles must take into account this initial variation in shape and composition, and must be based on an assumed shape, and as OFTEDAHL has stated (1948, p. 486), "no conglomerates have pebbles with a shape which mathematically is a sphere".

Much of the metaconglomerate and intercalated meta-arkose succession shows little sign of deformation but it is apparent that some parts of the succession have been deformed more strongly than others. Even allowing for the control of shape by composition, pebbles vary from being rounded and ellipsoidal to markedly elongated and squashed. This elongation and flattening has taken place parallel to the bedding of the metasediments. The effects of the deformation have been concentrated in zones so that zones of conglomerate containing markedly elongated and flattened pebbles occur within less-deformed or non-deformed rock. The meta-arkoses and metaconglomerates have proved resistant to deformation as shown by the deficiency of folds and it seems that the stress has been released by a shearing parallel to bedding which has resulted in the elongation of pebbles and their development in zones.

Boudinage

Boudinage is a common feature in the basic granulite sills and in the calc-silicate granulite horizons which occur in the hypersthene-quartzo-feldspathic gneisses of the complex. The correlation of boudinage to phases of deformation is difficult since a sill or concordant layer in the gneisses may have suffered either boudinage or folding depending on its orientation relative to the deforming forces.

Shearing

Shearing was more common in the late phases of deformation although the presence of shear-planes in the gneisses filled by later felsic material demonstrates the existence of early movement-zones. In places shearing has produced wider zones of sheared rock separating disorientated blocks of gneiss (tectonic mélange).

Folds

Folds vary from small-scale, with wave-lengths and amplitudes of a few centimetres, to large-scale structures having amplitudes of up to 3 km. In attitude the folds vary from upright to overturned and flat-lying, and in type from open to isoclinal (recumbent). One nappe-like structure has been observed.

The folds exist mainly in the rocks of the lower structural level, *i.e.* pelitic rocks and the hypersthene-quartzo-feldspathic gneisses and associated rocks, where at least four phases of folding can be recognised. Some small-scale intrafoliational folds noted in the pelitic and hypersthene-quartzo-feldspathic gneisses are indicative of an early F_1 phase of folding but some of these folds could also be palimpsest structures due to slumping during sedimentation. Limbs of the folds are usually sheared parallel to foliation (S_1) so that they appear isolated. S_1 is usually parallel to the axial surfaces of such folds. A relic pre-S_1 foliation is represented in such folds, being truncated by S_1. Little is known about this fabric which may be a relic sedimentary fabric (S) or even a metamorphic schistosity (S_0) parallel to S.

F_2 folds are the most important structures in the complex and they have affected the regional foliation (S_1) producing the general NE trend of the complex and areas, particularly towards the east, of flat-lying to shallow-dipping structure. Both small- and large-scale folds exist having axes varying from NNE–SSW to NE–SW. Due to later F_3 folding these axes in places vary in direction as for example on the peninsula to the north-west of Frederiksdal. Axes, now generally shallow, plunge either to the NE or SW due to the later F_3 folding. F_2 folds vary from open to close and isoclinal, and from upright to overturned and recumbent. The overturning is invariably to the south-east (Fig. 28) so that axial planes show a range in dip from vertical through dips to the NW to flat-lying. Small-scale folds are commonly isoclinal (Fig. 29) and occasionally chevron in style. Many are plastic to semi-plastic in type (Fig. 30).

A series of well-exposed folds occurs in the high mountains on the east coast of Igdlukasîp tunua, the largest fold being the northernmost one. This fold is an isolated recumbent fold or nappe-like structure with an amplitude of over 3 km and a broad south-easterly-facing hinge

Fig. 28. Suggested large-scale F_2 isocline in hypersthene-garnet gneiss. Sub-concordant pegmatites indicate the general attitude of the gneiss foliation, a thicker pegmatite having developed in the axial-plane region of the fold. The structure is overturned towards the south-east and has a north-westerly dipping axial plane. View from the helicopter looking north-east over Kangikitsoq towards lake 340 m (right centre). The summit of the mountain displaying the fold (upper centre) is 1080 m. The highest peak in the mountain range to the right of this summit in the background reaches 1580 m while that in the left background reaches 1763 m. August 3rd, 1963.

(Fig. 31). The mountain Natsingnat (1080 m) is formed from the single fold, the summit lying on the upper limb, the lower overturned limb forming the lower parts of the mountain. The structure seems to have been "pushed" towards the south-east so that immediately in front of the hinge-zone the gneisses have an anomalous structural pattern. Folds farther away to the south-south-east in the section are open to isoclinal in style but not conspicuously overturned.

F_3 folds are represented by two types of folds which both post-date the F_2 folding but which in fact may differ somewhat in age. They are here denoted as F_3I and F_3II folds. Some plastic type small-scale folds (F_3I) have been recognised refolding the basic granulite sills in the hypersthene-quartzo-feldspathic gneisses (Fig. 27). These folds have been formed by a flattening of the isoclinally folded gneiss pile and are only of a small scale. The axes of the folds vary according to the trend of the basic granulite sills in the gneisses resulting from the F_2 folding, (see p. 60). Amplitudes of the folds are all below 3 m. Wave-lengths vary up to 4 m but the majority are below 2 m.

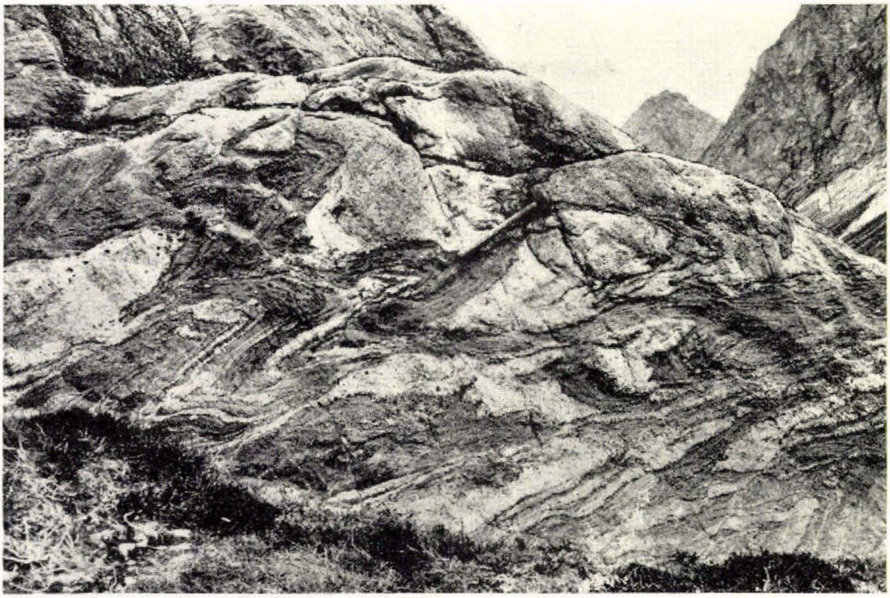

Fig. 29. Small-scale F_2 isoclinal folds in hypersthene-quartzo-feldspathic gneiss. Such folds are characteristically overturned to the south-east and here have axial planes parallel to the hammer shaft. South coast of Amitsuarssuk.

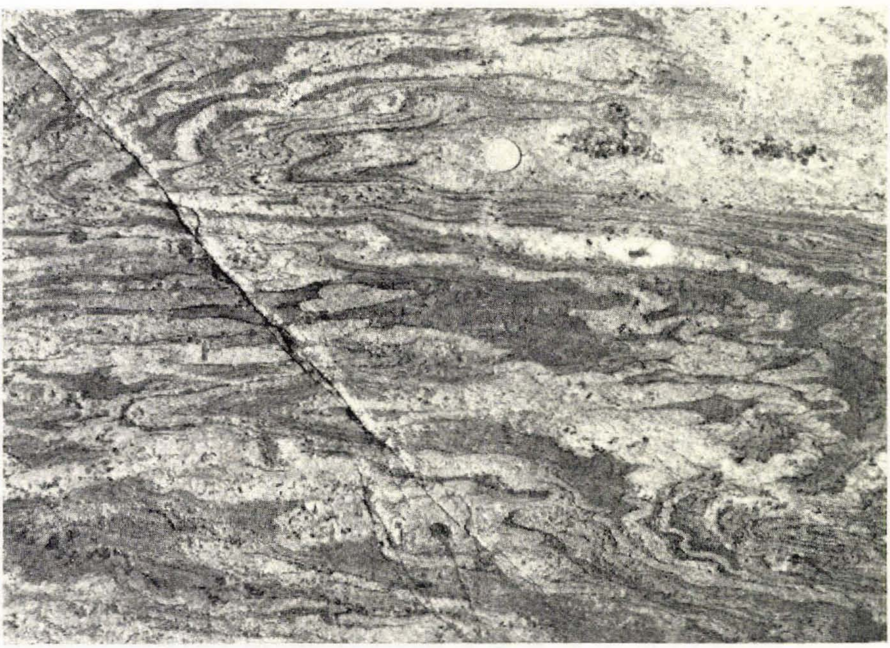

Fig. 30. Small-scale F_2 isoclinal folds in charnockitic gneiss characterised by a plastic style. The diameter of the coin is 2.8 cm. West coast of Kangikitsoq.

Fig. 31. Large-scale F_2 recumbent isocline in hypersthene-garnet gneiss in the western face of Natsingnat (1080 m). The flat-lying axial plane of the fold is parallel to regional foliation which to the east of Tasiussaq is flat-lying to shallow-dipping. Pegmatites, sub-concordant to the foliation, outline the fold. View taken from the helicopter looking approximately north-east along the axis of the structure. The height from the sea to the crest of the ridge on the left is approximately 350 m.

Other F_3 folds (F_3II) are open in style and vary from small- to large-scale (Fig. 4). Some structures are characterised by the presence of well developed sets of parasitic folds on the limbs. Both large- and small-scale folds have been seen refolding F_2 isoclinal folds; where the folding has been superimposed on flat-lying isoclinal F_2 folds, open antiforms and synforms with general NW axes have developed, but where the F_2 folds were originally more open in style, F_3 domes and basins have been produced by the superimposed folding. A shallow dome structure is present in the extreme west of the area at Pingoq while a basin structure exists in Putôrugtoq mountain to the south of Taserssuaq where dips of the structure reach up to 40°.

F_4 folds are represented by a few broad, shallow folds up to 2.5 km in wave-length. Such folds are well seen in the high dissected terrain in the eastern part of the area especially north of Kangikitsoq. Axes are difficult to estimate due to the large scale of the folds but they seem to vary from E–W to NE–SW.

Structural development

It is not known whether gneisses of more than one age exist in the Tasiussaq area (see p. 85) and thus an interpretation of the structural development of the complex has some limitations. The two structural units recognised, a lower folded unit composed of the pelitic and hypersthene-quartzo-feldspathic gneisses and an upper unit composed of the meta-arkoses, metaconglomerates and metavolcanics, may represent two tectonic units of a single supracrustal pile or they may be composed of rocks of two different ages, so that the meta-arkoses, metaconglomerates and the metavolcanics (with or without certain parts of the gneisses) represent part of a supracrustal series younger than the underlying gneisses. In the latter case it is theoretically possible that some structures recognised in the gneisses pre-date the deposition of the rocks of the upper unit, but on the other hand, the absence of one type of structure in the upper unit and its presence in the rocks of the lower unit cannot be used as evidence that the upper unit is of a younger age. Such a situation can be expected in the development of two tectonic units from a single supracrustal pile. Evidence is inconclusive and equivocal, and the exact relationship between the F_2, F_3 and F_4 fold phases and the rocks of the upper unit is not clear. The importance of the few F_1 folds represented by small-scale intrafoliational folds may be greater than previously recognised in this part of South Greenland. The folds may provide evidence for the presence of older crystalline rocks in the complex, but since very few of these folds have been observed, interpretations based on them are essentially tentative.

The main recognisable phase of folding (F_2) resulted in repetitions in the rock succession. Large-scale recumbent folds can be detected and in many parts of the gneiss terrain it is possible to recognise areas of gneiss which are overturned. Such repetitions are not revealed in the geological map (Map 1) since these major structures have all been detected in the hypersthene-quartzo-feldspathic charnockitic gneisses, which form a monotonous rock sequence devoid of a mappable stratigraphy. Whether such structures exist in the meta-arkoses of the upper unit is not known. On a small scale the repetitions are detectable by the occurrence of basic granulite sills in pairs (Fig. 27).

The main stress component during this F_2 folding was orientated east–west to north-west–south-east since isoclinal folds, both small- and large-scale have NNE to NE axes and are overturned to the east and south-east, so that the hinges of recumbent folds are east- or south-east-facing. During this folding some thrusting and shearing in the higher, more brittle parts of the rock succession probably occurred and this may have produced the sheared nature of some of the metaconglomerates and the elongation and deformation of the pebbles in certain zones in these rocks (see p. 54). Furthermore, this thrusting may have resulted in the rocks of the upper unit having been carried over the pelitic and hypersthene-quartzo-feldspathic gneisses. Such a tectonic transport may have been connected to large-scale nappe-like structures, the presence of which in the gneisses indicates a tectonic transport towards the east and south-east.

The small outlier of meta-arkoses and metavolcanics occurring in the Tasiussaq area to the west of Tasiussaq might represent a relic of such a large overriding cover, the lower parts of which were folded and disrupted during the movement. Such folding exists in the contact area of the meta-arkoses with the gneisses, a contact which has subsequently been migmatised. However, the exact age of such a thrusting is not known and movement of the upper unit may have occurred or continued to occur during later folding phases of the complex. Furthermore, how far the present situation of the outlier is due to down-folding is not clear.

In places it is clear that this F_2 folding affected rocks which were semi-plastic to plastic in nature, resulting in the plastic type of some small-scale folds in the gneisses (Fig. 30). This tendency for plastic conditions in the hypersthene-quartzo-feldspathic gneisses may have also produced a set of small-scale folds (F_3I) which probably pre-date the development of the NW-trending F_3 folds (F_3II). Evidence for this is provided by the basic granulite sills within the gneisses which, as a result of the F_2 folding, show a tendency to appear in pairs. These pairs of sills vary in attitude from flat-lying where recumbent folds exist to vertical where upright isoclinal folds exist. The thickened pile of rocks (due to the F_2 folding) would during the prevalence of plastic or semi-plastic conditions tend to be flattened or squeezed to regain structural stability. This flattening would tend to produce boudinage in the flat-lying to shallow-dipping basic granulite sills and folds in the vertical or steeply inclined sills (Fig. 27). Such folding is in places semi-plastic to plastic in style and the folds clearly refold the pairs of sills derived from F_2 folding. Such a flattening would tend to tighten pre-existing recumbent folds and to obliterate the hinges of such structures with the result that without the basic granulite sills repetitions in such rock sequences would not be detectable.

The time relationship between the production of such small-scale folding (F_3I) and the NW F_3II folding is not clearly known and it is quite possible that the development of the semi-plastic to plastic type of folding seen in the basic granulite sills (F_3I), pre-dated or only coincided in part with the forces which gave rise to the NW F_3 folds (F_3II). In such a case the F_3I folds might be regarded as a separate folding phase but it is possible that some overlapping in time occurred between the formation of the two fold types. Both types of F_3 structures clearly refold the F_2 isoclinal folds and the S_1 foliation but, whereas the F_3I small-scale plastic folding seen in the basic granulite sills had no regional effect on the foliation pattern of the complex, the F_3II folding produced variations in the strike direction of the regional foliation and produced large-scale folds and some basin and dome structures.

F_4 folding resulted in a broad warping of the complex producing large, open structures which have little effect on the pre-existing fold pattern or trend of the regional foliation. Such folding seems to have affected the gneisses of the complex at a time when the plastic or semi-plastic conditions had ceased and some fracturing seen in the complex may be connected with this late folding phase.

Metamorphism of the complex

Metamorphic grade

The lithological diversity of the rocks of the complex and the corresponding variety of mineral assemblages allow the metamorphic grade of the complex to be firmly established. The mineral assemblages indicate a major high-grade regional metamorphism under granulite facies conditions followed by local retrogressive metamorphism. The occurrence of abundant hypersthene in the paragneisses and basic rocks associated with such features as grey or bluish quartz, grey or greenish feldspars and the existence of microperthite and antiperthite, is conclusive evidence of granulite facies metamorphism. Hornblende has a very sparse distribution in the complex: in the metavolcanics and basic schists it is commonly a brown variety typical of the granulite facies and only in the ultramafic rocks and in some of the basic schists does green hornblende exist. All the paragneisses, the metavolcanics and the basic granulites of the complex, ranging from the highly transformed and migmatised gneisses to the meta-arkoses, metaconglomerates and metavolcanics bearing primary structures, and also possibly the G_1 granites, have been metamorphosed under granulite facies conditions.

Retrogression of the gneisses has not been severe. Evidence for retrogression is seen mainly in the west of the area towards Tasermiut fjord although even here the retrogression has only affected local areas, and in the majority of rocks fresh hypersthene still prevails. However, the metavolcanic rocks in the outlier to the west of Tasiussaq show evidence of stronger retrogression and the higher grade assemblage brown hornblende-orthopyroxene-clinopyroxene-plagioclase exists along with the lower grade assemblage epidote-green hornblende-tremolite-cummingtonite-plagioclase. Some retrogression has occurred in connection with shearing, fracturing and mylonitisation of the gneisses with a tendency for the development of chlorite, mica and epidote.

Mineralogically the retrogression is seen in the replacement of hypersthene and clinopyroxene by green hornblende, biotite and epidote, garnet by biotite and chlorite, hornblende by biotite, and biotite by chlorite. Furthermore, garnet has also recrystallised in the gneisses and it replaces hypersthene and grows over biotite. It is considered probable that the appearance of green hornblende, secondary biotite and garnet is indicative of a retrogression during the genesis of the gneiss complex but the replacement of garnet and biotite by chlorite, the occurrence of muscovite, epidote and the sericitisation and saussuritisation of the feldspar, all indicative of further retrogression, are probably connected to the genesis of the granitic complex and to late hydrothermal activity.

Subdivision of the granulite facies

The subdivision of the granulite facies into subfacies is a matter of some dispute. The granulite facies was erected as a separate facies by ESKOLA (1939) who considered hornblende to be unstable in the mineral assemblages of the facies. This is not now upheld and FYFE, TURNER and VERHOOGEN (1958), and TURNER and VERHOOGEN (1960) recognised two subfacies based on the stability of hornblende and biotite: a lower grade hornblende-granulite subfacies in which hornblende and/or biotite are stable and a higher grade pyroxene-granulite subfacies characterised by the absence of biotite and hornblende. This subdivision has received wide support but other subdivisions have been proposed: *e.g.* HSU (1955) suggested a biotite-granulite subfacies on the basis of the stability of biotite in the San Gabriel mountains, California, while DE WAARD (1965) has suggested a four-fold subdivision of the facies based on the stability of garnet in "granulite-facies terrane of the Adirondack Highlands". According to TURNER and VERHOOGEN (1960, p. 544) the pyroxene-granulite subfacies represents higher temperatures than the hornblende-granulite subfacies at equal H_2O pressures, but BUDDINGTON (1963) and others have stressed the importance of H_2O rather than temperature and pressure as the critical factor controlling the assemblages of the two subfacies (see also YODER, 1955). DE WAARD (1965, p. 186) has noted that confining pressure and water pressure commonly fluctuate independently from each other in granulite facies terrain and accordingly has proposed a subdivision of the pyroxene-granulite subfacies of FYFE, TURNER and VERHOOGEN (1958) into an orthopyroxene-plagioclase subfacies and a clinopyroxene-almandine subfacies, and the hornblende-granulite subfacies into a hornblende-orthopyroxene-plagioclase subfacies and a hornblende-clinopyroxene-almandine subfacies. The division is based on the observation that the pair orthopyroxene-plagioclase becomes unstable at high confining pressure and is replaced by the association clinopyroxene-garnet-quartz in rocks with excess silica.

Following the facies subdivision of TURNER and VERHOOGEN (*op. cit.*) the high-grade gneiss complex of the Tasiussaq area contains mineral assemblages of both the hornblende- and the pyroxene-granulite subfacies. The majority of the rocks of the complex (excluding the granitic rocks) contain some biotite and where this is absent, as in the metalavas, hornblende occurs. However in the meta-arkoses and metaconglomerates both biotite and hornblende may be lacking. In the hypersthene-quartzo-feldspathic charnockitic gneisses forming the largest area of gneiss in the complex, pyroxene-biotite-garnet-potash feldspar-plagioclase-quartz assemblages are intercalated on both a small and large scale with the quantitatively less important assemblage pyroxene-garnet-potash feldspar-plagioclase-quartz. Thus, adopting TURNER's and VERHOOGEN's facies classification (1960), rocks of the hornblende-granulite subfacies (biotite-bearing) are intercalated with rocks of the higher grade pyroxene-granulite subfacies (biotite-lacking) and as such should be interpreted as having been formed at different temperature-pressure conditions.

Both progressive and retrogressive relationships between metamorphic assemblages indicative of varying grade have been described from areas of granulite facies gneisses and metamorphic rocks. For example, GROVES (1935) from Uganda, BUGGE (1943) from southern Norway and MOREL (1958) from Malawi have described progressive relationships between the mineral assemblages, while WILSON (1958) from Australia, RILEY (1960) from northern Canada, COORAY (1962) from Ceylon and HEPWORTH (1964) from Uganda have demonstrated the retrogressive development of lower grade assemblages from higher. However, in the hypersthene-quartzo-feldspathic gneiss terrain of the Tasiussaq complex, it is clear that the gneisses bearing biotite are not the retrogressed variants of rocks of the pyroxene-granulite subfacies, nor does the relationship seem to be one of progressive metamorphism from rocks of the hornblende-granulite subfacies to rocks of the pyroxene-granulite subfacies. Both types of assemblages were apparently formed under the same temperature conditions and possibly under similar pressure conditions. The appearance of biotite is most probably controlled by the compositional variations in the psammitic rocks from which the gneisses were derived. Thus the subdivision into the hornblende-granulite subfacies and pyroxene-granulite subfacies clearly has some limitations.

An indication of the strong control of biotite distribution by the composition of the pre-metamorphic rock type is supplied by the ubiquitous and abundant presence of the mineral in the pelitic rocks (biotite schistose gneisses and schists, biotite-garnet gneisses and biotite gneisses) and its lesser abundance or absence in the hypersthene-quartzo-feldspathic gneisses derived from psammitic rocks. The pelitic rocks and the hypersthene-quartzo-feldspathic gneisses are the two most important gneiss groups of the complex; the rocks rich in biotite (pelitic rocks) have a restricted occurrence in the west of the area and the gneisses containing smaller amounts of biotite (hypersthene-quartzo-feldspathic gneisses) are restricted to the east. The slight retrogression of the high-grade gneiss complex is more marked in the west of the area than in the east but it is clear that this retrogression is not the cause for the present distribution of biotite in the two parts of the complex although undoubtedly secondary biotite was produced. The aerial distribution of biotite is not an expression of metamorphic grade.

Furthermore modal hornblende, the other critical mineral in the subfacies classification, also is greatly influenced by host rock composition and in the high-grade gneiss complex biotite and hornblende are generally antipathetic. Where biotite occurs in the pelitic rocks and in the charnockitic gneisses, hornblende, apart from traces of secondary hornblende, is absent, and in the few places where brown hornblende exists, *i.e.* in the metalavas and pyroclastics, biotite is absent. There are few examples where the two minerals occur together but in these the rocks have no free quartz, *e.g.* the ultramafic rocks and the basic schists. The instability of hornblende with biotite and quartz can be referred to the reaction:

$$\underset{\text{Hornblende}}{NaCa_2(Mg,Fe)_4Al_3Si_6O_{22}(OH)_2} + \underset{\text{Biotite}}{K(Mg,Fe)_3AlSi_3O_{10}(OH)_2} + \underset{\text{Quartz}}{7SiO_2}$$

$$\rightleftharpoons \underset{\text{Plagioclase}}{NaCaAl_3Si_5O_{16}} + \underset{\text{Diopside}}{Ca(Mg,Fe)Si_2O_6} + \underset{\text{K feldspar}}{KAlSi_3O_8} + \underset{\text{Hypersthene}}{6(Mg,Fe)SiO_3}$$

while the incompatibility of hornblende and quartz can be illustrated by the following reactions, all given by RAMBERG (1948):

(1)

$$\underset{\text{Hornblende}}{NaCa_2(Mg,Fe)_4Al_3Si_6O_{22}(OH)_2} + \underset{\text{Quartz}}{4SiO_2} \rightleftharpoons$$

$$\underset{\text{Plagioclase}}{NaCaAl_3Si_5O_{16}} + \underset{\text{Hypersthene}}{3(Mg,Fe)SiO_3} + \underset{\text{Diopside}}{Ca(Mg,Fe)Si_2O_6} + H_2O$$

(2)

$$\underset{\text{Hornblende}}{NaCa_2(Mg,Fe)_4Al_3Si_6O_{22}(OH)_2} + \underset{\text{Garnet}}{(Mg,Fe)_3Al_2Si_3O_{12}} + \underset{\text{Quartz}}{5SiO_2}$$

$$\rightleftharpoons \underset{\text{Plagioclase}}{NaCa_2Al_5Si_7O_{24}} + \underset{\text{Hypersthene}}{7(Mg,Fe)SiO_3} + H_2O$$

The evidence from the Tasiussaq area is supported by that of BANNO, TATSUMI, OGURA and KATSURA (1964, p. 411), who point out that "the sub-facies classification of the granulite facies may often lead to erroneous conclusions, because the sub-facies classification has had to be based on the difference in abundance of hornblende or biotite in the modal composition and that may be greatly influenced by the compositions of the host rocks". They conclude that biotite and hornblende (albeit hornblende and biotite of restricted compositions) can be stable in both subfacies. Furthermore RILEY (1960), through his study of high-grade gneisses from Baffin Island, has shown that amphibole and biotite are stable in charnockites and granulites at temperatures and pressures at least as high as those considered to form rocks of the pyroxene-granulite subfacies.

Therefore it becomes difficult to assign the rocks of the Tasiussaq high-grade complex to a particular subfacies of the granulite facies. Even if the subdivision of the granulite facies as proposed by DE WAARD (1965) is employed, rocks bearing mineral assemblages of all four subfacies occur in the complex, the orthopyroxene-plagioclase subfacies and the hornblende-orthopyroxene-plagioclase subfacies being represented respectively by the rocks mentioned above as belonging to the pyroxene-granulite subfacies and the hornblende-granulite subfacies of TURNER and VERHOOGEN (1960). The mineral assemblage diopside-garnet-quartz, which is characteristic of some of the meta-arkoses, represents the clinopyroxene-almandine subfacies while this mineral assemblage is joined locally by biotite, an association of the hornblende-clinopyroxene-almandine subfacies. The critical mineral assemblages are the result of the original composition of the host rocks combined with the varying amounts of H_2O and water pressure existing in various parts of the complex at different times. A possible indicator of grade, however, might be found in the stability of the hornblende and biotite in the presence of quartz. According to TURNER and VERHOOGEN (1960) and WINKLER (1965), and as recently shown by HAPUARACHCHI (1967), hornblende and biotite can be stable together in the presence of quartz in the lower subfacies of the granulite facies as they are in the lower grade (almandine-)amphibolite facies. The antipathetic relationship of hornblende and biotite and the non-stability of hornblende and quartz in rocks of the high-grade gneiss complex might be taken to indicate a higher grade, perhaps in part corresponding to an upper subfacies in the granulite facies.

Nature of the metamorphism

The high-grade gneisses of the complex have been derived through the metamorphism of original sedimentary and igneous rocks. The hypersthene granites, enderbites and associated granitic rocks of the complex (G_1) are also considered to have received their high-grade mineralogy through granulite facies metamorphism. The G_1 granites form layers in the hypersthene-quartzo-feldspathic gneisses and they are mineralogically similar to the gneisses. The granites either represent early igneous granitic intrusions into the gneisses (or rocks from which the gneisses have been derived), or they represent compositional variations in the sedimentary series. The homogeneous nature of some of the hypersthene granites and enderbites tends to support the first view, although the transitions from the granites to the gneisses both in composition and in character, plus the fact that composition varies considerably within each layer of G_1 granite, are suggestive of the latter. Temperatures during the granulite facies metamorphism must have been high for the formation of hypersthene in association with the perthitic feldspars. Signs of anatexis in the gneisses are small and local. Some parts of the charnockitic gneisses in the east of the area and some areas in the pelitic gneisses of the west display a style of folding which is semi-plastic to plastic, and in small local parts of the charnockitic gneisses blurring of gneissic structure is probably the result of some fusion and a tendency to melting. However, there was no tendency for the production of a widespread anatectic melt.

According to WINKLER and VON PLATEN (1958) and WINKLER (1961), a partial anatectic melt may form at a temperature of $700 \pm 40°C$ at 2000 atm. of water pressure in a quartzo-feldspathic rock while TUTTLE and BOWEN (1958) consider that melting in the average granite will commence at 640°C at a pressure of 4000 kg/cm² if sufficient water is present. VON PLATEN (1965, p. 217) concludes that at a water pressure of at least 2000 bars an anatectic melt would be generated in nature at temperatures of $700 \pm 30°C$ if the rock contains plagioclase, quartz and a potassium mineral (*e.g.* biotite, muscovite, potash feldspar). Temperatures in the Tasiussaq area were presumably at least as high as these and since there are no clear signs that large-scale anatexis has occurred it is concluded that water was not available in excessive quantities at this level of the crust during the time span of the granulite facies metamorphism.

On the other hand the gneisses of the high-grade gneiss complex are not characterised by the absence of hydrous minerals and the majority of the gneisses contain some biotite. Where this is absent, as in the metalavas, hornblende may exist, although elsewhere the gneisses contain neither biotite nor hornblende. For the crystallisation of such hydrous minerals water must have existed in the system at some stage during the metamorphism and must have been readily available to combine with other elements to produce biotite and hornblende. It is debatable whether hypersthene could have crystallised at the same time as the biotite and hornblende under such conditions since it might be expected that hornblende or another amphibole would have crystallised instead of the orthopyroxene. Since this has not been the case, it is considered that the formation of biotite and the small amount of hornblende occurred or continued after an essentially anhydrous phase in which hypersthene crystallised, *i.e.* some water was introduced. In the realms of such high metamorphic conditions where temperature exceeds the minimum necessary for anatexis to develop and where water pressure is lower than confining pressure, only minor fluctuations in the three variables temperature, confining pressure and water pressure appear necessary to critically affect the mineral stability field. Even anatexis itself can take place over such a small critical temperature range as 30–50°C (von Platen, 1965) and it seems also likely that fluctuations in the availability of the H_2O component at the same temperature and pressure conditions within the granulite facies is capable of controlling the appearance of the hydrous minerals biotite and hornblende. In this way it would seem possible for biotite and hornblende to occur in equilibrium with anhydrous minerals, such as hypersthene.

This conclusion is very similar to that drawn by Hsu (1955) from his metamorphic study in the San Gabriel Mountains of California, where the granulites "are composed of typical granulite minerals and crystalloblastic biotite", a mineralogy remarkably similar to the hypersthene-quartzo-feldspathic gneisses and pelitic rocks of the Tasiussaq area. Hsu concludes (p. 223) that "biotite is probably in equilibrium with the anhydrous granulite minerals in the presence of the H_2O component".

The question of water has played an important part in granulite facies metamorphic studies and Eskola (1939, 1952) posed the question whether the relative deficiency of minerals containing an H_2O component in granulites was due to the prevailing high temperature of the metamorphism which drove away and expelled water from the system, or whether it was due to the original "dryness" of the pre-metamorphic rocks. Yoder (1955) in a paper dealing with the role of water in metamorphism has shown that mineral assemblages of several metamorphic facies can be formed under the same pressure-temperature conditions by variation in the amount of water available in the system and thus (p. 518) "granulites, for example, may have formed in the absence of water at low temperatures, or at high temperatures in the presence of water above the stability range of the hydrous minerals".

Ramberg (1952, p. 158), Hsu (1955, p. 242) and Buddington (1963, p. 1162) among others have further commented on the control of biotite in the granulite facies by H_2O variations in the rocks, either the actual amount of water present in the system or in terms of water vapour pressure. The author holds the opinion that the presence of biotite and brown hornblende in the high-grade gneiss complex along with characteristic anhydrous mafic minerals such as hypersthene and clinopyroxene does not necessarily indicate disequilibrium in the mineral assemblages. The brown hornblende and the biotite are possibly chemically water-poor and it is considered possible that in an essentially water-deficient system enough water can exist to permit the crystallisation of a hydrous mineral phase without entering the realms of anatexis. Evidence that some hydrous minerals continued to crystallise after the anhydrous hypersthene and clinopyroxene is provided by some replacement of the pyroxene by amphibole and biotite. These mineral replacements are not regarded as solely dependent on variations in the temperature and pressure conditions affecting the rocks but rather on variations in the H_2O component within the system.

Following this initial granulite facies metamorphism some retrogressive mineral changes took place (see p. 62), the complex was subjected to migmatisation, and a group of dolerite dykes was intruded. It is considered that during the development of these events, metamorphic conditions did not decline appreciably and the slight retrogression seen in the gneisses is not the result of a separate and disconnected set of metamorphic conditions affecting the complex following the initial granulite facies metamorphism. On the contrary, plutonic conditions continued without an apparent break but with fluctuations in the availability of water and water pressure and with small variations in temperature and pressure.

The concept of "retrograde metamorphism" as first outlined by BECKE (1909) and later amplified by HARKER (1939) is used by TURNER and VERHOOGEN (1960, p. 452) when "a high-temperature metamorphic mineral assemblage is converted to an assemblage (usually more hydrous) stable at lower temperature". In view of the important role played by water in metamorphism, it becomes very difficult, and perhaps often leads to erroneous conclusions, to apply such a restricted definition to metamorphic rocks in nature, especially since it has been experimentally demonstrated that a high-grade mineralogy can be changed to assemblages considered to be of lower grade solely by variation in H_2O conditions. Hence YODER (1952, p. 569) through experimental petrological studies remarks "the conclusion is reached that the presence of an 'excess' or 'deficiency' of water vapor greatly influences the mineralogy of a metamorphic rock", so that mineral assemblages usually considered as representative of several metamorphic facies may develop under the same temperature and pressure conditions. Thus "retrograde metamorphism need not mean a change in pressure or temperature conditions, but may mean access of water vapor". TURNER and VERHOOGEN (1960) in their definition of "retrogressive metamorphism" note that minerals resulting from a downgrading are "usually more hydrous" but it is difficult to recognise which hydrous minerals are due to a fall in temperature and which are due to an increase in the availability of water at constant temperatures.

A broader concept of retrograde metamorphism, and one which is more acceptable to the present author, was proposed by SCHWARTZ and TODD (1941) who stress the importance of H_2O pressure and water content in some retrograde metamorphic changes and suggest (p. 189) "that, in order to avoid confusion in the future, the term 'retrograde' and its synonyms be divorced from all theory regarding the cause of retrogression and be used with reference to any metamorphic mineral which apparently has been formed at the expense of a higher-grade metamorphic mineral". Recently CARPENTER (1968) has demonstrated that pressure gradients brought on by deformation during regional metamorphism can lead to the migration of water into low-pressure areas thereby rendering a "high-grade" anhydrous mineral assemblage unstable relative to a more hydrous "low-grade" assemblage.

It is suggested that the slight so-called retrogressive changes in the high-grade gneiss complex may primarily be due to changes in H_2O content, accompanied by minor fluctuations in P–T conditions, and that the retrogression is not the result of a major and significant decrease in temperature during the plutonic history of the complex.

Towards the end of the evolution of the complex, a set of dolerite dykes was intruded into the granites and gneisses. These dykes recrystallised to pyroxene-metadolerites bearing a high-grade mineralogy, and they provide an example of the "granulite trend" in the metamorphism of dolerites (DAWES, 1968). The high-grade nature of the dykes corresponds to the nature of the charnockitic rocks which the dykes truncate. This relationship is not coincidental. A survey of the literature of basement areas composed of charnockitic rocks cut by high-grade basic dykes led the author to believe in the possibility that such pyroxene-metadolerite dykes could be formed at temperature and pressure conditions below those usually considered as typical of the granulite facies provided the environment was water-deficient. Consequently a theory of dipsenic metamorphism was proposed. However, it seems likely that the position of the thermal front during the metamorphism of the dolerites did not differ greatly from its position during the entire plutonic evolution of the high-grade gneiss complex.

Relation of migmatisation, metamorphism, deformation and granitisation

Relations in the Tasiussaq complex

Granulite facies metamorphism is one of the earliest plutonic events recognisable in the complex. The high-grade metamorphism initiated early in the deformational history of the complex and is essentially pre-migmatitic. Brown biotite, intimately associated with hypersthene and garnet, is folded by the F_2 isoclinal folds indicating the production of high-grade minerals before or during the early stages of F_2 deformation. This early hypersthene and garnet is replaced and truncated by the early migmatitic veins M_1 and M_2 indicating a pre-migmatitic age for the metamorphism. The M_1 veins are affected by the F_2 isoclinal folds. Some veins are apparently folded more severely than others and thus it seems that migmatisation started during F_2 folding. This is supported by the fact that some of the F_2 folds are semi-plastic to plastic in style. (These relationships can only be established in the folded migmatitic rocks of the complex and thus theoretically the high-grade mineralogy of the meta-arkoses, metaconglomerates and the metalavas could be of a later age—see p. 74).

M_2 veins are tentatively correlated with the formation of the G_2 granites which are mineralogically similar and which clearly post-date M_1 but pre-date M_3. The M_2 veins post-date the F_2 folding but they are folded in places by F_3. It is suggested that M_2 migmatisation occurred during F_3 folding and the development of the G_2 granites took place during or towards the end of F_3. Garnet and hypersthene have recrystallised in connection with the G_2 granites and the former mineral commonly occurs in M_2 veins and flecks. Orthopyroxene also occurs in some of the M_2 veins and the few charnockitic veins noted are possibly all post-M_1 in age (see p. 50).

G_3 granites are of the same age as the third phase of migmatisation, and the granites and the M_3 pegmatites both clearly post-date the formation of the G_2 granites, F_2 and F_3 folds and M_2 felsic veins, but the relationship to the F_4 structures is not known. Garnets are common in both the pegmatites and the G_3 granites.

The pyroxene-metadolerite dykes post-date F_2 and F_3 fold phases and the migmatisation phases M_1, M_2 and M_3. However, the relationship of the dykes to F_4 is not clear but since some folded pyroxene-metadolerite dykes have been noted, it seems possible that the dyke intrusion and the metamorphism producing the high-grade mineralogy pre-date the F_4 folding of the complex. The possibility that the folding of the dykes is connected to the genesis of the later granitic complex cannot however be dismissed.

It has been argued in the section on metamorphism (p. 69) that high-grade metamorphic conditions prevailed during the genesis of the complex and that retrogressive effects due to temperature fluctuations are minor. On the other hand migmatisation of the rocks of the complex has been widespread and in places severe. It is clear that an early granulite facies mineral assemblage pre-dates the migmatisation since orthopyroxene, garnet and associated biotite are clearly replaced and truncated by M_1 and M_2 migmatite veins. However, the granulite facies mineralogy of the rocks is little affected by the migmatisation and there is no evidence for a general regional downgrading of the complex. This stability of the high-grade mineralogy of the gneisses suggests that the migmatisation took place under metamorphic conditions approaching, or as high as, those which produced the initial mineralogy, *i.e.* granulite facies conditions. This is supported by the facts that hypersthene and garnet occur in the M_2 migmatitic veins, hypersthene occurs in the G_2 granites where the mineral has been inherited from the replaced country rocks and has recrystallised, and that charnockite veins clearly post-date the early high-grade mineralogy of the gneisses. Clearly, high-grade minerals of more than one generation occur in the complex: even the late cross-cutting basic dykes (pyroxene-metadolerites) which post-date the orthopyroxene-bearing M_2 veins and probably the charnockite veins display a high-grade mineralogy.

It might be argued however, that since late high-grade metamorphic condition saffected the complex, the stability of the high-grade mineralogy referred to above might be apparent and that the present mineralogy of the complex is essentially due to a post-migmatisation granulite facies metamorphism. While later high-grade metamorphism has no doubt led to the recrystallisation and production of high-grade minerals, the fact that an early pre-migmatisation high-grade mineralogy can be recognised and the fact that granulite facies minerals are associated with the M_2 and charnockite veins suggests a migmatisation under metamorphic conditions not greatly different from those of the granulite facies. Further, no relics of downgraded rocks occur in the complex and no examples of hypersthene developing from lower grade minerals have been seen, a situation which might be expected and one which has been described by HEPWORTH (1964) from the Southern West Nile, Uganda, where granulite facies rocks in places have been retrogressed and **have** been affected later by a post-migmatisation high-grade metamorphism with the production of late hypersthene from lower grade minerals.

Correlations with adjacent areas

The Tasiussaq area was the only area of granulite facies rocks in South Greenland mapped by GGU in the years up to 1963. The basement to the north-west is characterised by amphibolite facies or lower grade rocks. Following investigations in 1965, BRIDGWATER, SUTTON and WATTERSON (1966) concluded that the gneisses to the east and south-east of the Tasiussaq area are mainly amphibolite facies in grade although further mapping in 1967 (SUTTON and WATTERSON, 1968) has revealed some areas of granulite facies rocks in the Kap Farvel region. Thus, since the Tasiussaq area and parts of the immediate area to the north mapped by WALLIS (1966) are apparently flanked by areas of mainly lower grade rocks, the possibility arises that the Tasiussaq area is a relic area of granulite facies rocks.

ESCHER (1966) has given an account of the deformation and granitisation of the rocks on the Nanortalik peninsula to the west of the Tasiussaq area and this has been adopted as a standard chronology for this part of South Greenland (ALLAART, 1964). ESCHER (*op. cit.*) recognises three periods of folding: the first deformation produced both large- and small-scale isoclinal folds with axes trending north-north-east to north-east, the second deformation resulted in more open folds with north-west axes and the third period of folding resulted in very large open folds with east-north-east to north-east trending axes. These fold phases can probably be correlated with the F_2, F_3II and F_4 of the Tasiussaq area. WINDLEY (1966b) has dealt with the correlation of fold phases in the area between Julianehåb and the Tasiussaq area and terms the first, second and third periods of folding on the Nanortalik peninsula F_2, F_3 and F_4. This terminology was later adopted by ESCHER (1967).

ESCHER (1966, p. 41) records that the first phase of metamorphism took place during the second and principal folding period (*i.e.* F_3 according to the later accepted terminology), and that a "second phase of metamorphism was effective after the second and probably at the beginning of the third deformation phase", *i.e.* F_4. ESCHER (1966, pp. 43–44) reports the first signs of migmatisation as due to regional metamorphism post-dating F_2 and coinciding with F_3, and that (1967, p. 146) "the F_2 deformation was thus essentially pre-migmatitic". The early granulite facies metamorphism of the Tasiussaq area was essentially pre-migmatitic and is pre- to early F_2 in age and thus it appears to predate the main metamorphism of the Nanortalik peninsula. The slight retrogressive effects seen in the high-grade gneisses of the Tasiussaq complex possibly coincided in time with the main metamorphism of the Nanortalik peninsula.

On the south-western part of the Nanortalik peninsula the degree of metamorphism increases with depth and ESCHER (1966) describes a continuous transition from low granulite facies gneisses at the base of the succession, through amphibolite facies and albite-epidote amphibolite facies rocks in the middle part of the rock series, to low-grade metavolcanic rocks at the top of the succession. The main metamorphism producing the mineral assemblages of this continous metamorphic sequence, including the formation of sillimanite, cordierite and hypersthene in the high-grade gneisses at the base, apparently post-dates the F_3 folding although early biotite, the only indication of the first phase of metamorphism, was probably formed during F_3 (ESCHER, 1966, p. 41).

In the Tasiussaq area, all the rock types of the succession from pelitic, psammitic and psephitic to volcanic have been metamorphosed under granulite facies conditions, and since some psammites to the west of Tasermiut fjord belonging to the "quartzite unit" (ESCHER, 1966, p. 38) plus many rocks to the north of the Tasiussaq area (WALLIS, 1966) are also of granulite facies, there seems to be a discordance between metamorphic isograds and lithostratigraphic units in the north and eastern part of the Tasermiut fjord region. Furthermore, since there is a time difference between the early granulite facies metamorphism of the Tasiussaq area and the main metamorphism to the west of Tasermiut fjord, it is conceivable that an early granulite facies metamorphism also affected a larger area than is now directly apparent.

What is clear is that the production of a high-grade mineralogy of the gneisses in the Tasiussaq area occurred early in the metamorphic history of the Tasermiut fjord region and unless it is envisaged that the extent of the granulite facies metamorphism was restricted to the Tasiussaq area and the immediate area to the north, then parts of the surrounding gneisses must also represent retrogressed rocks. WEGMANN (1939) had earlier recognised high-grade gneisses (granulites) characterised by hypersthene and garnet in the Kap Farvel area and had contrasted the rocks with the lower grade gneisses to the north-west of Tasermiut fjord. These granulite facies rocks are part of the flat-lying metamorphic complex of the Kap Farvel area, a complex which developed early in the history of the area (SUTTON and WATTERSON, 1968). Their presence adds some weight to the suggestion that granulite facies rocks had a much wider distribution than at present, and it is tentatively suggested that the Tasiussaq area and the immediate area to the north represent a relic granulite facies area which has escaped the retrogression which affected the areas to the north-west, east and south-east. How much of the basement of South Greenland has passed through granulite facies conditions must remain a matter of conjecture.

Comments on the relationship between metamorphism and structure of the complex

As noted earlier (p. 70) the relationship between metamorphism, and migmatisation and deformation can only be established in the gneisses and schists of the complex. Thus an important consideration is whether the high-grade mineralogy of the meta-arkoses and meta-conglomerates is the product of a metamorphism later in age than the early granulite facies metamorphism which affected the gneisses of the complex. The basic problem here, however, is that the critical structural relationship between the relatively undeformed and unmigmatised meta-arkoses and metaconglomerates, and the pelitic and hypersthene-quartzo-feldspathic gneisses is not clear (see pp. 78, 85), and a possibility exists that there is material of more than one age in the complex. Thus theoretically the meta-arkoses and metaconglomerates might have been subjected to one granulite facies metamorphism whereas the gneisses and schists are polymetamorphic. In this case the situation in the Tasiussaq complex may be similar to that suggested by BONDESEN (1966) for the Agto area of West Greenland where an older polymetamorphic basement is considered to be overlain by "a younger cover, represented by the high-grade metasedimentary sequence" composed of "granulites" containing well-preserved sedimentary structures.

Whatever the original relationship between the folded and migmatised gneisses and the meta-arkoses and metaconglomerates, granulite facies metamorphism did affect original sedimentary rocks to produce the high-grade meta-arkoses and metaconglomerates, a metamorphism which caused little destruction of primary sedimentary structures. The rocks have recrystallised at high temperatures under static conditions in an environment lacking in water; the metamorphism was essentially isochemical. The presence of metasedimentary rocks retaining primary structures is not an uncommon feature in granulite facies terrain (*e.g.* Ceylon, COORAY 1962; Antarctica, MCCARTHY and TRAIL 1964; Australia, WILSON 1958; West Greenland, BONDESEN 1966), where metasediments may be intimately associated with gneisses, granites and schists of similar grade. It has already been argued that the meta-arkoses, meta-conglomerates and gneisses of the Tasiussaq complex contained insufficient water for a partial anatectic melt to form. Thus it is probable that some water was expelled from the system, particularly from the pelitic rocks. In the case however of the meta-arkoses and metaconglomerates and also the quartzo-feldspathic gneisses derived from psammitic rocks, it may not be necessary to evoke a large-scale expulsion of water since the original sediments (arkoses, conglomerates, sandstones) were possibly composed of anhydrous minerals that were dehydrated on compaction.

The time and nature of the onset of the "dry" conditions are not clear, but according to WINKLER (1965, p. 122) water-deficient conditions might be achieved "when the thermal effect of the metamorphism continued for an exceptionally long time or when an already metamorphosed gneissic complex is subjected to a second episode of very high grade metamorphism, long after the first one". It appears unlikely in view of the presence of the excellently preserved sedimentary structures that the meta-arkoses and metaconglomerates have been under granulite facies conditions "for an exceptionally long time" and this might lend support for the idea that areas of older basement gneiss exist in the complex. If this hypothesis turns out to be correct it is conceivable that most of the water close to the dry and older gneiss areas was driven out from the meta-arkoses and metaconglomerates with the result that granulite facies conditions existed in such rocks.

Nature and structure of the stratigraphical rock types

It is tempting to conclude that the metamorphic rocks of the complex have been derived from a single supracrustal pile composed of sedimentary and volcanic rocks. However, the possible existence of reworked areas of older rocks cannot be overlooked particularly since crystalline pebbles having a variety of compositions exist in the metaconglomerates, rocks which together with the meta-arkoses form a higher unit than the migmatised and deformed gneisses and schistose rocks. The question of a "basement-cover" relationship and the possible origin of the metaconglomerates and associated psammites is raised later (p. 84).

Nature of the pre-metamorphic rock types

The rock types of the complex indicate that a variety of lithological "primary" rocks have existed in the Tasiussaq area: argillaceous (giving rise to the different pelitic rocks), arenaceous (meta-arkoses, schistose gneisses, hypersthene-quartzo-feldspathic gneisses, migmatised psammites), rudaceous (metaconglomerates), calcareous (calc-silicate granulites) and volcanic (metalavas, basic schists). The graphite in some of the pelitic rocks is taken to represent organic material.

The calcareous rocks, sulphide-bearing rocks and the graphitic rocks are only of minor importance forming small layers in other rock types while the metaconglomerates are found intimately associated with the meta-arkoses. Thus it is necessary to establish the stratigraphical relationship and significance of three main rock types: the pelitic rocks, the psammitic and psephitic rocks, and the metavolcanic rocks, although the last have but a local occurrence. Metamorphism, deformation, migmatisation and the development of the granitic complex have obscured many critical relationships between these rock groups.

Relationships of the rock types

Pelitic rocks are restricted to the west while the remaining part of the area (excluding the rocks of the granitic complex and the G_1, G_2 and G_3 granites) is composed of rocks derived mainly from psammites and subsidiary psephites. The exact relationship between the unmigmatised psammites and psephites (meta-arkoses and metaconglomerates) and the hypersthene-quartzo-feldspathic gneisses has been obliterated by the younger biotite rapakivi granite. Any statement about their relationship must be one of conjecture. However, the hypersthene-quartzo-feldspathic gneisses contain intercalations of psammitic rocks which have only been slightly migmatised (migmatised psammites) and some lighter-coloured layers exist in the higher parts of the mountain Putôrugtoq in the northern part of the area which are probably meta-arkosic. Such features suggest that the hypersthene-quartzo-feldspathic gneisses have been derived from psammitic rocks but they far from prove that the gneisses and unmigmatitic meta-arkoses (and associated psephites) are part of the same stratigraphical series. The exact relationship of the meta-arkoses and the metaconglomerates to the pelitic rocks of the west is not directly apparent (see fig. 32) but from structural considerations the former rocks appear to be the younger.

The relationship of the hypersthene-quartzo-feldspathic gneisses to the pelitic rocks of the west is not one of simple stratigraphical concordance. The hypersthene-quartzo-feldspathic gneisses to the east of Tasiussaq shallowly dip to the north-west towards the pelitic rocks and although some of the gneisses have a position which could stratigraphically underlie the pelitic rocks, it seems that the pelitic rocks are not all younger than the hypersthene-quartzo-feldspathic gneisses. The boundary separating the (pelitic) biotite-garnet gneisses from the (psammitic) hypersthene-quartzo-feldspathic gneisses is arbitrary and no clear or sharp junction exists between the two types of gneiss. This junction may represent a metamorphosed sedimentary facies boundary between original argillaceous and arenaceous sediments of similar age, or the boundary may have another significance representing a discordance in the rock series.

The outcrop of metavolcanic and psammitic rocks to the west of Tasiussaq is incongruous with this stratigraphic pattern. The rocks form an isolated outcrop on top of the gneisses and yet occupy a lower level than some of the surrounding gneisses, the latter showing little sign of structural sympathy with the outlier. The relationship of the psammitic rocks to the metavolcanic rocks in the outlier appears to be one of stratigraphical concordance and a certain amount of mixing of volcanic and sedimentary material has occurred. By comparison with the area

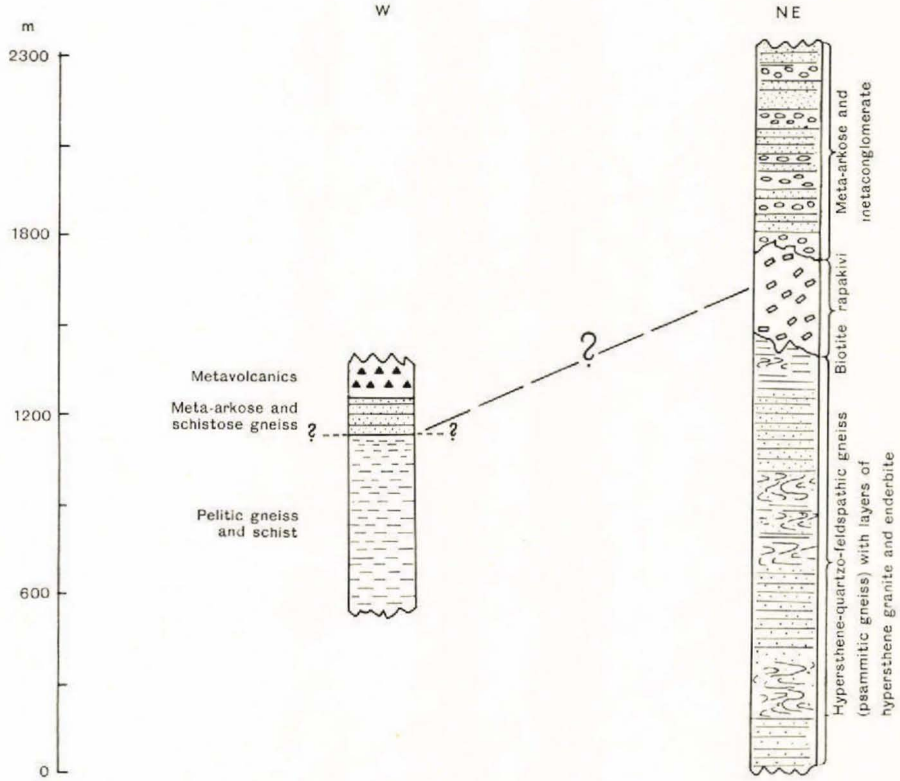

Fig. 32. Diagrammatical representation of the stratigraphical relationships between the western, and the eastern and northern parts of the high-grade gneiss complex. The possible correlation of the meta-arkoses and quartz schistose gneiss of the west with the psammitic and psephitic rocks of the north and east is suggested by comparison with the stratigraphy established to the north-west of Tasermiut fjord.

to the west of Tasermiut fjord, where similar rock types and relationships occur (Escher, 1966), it seems that the metavolcanic rocks might be regarded as the youngest supracrustal rocks in the complex being younger than the meta-arkoses and metaconglomerates in the north of the area. However, if the basic schist layers in the pelitic gneisses are the result of volcanicity contemporaneous with the deposition of the argillaceous sediments, then older metavolcanic rocks occur in the complex. Conversely, the basic schists may be the intrusive equivalent of the metalavas having been emplaced into the older pelitic rocks.

Outcrops of metavolcanic and psammitic rocks similar to the occurrence in the Tasiussaq complex occur to the north-west of Tasermiut fjord in the ground mapped by Escher (1966). The relationship of the psammitic and metavolcanic rocks to the underlying pelitic gneisses and schists in the Tasiussaq area is not directly apparent and a number of possibilities must be considered. These are:

1. The pelitic gneisses and the overlying meta-arkoses and metavolcanics are rock units of a single supracrustal series with the boundary between the gneisses and the meta-arkoses representing either:

a. a stratigraphical concordance reflecting a change in depositional conditions from argillaceous sedimentation to arenaceous although local movement along the boundary may have taken place (ESCHER, 1966), or

b. a tectonic boundary so that the meta-arkoses and the metavolcanics have been tectonically moved to the present position on top of the gneisses (DAWES, 1965).

2. The pelitic gneisses and the overlying meta-arkoses and metavolcanics are rock units of two series of different ages with the boundary between meta-arkoses and gneisses representing either:

a. an original sedimentary discordance between two rock units of different age, the actual discordance which may or may not have been modified by some local tectonic movement, or

b. a tectonic boundary so that the meta-arkoses and metavolcanics have been transported to their present position along a thrust or fault plane, possibly in connection with nappe tectonics.

It is not known whether the meta-arkoses in the outlier to the west of Tasiussaq are of the same age and part of the same rock series as the meta-arkoses and metaconglomerates of the north (see fig. 32), although by comparison with the stratigraphy erected by ESCHER (1966) in the area to the north-west of Tasermiut fjord, this would appear to be so. Thus the significance of the thickness variation of the meta-arkoses from 120 m in the outlier to over 700 m in the north may either be sedimentological (ESCHER, 1966, p. 17) or tectonic, or even both. In possibilities 1b and 2b above it might be expected that parts of the rock series are cut out or reduced in size as a result of tectonic transport and thus the small thickness of the psammitic rocks in the outlier might represent but a slice of the thick metasedimentary succession present in the north. Such a tectonic reduction might accentuate an original sedimentological variation.

Comparison with surrounding areas

The rock types of the Tasiussaq area have some similarities to those mapped by ESCHER (1966) to the west of Tasermiut fjord, although the hypersthene-quartzo-feldspathic charnockitic gneisses which form the largest area of gneiss in the Tasiussaq area do not occur to the west of Tasermiut fjord. These gneisses correspond to the "granulites" of

WEGMANN (1939) which were considered as part of the Kap Farvel granulite complex.

The pelitic rocks in the west of the Tasiussaq area can be reasonably correlated with the pelitic gneisses and schists occurring on the other side of the fjord which belong to the lowest units in ESCHER's stratigraphy and which correspond to WEGMANN's (1948) Amitsok series. Furthermore, the unmigmatised psammites and psephites occurring in the north of the area appear to belong to the same rock series as ESCHER's "quartzite unit" and to the "lower psammitic unit" containing conglomerates noted by WALLIS (1966) in the area bordering the Tasiussaq area on the north. Such rocks occur on both sides of Tasermiut fjord and also on the island Qeqertâraq in the fjord. These psammites and psephites can most probably be correlated with part of the thick psammitic and psephitic succession mapped by J. WATTERSON on "Fastlandet" which corresponds to WEGMANN's (1939) "quartzite gneiss" and the Sydsermilik series (WEGMANN, 1948).

WEGMANN (1939) originally recognised the complicated nature of the rock stratigraphy in the Tasermiut fjord region. He (1939, p. 198) observed the passage in the north-east of the region between the psammitic and psephitic rocks ("quartzite gneiss"—later named the Sydsermilik series) and the upper volcanic rocks, remarking that the "conglomerates give way to greenstone agglomerate and then to the volcanic rock series", but he also believed that the upper boundary of the psammites and psephites dipped steeply to the south so that the rocks were overlain by the gneisses on the south-western part of the Nanortalik peninsula which in turn were overlain by the volcanic series. He finally concluded (WEGMANN, 1948) that the Sydsermilik series (composed of ripple-marked sandstones, cross-bedded quartzites and conglomerates) formed the base of a succession, the series being overlain by the Amitsok series, composed of pelitic gneisses and schists containing sulphides and graphite which was itself overlain by the volcanic series (Arsuk series). ESCHER (1966, p. 15) following detailed mapping on the Nanortalik peninsula, has been able to demonstrate that the Sydsermilik series of WEGMANN overlies and not underlies the Amitsok series.

It is clear that WEGMANN believed that the stratigraphical units of the area immediately to the north-west of Tasermiut fjord also had a wider extent to the east and that the different character of the rocks of the Kap Farvel area (granulites) was due to the higher grade of metamorphism (WEGMANN, 1939, p. 207). WEGMANN (1939, p. 197) reported a transition between the quartzite gneiss (quartzites and conglomerates, later known as the Sydsermilik series) and the granulites, and (p. 200) remarks that the transition "throws light on the original nature" of the granulites (including in this respect the hypersthene-quartzo-feldspathic gneisses of the Tasiussaq complex), the difference between the granulites and the quartzite gneisses being essentially one of "the different kinds of metamorphism".

The author agrees with this in so much that the hypersthene-quartzo-feldspathic gneisses of the Tasiussaq complex and farther to the east (WEGMANN's granulites) have been derived generally from psammitic rocks, but questions the assumption that the main part of the granulites has been derived from the same psammitic material as WEGMANN's Sydsermilik series. The granulites of WEGMANN do not occur to the west of Tasermiut fjord in association with rocks of the Sydsermilik series, and in the northern part of the Tasiussaq area the later biotite rapakivi granite has obliterated the relationship between the hypersthene-quartzo-feldspathic gneisses and the meta-arkoses and metaconglomerates. Undoubtedly some gneissic rocks have been derived from the psammitic rocks of the Sydsermilik series producing transitions between the metasediments and gneisses, a situation noted by WEGMANN (1939) and which appears to exist in the area north of the Tasiussaq area studied by WALLIS (1966). However, it seems much more probable to the author that in this part of South Greenland two groups of psammitic rocks occur, those from which the majority of the hypersthene-quartzo-feldspathic gneisses of the Tasiussaq complex, and perhaps many of the granulites mentioned by WEGMANN (1939), have been derived, and part or all of the quartzites, meta-arkoses and metaconglomerates now recognisable in the Tasermiut fjord region, *i.e.* the Sydsermilik series of WEGMANN (1948), the "quartzite unit" of ESCHER (1966) and the meta-arkoses and metaconglomerates of the Tasiussaq complex.

The relationship of the metasedimentary rocks in the Tasermiut fjord region is complex and it may be audacious to make regional correlations when tracts of country near the head of the fjord remain uninvestigated. Thus it is not clear whether all the outcrops of meta-arkoses, quartzites and metaconglomerates mentioned above are part of a single rock series or whether the psammitic rocks of the south-west of the region can be correlated with the considerably thicker psammitic and psephitic succession in the north-east (see ALLAART, 1964, p. 16). For example, WALLIS (1966) has reported towards the head of the fjord to the north of the Tasiussaq area the existence of two thick psammitic units (the lower characterised by psephites) separated by a thick unit of pelitic rocks either, or both, of which might be correlatable with the "quartzite unit" of ESCHER (1966). Furthermore, whether the psammitic rocks giving rise to the hypersthene-quartzo-feldspathic gneisses are in part the same age as WEGMANN's Sydsermilik series or have been derived from older psammitic rocks, cannot be clearly elucidated at the present state of investigation in the Tasermiut fjord area.

Age of the complex and of the Tasermiut fjord region

There are no absolute age determinations on rocks of the high-grade gneiss complex but samples of noritic gabbro from the Frederiksdal rapakivi massif have given 1645 ± 50 m.y. (K/Ar on augite, see BRIDGWATER, 1965). This massif clearly post-dates the high-grade gneiss complex. An age of 1600 ± 30 m.y. has been obtained on pelitic gneiss from the Nanortalik peninsula (K/Ar on biotite, indicating an age of the Sanerutian metamorphism, see LARSEN and MØLLER, 1968), a rock correlatable with the pelitic rocks of the western part of the Tasiussaq area. While the author has previously assumed that the main metamorphism of the gneisses of the Tasiussaq area is of Ketilidian age (ca. 1700–2000 m.y.), the possible existence of pre-Ketilidian rocks has drawn comment through the presence in the Tasiussaq area of metaconglomerates containing pebbles of metamorphic and granitic rocks, and has been tentatively suggested (DAWES, 1968). However, the metaconglomerates, as stressed previously, only occur as inclusions in, and as a roof-zone to, the younger biotite rapakivi granite and field relationships to other rock types of the complex have been destroyed. Thus any discussion on the age of the rocks of the Tasiussaq area must embrace remarks on the geology of the whole Tasermiut fjord region where gneisses and supracrustal rocks exist in association.

Although considerable discussion has taken place on the division of the Precambrian and the terminology of the rock units recognised in South Greenland, relatively little has been directed at the Tasermiut fjord region. WEGMANN (1938a) used the old Norse name for Tasermiut fjord (Ketils-fjord) to describe the main geological cycle of events in the southern tip of West Greenland which included sedimentation, volcanicity and plutonism. He recognised Ketilidian metamorphosed sediments and volcanics in two areas; in the Ivigtut area and in Tasermiut fjord region. He also recognised pre-Ketilidian rocks in both areas, as gneisses as a base to the Ketilidian in the Ivigtut area (WEGMANN, 1939) and as pebbles in the metaconglomerates in Tasermiut fjord. Between the two areas of metasedimentary and metavolcanic rocks, WEGMANN recognised areas of gneiss and the Julianehåb Granite but also accepted the incorporation of areas of "Präketilidisches Kristallin" (WEGMANN, 1938b, p. 512) in the Ketilidian cycle. WEGMANN's interpretation of the geology of the Ivigtut area has been confirmed by recent detailed mapping by GGU and the Ketilidian suprastructure and infrastructure recognised by BERTHELSEN (1960) are now known to represent Ketilidian (cover) and pre-Ketilidian (basement) rocks respectively (HIGGINS and BONDESEN, 1966; HENRIKSEN, 1969).

Such a distinction between basement and cover is not recognisable to the south-east of the Julianehåb Granite although the gneiss complex of the region is overlain by an important supracrustal series (PULVERTAFT, 1968, p. 102; ALLAART, BRIDGWATER and HENRIKSEN, 1969). This supracrustal series is best preserved on the Nanortalik peninsula to the north-west of the Tasiussaq area where the succession is over 2000 m thick (ESCHER, 1966). In published accounts of the regional geology of South Greenland this supracrustal series of the Nanortalik peninsula has been assumed to be of Ketilidian age (ALLAART, 1964; BRIDGWATER, 1965; BERTHELSEN and NOE-NYGAARD, 1965; PULVERTAFT, 1968; ALLAART, BRIDGWATER and HENRIKSEN, 1969) and thus by inference to post-date the pre-Ketilidian basement of the Ivigtut area. The gneisses forming the country rocks to the south-east of the Julianehåb Granite have been generally accepted as having been derived from Ketilidian rocks (ALLAART, 1964; BRIDGWATER, 1965; BERTHELSEN and NOE-NYGAARD, 1965) although ALLAART, BRIDGWATER and HENRIKSEN (1969, p. 867) remark that "there is no proof that all the gneiss south of the Julianehåb Granite was formed from supracrustal rocks laid down at the same time as those of the Tasermiut fjord". The need for caution in the correlation of metasedimentary and metavolcanic rocks between the Ivigtut and Tasermiut fjord areas has been stressed by most authors and WATTERSON (1965) has noted that the supracrustal series of the Nanortalik region could theoretically be Sanerutian in age, concluding (p. 135) that "in view of this the continued correlation of supracrustal rocks occurring in the widely separated Ivigtut and Nanortalik regions would seem unwise, although by no means definitely incorrect".

ESCHER (1966) has recognised five lithostratigraphic units within the rock succession of the Nanortalik peninsula: pelitic gneiss, semipelitic schist, pelitic schist, quartzite and volcanic units. No unconformities are seen between the units and the idea has been favoured that the rocks represent members of a single supracrustal series, the lower units of which have been migmatised and pegmatised to produce an infrastructure, while the upper quartzite and volcanic units represent a suprastructure. A division into a folded and migmatised unit of pelitic gneisses and schists and an unmigmatitic and relatively undeformed unit of meta-arkoses, metaconglomerates and metavolcanics can be recognised in the Tasiussaq complex (see pp. 47, 59) where the possibility cannot be excluded that the units are not members of a single supracrustal pile. The boundary between the two units could theoretically represent a basement-cover relationship. The fact that no unequivocal or undisputable rock discordancies have been recorded between the underlying gneisses and the psammitic rocks in the Tasermiut fjord region, little affects the argument since original discordancies would show a tendency during plutonism to be transformed into concordance resulting in the parallelism between units, a situation noted from elsewhere in Greenland (Windley *et al.*, 1966; Henderson and Pulvertaft, 1967) and from other fold-belts (Eskola, 1949; Strand, 1952; Preston, 1954; Ramsay, 1958).

It is also possible that an original major break occurs lower down in the Nanortalik rock succession within the so-called infrastructure, since migmatisation and pegmatisation has resulted in a transition from basal massive gneiss into pelitic gneiss and pelitic schist, so that any original discordancies have been obliterated (Escher, personal communication). Thus the two tectonic levels recognised within the infrastructure (Escher, 1967, p. 43)—a lower level composed of gneisses and an upper level of pelitic and semi-pelitic schists—may also have some connection with a reactivated basement-cover relationship. Pulvertaft (1968, p. 103) has already remarked "how little stratigraphical significance can be attached to such a transition", in the light of relationships in the Ivigtut (see Henriksen, 1969) and Umanak (see Henderson and Pulvertaft, 1967) areas.

If a basement-cover relationship does exist in the Tasermiut fjord region then the author favours the idea that the two rock series involved are Ketilidian and pre-Ketilidian in age although Watterson's comment (1965, p. 135) that no evidence has unequivocally proved that the supracrustal rocks of the Tasermiut fjord region are not Sanerutian in age might be similarly employed to argue that any basement-cover relationship in the region could theoretically be between Ketilidian and Sanerutian rocks or between pre-Ketilidian and Sanerutian rocks. However, breaks or unconformities within a supracrustal series or between gneiss and metasediments may not necessarily represent divisions between major Precambrian geological cycles, *e.g.* pre-Ketilidian and Ketilidian, but between rock units of a single geological cycle (geological cycle used in the sense of Harpum, 1960). Thus in any reinterpretation of the geology of the Tasermiut fjord region on the basis of breaks within the gneiss and supracrustal series of the area, the possible significance of the conglomerates, first noted by Wegmann (1939), must be mentioned. Despite the fact that Wegmann's interpretation of the origin of these rocks varied (see Wegmann, 1939, 1948), he stressed the importance of the pebbles in the psephites as indicators of pre-Ketilidian history. There is however another possibility.

In the northern part of the Tasiussaq area the metaconglomerates contain smooth, rounded pebbles up to 70 cm in length composed of a variety of crystalline rocks: granites, gneisses, metasediments and basic rocks (see p. 30). These metaconglomerates are clearly sedimentary or epiclastic in origin in the sense of PETTIJOHN (1957) in contrast to cataclastic and pyroclastic conglomerates. (This statement does not exclude the possibility that some of the metaconglomerates in the Tasermiut fjord region, particularly those composed of angular fragments, are volcanic in origin). It is clear that the pebbles were of crystalline character at the time of incorporation into the conglomerate and the present high-grade mineralogy is clearly superimposed upon original metamorphic mineral assemblages of the pebbles. This is worth pointing out, since the conglomerates have been subjected to high-grade metamorphism and therefore it is to be expected that the pebbles now have a crystalline character. The metaconglomerates thus are not intraformational in nature (WALCOTT, 1894) formed by penecontemporaneous fragmentation and redeposition of the stratum in question, but extraformational *i.e.* containing pebbles of some stratum which was not being deposited at the time (PETTIJOHN, 1957). This implies that the conglomerates have not been derived from an older part of the same supracrustal series but from a sequence of older rocks which contained granites, gneisses, metasediments and basic rocks. Regarding the age relationship between the metaconglomerates and these parental rocks, two possibilities exist: either the pebbles represent basement rocks of an older geological cycle as WEGMANN suggested, *i.e.* pre-Ketilidian, or they represent Ketilidian crystalline rocks, so that the conglomerates and associated psammites represent late-Ketilidian deposits. The psammites and psephites in the Tasermiut fjord region have not been mapped in detail and thickness variations and the extent and persistence of the conglomerate units is not known accurately. Hence correlation of psammite and psephite units cannot be made easily (see p. 80). No detailed work has been carried out on transport direction of the sandstone and conglomerate material, which could provide evidence about the source area for the rocks, studies essential to the understanding of the existence and nature of older rocks forming the depositional basin or trough and the depositional environment of the sediments. Until these aspects of the geology are known, the idea that the metaconglomerates and psammites could represent a late-Ketilidian flysch-type sedimentation in a marginal trough cannot be overlooked. The source area for the rocks would then be the older Ketilidian and pre-Ketilidian areas uplifted during the main Ketilidian diastrophism.

On the other hand, if the pebbles represent pre-Ketilidian rocks as WEGMANN (1939, 1948) suggested, the question of the extent and outcrop of the basement remains open. The facts that the metaconglomerates occur as thick units (see p. 30) and occur over a large area, high in the established stratigraphy of the region being underlain by at least 1500 m of gneisses and schists, and that the pebbles are relatively large in places, reaching 70 cm in length indicating a "local" source for the conglomerates, favour, if anything, the idea that the parental basement rocks occur somewhere in this part of South Greenland.

In view of the discussion above little definite can be said on the age of the rocks of the high-grade gneiss complex. Four possibilities exist:

1. The metasediments, metavolcanics and gneisses of the complex could all be of the same age (Ketilidian) representing members of a single, but *complex*, supracrustal pile.

2. The rocks of the complex may all be Ketilidian in age and part of a single geological cycle but the meta-arkoses and metaconglomerates may represent late-Ketilidian flysch-type sediments younger in age than the underlying early-Ketilidian gneisses and schists.

3. If a major break occurs in the rock succession of the Tasermiut fjord region so that the psammites, psephites and metavolcanics represent rocks of a younger geological cycle than the underlying gneisses and schists, then most of the Tasiussaq complex may be composed of pre-Ketilidian material. However, the true stratigraphic position of the charnockitic gneisses forming the largest area of gneiss in the complex, is obscure (see p. 80) and parts or all of the gneisses might have been derived from psammitic sediments equivalent in age to the "quartzite unit" of the Nanortalik peninsula and to the meta-arkoses and meta-conglomerates of the Tasiussaq complex. Thus Ketilidian rocks might still have a wide extent.

4. If a major break occurs low down in the Nanortalik peninsula succession then the Tasiussaq complex is probably composed of both pre-Ketilidian and Ketilidian material although the actual distribution of the basement rocks would again depend on the age of the psammitic rocks from which the charnockitic gneisses have been derived (see 3 above).

Reference has already briefly been made (see p. 45) to the possibility that parts of the G_2 pyroxene-bearing granites are older than some of the hypersthene-quartzo-feldspathic gneisses, a relationship which may be indicative of older basement rocks. Furthermore, the presence of basic granulites which tend to be concentrated in certain tracts of gneiss might provide direct evidence of older crystalline rocks. It was originally considered that all the basic granulites were connected to the volcanicity producing the metavolcanics of the area (Dawes, 1965) since some sills occur in the metavolcanic sequence. However, where the basic granulites appear as distinct sills within the charnockitic gneisses, it is perhaps easier to envisage intrusion into migmatitic or gneissic host rocks rather than into sedimentary rocks from which the gneisses have been derived since such a form might be expected to be disrupted and altered during the change from sediment into paragneiss. The basic granulites clearly predate the F_2 folding (see p. 52) and are migmatised. Unfortunately no basic granulite sill has been seen to cut folds in the gneisses—a relationship which would be very significant; however, since the sills clearly represent early intrusions and were probably emplaced into crystalline rocks, their pre-Ketilidian age is a possibility. On the other hand, the crystalline rocks into which the intrusions were emplaced may be early-Ketilidian in age and the intrusions inter-Ketilidian.

ESCHER (personal communication) is considering a possible reinterpretation of the geology of the Nanortalik peninsula where the gneiss and supracrustal succession is best preserved. If evidence revealed by further field work suggests that the relation between the pelitic gneisses and schists, and the overlying metasediments and metavolcanics is not truly one of stratigraphical concordance, or if suggestions of a break occur lower down in the rock succession, then a reinterpretation would appear justified. Furthermore, further field work might reveal the relationship between the charnockitic gneisses and the unmigmatitic psammitic and psephitic rocks. Both groups of rocks outcrop in the high mountains to the north-east of the Tasiussaq area and if the relationships of these rocks have not been destroyed by the younger rapakivi granites, a key to the age of the gneisses might be found there.

The recognition of a basement-cover relationship in the Tasermiut fjord region would have some repercussions in the age of the gneisses to the north-west between Tasermiut fjord and the Julianehåb Granite and also to the south-east of the Tasiussaq area. In such a case parts or most of the gneisses mapped previously as Ketilidian (WINDLEY, 1966a; BERRANGÉ, 1966; PERSOZ, 1969) may be essentially pre-Ketilidian in age having been reworked in Ketilidian time while pre-Ketilidian elements may be incorporated in the metamorphic complex of the Kap Farvel area which has been assumed to have been derived from Ketilidian sediments and volcanics (WEGMANN, 1939, 1948; SUTTON and WATTERSON, 1968). ALLAART (1967) has suggested that parts of the Julianehåb Granite are relic areas of pre-Ketilidian basement and WINDLEY (personal communication) favours the idea that the F_1 intrafoliational folds of the Sârdloq area, described previously as Ketilidian structures (WINDLEY, 1966b) are pre-Ketilidian in age. The Sârdloq area borders the Julianehåb Granite on the south-east and ALLAART, BRIDGWATER and HENRIKSEN (1969, p. 872), when reviewing the area south-east of the Julianehåb Granite, remark that "the gneisses in the vicinity of the Julianehåb Granite may contain considerable amounts of reworked pre-Ketilidian basement". The extent of this "pre-Ketilidian basement" could be much larger than hitherto stated.

GRANITIC COMPLEX

Granites of three main ages post-date the formation of the high-grade gneiss complex. These are: an autochthonous microcline granite (G_4) formed by reactivation and granitisation of the gneisses of the high-grade gneiss complex, rapakivi granites (G_{5a-b}) which have a complicated history and display both magmatic and metasomatic characters, and a later allochthonous microgranite (G_6). Pegmatites and aplites are associated with each granite phase. During the genesis of these granitic rocks, basic magma was intruded into the granites of the complex and also into the high-grade gneiss complex producing basic intrusions of three ages. The first intrusion episode gave rise to dykes and small bodies connected in time with the development of the biotite rapakivi granite (G_{5a}). Many of these intrusions are composite in nature and display a veining of basic material by acid. The small bodies are mainly connected in space with the biotite rapakivi granite but the dykes have a wider distribution and penetrate out from the granitic rocks into the high-grade gneiss complex. The second episode resulted in basic dykes which post-date the rapakivi granites but probably pre-date G_6 formation while at least two generations of diorite sheets represent a third episode of intrusion. Some ultramafic dykes were also intruded and although these dykes have only been noted cutting rocks of the high-grade gneiss complex, they were probably intruded in the late development stages of the granitic complex.

Granitic rocks

Microcline granite (G_4)

This granite outcrops to the west, south-west and east of Tasiussaq bay forming an almost complete girdle around the younger biotite rapakivi granite (see Map 1). Inclusions of the granite are quite common in the rapakivi.

The granite is grey in colour, medium-grained, commonly containing many gneissic inclusions in all stages of transformation. These inclusions vary from a few centimetres to metres in length and they are orientated producing a structure within the granite (Fig. 33). Elsewhere the granite is homogeneous with biotite flakes having a random orienta-

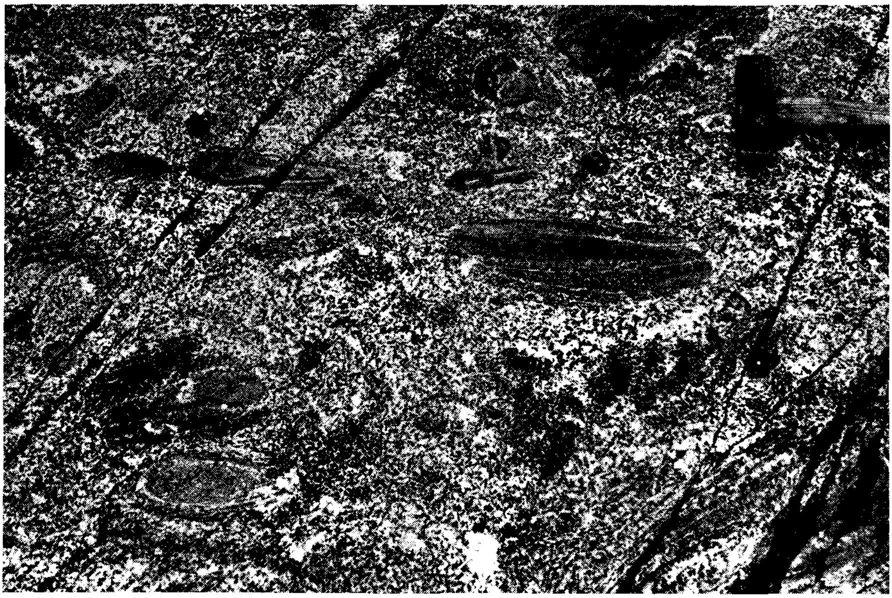

Fig. 33. Microcline granite (G_4) displaying an orientation of pelitic enclaves which are in various stages of transformation. The granitic material is characterised by the random orientation of biotite flakes. North-west of Tasiussaq village.

tion. In many places disorientated inclusions of gneiss have sharp contacts with the granite although many examples exist where disorientated inclusions are in a state of transformation. In places the granite is porphyroblastic, crystals of microcline reaching up to 3 cm in length. Composition varies according to the abundance of gneissic inclusions present.

Essential minerals are microcline, perthite, plagioclase (An_{7-15}), myrmekite, quartz and biotite with or without garnet. Accessories include zircon, apatite and ore, with epidote, penninite and sericite forming the main alteration minerals. Textures are xenomorphic-inequigranular but in enclave-rich granite biotite displays a foliation. Some biotite flakes display curved cleavage traces and quartz in places shows undulose extinction. Some deformation apparently accompanied the granite formation and occasionally quartz has recrystallised.

Contacts of the granite with the gneisses of the high-grade gneiss complex are typically diffuse varying from concordant to sub-concordant to the foliation of the host rocks, but three types can be recognised. No sharp, cross-cutting contacts have been noted.

1) An agmatisation of the older rocks by streaks and veins of microcline granite (Fig. 34). This feature is common where the host rocks are homogeneous.

Fig. 34. Agmatite produced by veining of basic metavolcanic rocks by microcline granite (G_4). A later apophysis of biotite rapakivi granite (G_{5a}) cuts the agmatite. West of Kûgssuaq.

2) A transition from host rock gneisses through granitic gneiss, migmatitic granite, foliated granite to granite. Such a transition series is well shown by the passage from biotite-garnet gneiss to granite on the Tasermiut fjord coast to the south-west of Tasiussaq village and also east of Tasiussaq bay.

3) A growth of feldspar porphyroblasts in the host rock gneisses either singly or in clouds.

The granite has been formed by a granitisation and replacement of the gneisses of the high-grade gneiss complex, and it is essentially autochthonous in type merging to parautochthonous since some mobility of granitic material has occurred. The granite is probably the same age as the "microcline-biotite-granite" or "granite A" mapped by Escher (1966) on the Nanortalik peninsula to the north-west of the Tasiussaq area.

Biotite rapakivi granite (G_{5a})

The main area of biotite rapakivi granite in the Tasiussaq area occurs in the north bordering Tasermiut fjord. Smaller outcrops occur in the south of the area but these form outliers on the older gneisses having

Fig. 35. Biotite rapakivi granite (G_{5a}) illustrating the variation in the size and shape of microcline megacrysts, the mantling of the megacrysts by plagioclase and the concentric zoning of plagioclase within some megacrysts (*e.g.* in centre). The rapakivi is cut by an AP_5 aplite (top left) and by later shearing. The chisel is 22 cm long. Tasermiut fjord coast, east-north-east of the island Qeqertaq.

been originally connected to the Frederiksdal massif which borders the Tasiussaq area on the south.

The main outcrop of rapakivi is the southern part of a much larger granite massif which stretches towards the head of Tasermiut fjord and which has been investigated by WALLIS (1966). The western boundary of the massif mainly lies in Tasermiut fjord but outcrops of the same rapakivi have been mapped to the west of Tasermiut fjord (ESCHER, 1966).

The rapakivi cuts the microcline granite (G_4) (Fig. 34) and the gneisses of the high-grade gneiss complex but it is cut by the micro-granite (G_6) (Fig. 44). The internal primary structures of the rapakivi are cut by a group of basic dykes and bodies but the formation of the rapakivi texture of the granite post-dates these basic intrusions.

The massif in the Tasiussaq area is ethmolithic in form with inward-dipping contacts varying from near vertical to 60°. Small outliers of rapakivi exist on the mountain Putôrugtoq. Contacts vary from sharp and distinct to more diffuse and agmatitic. Abundant evidence exists to demonstrate a late potash metasomatism of the granite and in many

Fig. 36. Rapakivi texture in biotite rapakivi granite (G_{5a}). Microcline perthite megacrysts are mantled by plagioclase. Tasermiut fjord coast, west of Kûgssuaq. Photo: Preben Christensen.

places original sharp contacts have been obliterated and made diffuse by growths of potash feldspar porphyroblasts. Apophyses from the granite are few although several exist north of Taserssuaq cutting the meta-arkoses which form a roof-zone to the massif. They are notably discordant to structure and a dilatational mode of formation is indicated.

Few contact-metamorphic effects are discernible in the adjacent country rocks of the rapakivi and a conspicuous contact aureole is absent. Slight retrogression has occurred in the high-grade gneisses in the immediate vicinity of the rapakivi and pyroxene has been replaced by amphibole and, in places, chlorite. The later potash metasomatism, however, has affected the country rocks adjacent to the rapakivi and potash feldspar megacrysts may occur in the host rocks tens of metres from the actual contact.

The main rock type of the massif is a porphyritic coarse- to very coarse-grained biotite granite with feldspars which vary from mesostasis size up to microcline megacrysts 17 cm long (Fig. 35). The feldspars vary from cream and buff in colour to various shades of pink and red. The rapakivi characteristically breaks down into a brown gravel ("moro") which makes sampling difficult. The megacrysts vary in shape from ovoid to euhedral and they are commonly mantled by oligoclase-albite

Fig. 37. Allogenic enclaves in the biotite rapakivi granite (G_{5a}). The enclaves are composed of metaconglomerate (centre) and meta-arkose, and they are concentrated to form an oligomict breccia. Disorientation of the enclaves is marked. In the top left the inclusions are tightly packed together and little granitic matrix separates them. The chisel is 22 cm long. Tasermiut fjord coast, west of Tasiussaq bay.

(Fig. 36). Concentric rings of oligoclase-albite are occasionally present in both ovoid and well-shaped megacrysts (Fig. 35). In places megacrysts are so abundant that they have coalesced to form a continuous network with the groundmass of the granite filling the interstitial spaces. The microcline of the megacrysts is perthitic but at least one earlier generation of potash feldspar in the groundmass is non-perthitic. Albite and albite-oligoclase occur in the groundmass and at least two generations of quartz exist, the youngest generation being in myrmekite in association with the microcline megacrysts. Accessory minerals include apatite, sphene, tourmaline, allanite, zircon, clinopyroxene, orthopyrox-

Fig. 38. Polymict breccia composed of quartzo-feldspathic gneiss, biotite gneiss, meta-arkose and basic rock within biotite rapakivi granite (G_{5a}). The enclaves are elongated parallel to their internal foliation and they have been closely packed together and aligned. Little granitic matrix occurs between the enclaves. Some potash feldspar megacrysts exist within enclaves (top left). The chisel is 22 cm long. The island Qeqertaq in Tasermiut fjord.

ene, hornblende, fluorite, monazite, ilmenite, magnetite and haematite, while epidote, sericite, penninite, kaolin and antigorite are the main alteration products.

Within the massif there are small areas of granite which differ from the normal biotite rapakivi in composition and texture. The main types are: blue quartz granite characterised by smoky-blue quartz; a non-

Fig. 39. A meta-arkosic enclave within the biotite rapakivi granite (G_{5a}) showing evidence of solid state growth of potash feldspar megacrysts. Suspicion of mantling occurs in some of the megacrysts. A small pegmatite vein cuts the enclave. Tasermiut fjord coast, west of Tasiussaq bay.

porphyritic leucocratic granite which is medium-grained, buff to light brown and lacks the microcline megacrysts; a hornblende rapakivi granite in which hornblende becomes dominant over biotite; a dark grey porphyritic granite which has a high percentage of biotite in the groundmass and which has fewer feldspar megacrysts than the normal biotite rapakivi, and areas of syenite and quartz syenite where the quartz percentage of the rapakivi falls outside the granite field. All these types grade into the normal biotite rapakivi granite and passages from syenite through quartz syenite to granite are common but difficult to map accurately.

The rapakivi is devoid of a major foliation or lineation and the main internal structure is represented by widespread occurrences of polymict and oligomict inclusion breccias. Local examples of basin-shaped mafic layering exist and also small areas where the microcline megacrysts show a preferred orientation. It is significant that some of the megacrysts displaying rapakivi texture were formed later than the mafic layering since crystals grow into and across the layers.

Allogenic xenoliths are distributed throughout the granite but they occur in concentrations (Fig. 37) so that large areas of the granite are

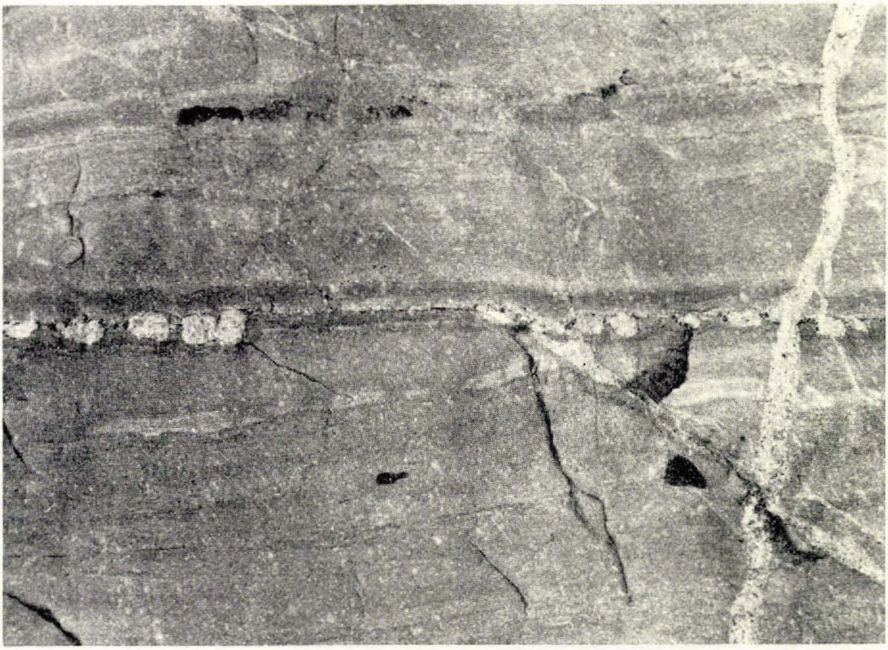

Fig. 40. Microcline megacrysts in meta-arkose aligned along a bedding plane. The meta-arkose forms the roof-zone to the biotite-rapakivi granite (G_{5a}). The megacrysts have grown in the solid state, their alignment having been controlled by the structure of the host rock. The meta-arkose and a megacryst are cut by an AP_5 pegmatite. The diameter of the coin (extreme left) is 2.2 cm. North of Taserssuaq lake.

quite homogeneous with few or no enclaves. They vary in size from centimetres up to the largest of 2×1.3 km. A mass of meta-arkose at least 700 m thick composes the mountains immediately to the north of Taserssuaq and this forms a roof-zone to the rapakivi. Further large areas of undisturbed meta-arkosic material north of Tasiussaq bay suggest that the level of the present exposure corresponds to the highest part of the massif.

The polymict and oligomict breccias are indicative of strong movement and in the polymict types inclusions of different rock types (quartzo-feldspathic gneiss, meta-arkose, metaconglomerate, pelitic gneiss, basic schist, basic granulite and pegmatite) have been collected together and aligned (Fig. 38). The inclusions are commonly elongated or spindle-shaped and are orientated with their long axes parallel. Polymict inclusion breccias may occur in layers in the normal rapakivi or in irregular patches. The preferred orientation of the inclusions in the breccias is truncated by normal biotite rapakivi demonstrating that a mobile phase aligning the inclusions existed early in the history of the rapakivi before the formation of the present texture of the granite.

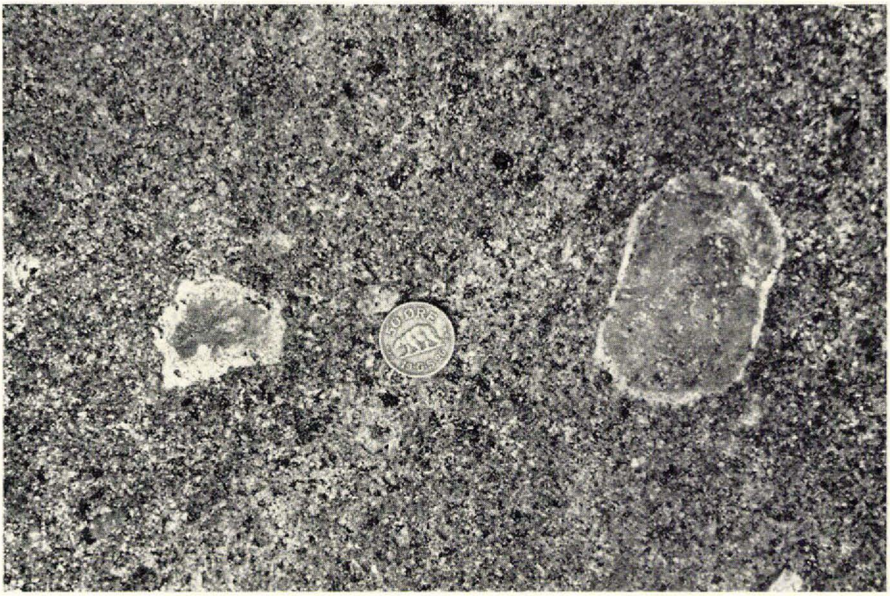

Fig. 41. Two mantled potash feldspar megacrysts in granitised dolerite within the biotite rapakivi granite (G_{5a}). This dolerite was intruded into the rapakivi massif after the initial mobile phase of the granite and then was subsequently granitised. Later potash metasomatism has resulted in the growth of the feldspar megacrysts and the rapakivi texture. Diameter of the coin is 2.2 cm. East coast of Tasiussaq bay.

Microcline megacrysts, some showing rapakivi texture, are commonly present in the inclusions within the rapakivi (Fig. 39), in the country rocks adjacent to the rapakivi (Fig. 40) and in basic intrusive rocks associated with the rapakivi (Fig. 41). They have developed at a late stage in the history of the granite.

Following the formation of the feldspar megacrysts of the granite, numerous later alterations took place. Crush- and shear-zones are widely distributed throughout the granite, hydrothermal alterations resulted in widespread epidotisation and in places chloritisation, and aplites, pegmatites and quartz veins were formed (see p. 104). The shearing and crushing have led to a reduction of the grain size of the granite. Feldspar megacrysts have become shattered and they form angular fragments situated in a fine-grained matrix of quartz, plagioclase, epidote, chlorite and mica. Growth of pistacite, penninite and sphene is directly related to movement-zones and these minerals produce a green to yellowish colour in shear- and crush-zones.

The hydrothermal alteration has resulted in the feldspars changing colour from white and cream to shades of orange, red and pink. The microcline and plagioclase are altered to different shades of colour and this accentuates the rapakivi texture and the concentric rings of plagioclase within microcline megacrysts, and gives a guide to the relative percentages of the two feldspars within the rapakivi.

Hornblende rapakivi granite (G_{5b})

Two bodies of hornblende rapakivi occur in the east of the area to the east-south-east of Tasiussaq bay but granite of this type appears

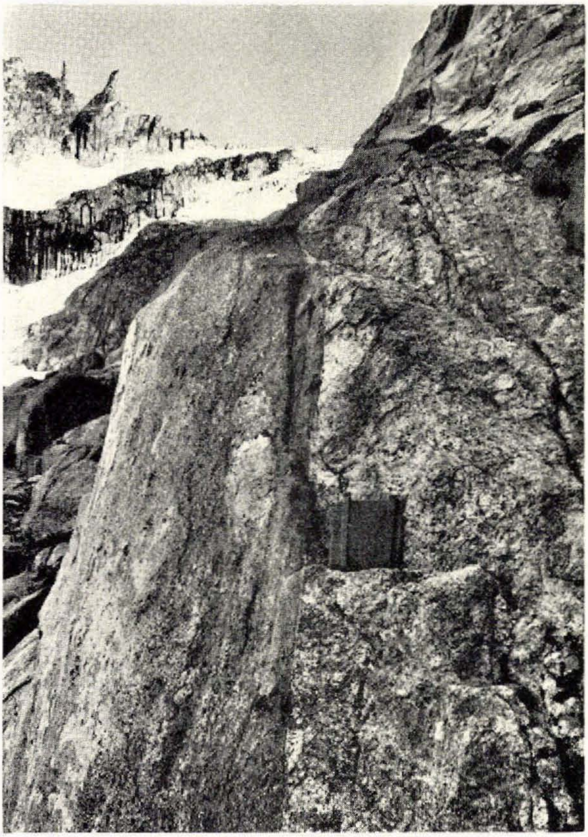

Fig. 42. Sharp contact of the hornblende rapakivi granite (G_{5b}) with G_2 granite and included gneiss of the high-grade gneiss complex. Contact is knife-sharp, steep and discordant to the structure within the G_2 granite (right). The map-case is 30 cm in length. South of Itivdlerssuaq.

to have a much wider extent on the peninsula to the north of Frederiksdal, to the east of the Tasiussaq area. The bodies have knife-sharp, discordant contacts to the gneisses and granites of the high-grade gneiss complex, dipping from 75° to vertical, generally inwards (Fig. 42).

The age relationship of this rapakivi to the biotite rapakivi is based on evidence from basic dykes since the two rapakivis outcrop in different areas. NW-striking basic dykes are sharply truncated at the con-

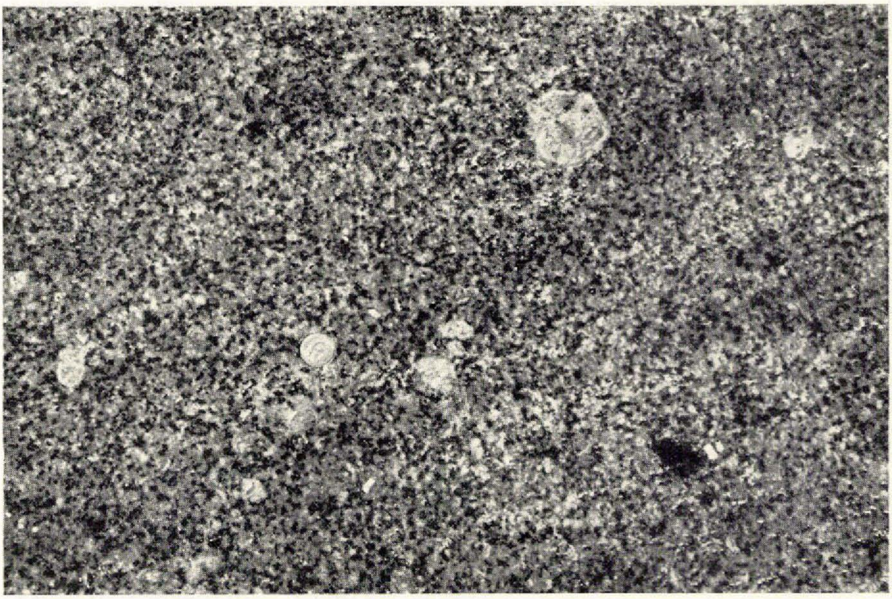

Fig. 43. Hornblende rapakivi granite (G_{5b}) illustrating the sparse distribution of potash feldspar megacrysts, some of which are mantled by plagioclase. The diameter of the coin is 2.8 cm. South of Itivdlerssuaq.

tact of the hornblende rapakivi (see DAWES, 1968, fig. 14) but dykes of the same swarm penetrate the biotite rapakivi. The hornblende rapakivi is thus considered to be a younger phase of intrusion than the biotite rapakivi.

The granite is porphyritic and medium-grained (Fig. 43) varying in colour from dark grey on fresh surfaces to brown on weathered surfaces. Hornblende is the dominant mafic mineral but it can be exceeded in volume percentage by biotite and less frequently by orthopyroxene. Potash feldspar, perthite, plagioclase and quartz are the main felsic minerals while clinopyroxene, apatite, sphene and ore form the main accessories. The porphyritic texture is produced by microcline megacrysts which are perthitic and which have an irregular distribution throughout the granite. Rapakivi texture is locally displayed. Local saucer-shaped mafic layering exists. Inclusions of country rock are few but large xenoliths of gneiss up to 250 m in length occur in some of the higher parts of the rapakivi.

Samples of similar granite, somewhat more greenish grey in colour and characterised by hypersthene, were collected during reconnaissance mapping of the west side of Torssukátak fjord to the east of the Tasiussaq area.

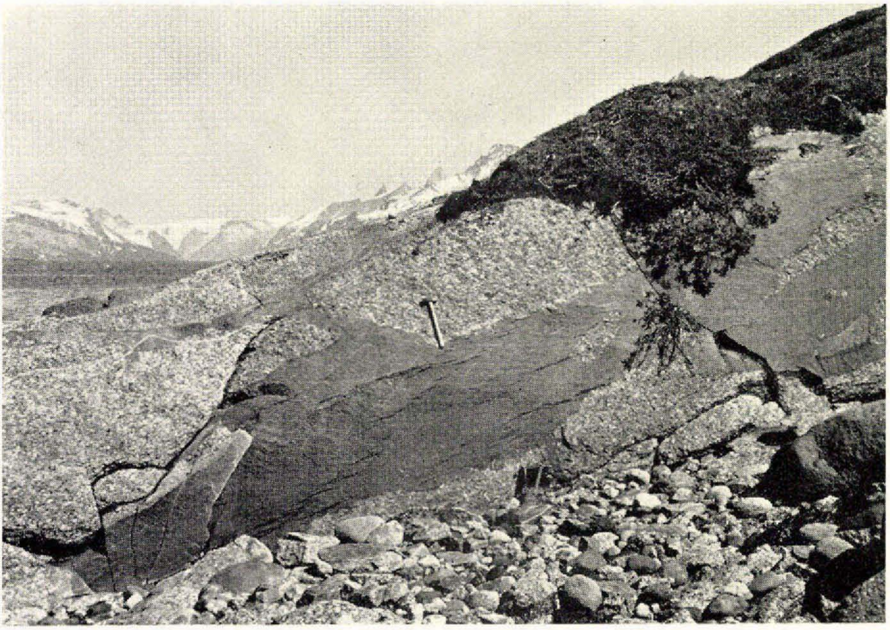

Fig. 44. Irregular branching intrusion of microgranite (G_6) within the biotite rapakivi (G_{5a}). Contacts are sharp and the microgranite cuts the potash feldspar megacrysts of the rapakivi but is cut by small AP_6 leucocratic veins (top left). Tasermiut fjord coast, west of Tasiussaq bay. The Inland Ice at the head of the fjord can be seen in the distant background (top left).

Microgranite (G_6)

There are small occurrences of microgranite in the northern part of the area along the Tasermiut fjord coast. The granite forms sheets, dykes and irregular branching masses characterised by sharp, discordant contacts to the host rocks (Fig. 44). On Naujat nûat the granite forms three parallel dykes, the largest of which is 50 m in width. Apophyses are common. The dykes have cross-cutting relationships with the gneisses of the high-grade complex and perpendicular offsets at oblique intersections indicate a dilatational mode of emplacement.

The microgranite is fine-grained and homogeneous but in places it is slightly porphyritic with small feldspar crystals up to 1 cm. Main minerals are microcline, plagioclase and quartz with smaller amounts of mica. Zircon and apatite form accessories.

The microgranite post-dates the formation of the biotite rapakivi granite (G_{5a}), and bodies of the microgranite cut the late feldspar megacrysts of the rapakivi, the associated pegmatites, aplites and quartz veins (AP_5) and the late NNE-NE shearing of the rapakivi.

The outcrops in the Tasiussaq area are the easterly extensions of a much larger body of microgranite which exists on the Nanortalik peninsula and which has been mapped by ESCHER (1966) and called "granite C".

Aplites and pegmatites

At least three phases of aplite and pegmatite formation (here termed AP_4, AP_5, AP_6) occurred during the genesis of the granitic complex, each phase being genetically connected to an episode of granite formation. This can be established from the age relationships of the pegmatites and aplites and the G_4, G_5 and G_6 granites. However, only the age of a limited number of pegmatites and aplites can be established with any certainty since a pegmatite cutting the G_4 granite may be of AP_4, AP_5 or AP_6 age unless it can be shown to pre-date either G_5 or G_6. Furthermore, the aplites and pegmatites are not spatially restricted to rocks of the granitic complex and many occur cutting the high-grade gneiss complex. In these cases it is difficult to establish the age of the aplites and pegmatites unless individual dykes can be traced from older gneisses into rocks of the granitic complex or can be seen in cross-cutting relationship with basic dykes of known age.

Four main genetic types of aplites and pegmatites can be recognised: dilatational types, replacement types, recrystallised shear-zone pegmatites, and composite aplite-pegmatite dykes and sheets. Numerous generations of each type exist and local chronologies of aplites and pegmatites can be erected from cross-cutting relationships. All the pegmatites and aplites occur as dykes or sheets all below 5 m in width; they are composed of potash feldspar, plagioclase, quartz, with or without micas and garnet. Beryl, tourmaline, sphene, apatite and magnetite have been noted as accessory minerals.

The dilatational types are characterised by sharp, straight, discordant contacts to the host rocks. Perpendicular offsets at oblique intersections typify these pegmatites and aplites and it is considered that they have been formed by intrusion of acidic fluid into dilating fissures. The presence of the pegmatites and aplites in local swarms supports the view that the introduction of material took place along fracture-zones.

Replacement pegmatites and aplites tend to be more irregular than the dilatational types but some straight dykes and sheets occur. These are, however, characterised by the absence of perpendicular offsets at oblique intersections. The replacement nature of the pegmatites and aplites is supported by basic rims which flank some of the dykes or sheets and which are indicative of extraction and diffusion of felsic material from the host rock into the pegmatite or aplite zone (Fig. 45).

A few pegmatites can be demonstrated to have resulted from the recrystallisation of shear-zones and the transition from a shear-zone,

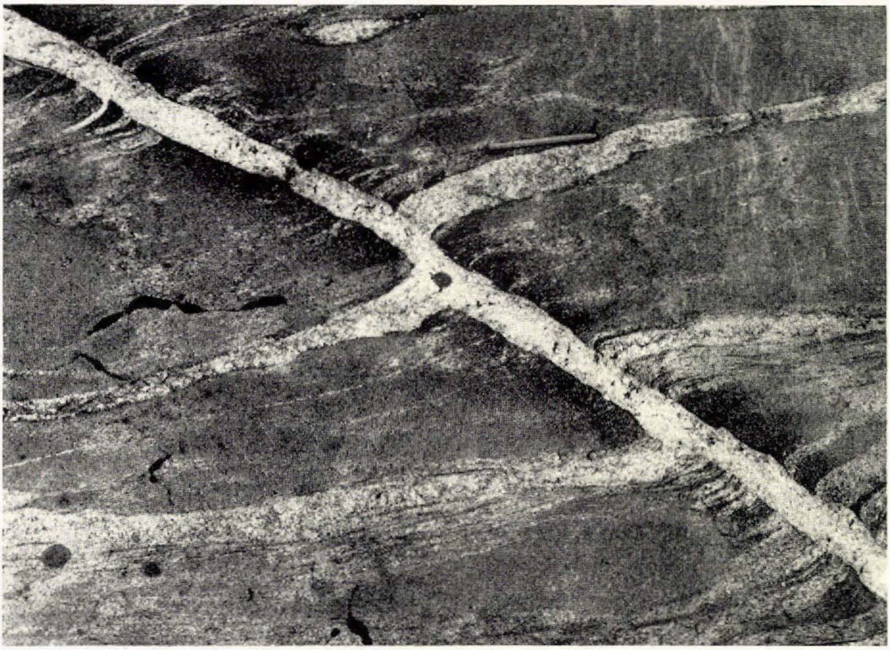

Fig. 45. A replacement pegmatite which has developed in the axial region of a shear-fold in the charnockitic gneiss. The dark tract in the gneiss adjacent to the pegmatite is the result of the extraction and diffusion of felsic material from the gneiss into the pegmatite. East of Tasiussaq bay.

Fig. 46. Recrystallised shear-zone pegmatite in microcline granite (G_4). Relic tracts of the shear-zone can be seen within the pegmatite. Lens-cap has a diameter of 4.5 cm. South of Tasiussaq bay.

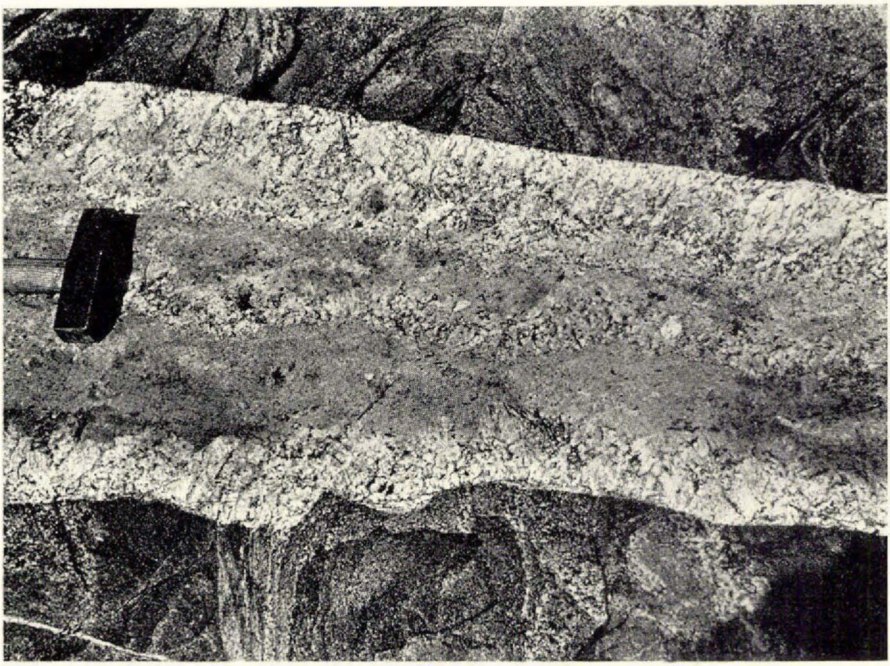

Fig. 47. Composite aplite-pegmatite cutting biotite gneiss. The dyke is composed of two pegmatite margins and a central zone; the aplite forming two strips. Note the relic patches of aplite (top left) within the pegmatite and the encroachment of feldspar megacrysts into the aplite. The dark mineral is biotite. Pingoq.

through a feldspathised zone in which sheared rock forms relic tracts (Fig. 46), to a pegmatite with sharp contacts and no sign of sheared material, can be photographed along the strike of a single shear-zone.

Composite aplite-pegmatite dykes and sheets resemble the dilatational types in that they have sharp, straight, discordant contacts to the host rocks and commonly occur in local swarms. They are composed of both aplite and pegmatite material, the spatial relationship of which varies in different dykes and sheets or even in a single body. The most frequent relationship is the presence of margins of pegmatite flanking an aplite core although a central pegmatite strip may occur in addition (Fig. 47). The internal features of the dykes and sheets indicate that the pegmatite has replaced the aplitic material and that the aplitic material was formed by injection of a fluid phase into fractures. Salient criteria are:

1. Movement perpendicular to the length of the body has taken place at oblique intersections. Where two composite sheets or dykes intersect the contact has been healed by the later growth of pegmatitic material (Fig. 48).

2. Disorientated inclusions of host rock within the aplitic material are common (Fig. 49).

Fig. 48. Cross-cutting aplite-pegmatite sheets in biotite gneiss, the younger sheet (centre left to bottom right) having caused perpendicular offset at the oblique intersection with the older sheet. Later growth of pegmatitic material has partially healed the original junction between the two sheets. South of Pingoq.

Fig. 49. Composite aplite-pegmatite dyke in charnockitic gneiss. Pegmatite composes the left half of the dyke except for a marginal zone of aplite. Aplite forms the right half of dyke and also occurs as relic tracts in the pegmatite. The dyke is discordant to the foliation of the gneiss (parallel to the hammer shaft) and a disorientated gneissic inclusion occurs in the aplite. East coast of Tasiussaq.

Fig. 50. Composite aplite-pegmatite dyke cutting charnockitic gneiss. Two main tracts of pegmatite and a smaller central one exist in an aplite dyke. Mafic layering in the aplite material (centre left) is sharply truncated by the pegmatitic material. East coast of Tasiussaq.

3. The encroachment on the aplite by the pegmatitic feldspars (Fig. 47).

4. Relic patches of aplite within the pegmatite (Fig. 47).

5. Primary mafic banding (biotite) in the aplitic material is truncated and overgrown by pegmatitic material (Fig. 50).

The formation of the composite aplite-pegmatites is considered to have been very similar to that suggested by STONE and AUSTIN (1961) for the aplite-pegmatite relationships in the granitic rocks of western Cornwall.

A main phase of pegmatite and aplite formation (AP_4) was connected with the reactivation of the gneisses of the high-grade gneiss complex and the genesis of the microcline granite (G_4). Examples of all four types of aplite and pegmatite exist of this age. AP_5 aplites and pegmatites post-date the intrusion of the rapakivi granites (G_5) and the development of the microcline megacrysts of the granites. Aplites are more abundant than pegmatites and some pure quartz veins exist but numbers are small compared to the AP_4 aplites and pegmatites. AP_6 pegmatites and aplites post-date the formation of the microgranite

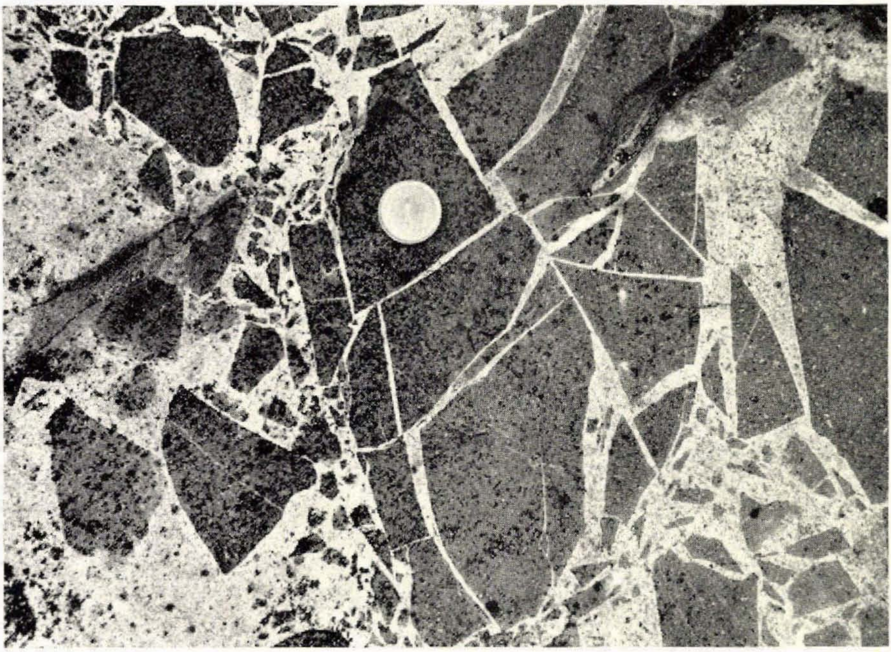

Fig. 51. Brittle agmatisation caused by the introduction of a liquid acid phase into consolidated dolerite. The dolerite has acted in a competent way. The basic blocks are angular in shape and the acid veins regular (*cf.* fig. 52). The lens-cap is 4.5 cm in diameter. East coast of Tasiussaq bay.

(G_6) but due to the restricted occurrence of this granite few dykes of AP_6 age can be established. Many veins and dykes cutting the biotite rapakivi granite (G_{5a}) may be of AP_6 age.

Basic intrusions

Basic rocks associated in time with the biotite rapakivi granite

A suite of basic rocks including gabbro, dolerite, norite and diorite, was emplaced after the initial intrusion of the biotite rapakivi granite but before the late stages in its genesis. The basic rocks form small bodies and dykes. They are not restricted spatially to the biotite rapakivi massif and some dykes cut the rapakivi contacts and penetrate out from the granite into the surrounding high-grade gneisses.

Approximately thirty-five dykes have been recorded and these lie in two main directions: N 10–30°E and N 10–30°W. Within the high-grade gneisses the dykes are continuous and have sharp, chilled contacts but in the rapakivi they have a less continuous and more irregular form (see Dawes, 1968, fig. 15). The dykes vary in width from 10 cm to 6 m

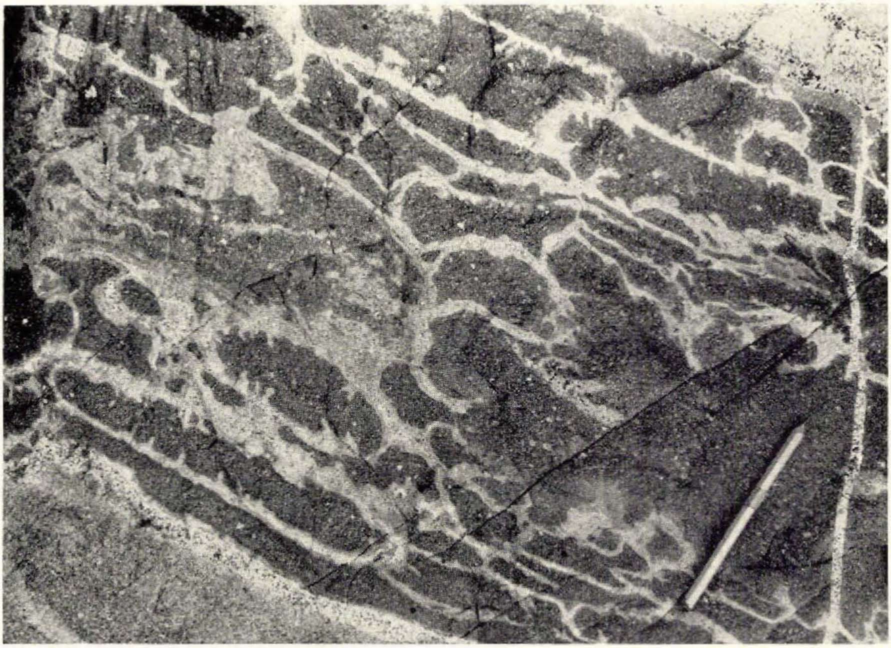

Fig. 52. Malleable agmatisation or "net-veining" caused by the introduction of a liquid acidic phase into unconsolidated dolerite. The acid veins are characteristically irregular and dentate and the enclosed blocks of dolerite are rounded (*cf.* fig. 51). South-east of Tasiussaq bay.

and in attitude from 70° to vertical. Thirteen bodies have been mapped, all except one occurring within the biotite rapakivi massif. The bodies are either flat-lying lens-shaped masses or true discordant bosses. The largest has an approximate area of 2.5 km.

The basic rocks vary from unmetamorphosed fresh rock types with no sign of alteration of the original pyroxene, through rocks showing stages of replacement of the pyroxene by hornblende into hornblende-plagioclase rocks. Sub-ophitic, ophitic and microporphyritic textures are preserved in all stages and no secondary lineation or foliation is developed in the amphibolitised rocks. The dykes in the south-south-east farthest away from the biotite rapakivi granite are fresh dolerites which become progressively more metamorphosed towards the rapakivi. The same transition exists in some of the larger bodies within the rapakivi where fresh gabbro and norite in the centre of the bodies become transformed to metagabbro and metanorite and finally to amphibole-plagioclase rocks in the peripheral zones of the bodies. A few dykes in the rapakivi display granular textures caused by intrusion and crystallisation of the basic rocks in unconsolidated granite. The nature and metamorphism of the basic rocks has been dealt with in some detail elsewhere (DAWES, 1968).

Fig. 53. Granite pipes in dolerite of the biotite rapakivi granite (G_{5a}). The surface of exposure is approximately at right angles to the vertical extent of the pipes, so that shapes represent true cross-sections of the pipes. The pipes are genetically connected to the "net-veining" which occurs in the basic rocks of the rapakivi and transitions from pipes to "net-veins" can be seen in places. North-west of Tasiussaq bay.

The basic rocks both inside and outside the rapakivi massif have been agmatised by acid material in a pattern which has been controlled by the state of consolidation of the basic material at the time of introduction of the acid. Two main types can be recognised, transitions occurring between them. Brittle agmatites formed where the basic material had consolidated at the time of acid intrusion (Fig. 51) while in the typical "net-veined" basic rocks (malleable agmatites) the basic material was unconsolidated (Fig. 52). Abundant evidence exists for the coexistence of basic and acid liquid phases and in places the basic material has chilled against the acid. Occurrences of granite pipes in dolerite are common (Fig. 53) and these are interpreted as due to the upward rise of the acid liquid phase into the denser unconsolidated dolerite. In some cases chilling has taken place in the material around the pipes. The pipes simulate those described by Elwell (1958) from Slieve Gullion and by Elwell, Skelhorn and Drysdall (1960) from Guernsey. The mechanism of "net-veining" and the various contact effects the basic rocks adjacent to the acid (*i.e.* chilling, pseudochilling, mineral alterations) will be dealt with in detail in a later paper on the rapakivi granites of the Tasiussaq area.

A significant point is that these basic rocks were emplaced into the rapakivi granite after the formation of the inclusion breccias and the early magmatic stage of the granite but pre-date the formation of some of the feldspar megacrysts of the rapakivi (Fig. 41). Feldspars grow into the basic rocks within the rapakivi and in places have obliterated the contacts between the dykes and the granite, and also between the agmatising acid veins and the basic rock. Thus the basic rocks provide a means of separating the rapakivi into two stages, a magmatic stage giving rise to its form with discordant and sharp contacts, inclusion breccias and mafic layering, and the non-magmatic metasomatic stage producing the present rapakivi texture.

Metamorphosed basic rocks later than the rapakivi granites

Two types of basic intrusions can be recognised in the Tasiussaq area which post-date the rapakivi granites but which are considered to be part of the granitic complex. The two types have not been seen in contact and their relative ages are unknown. The metadolerite dykes are considered to pre-date G_6 since acidic veins of presumed G_6 age cut the dykes. The diorite sills could be of the same age. Both types are taken to represent late intrusions connected to the basic rock suite described above.

Metadolerite dykes

Few basic dykes of this type exist in the Tasiussaq area but some dykes cutting the high inaccessible peaks to the east of the Tasiussaq area are probably of a similar age. Other dykes cutting rocks of the high-grade gneiss complex may be of a comparable age but it is difficult to establish accurately their age unless there are cross-cutting relationships with other basic dykes or aplites and pegmatites of known age.

The dykes are partially recrystallised dolerites being fine- to medium-grained and bearing relic sub-ophitic textures. They vary from grey to greenish in colour and occasionally display fine-grained chilled margins. All the dykes have a trend between N 15° W and N 60° W and are steep or vertical in attitude. Some dykes have been altered more than others and in places the dyke rock has been severely sheared and chloritised producing a distinct greenish colouration, a colour typical of most of the dykes. Pyroxene has been partially or completely replaced and altered to amphibole or chlorite and the plagioclase has been altered, sericite and saussurite forming the main alteration products. Biotite is an important mafic mineral in some dykes while magnetite and apatite are accessory.

The dykes cut the high-grade gneiss complex, the microcline granite (G_4) and the hornblende rapakivi granite (G_{5b}) but all the dykes are cut by epidote shears, and near the Tasermiut fjord coast by small irregular acid veins. The acid veins are probably of G_6 age although they cannot be directly connected with the outcrops of microgranite (G_6) which occur on the Tasermiut fjord coast. The dykes seem to have been intruded prior to the NNW–NW shearing movements now present throughout the biotite rapakivi massif as epidote-filled shears which post-date the formation of the feldspar megacrysts of the rapakivi.

Diorite sheets

At least two generations of diorite sheets are present in the Tasiussaq area, both having similar morphological and compositional characters. One cross-cutting dilatational relationship has been noted to the north-east of Tasiussaq bay where two parallel sheets, thought originally to be of the same generation, vary in attitude and one sheet cuts the other.

The sheets cut the gneisses of the high-grade gneiss complex and clearly post-date the microcline granite (G_4), the biotite and hornblende rapakivi granites, and the associated basic intrusions of the granitic complex. The relationship to the microgranite (G_6) is not seen but the sheets are cut by the NNE dolerite dykes (see p. 112) and by later NW fault movements. The diorite in places is heavily sheared.

The sheets outcrop in two main areas: to the south-west of Tasiussaq bay and to the east of Usuk. They form branching sheets which characteristically form topographical terraces at the back wall of which the diorite is exposed. The sheets vary in thickness from a few centimetres to 6 m and they are characteristically flat-lying varying from horizontal to slightly dipping up to 25°. In one locality a 45° dip is found. The sheets have discordant, sharp contacts to the host rocks and in places distinct, chilled margins.

The diorite varies from fine- to medium-grained and in colour from grey to greenish grey, and in sheared rock to green. Locally coarser-grained patches exist in the diorite where hornblende crystals reach 1 cm in length. The diorite is composed essentially of hornblende and plagioclase with lesser amounts of pyroxene and biotite, while magnetite, sphene and quartz form the accessories. Both brown and green hornblende occur commonly replacing the pyroxene. The plagioclase displays albite twinning and is often zoned. It forms randomly orientated crystals and no sub-ophitic texture has been seen even in the types bearing primary pyroxene. In some contact samples the hornblende crystals and also the plagioclase laths are aligned parallel to the contact; elsewhere the plagioclase is situated interstitially between the pyroxene and hornblende. Zeolite cavities exist in some types.

Three types of variation occur in the macrostructure of the diorite:

1. A crude rhythmic layering in the larger sheets parallel to the contacts of the sheets. Individual layers vary from 12 cm to 25 cm in width and commonly show an increase in grain size from top to base. The bases of such layers are more melanocratic than the top. The contacts of the layers are usually distinct.

2. Fine-grained bands and tracts with sharp contacts within normal medium-grained diorite.

3. An anastomosing veining of the diorite by small leucrocratic veins.

The sheets have been emplaced along horizontal fractures or shear-zones in the host rocks and the irregular forms seen in some places illustrate the control of shape by the pre-existing fracture pattern. It is considered that multiple intrusion produced the distinct fine-grained bands within the sheets, the rhythmic layering being the result of a gravitational crystal settling in the intruded basic magma. It is also possible that different pulses of magma were connected with the layer formation. The leucocratic veining represents a late stage feature in the formation of the sheets, probably representing the late residual acidic fluid from the magma.

Ultramafic dykes

Ultramafic dykes have been mapped at five localities on the peninsula to the north-west of Frederiksdal. The dykes tend to weather readily being exposed in scree-filled gullies. Hence exposures are poor. The dykes have trends varying between NNW and NNE with steep or vertical attitudes and they vary from 20 cm to 5 m in width. They have sharp, discordant contacts but are generally impersistent.

The rock varies from green to greyish green in colour and from medium- to coarse-grained. Samples from some dykes have a slight flecked appearance due to the presence of plagioclase clusters. No layering has been observed. The dyke rock varies in mineral composition from dyke to dyke and some dykes have been severely altered. All the dykes are ultramafic in composition and plagioclase never reaches more than 10 $^0/_0$ of the total volume; usually it is absent. Main minerals are hornblende (both brown and green varieties), colourless amphibole, clinopyroxene and biotite with lesser amounts of calcite, chlorite, sericite, epidote and ore. Some calcite veins cut the dykes.

The chronological position of the dykes is obscure. They cut the gneisses and G_3 granite of the high-grade gneiss complex but the relationship to the rocks of the granitic complex is not clear. They are however cut by late NW faults and they have been sheared and epidote veined. It is tentatively suggested that they post-date the formation of the rapakivi granites (G_5) and are late-stage ultramafic intrusions possibly of similar age to the ultrabasic rocks mapped by BERRANGÉ (1967) and commented on by WALTON (1967).

Age of the granitic complex

As stated earlier (p. 81) an isotopic age determination on augite from the Frederiksdal rapakivi massif has given an age of 1645 ± 50 m.y. This rapakivi massif borders the Tasiussaq area on the south and outliers of the rapakivi occur within the Tasiussaq area as a result of the shallow inward-dipping contact of the massif. The Frederiksdal rapakivi is of the same suite and is the same age as the biotite rapakivi granite of the Tasiussaq area (G_{5a}, see p. 89) and thus, following the proposed age divisions of the Precambrian of South Greenland (see BRIDGWATER, 1965), the biotite rapakivi of the granitic complex is Sanerutian in age. It is considered that the genesis of the granitic complex of the Tasiussaq area took place in Sanerutian time. Using the chronological terminology given by ALLAART (1967, pp. 7,132) in which the Ketilidian has been extended to include the Sanerutian events, the genesis of the granitic complex can be correlated with the second episode of Ketilidian plutonism.

LATE EVENTS

Certain late events, divisible into two categories, can be recognised in the Precambrian history of the Tasiussaq area and these are considered to post-date the main plutonic conditions which prevailed during the genesis of the granitic complex. These events are the intrusion of basic magma into fractures giving rise to dolerite dykes, and tectonic events including faulting and mylonitisation which in places are associated with some secondary mineral alteration. Whether such events took place during true cratogenic conditions cannot be clearly elucidated.

Dolerite dykes

Two main directions of dolerite dykes can be distinguished in the Tasiussaq area cutting rocks of both the high-grade gneiss complex and the granitic complex. A NNW set occurs mainly to the west of Tasiussaq and a NNE set exists to the east of Tasiussaq. No intersections of these have been observed and thus the age relationship of the two directions is not directly apparent. Three dykes exist which have trends varying from WNW to WSW. The NNW and NNE dykes were intruded between two phases of faulting and both dyke directions are cut by the main ENE–WSW to E–W faults but not by the NW faulting movements. The relationship of dykes of the third direction to such faults is not known. By reference to adjacent areas where intersections between similar sets of basic dykes occur (Oen, 1961) the following dyke chronology is tentatively suggested:

NNW dykes
NNE dykes
oldest — WNW to WSW dykes

All dykes have straight, constant strike directions, parallel sides, and chilled margins to the adjacent country rocks. They are characterised by a brown to reddish-brown colour which accentuates their presence especially where they are heavily weathered. The dykes are usually regularly jointed and commonly display spheroidal weathering. The rock type of the dykes is a homogeneous dolerite but in some dykes plagioclase phenocrysts up to 3 cm exist. No internal layering has been observed. Inclusions of country rock have been noted in a NNW dolerite at one locality.

Seven NNW dykes exist varying in width from a few centimetres to 15 m. The majority are vertical in attitude (Fig. 5) but some are inclined steeply to the west. Five dykes with a general NNE trend have been noted and these vary in width from 20 cm up to 15 m. They are mainly vertical in attitude or steeply inclined to the east. The dykes varying about a general E–W direction are small, all below 1.5 m in width.

The rocks vary from fine- to medium-grained and vary in texture from sub-ophitic to ophitic. Essential minerals are augite (brown and colourless varieties) and plagioclase; occasionally there is abundant magnetite. One of the E–W dykes is olivine-bearing. Main accessories are apatite and ore. Sphene is rare. In the NNW dykes calcite amygdules occur. Chlorite and some amphibole (uralite), derived from pyroxene, exist in the oldest dykes (WNW) and occasionally in the NNE dykes. Plagioclase in these dykes is cloudy and small amounts of epidote and sericite exist in places as hydrothermal alterations of the feldspar.

Faulting and mylonitisation

Two main fault directions are recognisable in the Tasiussaq area together with other less conspicuous fault trends. A main phase of NW faulting occurred prior to the intrusion of the NNW and NNE dolerite dykes and a second main phase producing mainly ENE faults, occurred after the dyke intrusion. Both these fault phases have produced morphological features in the area and also some lateral displacement of geological features. Mylonites and crush-zones are common especially in connection with the NW faulting. Some epidotisation of fault-zones has occurred as well as the development of chlorite in thin films on small fault surfaces. Occasional quartz veining has been noted in crush- and mylonite-zones and in places some iron mineralisation is marked by haematite staining.

A NW fault has controlled the square end of Tasiussaq providing a striking example of a fault-controlled fjord. Many valleys and depressions owe their existence to faults, the long valley Kûgssuatsiaq extending across the area being an example. Movements along the faults seem to have been generally vertical but some transcurrent movement is indicated where suitable marker horizons exist. The largest lateral movement detectable is about 1.7 km shown by the sinistral displacement of the G_1 granite horizon along a NW fault on the peninsula to the north-west of Frederiksdal. Dextral movement has occurred along some of the ENE faults and this is well illustrated by the offsets of the NW dolerite dykes in the west of the area, south of Pingoq. The largest displacement here is approximately 130 m.

Lack of suitable marker horizons in the gneisses and the scarcity of late basic dykes prevent an accurate survey of the movement statistics of the faults and it is not known whether some faults were initiated during the development of the high-grade gneiss complex or the granitic complex and have been rejuvenated.

Age of the late events

No absolute age determinations have been carried out on the younger basic dykes of the Tasiussaq area. However the late events most probably correspond with the Gardar dolerite dyking and tectonism which have been recognised and dated further to the north-west in South Greenland. One sample of dolerite from a NNW dyke in the Tasiussaq area has been investigated palaeomagnetically and TARLING (personal communication, see also 1966) has found a remanent magnetic direction corresponding to that known from other Gardar dykes. Gardar activity in South Greenland occurred approximately between 1150 m.y. and 1430 m.y. ago (BRIDGWATER, 1965). Whether any of the fault movements are younger in age than the Gardar, possibly connected to the Grenville activity of the Canadian Shield, cannot as yet be established.

NATURE OF THE PRECAMBRIAN EVOLUTION

One of the main conclusions to emerge from the study of the Precambrian rocks of the Tasiussaq area is that the geology of the Tasermiut fjord region is complex and by no means fully understood. Geological investigations in this part of South Greenland have reached a stage which allow only for a number of possibilities to be put forward regarding the relationship of the rock types, but which do not permit a straightforward statement on the nature of the Precambrian evolution. Further basic field work in the region, supported by sedimentological studies and a programme of radioactive age determination, is essential before any of these possibilities can be safely eliminated. Although the author believes that the granitic gneisses, gneisses, schists, metasediments and metavolcanic rocks of the Tasermiut fjord region have not all been derived from a *single* supracrustal pile, the precise age, position and status of older crystalline rocks cannot be firmly advanced with the present state of knowledge. Since these older crystalline rocks may be pre-Ketilidian or Ketilidian in age and supracrustal rocks of more than one age may occur (see p. 85), regional correlations of gneiss and supracrustal rocks in South Greenland should be made with extreme caution.

The Precambrian evolution of the Tasiussaq area included at least one period of sedimentation and volcanism followed by plutonic activity including metamorphism, deformation, migmatisation, granitisation and the intrusion of granites and syn-plutonic basic and ultramafic rocks. Late basic dyking and fracturing indicate a trend in the crust towards brittle non-orogenic conditions towards the end of the Precambrian history.

It can be assumed that arenaceous and argillaceous material together with subsidiary calcareous rocks were deposited on an old crystalline basement. Volcanicity also occurred during this depositional phase (represented by basic schists in the pelitic rocks) and it might be considered that both the sediments and volcanics were formed during geosynclinal conditions although the geosynclinal nature of these rocks is an assumption. Little is known about the depositional conditions of the supracrustal rocks of the Tasermiut fjord region and the possibility of the psammitic and psephitic rocks representing flysch-type deposits has already been mentioned (see p. 84).

Some basic sills and dykes were intruded following the formation of the youngest supracrustal rocks of the region and these are represented by basic granulite sills in the Tasiussaq complex and by basic sills in the upper part of the supracrustal succession on the Nanortalik peninsula (ESCHER, 1966). Some of these intrusions may be the same age as the metavolcanic rocks of the Nanortalik peninsula while some basic granulites in the Tasiussaq complex may pre-date this volcanicity having been intruded into older crystalline rocks (see pp. 85–86).

Plutonic conditions ("plutonic" in the sense of READ, 1957) prevailed after the formation of the supracrustal rocks and transformed the pre-existing rocks (supracrustal rocks and older basement) into the para-gneisses, metavolcanics and basic granulites of the high-grade gneiss complex. (If the psammites, psephites and overlying rocks in the Tasermiut fjord represent flysch-type sediments this plutonic activity may pre-date the deposition of these rocks). During the plutonism, metamorphism reached granulite facies conditions producing the high-grade mineralogy characterised by hypersthene. This metamorphism also most probably affected the G_1 granites of the complex, producing the ubiquitous hypersthene in rocks varying from charnockite to enderbite in composition. Fluctuations in the availability of H_2O in connection with some temperature changes occurred following the establishment of the high-grade mineralogy to produce local retrogressive mineral alterations.

During the plutonism widespread metasomatism and migmatisation of the gneisses occurred. Diffusion and migration of elements, mainly Na, Si and K, took place both on a local and large scale to give rise to the leucosome of the gneisses (AP_1, AP_2 and AP_3) and the G_2 and G_3 granites. Deformation produced at least four phases of folds accompanied by flattening, shearing and fracturing. The early phase of deformation (F_2), giving rise to both small- and large-scale isoclinal and recumbent folds and also nappe-like structures, was the most important. Thrusting and the tectonic transport of large masses of rock possibly occurred in the more brittle, upper parts of the supracrustal succession and there is a possibility that the meta-arkoses, metaconglomerates and meta-volcanics were carried over the older rocks (pelitic and hypersthene-quartzo-feldspathic gneisses) but the exact age of such thrusting relative to the early recumbent folding is not known. Some folding took place under plastic conditions.

A set of dolerite dykes, now represented by the pyroxene-meta-dolerite dykes, was intruded into the granites and gneisses of the high-grade gneiss complex following the main phases of migmatisation of the gneisses. These dykes were intruded into country rocks which were not cold at the time of intrusion, *i.e.* they are not comparable in type with true cratogenic or anorogenic dykes. The dykes were subsequently metamorphosed to pyroxene-metadolerites with the production of a high-grade metamorphic mineralogy characterised by the absence of hornblende and the presence of pyroxene in all stages of the transformation to metadolerite. It has been mentioned elsewhere that this metamorphism was essentially dipsenic in type and that it took place without appreciable deformation at temperatures which could be theoretically somewhat lower than those usually accepted as being typical for the granulite facies but in a water-deficient environment (see p. 40).

The genesis of the granitic complex is considered to have followed the development of the high-grade gneiss complex without lapse of a major period of time and no evidence exists in the Tasiussaq area for a major break in plutonic conditions separating the two complexes, *i.e.* no evidence of cratogenic basic dykes. However if the psammites and psephites in the Tasermiut fjord region, which clearly pre-date the granitic complex, represent syn-orogenic flysch-type deposits then clearly uplift of certain areas in this part of South Greenland has taken place prior to the development of the granitic complex. The age relationship of the meta-arkoses and metaconglomerates and the pyroxene-metadolerite dykes of the Tasiussaq area is not known but if flysch-type deposits do occur then the contemporaneity of the basic dykes and the sediments is not improbable. Both dykes and sediments now display high-grade mineralogy which may be the result of the same metamorphism due to a rise in the thermal front.

A rise in the position of the thermal front, coupled with a possible influx of water, was responsible for a general reactivation of the gneisses and the generation of the granites of the granitic complex. This plutonism is not considered divorced from the plutonic activity which took place during the genesis of the high-grade gneiss complex but rather as complementary plutonic activity. The granitic complex represents a culmination of the plutonism in the form of anatexis and the passage of the resultant magmas to higher levels in the crust producing granitic types in a "granite series". This anatexis took place in conditions unlike those that existed at the present level of exposure during the genesis of the high-grade gneiss complex and water must have been available for such granitic melts to have formed.

The microcline granite (G_4) of the granitic complex is an autochthonous granite having developed through granitisation and replacement of the gneisses of the high-grade gneiss complex, but in places it might be regarded as parautochthonous because there has been some mobility with inclusions of gneiss having been rotated and conspicuously disorientated in a way as to suggest local granitic intrusion. The rapakivi granites are late-plutonic in age and are essentially post-tectonic in nature, although they may have been accompanied by deformation at lower depths. The genesis of the rapakivi granites can be divided up into two stages: a magmatic stage, responsible for the emplacement of the massifs and their internal structures, and a later non-magmatic, metasomatic stage. The later microgranite (G_6) represents a late granite in the "granite series" being allochthonous in type. During the genesis of the three granite types a notable feature was the migration and diffusion of potash. As a result of potash metasomatism, microcline porphyroblasts in the microcline granite and the rapakivis are common in non-magmatic environments (*cf.* ESKOLA, 1956). Further evidence for the potash metasomatism is provided by the late development of potash feldspar in the composite aplite-pegmatite sheets. Migration of Na and Si was also of importance during the genesis of the granites.

No contact metamorphic aureole exists in the country rocks around the rapakivi granites, granites which were undoubtedly emplaced by magmatic intrusion. This might be taken to indicate that the granite and the country rocks were in approximate thermodynamic equilibrium at the time of emplacement, implying that the country rocks were at elevated temperatures under plutonic conditions (*cf.* SYLVESTER, 1964). Basic magma was intruded into such country rocks and also into the rapakivi in the form of small bodies and dykes. The basic magma in some cases was followed to the present level by acid material which intruded the basic producing composite dykes and bodies. During the waning of the plutonic conditions prevailing during the genesis of the granitic complex, further basic intrusion took place. A significant feature during the development of the granitic complex was the availability in the crust of both basic and acid magma and the ability of both types of magma to be intruded as dykes and bodies, often composite in nature, under plutonic conditions. A series of granitic and basic rocks of ages similar to those in the Tasiussaq area has been described by PERSOZ (1969) from an area some 50 km to the north-west of the Tasiussaq area.

The rapakivi granites of the Tasiussaq area are part of a suite of "New granites" present throughout South Greenland, some of which have been reviewed by BRIDGWATER (1963). BRIDGWATER, SUTTON and WATTERSON (1966) record "charnockitic" granulite facies rocks in connection with some of the rapakivi granites in the Kap Farvel area to the south-east of the Tasiussaq area. They record 1–2 km wide contact aureoles to the larger intrusions and state (p. 54) that the charnockitic border zones are "thought to be closely related to the emplacement of the rocks of the rapakivi suite". In the Tasiussaq area no such relation to charnockitic or granulite facies rocks exists. This apparent difference in contact characters of granites of the same suite may be due in some way to the varying amount of water present in the rocks as suggested by Bridgwater *et al.* (*op. cit.*), which may in part be connected to the depth of exposure. In the Tasiussaq area parts of the roof-zone of the rapakivi have been preserved and the highest levels of rapakivi development are exposed. No development of charnockitic rocks occurs to the north-west of the Tasiussaq area in connection with rapakivi granites, either in the Sydprøven granite mapped by Bridgwater (1963), in the granite mapped by Persoz (1969) or in the rapakivi on Sermersôq (Oen, 1961), nor do such rocks occur at the other rapakivi contacts existing in the Tasermiut fjord region (Escher, 1966; Wallis, 1966). Furthermore, the appearance of contact aureoles in which charnockitic rocks have developed from host rock gneisses may be controlled to some extent by the time relationship between the early granulite facies regional metamorphism and the intrusion of the rapakivi granites. It is possible that a large time interval separates the intrusion of various members of the so-called rapakivi suite in South Greenland so that all members are not of the same age. In the Tasiussaq area the rapakivi granites clearly post-date the early granulite facies regional metamorphism of the gneisses; in the Kap Farvel area the rapakivi granites are probably closer connected in time to the granulite facies regional metamorphism, with the result that the distinction between an early high-grade regional metamorphism and the intrusion of the rapakivi granites and associated rocks is less clear.

Following the plutonic conditions prevailing during the development of the granitic complex, it is assumed that a gradual change took place in the crust towards non-plutonic conditions by withdrawal of the thermal front. It is thought most likely that the dolerite dyking and fracturing took place after this withdrawal but little can be said about the actual crustal conditions during basic intrusion.

The early metamorphism, folding, migmatisation and granitisation might be referred to a compressive stage of a developing orogeny with the sequence granulite facies metamorphism-retrogressive mineral changes-migmatisation-granitisation-basic dyking-dipsenic metamorphism, recognised in the high-grade gneiss complex, representing a tendency for the crustal segment to rise due to isostatic recovery. The granitic complex represents the tendency for acidic magma, generated at depth by palingenesis during the early stages of the developing orogeny, to rise to higher levels in the crust, at the same time as which basic magma was also available to be intruded at high levels as dykes and small bodies. The late dolerite dyking and faulting are considered to be essentially post-plutonic or epeirogenic events occurring at the end of the orogeny. However, any strict analogy to the concept of a geological cycle (HARPUM, 1960) must be limited, since as stressed previously, the importance and role of older basement rocks, and even the exact age of the supracrustal rocks in this part of South Greenland, are not accurately known, and some rocks and events thus considered as part of a single geological cycle might well turn out to be part of an older cycle of events.

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Færdig fra trykkeriet den 5. december 1970

Table 1

Diagrammatical representation of the main Precambrian geological events of the Tasiussaq area. An attempt has been made to show the relative ages of events based on field and petrological relationships in the Tasiussaq area. No consideration has been given to the absolute age of the individual events and thus the vertical scale of the table is of little significance.

	Deposition and volcanicity	Tectonism	Metamorphism	Migmatisation	Granite formation	Aplites and pegmatites	Basic and ultramafic intrusion
← LATE EVENTS →		ENE to E faulting					
							NNW dolerite dykes NNE dolerite dykes E–W dolerite dykes
			Epidotisation				
		NW faulting					
↑							Two generations of diorite sills ? NNW ultramafic dykes
		Shearing and NE faulting					
						Aplite and pegmatite veins – AP_6	
					Intrusion of microgranite – G_6		
							NNW–NW dolerite dykes
			Epidotisation				
		Shearing and faulting					
						Quartz veins Aplite and pegmatite veins and dykes – AP_5	

GRANITIC COMPLEX					Potash metasomatism of the rapakivi granites		
			Amphibolite facies metamorphism	Agmatisation of the basic rocks associated with the rapakivi granites	Granitisation of the basic rocks associated with the rapakivi granites		
					Intrusion of hornblende rapakivi granite – G_{5b}		NNE dolerite dykes
				Net-veining of basic rocks in G_{5a}			Dolerite, norite and diorite in dykes and small bodies
							NNW dolerite dykes
				Agmatisation of the country rocks adjacent to G_{5a} rapakivi	Intrusion of biotite rapakivi granite – G_{5a}	Aplite and pegmatite veins and dykes – AP_4	
			Reactivation of older gneisses and granites		Microcline granite – G_4		
HIGH-GRADE GNEISS COMPLEX		F_4 deformation: large-scale open folds. Boudinage	Dipsenic metamorphism			? Regional pegmatites younger than M_3	
	Volcanicity Psammitic and psephitic sediments						Pyroxene-metadolerite dykes
				M_3 migmatisation	G_3 alaskite and garnet alaskite	M_3 pegmatite layers	
			Retrogressive metamorphism				
		F_3 deformation: open to close large- and small-scale folds		M_2 migmatisation	G_2 pyroxene-, biotite-, garnet-bearing granite and granodiorite	M_2 pegmatite and aplite veins and layers	
	?	F_2 deformation: isoclinal large- and small-scale folds. Nappe-like structures		M_1 migmatisation		M_1 aplitic veins and streaks	
			Granulite facies metamorphism				
					G_1 enderbite and hypersthene granite		
		? F_1 deformation: small-scale folds					
	Volcanicity Psammitic and psephitic sediments						
	Pelitic, psammitic and calcareous sediments with volcanicity	? Geosynclinal sinking					Basic granulite sills and ultramafic intrusions
	? Basement of folded gneisses and granites and other metamorphic rocks represented by pebbles in younger metaconglomerates and by possible areas of reworked gneiss. ? F_1 structures.						

PLATES

Plate 1

Aerial photograph of the Frederiksdal-Tasermiut fjord region of South Greenland seen from the south-west to illustrate the general physiography of the Tasiussaq area. Tasermiut fjord is on the left bordering the Tasiussaq area on the west and the entrance of Narssap sarqâ is on the right. The Inland Ice and nunataks are visible at the head of Tasermiut fjord and in the distant background. The low-lying islands in the foreground are part of the Frederiksdal rapakivi massif. They correspond to the strandflat of WEGMANN (1938a) and contrast markedly with the higher ground of the Tasiussaq area. The division of the Tasiussaq area into a lower western part and a much higher eastern and northern part is clearly seen. 5th June, 1948. Copyright, Geodetic Institute, Copenhagen, Denmark.

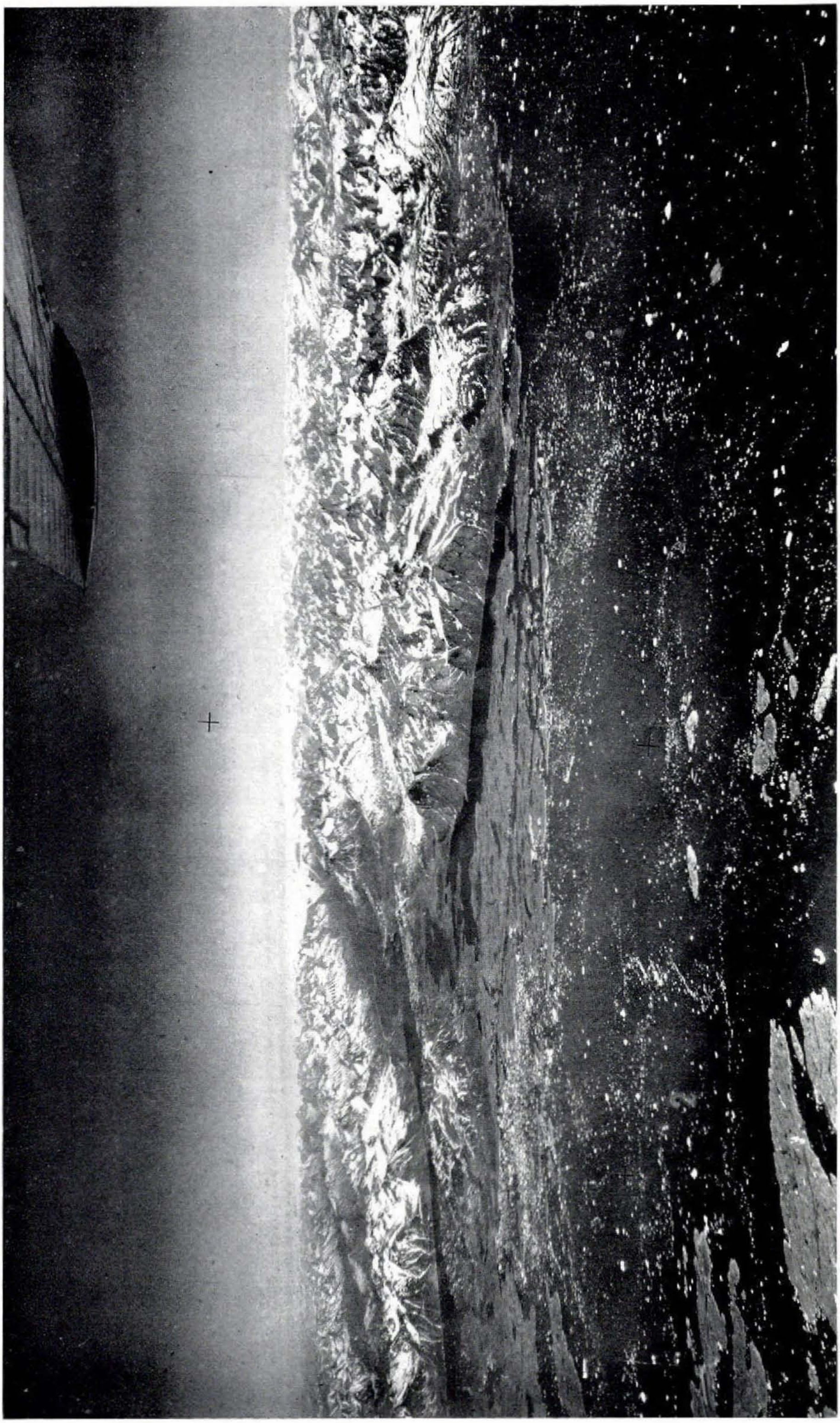


Plate 2

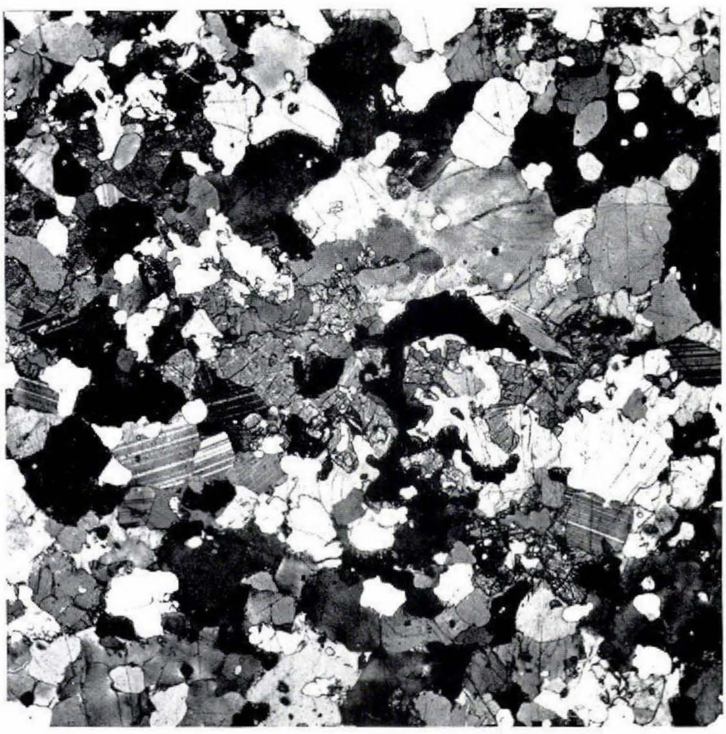
a.

Photomicrograph of biotite schistose gneiss showing a well-defined foliation of biotite flakes and a xenomorphic-granoblastic mesostasis of quartz and plagioclase. Garnet (top right) also occurs. GGU 58517, plain light, X 27. Text reference page 19.

b.

Photomicrograph of hypersthene-garnet-quartzo-feldspathic gneiss showing a typical felsic matrix of quartz and feldspar together with hypersthene and garnet (black centre). GGU 67163, crossed nicols, X 15. Text reference page 24.

a.



b.

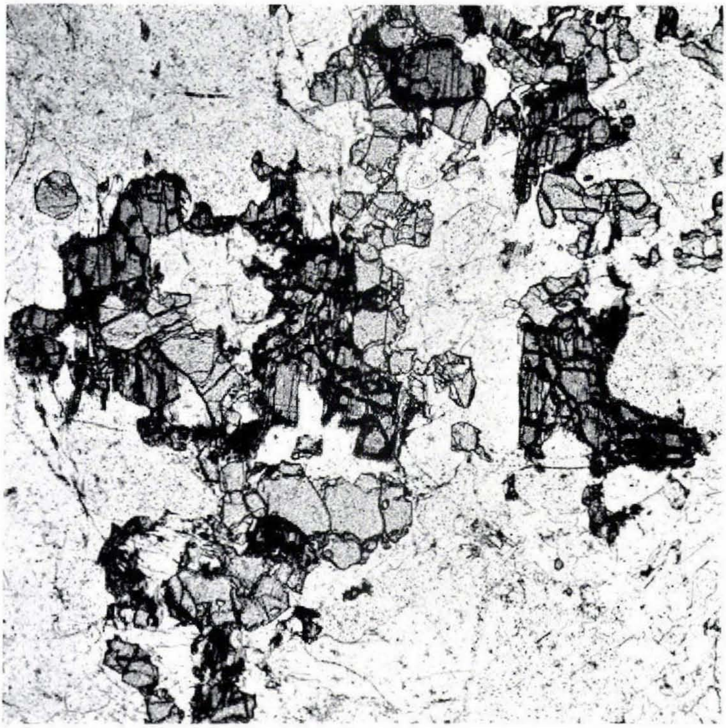
Plate 3

a.

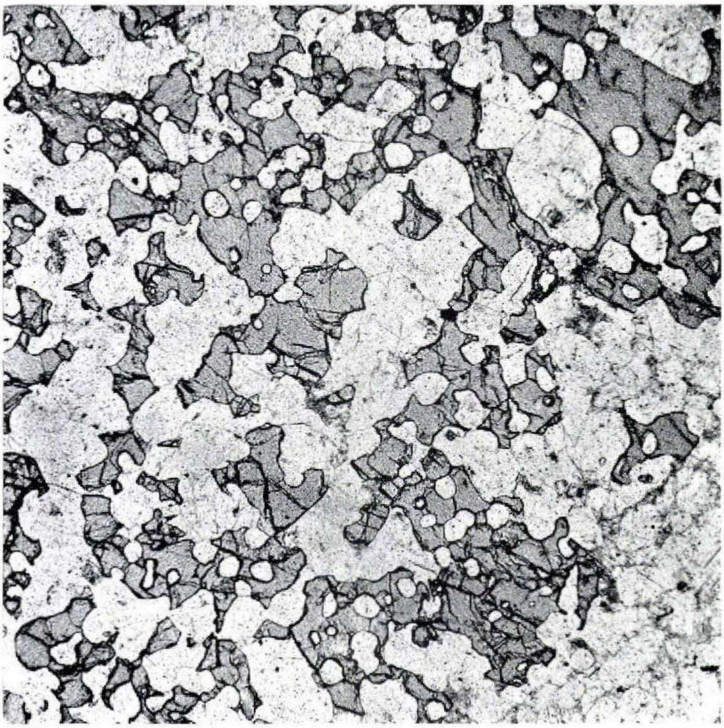
Photomicrograph showing the intimate association of garnet and hypersthene (dark) in hypersthene-garnet-quartzo-feldspathic gneiss. The white matrix is composed of perthite, plagioclase and quartz. GGU 58682, plain light, X 32. Text reference page 24.

b.

Photomicrograph of hypersthene-garnet-quartzo-feldspathic gneiss showing poikiloblastic texture of garnet. The rounded inclusions are mainly quartz but feldspar is common in the felsic matrix. GGU 58700, plain light, X 22. Text reference page 25.



a.



b.

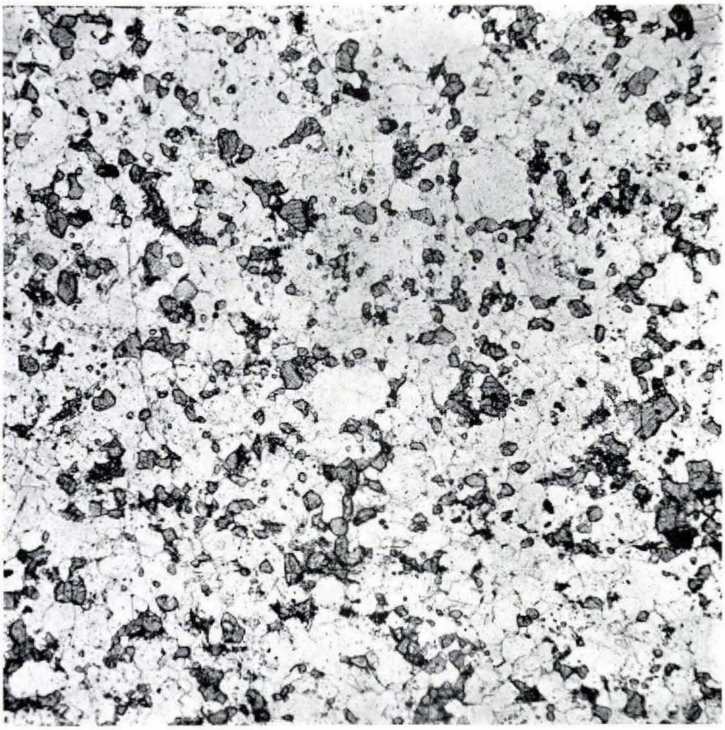
Plate 4

a.

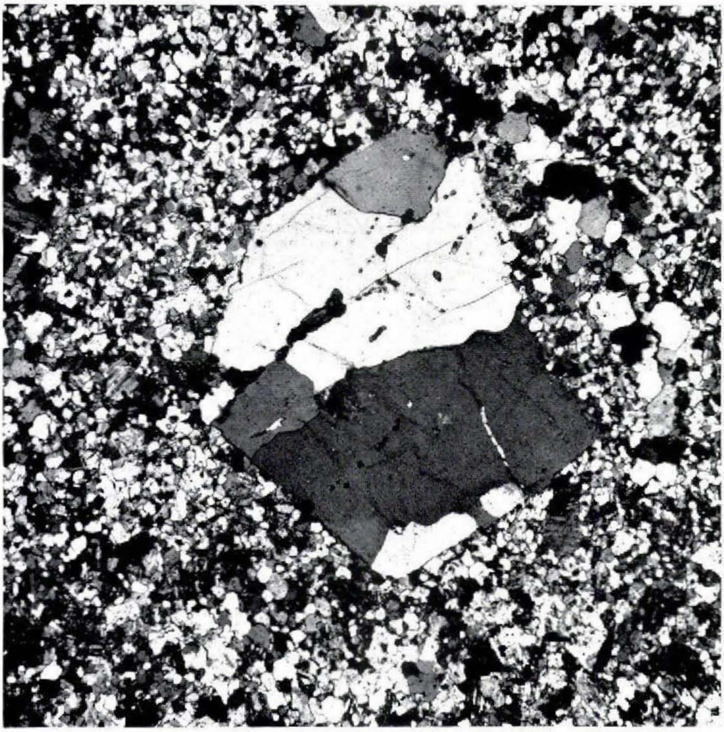
Photomicrograph of meta-arkose showing blastopsammitic texture composed of grains of clinopyroxene within a quartz-feldspar matrix. In places the clinopyroxene crystals clearly outline the original shape of the clastic grains of the arkose, *e.g.* centre. GGU 66951, plain light, X 25. Text reference page 30.

b.

Photomicrograph of metaconglomerate showing a single quartz pebble in a finer-grained matrix of quartz, potash feldspar and plagioclase. The pebble has recrystallised into sub-grains. Such isolated pebbles are common in the metaconglomerates and are also found in the meta-arkoses. GGU 66892, crossed nicols, X 23. Text reference page 31.



a.



b.

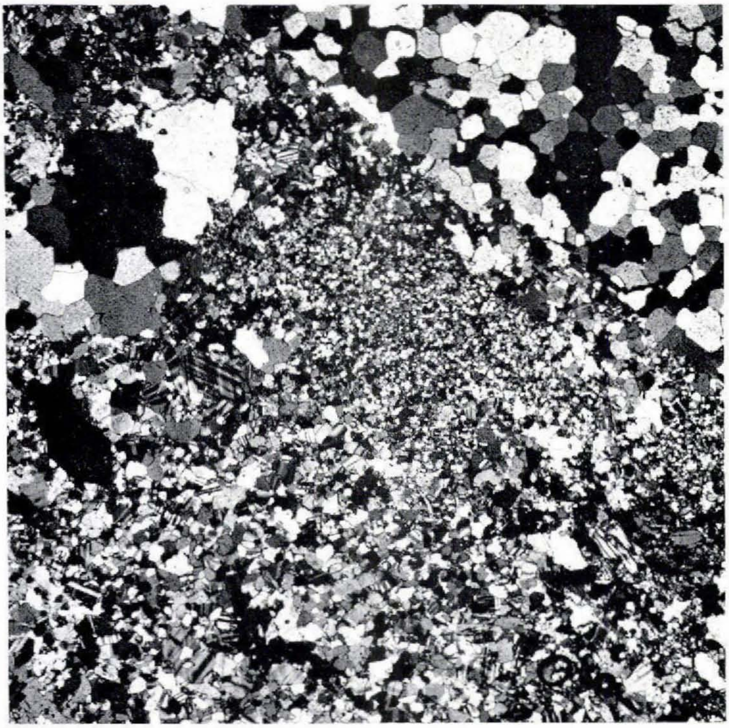
Plate 5

a.

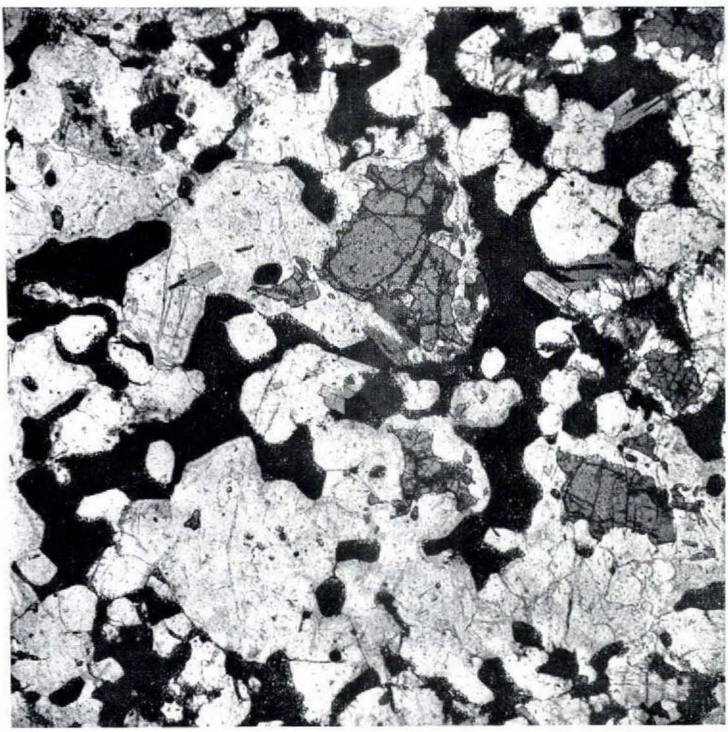
Photomicrograph of a polymictic metaconglomerate showing typical blastopsephitic texture. The original positions of the pebbles in the conglomerate are depicted by the conspicuous range in grain size of the felsic minerals. The pebbles are also of varying compositions, *e.g.* the matrix in the lower left is a granitic pebble containing abundant feldspar but those in the upper right and left contain no feldspar and appear to be quartz pebbles. Clinopyroxene, although not conspicuous under crossed nicols, forms prominent small grains. GGU 66967, crossed nicols, X 15. Text reference page 32.

b.

Photomicrograph of sulphide-bearing gneiss. The main sulphide is pyrrhotite which forms irregular grains and a crude incontinuous network. Large crystals of garnet and flakes of biotite are common, and the felsic matrix is composed of plagioclase and quartz. GGU 58695, plain light, X 22. Text reference page 33.



a.



b.

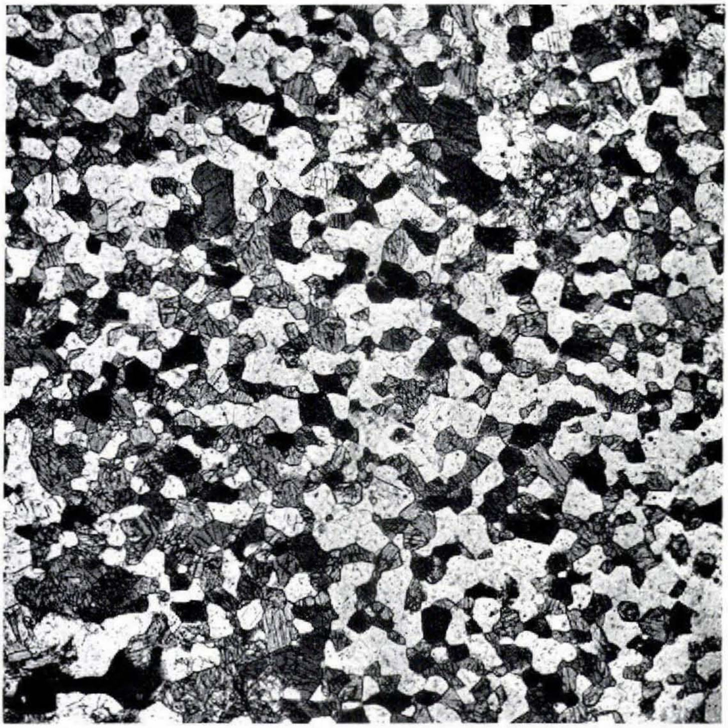
Plate 6

a.

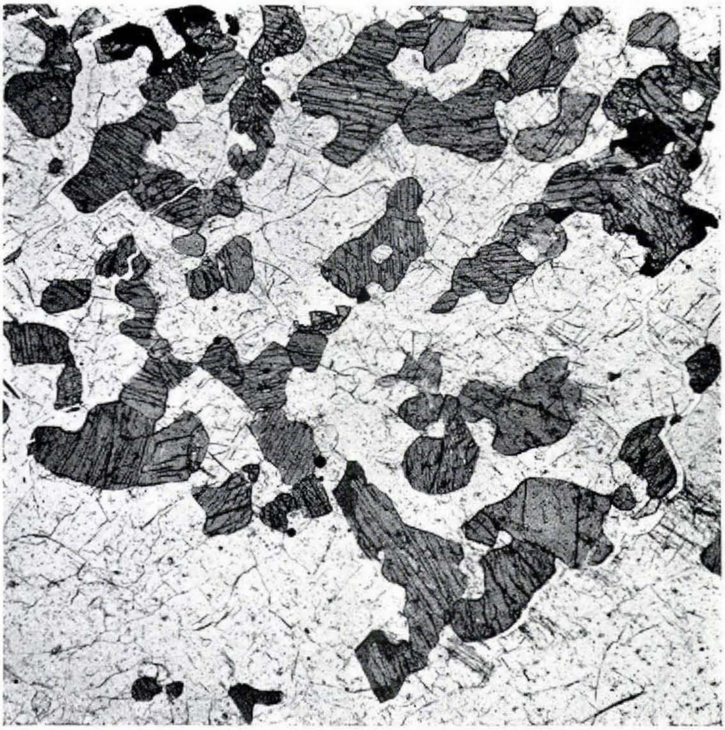
Photomicrograph of metalava showing granoblastic-xenomorphic texture composed of brown hornblende (dark), clinopyroxene and plagioclase. GGU 67129, plain light, X 23. Text reference page 35.

b.

Photomicrograph of metalava. Scapolite (white) forms a coarse-grained mesostasis in which clinopyroxene crystals are situated. In places green hornblende has replaced the pyroxene on the outer edges of the crystals. GGU 67132, plain light, X 23. Text reference page 36.



a.



b.

Plate 7

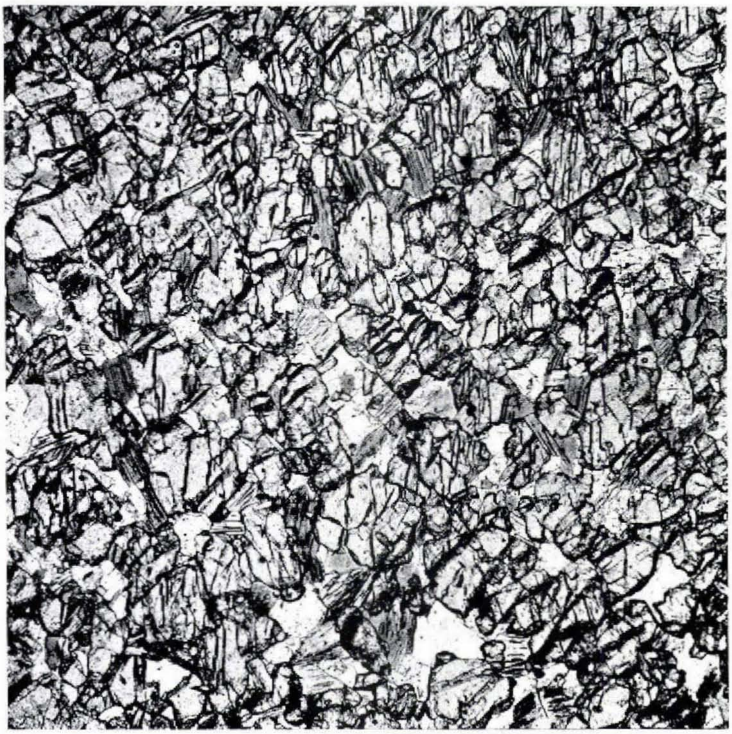
a.

Photomicrograph of basic granulite showing directional texture produced by the elongation of both the pyroxene and plagioclase grains. GGU 67168, plain light, X 23. Text reference page 39.

b.

Photomicrograph of basic granulite showing part of a large crystal of hypersthene which has recrystallised into sub-grains and which contains inclusions of biotite and plagioclase. The biotite forms a foliation. GGU 58686, plain light, X 17. Text reference page 39.

a.



b.

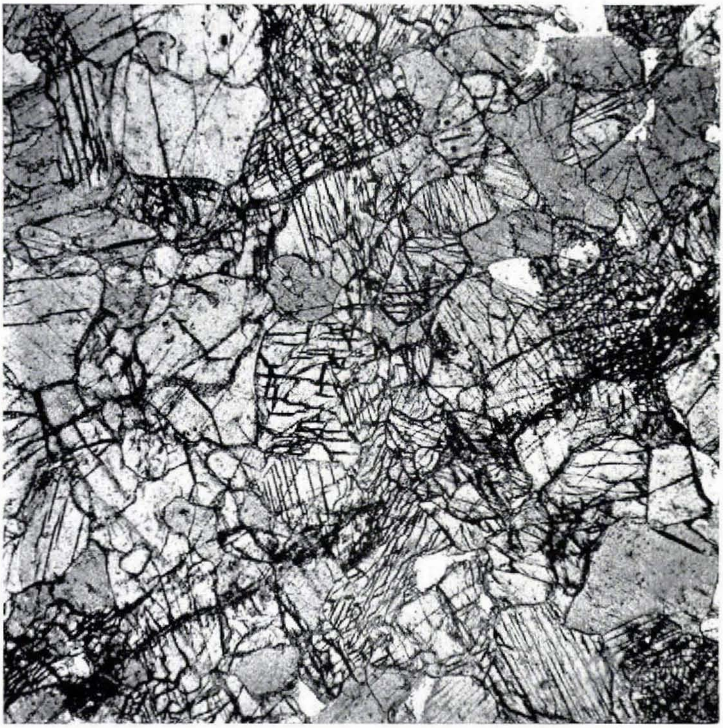
Plate 8

a.

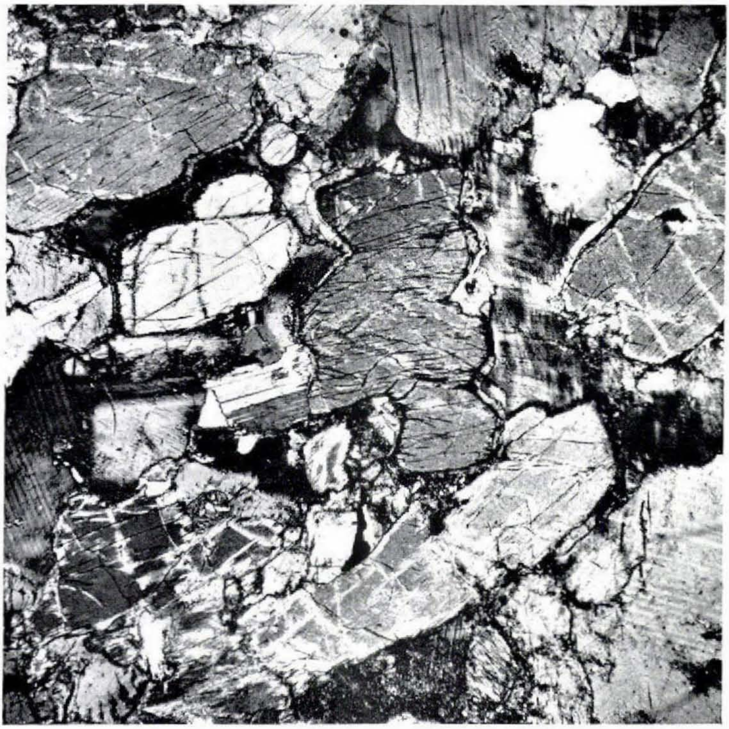
Photomicrograph of ultramafic rock from a small body in the hypersthene-quartzo-feldspathic gneisses. Rock composed of hornblende, orthopyroxene and clinopyroxene with some small grains of plagioclase. An alteration zone cuts the rock. GGU 67175, plain light, X 23. Text reference page 39.

b.

Photomicrograph of charnockite from a small vein cutting basic granulite. Large hypersthene crystals occur in association with potash feldspar, quartz and plagioclase. GGU 67122, crossed nicols, X 31. Text reference page 50.



a.



b.

MEDDR GRØNLAND BD. 189 NR. 3 (PETER R. DAWES)

GRØNLANDS GEOLOGISKE UNDERSØGELSE
THE GEOLOGICAL SURVEY OF GREENLAND

MAP 1