GRØNLANDS GEOLOGISKE UNDERSØGELSE
BULLETIN No. 54

TEXTURAL AND FIELD RELATIONSHIPS OF BASEMENT GRANITIC ROCKS, QAERSUARSSUK, SOUTH GREENLAND

BY

W. STUART WATT

WITH 15 FIGURES IN THE TEXT
AND 1 PLATE

Reprinted from
Meddelelser om Grønland, Bd. 179, Nr. 8

KØBENHAVN
BIANCO LUNOS BOGTRYKKERI A/S
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Abstract

Some textural and field relations of a basement granite in South Greenland are described. The main granite has been formed by granitization but locally there has been later mobilization, mobilized granite intruding non-mobilized granite as dykes. A cross-cutting contact of the granite against a strip of meta-sediment is interpreted as a local late reactivation of the granite that has flushed out the contact to give the sharp, cross-cutting relationship and an intermediate zone formed by contamination.

The textures described, principally myrmekite, and albitic rims to plagioclase and myrmekite, are interpreted as due to local changes in ionic concentration resulting in albite growing at the expense of microcline and *vice versa.* An overall potash-metasomatism has led to the formation of the large potash-feldspar porphyroblasts in a granodioritic matrix.

Zusammenfassung

Die geologischen und texturellen Verhältnisse eines Grundgebirgsgranits aus Südgrønland werden beschrieben. Die Hauptmasse des Granits entwickelte sich durch Granitisation. Es haben auch spätere Mobilisationen stattgefunden. Das mobilisierte Material intrudierte als Gänge in den nichtmobilisierten Granit. Ein quergriffiger Kontakt des Granits gegen einen Streifen von Meta-Sedimenten wird als lokale späte Reaktivierung angesehen. Diese war imstande, einen scharfen, quergriffigen Kontakt zu bilden, sowie eine Zwischenzone durch Kontamination.

Die untersuchten Texturen (hauptsächlich Myrmekit, albitische Hüllen gegen Plagioklas und Myrmekit) werden als lokale Änderungen der Ionenkonzentration erklärt, die im Wachstum von Albit auf Kosten von Mikroklin, und umgekehrt, ihren Niederschlag fanden. Eine grossräumige K-Metasomatose führte zur Bildung von Kalifeldspatporphyroblasten in granodioritischer Matrix.

CONTENTS

INTRODUCTION

Textural features of granites are important as petrogenetic evidence but are no substitute for unambiguous relationships demonstrable in the field. Their interpretation is often highly ambiguous and cannot be extended to cover the ultimate origin of the rock concerned. From textural evidence alone DRESCHER-KADEN (1948) has argued that granites were due to extensive recrystallization. TUTTLE (1952) ascribes importance to recrystallization, but refers it to a late stage in the history of originally magmatic granites. The writer believes that many textural features (myrmekite, sutured grain boundaries, secondary perthites etc.) testify to the importance of recrystallization but, in the absence of other evidence, cannot be used as proof that a rock originated simply by recrystallization.

This paper describes various textural features and field relations within part of the Julianehåb granite. The term "Julianehåb granite" here applies to the area of Precambrian granitic rocks between Kobberminebugt and Sârdloq without precise implications as to age or type of granite.

Fig. 1 is a generalized geological map covering the area of the Julianehåb granite. Qaersuarssuk, the island on which the following study has been made, occurs centrally within the granite area.

Published descriptions of the Julianehåb granite are scanty. There are a few brief early descriptions of particular areas where geologists landed while sailing through the area. The only two previous descriptions of Qaersuarssuk are those of GIESECKE (1878 p. 176 or 1910 p. 221) who sailed past in 1809, and USSING (dating from 1908, unpublished diaries). A general description of the Julianehåb granite is given by USSING (1912). In 1936 WEGMANN made a thorough reconnaissance of the area from Arsuk south to Frederiksdal. He gave the most comprehensive available description of the granite and introduced most of the terminology now used for the different geological episodes in the area (WEGMANN, 1938). Brief, recently published descriptions of particular areas are given by UPTON (1962) for Tugtutôq, HARRY and PULVERTAFT (1963) for Nunarssuit and by HARRY and OEN (1964) and WATTERSON (1965) for Kobberminebugt.

The granitic rocks of the Julianehåb district cover a minimum area of 8000 sq. km. In this area at least two periods of activity in the base-

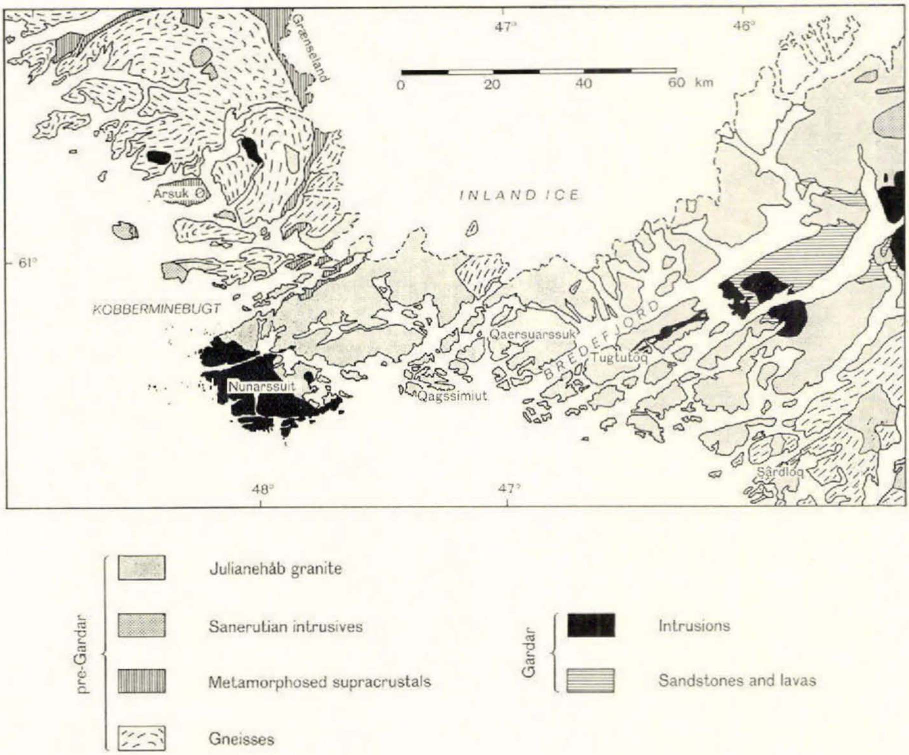

Fig. 1. Geological sketch map showing the extent of the Julianehåb granite in South Greenland. From maps compiled by T. C. R. Pulvertaft and J. H. Allaart.

ment are recognized; these are frequently referred to as the Ketilidian and Sanerutian. Relics of pre-Ketilidian rocks have not been proven. The Ketilidian granites are mainly the result of transformation and granitization of sedimentary and volcanic deposits of geosynclinal dimensions and character still visible on Arsuk Ø, at Grænseland and in Kobberminebugt. The Sanerutian is largely a period of reactivation of these older granites. It is also marked by the metamorphism and migmatization of a set of basic dykes which separate the two periods. The granitic rocks east of Qagssimiut appear to have been largely affected by this later recrystallization and locally display intrusive contacts.

The main period of granitization in Qaersuarssuk was probably Ketilidian and was accompanied by folding and the extensive formation of gneisses. The nebulitic granite-gneiss along the Bredefjord coast was probably formed at this time and possibly also the banded granite on Ûgarmiut. The nebulitic granite-gneiss and the non-granitized metasediments by Uniaríssat both have what is recognized as a Ketilidian trend (N.E.–S.W.). The Sanerutian was probably more in the nature of a reactivation and homogenization of the Ketilidian granites and gneisses.

GENERAL FEATURES OF THE BASEMENT IN QAERSUARSSUK

Meta-sediment

Meta-sediment occurs at Uniaríssat, in the north-west part of Qaersuarssuk, and at two other localities (plate 1). At Uniaríssat it is mostly a banded semi-pelite with a layer of psammite. This strip of meta-sediment is bounded to the north by granite: the sharp contact between these is described on p. 29. It grades south into gneissose-granite rich in amphibolitic horizons parallel to the foliation of the gneiss and sheathed by granitic material. The two other small areas of meta-sediment are mainly amphibolitic.

Gneiss and granite-gneiss

Gneiss and granite-gneiss occur in a number of places on Qaersuarssuk. Since inland exposures are poor many variations have been identified only in coastal sections. Sharp contacts between different types have not been seen and the contact of granite with gneiss on the northern side of Kangerdluarssuk avangnardleq can be mapped only for a short distance.

The most extensive granitic-gneiss area forms the high ground to the north of Kangerdluarssuk avangnardleq. This is a fine-grained, grey, streaky to homogeneous rock with a weak foliation and dictyonitic structures. A strip of aplitic granite occurs within it.

Near the meta-sediments in the northern part of Qaersuarssuk there is coarse-grained gneiss which contains numerous enclaves which are mainly amphibolitic to quartz-dioritic and are aligned with their lengths in a plane striking E.N.E.–W.S.W. and dipping north. They are often veined by granite.

Part of the western coast of the island provides considerable evidence that the gneisses were originally sedimentary. They display a marked foliation and some banding and fold axes are recognizable in folded amphibolitic bands, one of which shows refolding. A narrow calc-silicate band (garnet-quartz-?wollastonite) has been identified and there is a possible conglomerate (with foliation present only in the "pebbles") as well as enclaves of psammitic material.

Nebulitic granite-gneiss is prominent along the Bredefjord coast and occurs in patches on Mátâta nunâ. It also composes a large part of the area between the long lake 5 and Torssukátak.

On the Bredefjord coast it is regularly banded, owing to layers of different grain size and mafic mineral content, with a strike N.E. to E.N.E. and a moderate dip north-westwards.

Foliation of the mafic minerals is pronounced in some layers and may be absent in others. The aplitic layers are of course practically devoid of mafic minerals. Conformable to the banding there are occasional lenses of biotite-hornblende-schist which have parallel mineral layering. Some of the coarser-grained layers have numerous large feldspar porphyroblasts with an apparent random orientation. These bands resemble the big-feldspar granite.

Nebulitic granite-gneiss with the same strike and dip occurs on the southern side of Bredefjord, on Tugtutôq and Lille Tugtutôq, but across Lille Tugtutôq to the S.E. its dip changes to south-eastwards (UPTON, 1962 p. 10).

The banded nebulitic granite-gneiss along the coast of Bredefjord is interpreted as an originally supracrustal rock that has been extensively sheared but incompletely homogenized during granitization, the biotite-hornblende-schist being the least affected relic of the supracrustal rocks. The porphyroblasts appear to be confined to certain layers because other layers were less permeable to migrating material.

On Mátâta nunâ some of the nebulitic granite-gneiss is banded as on the Bredefjord coast but some relatively basic bands have been sheared to form hornblende-rich lenses in the granite.

A banded granite with regular mafic bands constantly striking 160° and dipping about 35° W is prominent on the western side of the island of Ûgarmiut. The dark bands are hornblende-rich, about 10 cm wide and are separated by leucocratic layers about 30 cm thick. This granite only very occasionally contains feldspar porphyroblasts.

Small patches of migmatite, agmatite, veined gneiss and micro-granite are scattered throughout the whole of the granite area.

Sporadic throughout the granite and gneiss are enclaves of biotite- and hornblende-schist, mafic schlieren and remnants of feldspathic gneiss and fine-grained quartz-diorite. Some of these dark enclaves are undoubtedly the granitized remains of old amphibolite dykes, but most are of unknown origin though they are very likely the remains of the original supracrustal rocks.

Enclaves of hornblende-schist are particularly common on the high ground to the south of Nûgssuaq where they have various orientations though some a few metres in length, lie with their long axes approximately north–south. This is not the main trend of the known occurrences of supracrustal rocks.

The structural evidence provided by the enclaves and small areas of gneiss is too scanty to construct a pre-granite structural picture of the area. Moreover, subsequent mobilization and "plastic" deformation must have disturbed that pattern considerably at a number of places. Most of the structures now seen in the granite were probably impressed during the Ketilidian and some may have suffered Sanerutian modification.

Big-feldspar granite

The main granite of Qaersuarssuk is a medium- to coarse-grained grey granite characterized by the presence of large microcline porphyroblasts. The porphyroblastic granite occurs on the Bredefjord coast of southern Tugtutôq (UPTON, 1962) and on the peninsulas on the northern side of Bredefjord east of Qaersuarssuk (J. H. ALLAART, MS maps, 1960). On the northern side of Qaersuarssuk the granite is not quite as rich in porphyroblasts and in places appears nebulitic. Along the Bredefjord coast homogeneous granite passes into nebulitic granite-gneiss.

The granitic rocks have not been subdivided into different types. This is because the big-feldspar granite is merely a variety of the medium-grained biotite-granite that covers the greater part of the area. There is no fundamental difference in groundmass between granites with and granites without porphyroblasts.

The porphyroblasts are normally about 3–4 cm, occasionally 5 cm, long. Zoning is seen in a few of the larger ones (cf. UPTON, 1962 fig. 3). A density count at one place indicated 9 megacrysts per 100 sq. cm. In the western part of the area the granite is similar but porphyroblasts are sporadic and commonly only about 1 cm in length.

Occasionally there is a weak preferred orientation of the microcline porphyroblasts. This is not accompanied by parallelism of dark minerals. Nowhere is the orientation pronounced and no regular regional pattern has been seen in it. In areas where porphyroblasts are infrequent concentrations of them may occur locally along old shear zones.

Pegmatite dykes

Though not abundant, pegmatite dykes occur throughout but mainly in the southern part of the area. The pegmatites (and aplites) are largely of replacement origin but a few pegmatite dykes, with or without zoning, are demonstrably dilational. Most of the zoned pegmatites are in the southern part of the area. Pegmatite dykes are seen to be later than aplite dykes where the two occur together and, with the exception of a single, thin pegmatite dyke (fig. 13), both are cut by the mobilized granite on Mátâta nunâ (figs. 10 and 13).

PETROGRAPHY OF THE GRANITE

The big-feldspar granite is characterized by the presence of large microcline porphyroblasts but the groundmass in which these are embedded is indistinguishable from the granite without porphyroblasts. Biotite is generally the dominant ferro-magnesian mineral. Hornblende sometimes accompanies biotite and in a few examples is the dominant dark mineral phase. The relative proportions of the three main minerals are extremely variable and the rocks vary from syenite to granodiorite in composition. The term granite is applied *sensu lato* to all these rocks. The rocks are often too coarse-grained for their mode to be determined from a 2.5×3.0 cm thin section. For modal analysis JACKSON and ROSS (1956) recommend a surface 100 times the area of the largest mineral, i.e. about 300 sq. cm flat surface of porphyroblastic granite is necessary for modal determination. Apart from the inadequancy of sampling for modal analysis there is considerable lack of uniformity in the granite itself. Its groundmass is granodioritic but its total bulk composition approaches that of an adamellite. Its lack of uniformity is also seen in the replacement processes that appear to have taken place. Some examples show all the possible replacement processes; others show none at all; most show poor examples of one or two.

Primary accessories

Sphene is the most prominent and abundant primary accessory mineral and often forms megascopic grains of typically idiomorphic form. It composes up to 2 % of the rock. In one example it appears to be replacing biotite, while replacement of sphene by plagioclase and microcline is also known.

Apatite, though an insignificant accessory, is present in most examples as tiny idiomorphic crystals. Opaque minerals are ubiquitous though never abundant. They are normally associated with the mafic constituents but appear to be of both primary and secondary origin.

Mafic constituents

Biotite is generally the dominant mafic mineral though locally hornblende may be abundant and occasionally dominant. It is frequently altered to chlorite (commonly penninite). Prehnite can occur within biotite flakes, forcing their cleavage planes apart. The hornblende may show alteration to biotite.

The biotite normally occurs as isolated plates but in some examples may be associated with opaque material. Both biotite and hornblende can form inclusions within microcline porphyroblasts.

The biotite is a common green variety with α pale yellow to pale green-yellow and γ olive green to yellow-green (rarely brown). The hornblende is also green with α yellow, β dirty green and γ dark green.

Biotite is sometimes replaced by plagioclase, quartz or perhaps microcline. In sample 45230¹) some of the biotite adjacent to hornblende grains develops a myrmekite-like texture due to quartz vermicules.

Quartz

The amount of quartz in a thin section varies considerably. It can be absent or it can be the predominant constituent of the rock. Between these two extremes all proportions occur. It forms grains generally about 1.6 mm but up to 3–4 mm in diameter. The large grains frequently show a fractured form and amoeboid shape with highly sutured boundaries against other minerals and adjacent quartz grains. They usually show irregular, undulose extinction indicating that most of them are aggregates of grains with similar, but not identical, orientation. LJUNGGREN (1954) suggests that such aggregates are partially recrystallized quartz aggregates. It is uncertain whether the Greenland examples can be explained thus. The extinction of some seems too regular though they contain isolated patches in which it is different. The impression gained from the Greenland rocks is that large quartz grains have been fractured with concomitant development of strain extinction and frequently also slight biaxiality. This impression is strengthened by the presence of tiny individual grains of quartz along boundaries between microcline megacrysts or large quartz aggregates and microcline megacrysts. These tiny grains are ascribed to granulation.

Besides large amoeboid aggregates and tiny granular grains quartz forms moderately small (0.3 mm diameter), well-rounded, perfectly clear grains with sharp extinction—drop quartz (HÄRME, 1959 p. 45). These may be present within the microcline porphyroblasts or large quartz aggregates. They are believed to be a later generation of quartz (quartz

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1) Samples in the collection of Grønlands Geologiske Undersøgelse, Copenhagen.

II) due perhaps either to complete recrystallization of existing quartz or to replacement. Apart from extinction the two generations of quartz differ most markedly in their outline against surrounding minerals. Quartz I always has highly sutured outlines against other minerals and quartz grains and also penetrates between them. Quartz II forms small, well-rounded grains even when within a quartz I aggregate.

Quartz replaces all the other main minerals. Fig. 2 shows it replacing plagioclase as long thin protuberances parallel to the polysynthetic twin planes. These protuberances have the same optical orientation as the adjacent large quartz grain. This suggests some activation of the quartz. Veins of quartz also cut the large microcline porphyroblasts.

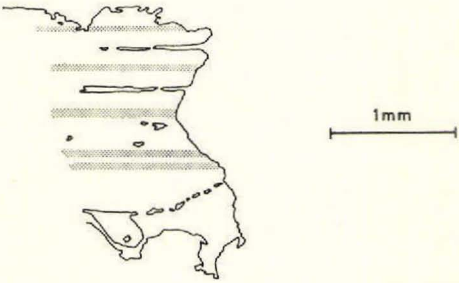

Fig. 2. Reactivation of the quartz has led to quartz veining the plagioclase. Drawn from sample 45201.

Quartz I occurs as inclusions within both microcline and plagioclase grains and may itself include all other minerals except microcline. The absence of microcline inclusions may be evidence of the age relationship between quartz I and microcline. Quartz II grains never contain inclusions.

Quartz may also be abundant as myrmekite. This will be discussed later (p. 18).

Plagioclase

Plagioclase (about An_{30}) is present in all the granitic rocks and is the dominant component of some thin sections. Carlsbad and polysynthetic twinning are common. It may occur as distinct grains in the groundmass, occasionally as small megacrysts, or as inclusions within the microcline porphyroblasts. Frequently, but by no means always, those plagioclase grains that abut against microcline, either as inclusions within the microcline or as separate grains surrounding microcline, have clear relatively sodic outer rims that resist alteration. These rims are thus particularly prominent where the rest of the plagioclase has been saussuritized.

Much of the plagioclase, particularly the inclusions within microcline porphyroblasts, shows signs of corrosion and replacement by microcline. Numerous examples show this; a few are illustrated as figs. 3 and 4. The corrosion of the plagioclase by microcline or quartz appears to be affected by its orientation relative to that of the replacing microcline (high misfit of VOLL, 1960 figs. 7a & b). This is clearly so in fig. 3a where alternate lamellae of polysynthetic twins are differently affected, resulting in step-like corrosion.

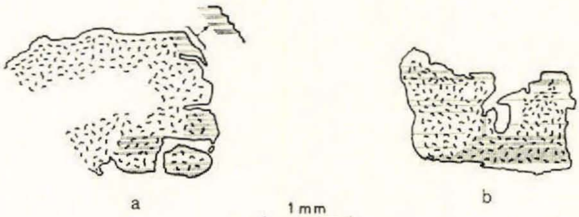

Fig. 3. Two inclusions of plagioclase, in and corroded by microcline, that show saussuritized interiors and clear rims. Differential corrosion of 3a has led to a step-like edge dependent on the twinning direction. Both drawn from sample 45201.

Plagioclase grains included in a microcline porphyroblast rarely have the same optical orientation as plagioclase grains surrounding the porphyroblast. Occasionally two or more inclusions within a porphyroblast will have the same orientation (fig. 5) and are interpreted as remnants of a single grain partially replaced by microcline.

Lobes of myrmekite sometimes penetrate microcline. Some of the lobes bear albitic rims and some show corrosion by the microcline.

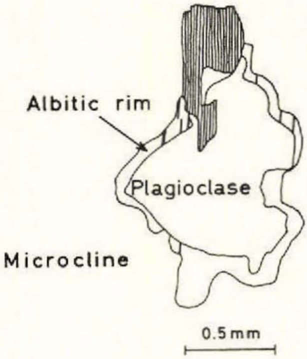

Fig. 4. A corroded, clear, albitic rim round a corroded crystal of plagioclase in a large microcline porphyroblast. Note the thin twin lamellae confined to the albitic rim (cf. fig. 8c). Drawn from sample 45380.

Very rarely a large plagioclase grain contains a few areas of clear microcline, always with very fine grid twinning invariably with one twinning direction parallel to that of the plagioclase (fig. 6). These inclusions are taken as showing early stages in replacement by microcline of the interior of the plagioclase crystal. Such "concordant" microcline within plagioclase is never surrounded by an albitic rim.

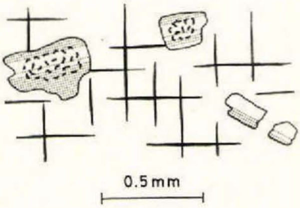

Fig. 5. Two pairs of plagioclase inclusions, each pair having exactly the same orientation. Whether or not the pairs are joined in the third dimension does not affect the argument that these are corroded remnants of larger plagioclase crystals. Drawn from sample 45219.

Occasionally there are large (4 mm) euhedral crystals of albitic plagioclase. These sometimes show one or two well-developed faces (see fig. 7) but more sodic overgrowths usually render the plagioclase crystals anhedral.

Zoning is rare. Some of the larger plagioclase crystals display a faint zoning, and this probably indicates that they have not been recrystallized. Plagioclase throughout all the granites is frequently saussuritized. Embayments of quartz in plagioclase suggest replacement

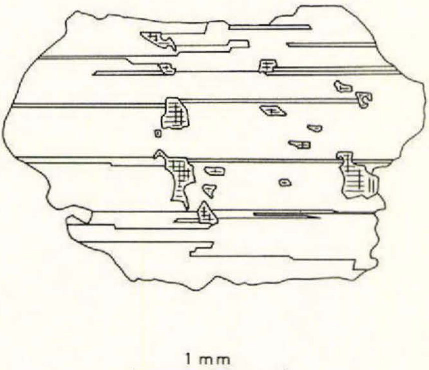

Fig. 6. A plagioclase grain with areas of microcline. The microcline has thin cross-hatch twinning with one twinning direction parallel to that of the plagioclase twins. The relationship is taken as representing an early stage in the replacement of plagioclase by microcline. Drawn from sample 45384.

of plagioclase by quartz. Plagioclase may also be veined by quartz (fig. 2).

In sample 45481 there is a large plagioclase crystal which contains areas that differ slightly in orientation from the main body of the crystal.

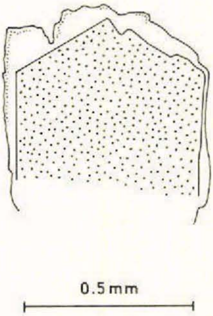

Fig. 7. A plagioclase inclusion in microcline. The early plagioclase shows crystal faces with an overgrowth of a later formed, clear albitic rim. Drawn from sample 45201.

Microcline

Microcline occurs in the granitic rocks both interstitially and as porphyroblasts. The porphyroblasts can be large and densely packed and have grown at the expense of plagioclase. Many of the larger ones are simply carlsbad twins. Since grid twinning is practically universal in the potash-feldspar, orthoclase is believed to be absent from these rocks. Megascopically, the porphyroblasts show rare apparent zoning (cf. Upton, 1962 fig. 3). No trace of this zoning has been seen in thin sections though J. H. Allaart (oral communication) has seen zones of perthitic plagioclase in similar feldspars from the Julianehåb granite in adjacent areas.

Inclusions of all the other mineral constituents occur in the porphyroblasts, in particular plagioclase and quartz but also occasionally biotite, opaque grains and sphene. The orientation of these inclusions normally bears no relation to the microcline lattice. With one possible exception (sample 45385, discussed below) they appear to be quite randomly arranged. The plagioclase inclusions show abundant evidence of corrosion by the microcline.

Lobes of myrmekite protrude between microcline and plagioclase grains. Rims of albitic plagioclase between microcline and plagioclase grains are common but by no means ubiquitous.

Vein, film and string perthites are extremely common. The albitic component of perthite in the microcline porphyroblasts composes any proportion up to about 35 % of a planar surface. With rare exceptions described below, the type, size, orientation and distribution of the perthite lamellae are not regularly related to the development of the albitic rims, myrmekite or inclusions. The lamellae clearly follow certain crystallographic directions in the microcline.

Occasional microcline "inclusions" with very thin cross-hatch twinning occur within grains of plagioclase (fig. 6), as mentioned above. Here the microcline and plagioclase are related in crystallographic orientation and are taken to represent an early stage in replacement of plagioclase by microcline. The main characteristic of this microcline is its very thin cross-hatch twinning.

The replacing microcline commenced growth in one plagioclase crystal which controlled its initial orientation. It eventually replaced not only that crystal but others which were adjacent and of different orientations. Thus a microcline porphyroblast was formed.

In sample 45385 cross-hatch twinning, film and vein perthite are well developed and there appears to be a slight negative correlation between cross-hatch twinning and perthite lamellae in their development. The microcline contains numerous small albitic "inclusions" with polysynthetic twin planes parallel to those of their host. The dominant set of twin planes in the "inclusions" is also parallel to the twin planes in the albite of the perthite lamellae and in a few instances the perthite lamellae are continuous with the "inclusions". The albitic "inclusions" are concentrated along the twin planes of the microcline and are often joined to lamellae in the adjacent perthite. As well as the orientated "inclusions" there are a few plagioclase "inclusions" of apparently random orientation which bear no apparent relation to the perthite lamellae. From the shape of the orientated plagioclase "inclusions" alone it could be argued that the microcline had replaced the plagioclase. But taking into consideration their constant orientation, and their relation to both the orientation of the microcline and the perthite lamellae these "inclusions" appear to be cogenetic with the perthitic albite lamellae. They possibly exemplify the late stage albitization described by ROBERTSON (1959), and many others.

Minor constituents

Minor constituents occasionally found include allanite, zircon, prehnite (in biotite) and haematite. Epidote is quite prominent in some samples which usually show other features of alteration or occur near shear zones.

Perthite

The microcline porphyroblasts are variably perthitic. The small areas of "concordant" microcline in plagioclase crystals and a few interstitial microcline grains show no sign of perthite structure under the microscope.

Vein perthite is by far the most common. The string perthite always lies at a small angle to the vein and film perthite. In sample 45268 it is flame-like and frequently reaches to the edge of the grain where it may even be more abundant than in the interior of the grain. Perthite is also known to stop abruptly against biotite or a microcline twin plane crossed by a film perthite lamella. The perthite lamellae on one side of the twin plane differ in orientation from those on the other side and this is taken as an indication that the orientation of the lamellae is controlled by the microcline host. In sample 45385 polysynthetic twinning is developed in the perthite lamellae.

On a planar surface the perthite lamellae lie with their long axes not quite at right angles to the trace of the twin plane. ROSENQVIST (1952 p. 81) distinguishes four types of perthite (described and named by ANDERSEN, 1929) which follow certain directions in the microcline: 1) vein perthites, which he says are the commonest type, vary in size, outline and orientation; 2) film perthites form thin films perpendicular to (010) and at an angle of – 75° with (001), though they are occasionally found orientated differently; 3) string perthites form regular strings which are parallel to (010) and at an angle of – 73° with (001) and in cross-section are oblong with the axes perpendicular to (010); 4) patch perthites are more irregular than the vein-type lamellae but tend to be orientated in the same direction.

ANDERSEN (1929) recognized the temperature control of perthites and his textural classification represents conditions of decreasing temperature. ROSENQVIST (1952) has found that there are different rates of diffusion in different crystallographic directions in the crystal and "It seems from the diffusion experiments [using Ra in alkali-feldspar] as if the symmetry of the diffusion ellipsoid decreases with increasing temperature. From these observations it may be concluded that perthite formed at high temperature must be of the string type, at intermediate temperatures of the film, or vein, type, and at low temperatures irregular, of the patch type." (p. 85). The experiments also showed that the rate of diffusion was very low but that "diffusion in the solid state may form perthites 0.2 mm in width within a reasonable length of time." (p. 86). He has in mind the order of 500 years.

The microcline of the Qaersuarssuk granites is believed to have been formed by replacement of pre-existing minerals, mostly plagioclase. At the elevated temperatures of granitization microcline can contain more sodium in solid solution than at lower temperatures and on cooling excess sodium may exsolve as albitic perthite lamellae.

While this may account for some of the perthite it cannot account for all of it. Perthite lamellae may compose anything from less than 1 % up to 35 % of the sectional area of a single microcline crystal. This upper limit exceeds that of albite held in solid solution in alkali-feldspar. SPENCER (1937) has shown that up to 30 % of the soda component can be held in solid solution by alkali-feldspar at temperatures of 750° C. These are probably considerably higher than temperatures during granitization.

Perthite formation has also been ascribed to simultaneous crystallization and replacement. On p. 21 it is suggested that the albitic rims round plagioclase and the myrmekite are both more or less contemporaneous with replacement of plagioclase by microcline. If the perthite lamellae and the albitic rims had the same origin connections between the two might be expected. Such connections have been seen but are rare. Thus simultaneous crystallization may have operated sometimes but seems to be the exception rather than the rule.

In a few examples flame-like perthite extends to the very edge of the crystal or abuts against an inclusion. These perthite lamellae seem of replacement not exsolution origin, for they are superimposed on evenly distributed vein or film perthite which does not become relatively scarce near them.

Myrmekite

Myrmekite occurs either as wart-like or cauliflower-like protuberances into microcline or as part of a larger plagioclase grain adjacent to a grain of microcline. It always has a crenulated border against the microcline (fig. 8). Only occasionally is the plagioclase of the myrmekite twinned. When the myrmekite is part of a larger plagioclase grain the twinning usually continues right through the area of the myrmekite and it is usually impossible to say whether the myrmekite has grown on to the plagioclase as a unit, or whether the quartz vermicules are formed by replacement; there is no apparent difference in composition between the plagioclase with quartz vermicules and the rest of the grain.

The proportion of myrmekite in the granite varies considerably; it appears to be scarcer in the plagioclase-rich granites and more abundant in those with microcline (cf. CHENG, 1944 p. 140).

In the myrmekite lobes the quartz occurs as long worm-like ramifications or vermicules arranged with their lengths approximately normal to the plagioclase—microcline border, or as short, irregular pear-shaped droplets. If it is of the droplet form the droplets are often smaller in the centres of the myrmekite lobes than at their margins (cf. OYAWOYE, 1959 and 1962 who shows the opposite relationship). The quartz does not always show the same orientation.

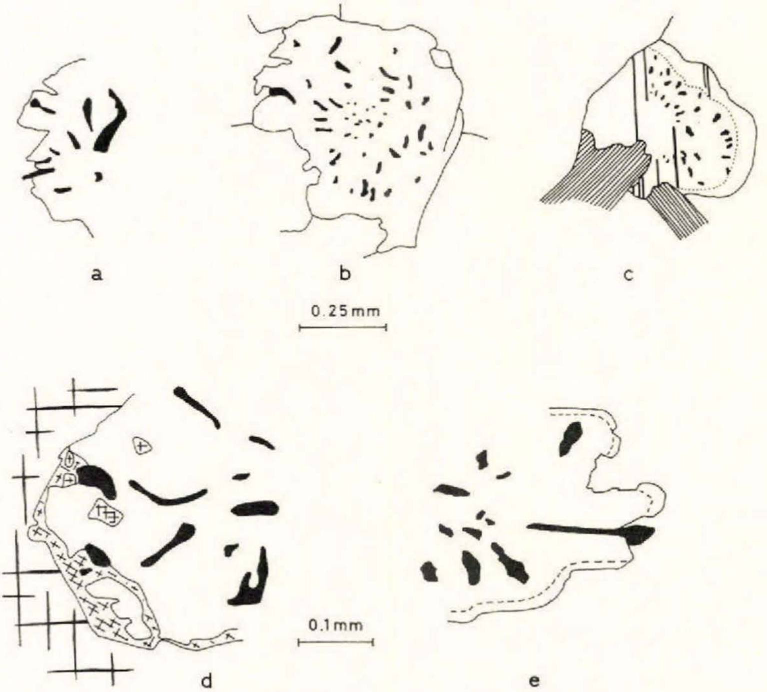

Fig. 8. Sketches of myrmekite in the granitic rocks. a, b and e are explicable as corrosion of the myrmekite exposing the quartz which now penetrates into the surrounding microcline in a and e. In d the myrmekite is partly surrounded by a finely twinned microcline (oblique crosses) different in orientation to the large microcline crystal surrounding the myrmekite (large, upright crosses). In c the twinning in the clear rim of the myrmekite does not continue into the myrmekite. For full explanation see the text. Solid black is quartz. a and b drawn from sample 45304, c and d from sample 45380, and e from 45284.

A few of the myrmekitic wart-like protuberances into the microcline have albitic rims like those round other plagioclase grains abutting microcline (cf. Cheng, 1944 p. 140).

The borders of the myrmekite are normally extensively crenulated. This is believed to be due to corrosion of the myrmekite at a later stage. In some examples the corrosion has continued until the quartz vermicules penetrate the microcline (figs. 8a and b). Sometimes a later albitic rim has repaired the corroded surface so that the quartz vermicles again lie entirely within plagioclase. Some cross the boundary between the myrmekite and the albitic rim though many end abruptly at that boundary. Albitic rims occasionally bear their own generation of vermicular quartz (cf. Eskola, 1914 p. 27 and Voll, 1960 figs. 7a & b). Even some grains of plagioclase included in microcline contain vermicular quartz.

The corrosion and replacement of myrmekite by microcline is possibly best seen in sample 45380 where there is an outer "rim" of microcline that is not in optical continuity with the main microcline grain (fig. 8d). The same slice bears evidence that quartz as well as plagioclase was affected by corrosion. A vermicule, presumably once entirely filled with quartz, is now partly filled with microcline (cf. Erdmannsdörffer, 1941 fig. 1b).

Shelley (1964 p. 43) has given a plausible explanation of the vermicular habit of quartz in myrmekite. Briefly it is that in a rock with cataclastic texture albite exsolved from monoclinic potash-feldspar will grow on plagioclase, and any quartz it traps will be under a confining pressure which is at a minimum in the direction of growth of the albite. On recrystallization the trapped quartz will therefore assume cylindrical forms normal to the growing surface of the albite. Myrmekite thus depends on the presence of a cataclastic texture and original quartz in the rock. In the absence of original quartz, albite rims will simply be formed.

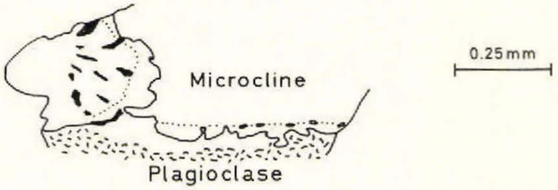

Fig. 9. Interpretation: The plagioclase and the myrmekite have clear outer rims possibly slightly more sodic than the rest of the plagioclase. The rim to the plagioclase has been corroded by the microcline. The myrmekite appears to have suffered corrosion exposing the quartz vermicules. Subsequently a sodic plagioclase was added and later also corroded. The formation of rims by addition is demonstrated in other sections, e.g. figs. 4, 7, and 8c. Drawn from sample 45328.

If all myrmekite were formed in this way all albite rims added to myrmekite should be indistinguishable from the myrmekite plagioclase. Figs. 8c, e and 9, however, illustrate examples of albitic rims differing slightly in composition from the myrmekite plagioclase. Admittedly this difference is not so great as that between the rims and the interiors of plagioclase inclusions in microcline (fig. 4) or separate plagioclase crystals (fig. 7).

According to Shelley, cataclastic texture is a prerequisite for myrmekite formation. Cataclastic textures occur in the Qaersuarssuk granites but show no apparent correlation with the presence of myrmekite. This does not mean that the Qaersuarssuk granites containing myrmekite never had a cataclastic texture. They could have had such a texture that now is obliterated by later recrystallization and growth of potash-feldspar.

Albitic rims to plagioclase and myrmekite

Plagioclase crystals, plagioclase inclusions in microcline, and myrmekite develop albitic rims where they abut against microcline porphyroblasts (figs. 7, 8c and 9) in some of the granitic rocks. The albitic rims are clear and sharply distinct from the adjacent (invariably more or less saussuritized) plagioclase. Plagioclase never develops albitic rims against small "concordant" microcline areas in plagioclase.

Albitic rims are normally of constant width following the embayments in corroded plagioclase. They usually have the same crystallographic orientation as the adjacent plagioclase but twin lamellae are thinner than, and not continuous with, thin twin lamellae that may sometimes be developed in the adjacent plagioclase. In two samples the rims are in optical continuity with perthite lamellae but not with the adjacent plagioclase. These are discussed later.

There are two possible origins for the rims; decalcification of plagioclase (favoured by DRESCHER-KADEN, 1948 p. 56), or growth of plagioclase at the expense of adjacent microcline.

The first process is not considered a feasible explanation of the examples described, for the following reasons.

1) The very thin twin lamellae in the rims are not continuous with the twin lamellae that may occur in the host plagioclase.
2) There is a sharp contact between the host and the rim.
3) In a few examples (fig. 7) the albitic rim is added to an euhedral crystal of plagioclase.

These three observations are consistent with an origin by addition at the expense of the adjoining microcline. The adjoining microcline is a prerequisite for the formation of the albitic rim.

Some albitic rims are of very irregular width. This may be due either to uneven growth or to further corrosion and replacement by microcline. Both explanations are compatible with the oscillation between growth and corrosion of plagioclase proposed on p. 22. Around some of the myrmekite the rims are less persistent.

Rare albitic rims contain a little vermicular quartz. This is a myrmekite later than that forming the main wart-like protuberances.

The relation of the rims to the perthite: The perthite itself is discussed on p. 16. With two exceptions there are no connections between the albitic rims and the perthite lamellae. One exception is seen in sample 45268 where the outer member of a double rim is in optical continuity with the perthite lamellae and has a refractive index slightly higher than that of the inner member.

In sample 45508 a myrmekite lobe adjacent to two microcline porphyroblasts is separated from one of the microcline grains by a margin of plagioclase continuous with the perthite in the microcline. This suggests that the marginal plagioclase and probably also the perthite are younger than the myrmekite.

Twinning in the albite rims around plagioclase and myrmekite is independent of twinning in the host (figs. 4 and 8c). The twins are very thin, irregularly spaced and terminate at the boundaries of the rim. Similar sparse polysynthetic twinning may occur through the myrmekite but does not continue into the albitic rim. The twinning in both the myrmekite and the albitic rims is probably secondary mechanical (glide) twinning induced by stress at different times (cf. VANCE, 1961; SHELLEY, 1964; EMMONS and GATES, 1943).

Origin of myrmekite, albitic rims etc.

It has been shown above that the albitic rims are additional and have been formed at the expense of the adjacent microcline though later eaten into by the microcline. Potash-metasomatism and the formation of microcline (mainly at the expense of plagioclase) will release large quantities of soda which may be locally concentrated along the margins of plagioclase grains. Replacement of potash-feldspar by plagioclase or *vice versa* will depend upon the relative amounts of potassium and sodium ions. This has been confirmed by the experimental work of O'NEILL (1948). Thus, if removal of soda does not keep pace with influx of potash, conditions will be reached where albite will grow at the expense of the microcline and form rims to the plagioclase. EDELMAN (1949) gives two equations (I and II p. 75) to summarize the process:

$$K + NaAlSi_3O_8 \rightleftharpoons KAlSi_3O_8 + Na$$
$$K + Si + CaAl_2Si_2O_8 \rightleftharpoons KAlSi_3O_8 + Ca + Al$$

As these are reversible equations depending upon the ionic concentrations, the main potash-metasomatism pushes them from left to right, replacing plagioclase by potash-feldspar with the release of sodium and calcium ions, while a concentration of sodium and calcium ions pushes them from right to left with the production of albitic rims and myrmekite replacing microcline.

This theory of myrmekite formation agrees with BECKE's (1908) and SEDERHOLM's (1916) results, without invoking both soda- and potash-metasomatism. Where quartz vermicules occur in parts of larger plagioclase grains some of the free calcium ions have probably replaced sodium in the plagioclase with the release of free silica.

Thus the production of myrmekite and albitic rims is complementary to the main granitization process involving introduction of potassium and removal of sodium and calcium.

Stages of alteration

1) original plagioclase.

2) potash-metasomatism leading to the formation of microcline, first interstitial and later replacing plagioclase.

3) the formation of microcline at the expense of plagioclase led to local excess soda forming albitic rims or, where the plagioclase was anorthitic, to excess lime and silica giving myrmekite.

4) continuation or renewal of potash-metasomatism with further corrosion of plagioclase including the albitic rims and myrmekite which led to albitic rims (rarely with quartz vermicules) around myrmekite.

Two other stages which took place are:

a) albite exsolved from microcline to form exsolution perthite;
b) replacement perthite was formed.

Temperature considerations

Edelman (1949) suggests that myrmekite was probably formed at low temperatures. Marmo and Hyvärinen (1935) go further and suggest that it was formed in the amphibolite facies (ca. 400°–700° according to Rosenqvist, 1952 p. 37 and 63). Kullerud and Neumann (1953) suggest, on the basis of the FeS content in sphalerite, that the temperature during granitization in the Rendalsvik area was 440 ± 25° C.

From experimental work granitic magma is not considered to exist below ca. 650° under hydrothermal conditions (Kranck and Oja (1960) 670°; Tuttle and Bowen (1958) 640°). Thus rocks undergoing granitization at ca. 450° could never be fused. The granitic rocks of the Qaersuarssuk area probably exemplify such a granitization.

Assuming a normal geothermal gradient, the temperatures of granitization mentioned above, will be reached at depths of 15 km (Tuttle and Bowen, 1958 fig. 62). Thus at shallower depths granitization can only take place during orogenic phases under locally heightened temperature and pressure.

During the Sanerutian reactivation, however, temperature and pressure must locally have been sufficiently high, not only for further granitization of second period discordant amphibolites—cf. Watterson,

1965) but also for the local mobilization of the granite seen at Uniaríssat and Mâtâta nunâ described on subsequent pages. Although the textures cannot be dated or directly related to the field relations the microcline porphyroblasts were formed before this mobilization took place (p. 25–28).

Orthoclase or microcline

Shelley's (1964) attribution of myrmekite and albite rims to exsolution of albite from monoclinic potash-feldspar raises the question of the state of the potash-feldspar, for it assumes the original presence of orthoclase. Since cross-hatch twinning is practically ubiquitous in the alkali-feldspar the alkali-feldspar is assumed to be microcline. On crystallographic grounds cross-hatch twinning in microcline is interpreted by Goldsmith and Laves (1954b p. 104) as evidence that microcline was originally monoclinic. On this view all the alkali-feldspar in the Qaersuarssuk granites has been primarily monoclinic; no primary microcline exists.

Marmo (1962 p. 59–60), however, disputes this conclusion and points out that while heating experiments clearly demonstrate that microcline is transformed to monoclinic sanidine under hydrothermal conditions at 525°C (Goldsmith and Laves, 1954a) the opposite transition on cooling has never been accomplished. Microcline has never been synthesized by direct crystallization (Goldsmith and Laves, 1954a p. 15); all attempts have yielded orthoclase. But microcline has been produced by the replacement of albite by potash-feldspar, admittedly without twinning, under hydrothermal conditions (Laves, 1951; Wyart and Sabatier, 1956); i.e. by potash-metasomatism.

Furthermore, Marmo (1962 p. 60) cites evidence that would also appear to contradict the experimental data. For example microcline, of high triclinicity and with good cross-hatch twinning, partly replaces plagioclase in synkinematic granites in a way that orthoclase never does. The alkali-feldspar in the granite on Qaersuarssuk has not been tested for triclinicity (Goldsmith and Laves, 1954a) but Nesbitt (1961) has carried out X-ray diffraction work on some megacrysts from the granite on the Julianehåb peninsula and has found that all those tested have a fairly high triclinicity. With the exception of one low value of 0.75 the triclinicity is greater than 0.9.

If it is postulated that the microcline was originally a monoclinic potash-feldspar, which can contain unlimited amounts of plagioclase in solid solution, the source of material to form the albitic rims to the plagioclase may be exsolved from the potash-feldspar (accepted by Gates, 1953 p. 58–59; Voll, 1960 p. 524). But it may also be derived from solutions outside the crystal.

MOBILIZATION OF GRANITE

Evidence that the granite has been locally mobile is well seen at three localities on the southern coast of Mátâta nunâ. The granite adjacent to meta-sediment east of Uniaríssat may also be interpreted as a mobile granite.

The island of Mátâta nunâ

In the neighbourhood of the small bay on the south coast of Mátâta nunâ and along the coast west of it nebulitic granite-gneiss appears to have been disrupted into large blocks some tens of metres in width and intruded by granite dykes. Relative displacement of different blocks of country rock is well illustrated on the western side of the bay (fig. 10). There a granite dyke 1 m wide separates tracts of nebulitic granite-gneiss differing in orientation and lithology. Three pegmatite dykes differing in trend are transected by the granite dyke and cannot be matched across it.

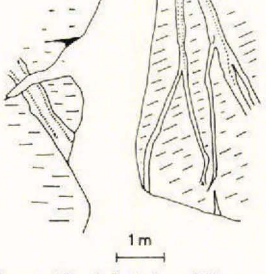

Fig. 10. An intrusive dyke of granite (plain) cutting gneiss (dashed lines) and zoned pegmatites (lines of dots within a pair of continuous lines). Bay on south coast of Mátâta nunâ.

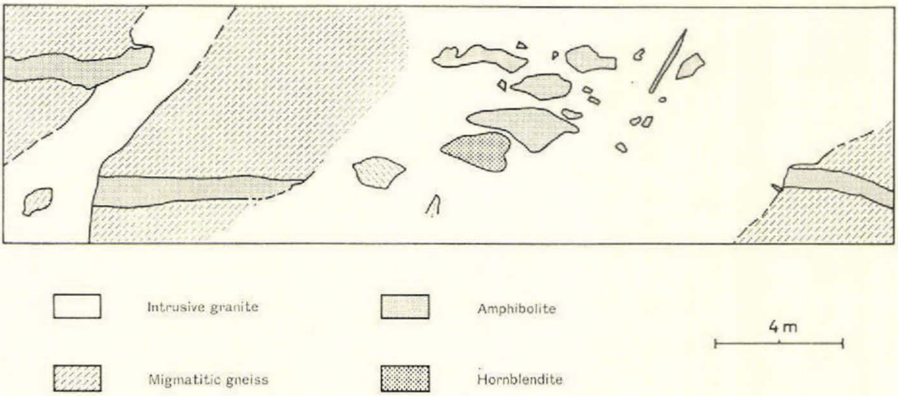

Fig. 11. An amphibolite dyke cut by later intrusive granite with numerous inclusions of the amphibolite and other rocks. Bay on south coast of Mátâta nunâ.

Nearby a discordant amphibolite dyke was used as a marker to measure the relative displacement between blocks of country rock. Where the amphibolite is continuous its dip is reasonably constant but its strike can vary. Fragments disrupted by granite differ in strike and dip however and provide some evidence of rotation about a horizontal axis.

The dyke is cut by intrusive granite dykes (fig. 11). The western granite dyke has a maximum width of two metres and is well-defined in the amphibolite dyke and adjacent nebulitic granite-gneiss. The eastern granite dyke is defined clearly against the amphibolite dyke but very vaguely against the nebulitic granite-gneiss. It is packed with inclusions derived mainly from the amphibolite but also from quartz-diorite and in one case hornblendite.

A third dyke of homogeneous granite is found on the eastern side of the small bay on Mátâta nunâ. At this locality three lengths of aplite can be approximately fitted back to give a single aplite dyke. Within the homogeneous granite separating and truncating the lengths of aplite there is a block of gneiss (figs. 12 and 13) that is quite distinctive compared with anything known in the neighbourhood. The gneiss block was extensively granitized before incorporation in the intrusive granite. Its foliation, accentuated by thin quartz-feldspathic segregation, its homogeneous layers, broken up by later granitic veins, and its more competent bands (now beads of epidote nodules) are all parallel. There are also concordant granitic layers with large feldspar porphyroblasts. A few thin pegmatite veins run sub-parallel to this layering. The whole assemblage is cut by two shear planes. Coarse-grained granite along the two shear planes appears to be younger than the porphyroblasts in the conformable layers, for the porphyro-

Fig. 12. A block of gneiss within the mobilized granite. The photograph is taken looking south. The area is shown diagrammatically in fig. 13. Only one length of the aplite is visible in the middle distance to the right of the photograph. The boundary between the intrusive granite and the country rock granite is just discernible between the nearer end of the aplite and the hammer and just beyond the far side of the block of gneiss. Bay on south coast of Mátâta nunâ.

blasts neither penetrate nor occur entirely within the granite. The structures in the gneiss block are totally absent from the surrounding homogeneous granite.

The country granite which is cut by aplite and the later homogeneous intrusive granite shows palimpsest structures. A thin late pegmatite dyke cuts both older granite and the intrusive granite.

The gneiss block is not likely to have been derived from the local nebulitic granite-gneiss for there are differences in structure and degree of granitization. Thus it must have been transported a considerable distance by the homogeneous granite indicating that this granite has been highly mobile.

Further examples of mobilization of granite are found on the N.W. corner of Ajatorfik (on the western side of Ûgarmiut) where there are three small inclusions of banded meta-sediment within a medium-grained homogeneous granite that cuts a country granite with palimpsest structure. These pieces of meta-sediment have no common orientation and do not show folding within themselves. None of them fits into the local structural pattern of the island.

The granite veins in discordant amphibolite dykes along the Bredefjord coast also indicate late activity of the granite.

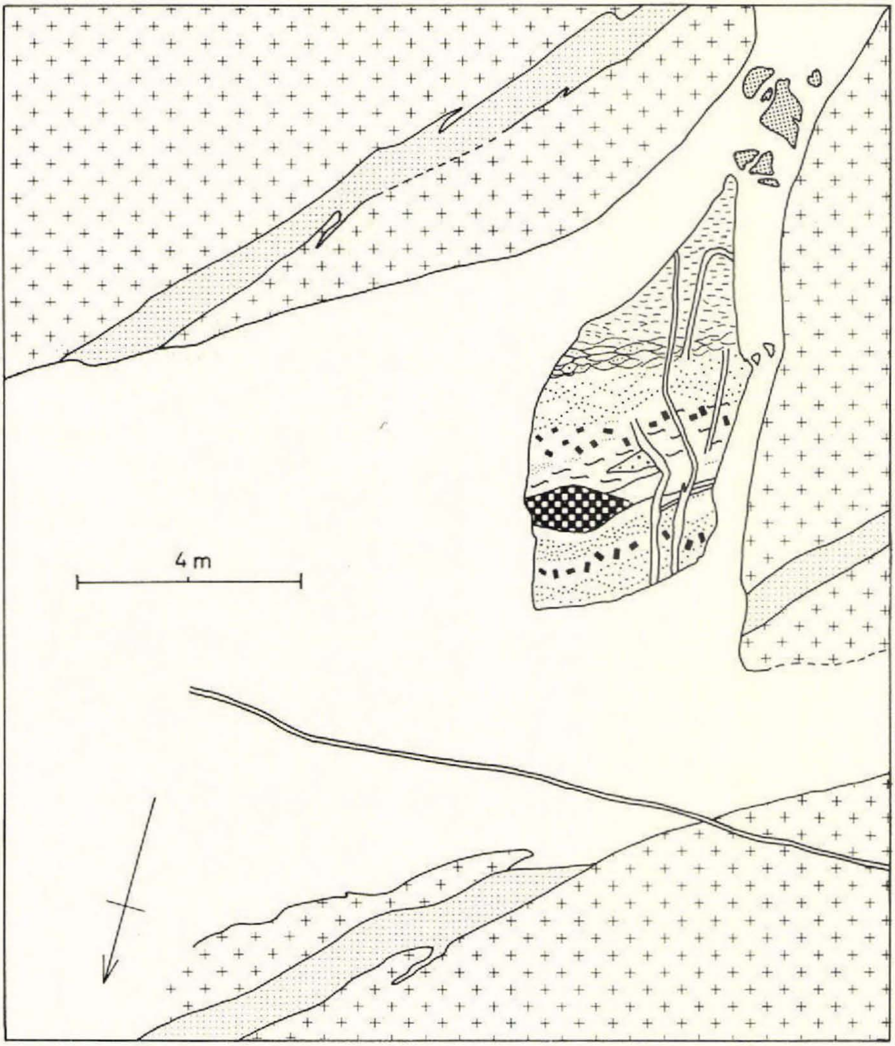

Fig. 13. A block of gneiss (seen in fig. 12) enclosed within homogeneous, intrusive granite (without ornamentation). The gneiss is streaky (schematically represented) with a lens of hornblendite (chequered) and leucocratic zones with microcline porphyroblasts (black rectangles). The outline of the gneiss block is sharp and clear-cut, particularly the N.W. and N.E. sides which are parallel and normal respectively to the lithological layering in the block. The well-defined sides and the lack of embayment by granite along layers of similar composition suggest that there has been little corrosion of the block and that the present edges are almost the original fracture surfaces. The gneiss was cut by thin pegmatites before it was emplaced in its present position. Small dioritic blocks (coarse dots) south of the main block were probably originally associated with the gneiss block. The intrusive granite cuts an older granite (crosses) and an aplite dyke (small dots) within it, but is transected by a later, thin pegmatite dyke. Bay on south coast of Mátâta nunâ.

Granite — meta-sediment contact to the east of Uniarîssat

East of Uniarîssat a sharp contact between a small area of meta-sediment and granite (figs. 14 and 15) dips steeply north-west at the only place where it is clearly exposed. Here too a well-defined 15–40 cm wide intermediate zone which disappears westwards, separates the granite and meta-sediment. The zone grades into granite over about 1 cm, but is knife sharp against meta-sediment. It cuts abruptly across the bedding of the meta-sediment (fig. 15) and, on the headland, sends an embayment into the meta-sediment.

In a very few places a second intermediate zone, gradational to the first, can be distinguished between the main intermediate zone and the granite. The main intermediate zone is fine-grained, light grey in colour and develops small hornblende crystals. The second intermediate zone is granitic in character. Its small feldspar porphyroblasts are similar in size to the feldspar in the granite but are foliated parallel to its contacts and its matrix is finer-grained than that of the granite.

Ptygmatic¹) veins are abundant in the meta-sediment near the granite and they cut the main intermediate zone. It has not been possible to determine whether this veining comes from, or cuts, the secondary intermediate zone and the granite. Epidote veins, often showing boudinage structure, are also frequent in the meta-sediment. No epidote vein penetrates the intermediate zone. Pegmatite is sometimes present along the contact between the meta-sediment and the main intermediate zone; it is rich in magnetite.

Preferred orientation of feldspar crystals is common in the second intermediate zone, but rare in the granite.

The following features are significant.

1) The intermediate zone is comparatively constant in width and closely follows the border of the meta-sediment (fig. 14).

2) The intermediate zone transects banding in the meta-sediment (fig. 15) and displays no trace of banding.

3) Ptygmatic veins cut the meta-sediment and the intermediate zone.

The contact can be interpreted in two ways, depending upon the bias of the observer and interpretation of the emplacement of the granite. There is no independent evidence of the mode of formation or emplacement of the granite.

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¹) Ptygmatic veins are small tortuously folded quartz-feldspar veins where the axes of the folds have *no* common orientation (cf. Read, 1931 p. 110—111; Dietrich, 1959 p. 358).

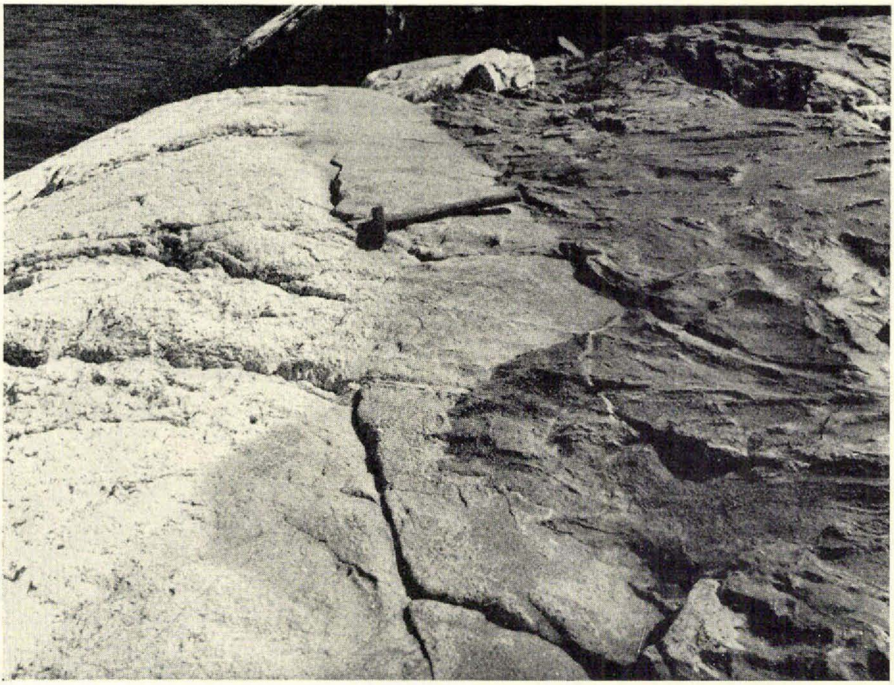

Fig. 14. The contact between granite and meta-sediment to the east of Uniaríssat. The intermediate zone is about 30 cm wide.

1) If the granite is assumed to be intrusive the intermediate zone could be ascribed to mixing of granitic material with meta-sediment. This mixture lubricated the contact and flushed it out to give the sharp, cross-cutting relationship. The ptygmatic veining is later. Its tortuosities depend upon the relative competencies of the injected material, the host rock and the less ductile adjacent rock which acted as a resister against which the vein buckled (Wilson, 1952).

2) If the granite is ascribed to replacement of supracrustal rocks the intermediate zone may be considered due to granitization of meta-sediment at the margin of activated granite. Replacement granites may lie in sharp contact with earlier rock. It may be argued that the ptygmatic veins were emplaced during deformation. Since there is no evidence of deformation after the formation of the intermediate zone, the ptygmatic veins must have been emplaced before or during the formation of the intermediate zone, where they survived granitization because of their low granitization potential.

Better evidence of late local mobilization of the granite is known from Mátâta nunâ not so far away. Thus it is feasible that late reactivation of the granite may have locally flushed out the contact and formed

Fig. 15. The contact between granite and meta-sediment to the east of Uniarissat. The intermediate zone cuts off epidote veins and banding in meta-sediment but is traversed by quartz-feldspar veins which terminate against the main granite.

the intermediate zone by mixing. The lack of an intermediate zone further to the west along the contact does not affect any of the argument; reactivation or mobilization may have been, and probably was, very local.

The pattern of behaviour of trace elements across the contact was investigated to see whether it would give any indication of the origin of the intermediate rock types. The results are inconclusive and difficult to interpret. However, the second intermediate zone shows considerable enrichment in copper and zinc and slight enrichment in lead compared to both the granite and the meta-sediment.

Discussion

The homogeneous, intrusive granite in the different localities described cuts the nebulitic granite-gneiss, which forms part of the country rock granite with palimpsest structure, and cuts a discordant amphibolite dyke, but is cut in one place by a thin pegmatite dyke. Thus it is probably Sanerutian (cf. Watterson, 1965).

The Sanerutian, though not a period of folding, does appear to be a time when there was an increase in the thermal gradient. Tuttle and Bowen (1958) discuss the conditions for incipient melting of average granite. With a thermal gradient of 30° C/km, which is most commonly accepted, melting will begin at depths of 12 to 21 km as long as there is sufficient water. The depth decreases to between 9 and 13 km if the thermal gradient is raised to 50° C/km. While this high gradient is rare it might be expected during a thermal event such as that during the Sanerutian. A granite need not be completely melted to become intrusive. Under certain temperature and pressure conditions it will melt completely in the presence of 2 % of water. If it is of eutectic composition under the same temperature and pressure conditions it will simply recrystallize if there is less water. Thus with deficient water a granite may fracture, while at the same pressure and temperature with sufficient water it will melt sufficiently to become mobile.

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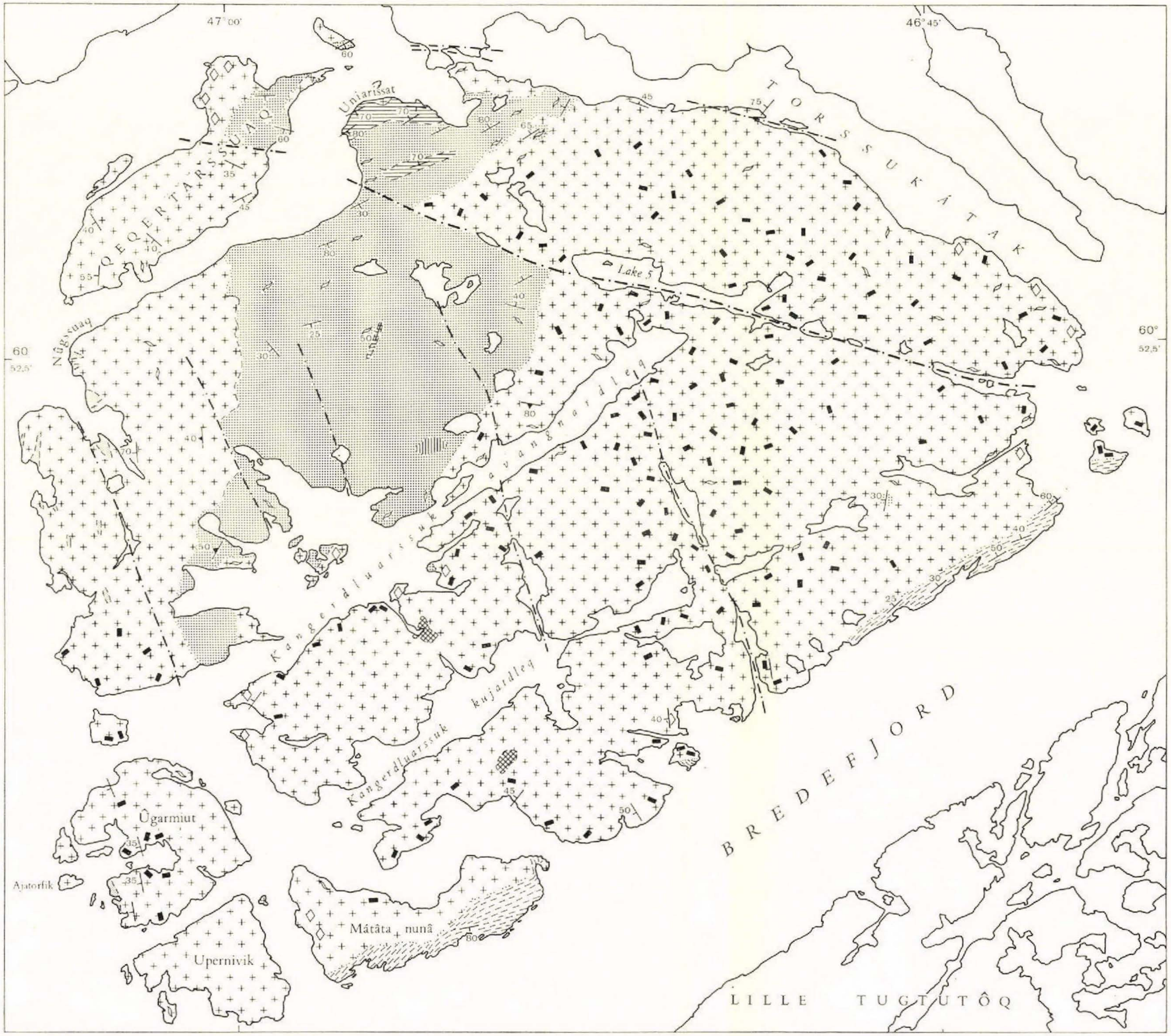
GRØNLANDS GEOLOGISKE UNDERSØGELSE
THE GEOLOGICAL SURVEY OF GREENLAND

MEDD. OM GRØNL. BD. 179 NR. 8 (W.S. WATT)

PL. 1



THE BASEMENT GEOLOGY OF
QAERSUARSSUK
SOUTH GREENLAND

+ Granite, medium-coarse grained
~ Basic lenses or enclaves
▮ Porphyroblasts
◇ Agmatite
▥ Diorite
▒ Gneiss, granitic gneiss
≋ Nebulitic granite-gneiss
▩ Amphibolite
▤ Meta-sediment
░ Fine-grained quartz-dioritic rock
–·–· Faults and crush zones
—— Established boundaries
– – – Inferred boundaries
…… Arbitrary boundaries
⊥ Lithological layering of any origin
▲ Foliation

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