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GRØNLANDS GEOLOGISKE UNDERSØGELSE
BULLETIN No. 59

THE DEFORMATION AND GRANITISATION OF KETILIDIAN ROCKS IN THE NANORTALIK AREA, S. GREENLAND

BY

ARTHUR ESCHER

WITH 44 FIGURES AND 9 TABLES IN THE TEXT
AND 6 PLATES

DANISH GEOLOGICAL CONTRIBUTION
TO THE INTERNATIONAL UPPER MANTLE PROJECT

Reprinted from
Meddelelser om Grønland, Bd. 172, Nr. 9

KØBENHAVN
BIANCO LUNOS BOGTRYKKERI A/S
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Abstract

The Nanortalik peninsula, situated between the fjords of Tasermiut and Sarqâ, is largely composed of Ketilidian schists, quartzites and volcanic rocks. All these rocks are more or less strongly folded. The folding took place probably in three successive phases during the Ketilidian period: A first deformation resulting in folds with NNE trending axes, was followed by a second major phase of folding with NW axes. This second folding was essentially plastic. A third deformation, acting probably on a more rigid mass, was characterised by the formation of fracture-cleavage. Third-period folds possess very long wavelengths; their axes are oriented NNE to NE.

Migmatisation started probably during the second deformation period resulting in the formation of many dykes and veins of pegmatite and aplite. Four generations of Ketilidian pegmatites can be recognised. Most of them appear to have been formed by metasomatic replacement.

It seems that during the Ketilidian orogeny, the evolution of the schists and gneissic schists tended to a granodioritic composition. Potassium metasomatism only became active at the end of the Ketilidian period.

In the NE part of the Nanortalik peninsula, three Sanerutian granites can be observed. These granites are similar in composition (quartz-microline-biotite), but possess different ages and textures. The time interval between the last Ketilidian deformation and the emplacement of the first Sanerutian granite was marked by the intrusion of several metadoleritic dykes.

The first and principal Sanerutian granite usually shows an indistinct foliation due to numerous oriented inclusions. Field evidence indicates that this granite was formed mainly by replacement of volcanic rocks. Chemical analyses show that large amounts of K, Si and Na have been supplied to produce the granitisation of the volcanic rocks.

The second Sanerutian granite is characterised by a coarse porphyroblastic texture and appears to have been emplaced partially by the intrusion of a melt and partially by a subsequent replacement of the host-rock. Finally, the last Sanerutian granite displays all the characteristics of a pure intrusive body. It is generally very fine-grained and forms many cross-cutting dykes.

Acknowledgements

I wish to express my hearty thanks to my former teacher, Professor H. BADOUX, for the support he has always given me during the accomplishment of this thesis. My special thanks are also due to Professor R. WOODTLI, who kindly supervised the petrographical and mineralogical part of this work.

I am very grateful to the Director of the Grønlands Geologiske Undersøgelse, Mr. K. ELLITSGAARD-RASMUSSEN, who made it possible for me to join the South Greenland expeditions for two summers.

It is also a pleasant duty for me to extend my best thanks to all persons who in one way or another helped me in Lausanne, Copenhagen and Greenland: To my colleagues, Dr. M. BURRI and Dr. M. WEIDMANN at the University of Lausanne; to Dr. J. ALLAART at G.G.U. in Copenhagen; to the skippers, Messrs. A. VIDSTEIN and H. VALENTIN; to my field companions, Messrs. A. HAVSTEEN-MIKKELSEN and M. PONTOPPIDAN for their help and patience.

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Finally I express the deepest gratitude to my wife, who helped me with the enlargement of the photographs and the drawing of the maps, and who typed the text.

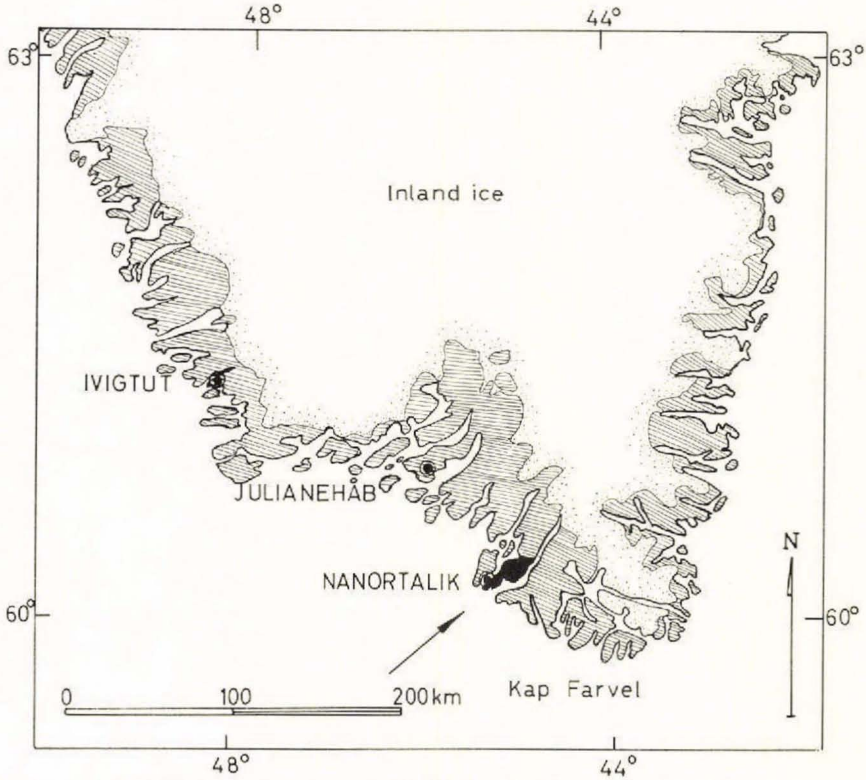


Fig. 1. Map showing the southern part of Greenland and the location of the area under consideration.

I. INTRODUCTION

Geographical situation

During the summer months of 1961 and 1962 the author was given the opportunity to participate in the geological expeditions organized by the Grønlands Geologiske Undersøgelse (G. G. U.) in S Greenland. The area that was examined during the two seasons includes the SW part of the peninsula situated NE of Nanortalik, between the fjords Tasermiut and Sarqâ-Sermilik. To the NE, the region is limited by the valley lying N of the Arpatsivip-qáqâ mountains (see the map fig. 1 and the geological map on plate 5).

The area adjoining to the NE has been investigated by J. Watterson during the same summers of 1961 and 1962.

Included in the territory is the island of Nanortalik and some small islands more to the SW. In the following text the term Nanortalik area or peninsula will be used for the whole of the examined territory. As this was far too large to be mapped and examined in detail in two summers, the entire area was mapped roughly during the first season (geological map on plate 5).

The second season was used to map and investigate two relatively small key areas situated in the central and NE part of the peninsula.

The geological mapping was executed with the help of topographical maps (scale 1:20.000) and aerial photographs from the Geodætisk Institut in Copenhagen.

Previous investigations

GIESECKE, who in the years from 1806 to 1813 carried out mineralogical studies along the west coast of Greenland, visited Nanortalik in 1806 on his journey to Lichtenau. He describes the rocks of Nanortalik island as granites with gneiss and syenite intercalations and mentions the presence of garnet and tourmaline. Like other early investigators, he was principally interested in the alkaline massifs.

In 1896 JESSEN visited Tasermiut fjord and mapped the "New granite" around Aniggoq mountain. In his description of the new granites he drew attention to the "Rapakivi-like" texture shown by the feldspars.

USSING compiled in 1912 the work of earlier geologists and named the new granites "syenites".

To WEGMANN (1938 and 1939) is owed most of the geological information and theoretical background concerning the area. He laid down the foundation of the presently adopted chronological division when he divided the Precambrian of southern Greenland into two main periods: the Ketilidian and the Gardar. WEGMANN discerned two principal groups in the Ketilidian formation: the Sermilik group, characterised essentially by pelitic and psammitic rocks, and the Arsuk group, by volcanic rocks. He visited Nanortalik island and Tasermiut fjord and gave several descriptions of the gneisses, quartzites, schists and granites belonging to the area. WEGMANN (1939 p. 201–202) also noticed the strongly folded layers in the Quvnerssuaq area and, although he did not observe them, he guessed the presence of pillowlavas and other volcanic rocks in the higher parts of the Nanortalik peninsula.

My regretted teacher and colleague, LORETAN, during the summer of 1960 started to investigate the whole of the peninsula situated to the NE of Nanortalik, including the area under consideration. Unfortunately, he was taken away by a fatal illness at the end of the same year. Several of his field notes were very valuable in the present investigation. The exact field terminology of the various rock-types in the Nanortalik area was for a large part laid down by LORETAN. His observations also enabled the whole of the region to be mapped during the first season on a large scale, and a preliminary stratigraphy to be established.

Summary of the general geology and chronology

The Ketilidian rocks are well represented in the Nanortalik area. On Nanortalik island and on the adjoining SW part of the peninsula there are essentially pelitic and semi-pelitic gneissic schists and schists. This monotonous series is overlain in the central part of the area by quartzitic schists and volcanic rocks. The quartzitic schists pass into massive quartzites in the NE of the region.

All these Ketilidian rocks are more or less strongly metamorphosed; the degree of metamorphism varies from low granulite facies for the basal formations, to greenschist facies for the upper volcanic deposits. The most common, however, is the amphibolite facies of metamorphism.

Folding took place in three successive phases during the Ketilidian period: A first, violent deformation, resulting in folds with NNE trending axes, was followed by a second major phase of folding with NW axes. This second folding was essentially plastic. A third deformation, acting probably on a more rigid mass, was characterised by the formation of fracture-cleavage. Third-period folds possess very long wavelengths; their axes are oriented NNE to NE.

Migmatisation started in the area probably during the second deformation period. Most of the schists and gneissic schists present strong migmatic features. They are particularly characterised by many dykes and veins of pegmatite and aplite. Four generations of Ketilidian pegmatites can be recognised (P1–P4).

During late-Ketilidian times, at the end of the third folding period, Si-Na-K metasomatism was particularly active, resulting in an albitisation and microclinisation of the rocks. The microclinisation continued after the end of the albitisation.

Relatively small bodies of granodiorite were probably formed by replacement during the second folding period. One of these granodiorites forms the core of an anticline S of Qagdlua mountain (geological map on plate 5).

At several places small amphibolite bodies or dykes (DA 1) were observed. These amphibolites show, when examined closely, a small but distinct discordance with the bedding. They are however folded and refolded together with the host rock. It appears therefore that they could represent the feeders for the volcanic rocks which form the upper part of the Ketilidian.

A second generation of amphibolite dykes (DA 2) was observed in three places on the Nanortalik peninsula. The DA 2 dykes are only slightly folded in places. They transect most Ketilidian formations and structures but are often partially replaced by late-Ketilidian pegmatites. Their age must be late Ketilidian. The term Sanerutian was introduced by Berthelsen (1960, 1961) together with the term Kuanitic to designate two distinct periods in the Ivigtut area between the end of the Ketilidian and the beginning of the Gardar. In that area the Kuanitic is characterised by the intrusion of numerous basic dykes, whilst during the Sanerutian several granites were formed. In the Nanortalik area however, two generations of basic dykes (DA 2 and DA 3) are separated by the emplacement of granite. Therefore there is no clear-cut distinction between a period of basic dyking and a period of granite emplacement and so the term Sanerutian is used to indicate the whole time interval between Ketilidian and Gardar.

The age of the large granite body (granite A) is determined by the fact that it transects and replaces the large P 4 pegmatites and the DA 2 dykes, and by the fact that it is itself cut by the DA 3 dykes. These observations place the granite A in the beginning of the Sanerutian period. This granite usually shows an indistinct foliation due to numerous oriented inclusions. Many observations show that the foliated granite was formed mainly by replacement of quartzitic schists and volcanic rocks. It occupies the higher regions in the centre and NE of the area.

Both DA 2 and DA 3 dykes are metadolerites, as shown by composition and texture. The intrusion of DA 3 basic dykes, which came after the emplacement of granite A, was followed by a period of NE faulting and shearing. The NE oriented shearplanes cut both the foliated granite and the dykes DA 3. Straight cross-cutting pegmatites and aplites appear to have been emplaced along the NE shearplanes.

The period of pegmatitisation and aplitisation was followed by the emplacement of a second post-Ketilidian granite (B). This granite seems to belong to the group of "New granites" (Bridgwater 1963), being characterised by a very coarse-grained porphyritic texture. It occurs in the NE part of the area in the Aniggoq and Arpatsivîp-qáqâ mountain ranges and is connected with the very important occurrence of new granite on the other side of Tasermiut fjord (Wallis 1962). Many observations show that this granite B was partly emplaced by intrusion and partly by replacement of the schists, quartzites and granite A.

The emplacement of the Sanerutian coarse-grained granite was followed by a period of pegmatitisation (large N-S pegmatites) and vertical shearing with faulting.

The end of the Sanerutian period came probably with the formation of a fine-grained granite (C) which occurs mainly in the Arpatsivîp mountains. It appears to have extended from there by a network of intrusive dykes which can be seen all over the Nanortalik peninsula. The main body of the microgranite seems to have been emplaced principally by intrusion.

Table 1. Chronological succession of geological events in the Nanortalik area.

Period	Deformation	Granitisation	Regional Metamorphism	Dyking
	Recent fracturing and faulting		Retrograde metamorphism	
Gardar				Dolerites (NNE and NNW).
	Vertical NW-faulting			
				Lamprophyre sheets
? ? ?	NNE-faulting and shearing			
Sanerutian		NE-aplites and pegmatites.		
		Microgranite (C)		
	Vertical NNE to NE shearing and faulting			DA 3 Meta-doleritic dykes
		Large N–S pegmatites		
		Coarsely porphyroblastic granite (B)		
	Tension?	Aplites and pegmatites along shearing zones		
	Deformation, NE-shearing			
	Tension?			
		Microcline-biotite-granite (A)		
? ? ?		P4 Pegmatites microclinisation	Retrograde metamorphism	DA 2 Meta-doleritic dykes
Ketilidian	Folding III (NE to NNE axes).	P3 pegmatites, microclinisation and albitisation	Maximum progressive metamorphism	
	Folding II (NW-axes)	Granodiorite, P2 pegmatites	Progressive metamorphism	
		P1 pegmatites		
	Folding I (NNE-axes)			
	Sedimentation and volcanism			DA 1 Amphibolite dykes as feeders for volcanic rocks

The Gardar period is only very poorly represented in the area. Some lamprophyre sheets belonging to this period can be observed cutting the quartzites and the different granites (including the last formed microgranite) in the Arpatsivîp-qáqâ range. In the centre of the territory, on Thomsens Ø and on Nanortalik island, a few late-Gardar dolerite dykes cut all the visible formations. Two successive generations of faults (NNE and NW) probably also belong to the Gardar period. Finally, in various points in the area, recent fracturing and faulting can be observed.

Table 1 gives the probable chronological sequence of the principal geological events in the Nanortalik area. In this area the Ketilidian formations and the Sanerutian granites are particularly well exposed, whilst the Sanerutian and Gardar basic dykes are almost lacking. It is for this reason that in the following pages only the Ketilidian rocks and Sanerutian granites will be examined in detail.

II. STRATIGRAPHY OF THE KETILIDIAN SUPRACRUSTALS

General

On the peninsula situated NE of Nanortalik, between the fjords Tasermiut and Sarqâ, extensive exposures of Ketilidian supracrustal rocks are found. The thickness of the sequence, so far as can be observed and has not been removed by erosion, amounts to a maximum of 2200 m.

The succession is largely made up of sedimentary rocks with only the uppermost 700 m being predominantly of volcanic origin. They are not too strongly metamorphosed or migmatised and thus it is possible to recognize their origin. Differences in lithology have enabled the rocks to be divided into the following five stratigraphic units (from top to bottom):

V) Volcanic unit.
IV) Quartzite unit.
III) Pelitic schist unit.
II) Semi-pelite unit.
I) Pelitic gneiss unit.

These five formations can be traced almost all over the area. The best exposures can be observed along the very steep walls forming most of the N and NW shores of Tasermiut fjord. On plate 2, four very simplified columnar sections are presented. They show the various units in four different places along Tasermiut fjord.

At first sight, the layers appear to be very regular and undisturbed. Detailed examination shows however that in many places, especially in the pelites and semi-pelites, repetitions occur due to isoclinal, recumbent folding. Therefore the measurable thickness of a rock formation in no way gives the actual thickness of the series before it was folded.

The pelitic gneiss unit (Unit I)

The pelitic gneiss unit is best represented on Nanortalik island and on the small peninsula of Tuapait (see geological map on plate 5). Its rocks form the lowest Ketilidian formation that can be observed in the area. They are essentially composed of massive, smallfolded, banded and commonly veined gneisses at the base and of gneissic schists at the top.

The basal massive gneisses can only be observed on the E and S coast of the small Tuapait peninsula. They are characterised by the presence of numerous crystals of sillimanite, cordierite and garnet. It seems probable that this lowest part of unit I does not represent a different lithological formation, but that it was derived from the gneissic schists by a more pronounced metamorphism. It has therefore no stratigraphical significance.

The top of unit I is represented by a 200 m thick series of biotite-rich gneisses and gneissic schists. This series is characterised by the presence of garnet and many amphibolitic horizons. The thickness of the intercalated basic horizons varies from a few cm to 8 m. It is possible that this top part of unit I was originally a pelitic-calcareous sediment.

The colour of all these basal pelitic gneisses and upper gneissic schists is generally light grey to blue-grey. The total apparent thickness of unit I reaches about 550 m.

The semi-pelite unit (Unit II)

The semi-pelite unit is best represented in the area of Tusardluarnâq mountain and on the small peninsula N of Nanortalik harbour. It is a monotonous sequence of dark grey, more or less quartzitic mica schists with common intercalations of quartzite horizons. The thickness of these quartzite layers varies from 50 cm to 20 m. A few amphibolite horizons and several pyrite- and graphite-bearing layers occur in this semi-pelite unit.

In the vicinity of Nanortalik, the formation is distinctly more metamorphosed and contains many augen-beds characterised by the presence of rounded isolated feldspar porphyroblasts. In these more metamorphosed and migmatised areas, the schist passes into a gneissic schist.

The rocks of unit II are separated from the underlying unit by a 50 to 100 m thick series of black biotite schists. This dark formation is particularly well exposed in the Qagdlua mountain range, where it is in contact with the anticlinal core-granite of that area. The basal biotite schists of unit II contain four characteristic pyritic and graphitic horizons, each about 3 m thick. The pyritic schist horizons are particularly conspicuous in the field by their rust-red colour, which is caused by iron oxides resulting from the weathering of the pyrite.

The apparent total thickness of the semi-pelite unit is very variable and can reach up to 500 m. It seems probable that the rocks belonging to this unit are mostly derived by metamorphism from argillaceous and arenaceous sediments.

The pelitic schist unit (Unit III)

The pelitic schist unit occurs nearly all over the area, except on Nanortalik island. Its upper and lower parts are composed of dark brown mica schists in which some pyritic and graphitic layers can be recognized. The middle part is essentially made of monotonous dark grey, slightly quartzitic mica schists. The upper part of these dark grey schists is characterised by the presence of numerous garnets. The thickness of the garnet-rich formation varies from 150 to 200 m. The total thickness of the whole unit amounts to 450 m.

It seems logical to infer that the schists belonging to unit III for their largest part were originally argillaceous sediments. The presence of garnets in some of the schists could indicate that they were also slightly calcareous in places. The sulfide- and graphite-bearing layers show moreover that here and there organic remains were mixed with the pelitic material.

The quartzite unit (Unit IV)

This unit is best represented in the NE part of the examined area, where the measurable thickness reaches over 800 m. It is composed principally of massive light grey to green quartzites containing numerous basic sills, intercalated mainly at the top and base. The massive quartzites often show a clear layering and, in some parts, the remains of current bedding and graded bedding. Even ripple marks have been observed in one place, where the rock was especially well preserved. Numerous conglomeratic horizons have been found. Hematite grains are frequent in all the quartzites.

The lithology and the location of the quartzites prove that they belong to the important quartzite formation extending towards the N (between Tasermiut and Sermilik fjords), which was described by WEGMANN (1939).

To the SW, the quartzite series gradually decreases in thickness and changes from a massive quartzite to a quartzitic schist. In the Qagdlua mountains these quartzitic schists are particularly well exposed.

The total thickness of the quartzite unit varies considerably from the SW, where it reaches only 150 to 200 m, to the NE where a maximum thickness of 850 m was measured. WEGMANN reports, still more to the NE, thicknesses up to 2000 m for the quartzitic sequence. In contradiction with WEGMANN's hypothesis that the quartzites underlie the Nanortalik schist and gneiss series, it seems now clear that the quartzites overlie the schists and are in direct contact with the volcanics which form the top of the Ketilidian succession.

The relatively large amount of hematite found in the rocks of unit IV indicates that they were no doubt originally red coloured sandstones.

Fig. 2. Western slope summit 1166 m (Qagdlua mountain) showing dark volcanic rocks in the upper part overlying massive quartzitic schists. The lowest part is formed of pelitic schists.

The volcanic unit (Unit V)

The rocks belonging to unit V are essentially volcanic: pillow-lavas, lapillis, tuffites. Interstratified with these are a few sedimentary horizons: quartzites, quartzitic slabs, green schists, breccias and conglomerates. Gabbro sills are common in the rocks of unit V. Their intrusion was probably due to the same volcanic activity which gave rise to the volcanic series.

The best exposures of these rocks are found around lake 410 m (N of Qagdlua). They also occur in the Quvnerssuaq mountain range and more to the north, where they form large-scale inclusions in the granite. The generally dark green colour of the formations belonging to unit V makes it very easy to recognize them even from great distances (fig. 2).

The lower part of the volcanic unit is essentially composed of epidotized pillow-lavas with intercalations of pyritic epidote-hornblende-schist. The unit generally starts at its base with some biotite-hornblende schists overlain by one or more thick (30 to 100 m) gabbro sills.

The upper part of unit V is predominantly composed of tuffites, lapillis and pillow lavas with common intercalations of quartzites, quartzitic slabs and thin greenschist horizons. Most of the pillows are considerably compressed and are often difficult to recognize.

The layers of unit V appear to be perfectly concordant with the underlying quartzites. Nowhere has a discordance been found. In some places a thin conglomerate horizon is intercalated in the basal schists. The top of the volcanic rocks has not been found. The visible thickness of the unit reaches about 700 m. The presence of large quantities of pillow lavas in the volcanic unit indicates a formation by subaqueous volcanic activity.

Discussion and conclusions

By interpreting the characteristic properties of each rock unit, an approximate picture of the origin of the observed formations can be obtained.

A thick series of sediments was laid down in a geosyncline. The source was probably made of very variable and complex crystalline rocks.

No basement of pre-Ketilidian rocks has been found; it is not known how the sedimentation started. However, it can be inferred that during a certain period of the geosynclinal cycle the sedimentation was essentially pelagic, giving birth to the units I, II, and III. The sediments must have been for the greatest part argillaceous with possibly some addition of calcareous material at the end of units I and III, as is indicated by the presence of amphibolitic layers and garnets. The presence of psammitic material in unit II suggests, moreover, a temporary dedrease in sedimentation depth, permitting the periodic arrival of detrital material. This could explain the alternation of pelitic and psammitic beds in unit II. The sulfide- and graphite-bearing layers, which occur in many places in units I, II and III, indicate that here and there organic remains were mixed with the argillaceous sediments.

After this essentially pelagic phase, the sedimentation became predominantly detrital, the result of which was the formation of the quartzitic rocks of unit IV. The presence in the quartzites of graded- and current-bedding structures as well as ripple marks, indicates that the psammitic material was laid down in a shallow sea. The enormous difference in thickness between the quartzites in the NE (800 m) and those in the SW (150 m), could be explained by sedimentation in the littoral zone at the border of the geosyncline.

Large quantities of detrital material were laid down near the border of the geosyncline, and much less, farther away from that border. This possibility is confirmed by the fact that the massive quartzites from the NE part of the area pass gradually into quartzitic schists when followed to the SW. According to this theory, the border of the geosyncline was situated at a certain period quite near the NE part of the region. The numerous conglomeratic horizons within the quartzites show that the sedimentation level was now and then near sea level.

After this detrital phase of deposition, during which the psammitic rocks of unit IV were formed, there probably followed a new deepening of the basin. This is corroborated by the presence of pelitic schists immediately on top of the quartzites. After the sedimentation of these argillaceous deposits, an intensive volcanic activity took place, resulting in the formation of the volcanic rocks of unit V. As the major constituent of units V is pillow-lava, it can be assumed that the eruptions took place on the bottom of the sea. The presence in the lower part of the volcanic unit of several interbedded pelitic schist horizons indicates, moreover, that the eruptions took place in deep sea during the early part of the formation of unit V. On the other hand the upper part of the volcanic rocks appears to have been formed later at shallow depth. This is suggested by the presence of many quartzite horizons.

Another result of the volcanic activity was the formation of numerous basic and ultrabasic sills and dykes within all the earlier formations. This appears to have been the case especially in the quartzites. Remnants of feeding dykes are locally found in most of the Ketilidian rocks, where they are always strongly folded and metamorphosed together with their host rock.

There seems to exist a striking similarity between the volcanic rocks of the Nanortalik area and those of Arsuk Ø as described by Wegmann (1938) and by Mulller and Froidevaux (1957). Common to both areas are pillows, breccias, tuffites, lapilli and intercalations of quartzitic slabs. It is tempting therefore to correlate the formations in the two areas.

III. THE KETILIDIAN STRUCTURAL EVOLUTION

General

Within the examined region the effects of at least three periods of folding have been recognized. During each period major folds accompanied by small-scale structures were developed.

The abundance of small structures has greatly assisted in the investigation of the larger folds, once the relation between large and small structures had been established.

The first and third phases of folding both gave birth to structures with NE to NNE trending axes. This often made it difficult to distinguish one from the other, especially in smallfolds.

The second Ketilidian phase led to the formation of well marked structures with NW trending axes. As a result several cross-folds were formed.

Due to the strong relief of the area and to the presence of several reliable bedding horizons, the reconstruction of the resulting structures proved to be relatively easy. This was particularly the case in the Qagdlua and Quvnerssuaq mountain ranges, where three dimentional mapping was made possible by high, well exposed cliffs and walls (see the detailed geological map of this region).

The general pattern of the major folds is presented on the structural map figure 3.

The first folding period

The oldest discerned phase resulted in the formation of many small- and large-scale isoclinal and often recumbent folds. The axes of these structures seem originally to have had a NE to NNE trend. As a result of later refolding their actual direction is variable. Their amplitudes can reach up to a 1000 m, but are generally not over 50 m. The axial planes of these folds generally dip to the NW.

Most affected by this first deformation are the lower gneisses and mica schists. The more rigid overlying volcanic series and quartzites seem to be deformed only to a minor degree by this first folding. Isoclinal smallfolds seem to be almost absent in the quartzites and volcanics.

Fig. 3. Simplified structural map of the central part of the peninsula situated between Tasermiut and Sarqâ fjords.

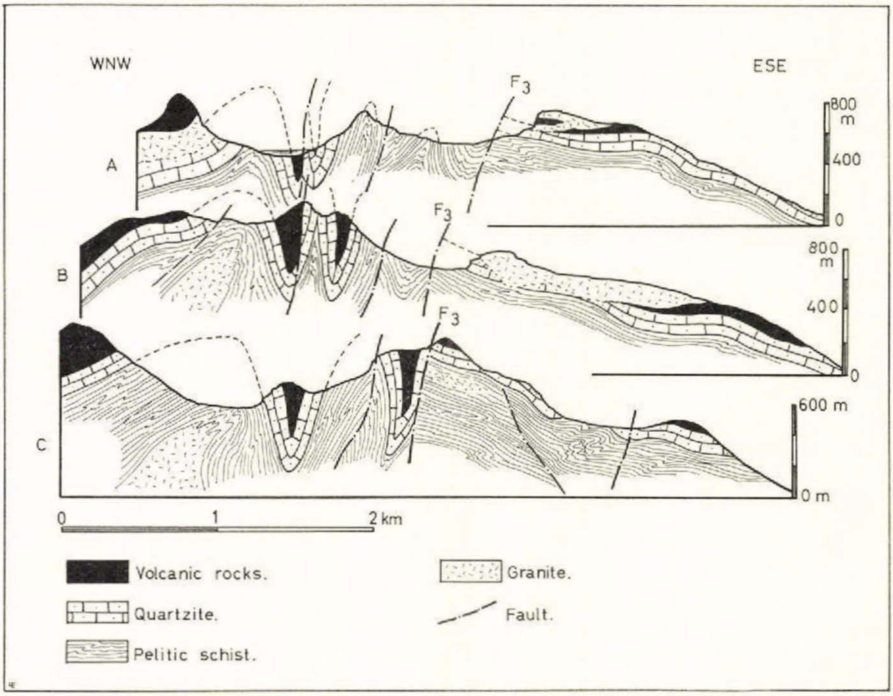


Fig. 4. Three hypothetical sections across the Quvnerssuaq mountain area. The exact position of the sections is represented on the map of fig. 3. The structures are a result of the first folding period.

These formations can thus be considered suprastructural compared with the more plastic, gneissic and schistose infrastructure (WEGMANN 1938).

However, in several places large-scale folds were observed in which the quartzitic and volcanic rocks are also strongly folded together with the underlying schists. This is the case in the area directly to the NW of Quvnerssuaq mountain, where a succession of steep, pinched synclines and anticlines, in places separated by thrust faults, was observed (fig. 4). The core of the synclines in these structures is generally formed by readily recognizable dark volcanic rocks. One of these synclines can be seen from the top of the Uitdlut shoulder (near Tasermiut fjord) looking towards the Quvnerssuaq mountains (fig. 5).

The fold-axes in this area NW of Quvnerssuaq, form a pattern of arcs with NNE, NE and WNW directions connected by tight bends (fig. 3 and plate 3). This bending of the axes must have been caused by the second NW folding which thus proves the early age of these structures.

In the south cliff of the Qagdlua mountains, viewed from Tasermiut fjord, an important anticline with a NNE axis can be observed, which

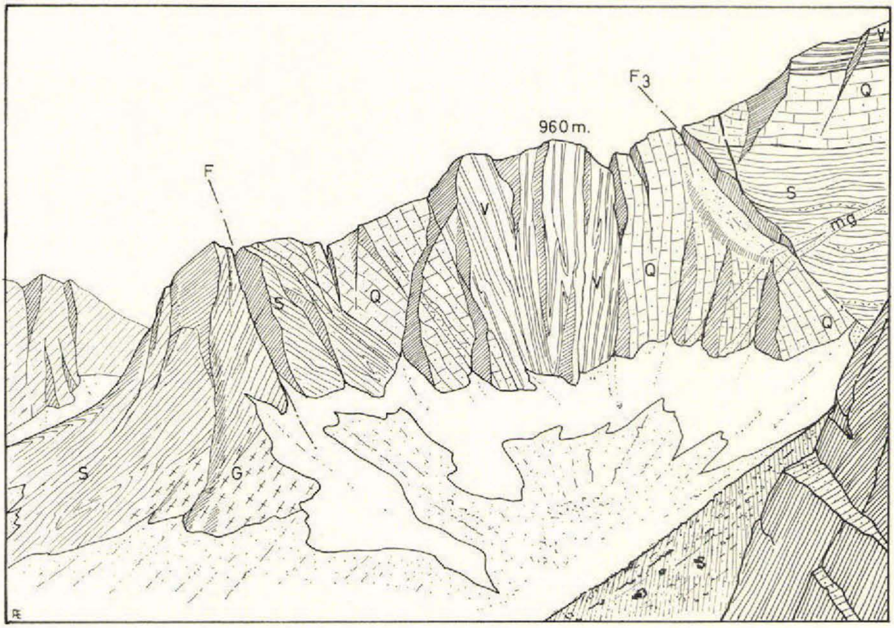


Fig. 5. View of the southern wall of the ridge coming down from Quvnerssuaq summit to lake 640 m. V = Volcanic rock; Q = Quartzites; S = Mica schists; G = Granite; F = Faults; mg = Microgranite.

is filled by a core granodiorite. This anticline must also be a result of the first folding period, for both limbs are slightly refolded by the second NW deformation.

The small-scale folds formed during the first phase in the infrastructure possess often strongly isoclinal forms. Moreover, they generally are recumbent and in some places develop axial-plane schistosity. All these facts make it very difficult to recognize them within the monotonous mica schists and gneisses. It was only with the help of the pyritic and graphitic schist horizons that in the apparently undisturbed isoclinal series the presence of such repetitions could be ascertained. A clear example of one of these repetitions has been observed in the garnet schist formation near lake 430 m, north of Quvnerssuaq (fig. 6).

It seems that most of the easily recognizable small-scale structures over the whole area are a result of the later, second and third folding phases.

The second folding period

The second phase of folding led to the formation of small- and large-scale structures with northwest-trending axes. This folding has affected on a large scale both the supra- and infra-structure. However, on a

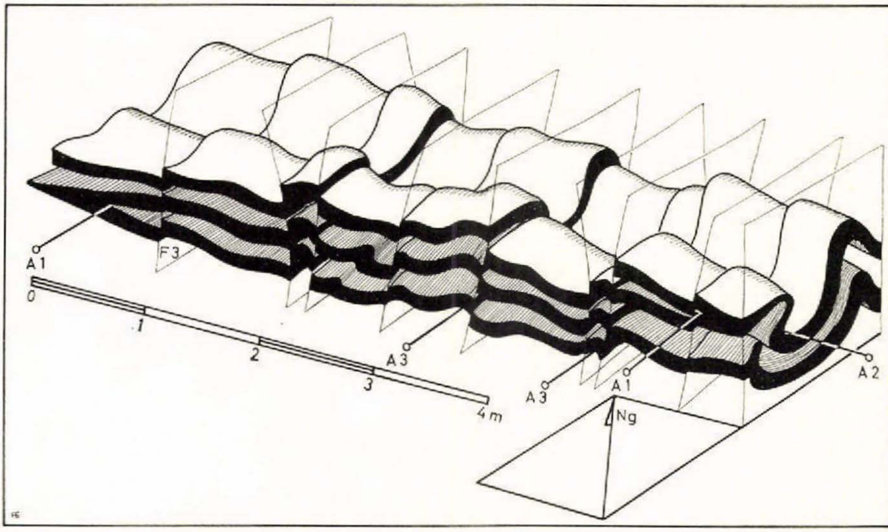


Fig. 6. Simplified reconstruction of a multiple-folded pyritic schist layer in the garnet-mica schist formation near lake 430 m (North of Quvnerssuaq). It is clear that the NW-folding (axis A 2) refolds the earlier formed isoclinal fold (axis A 1), while the coarse strain-slip-cleavage (F 3) cuts the whole structure.

small-scale, the gneisses and mica schists of the lower structural level are much more deformed by the NW folding than the overlying quartzites and volcanics. The resulting folds are generally open structures and their amplitude and wave-length reach up to a 1000 m.

It seems that the second deformation phase is the most important of the three. At least it gives at present the best recognizable structures in the field, in small as well as in large folds. One example of a large-scale NW structure can be seen near lake 560 m (north of Qagdlua), where the volcanic rocks, together with the underlying quartzites and schists, are folded into a large open anticline. Another anticlinal valley has been observed more to the northeast, between Aniggup qáqâ and Arpatsivîp-qáqâ mountains. This valley also has a NW direction. Many other similarly directed large-scale folds can be reconstructed (fig. 3).

Several of these folds appear to have deformed older first-period structures. For instance, this is the case in the region north of Quvnerssuaq. Here, as described in the preceding chapter, large isoclinal folds with originally NNE trending axes have been refolded and twisted by the second NW folding. Typical in such structures is the curved direction in a horizontal plane which the axes have after a second refolding (fig. 3 and plate 3).

Similar cross-folds have been described in Northern Scotland by Sutton (1960).

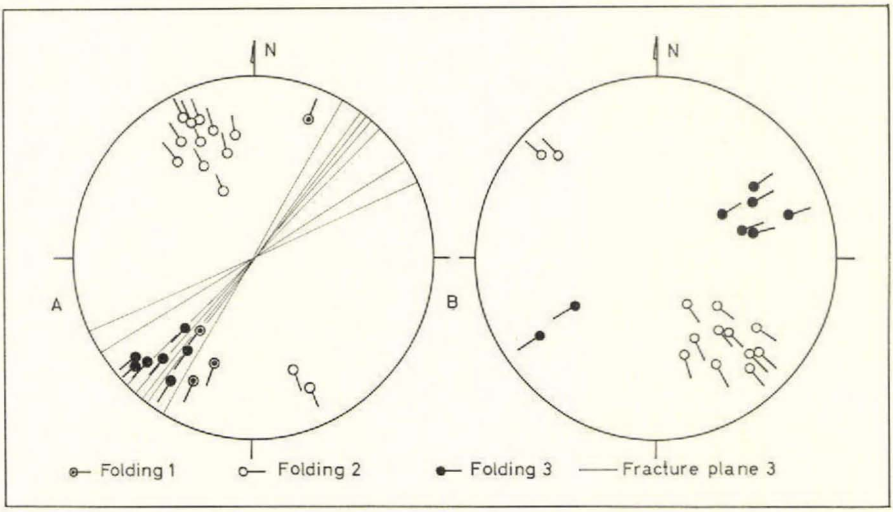


Fig. 7. A: Stereogram of smallfold-axes and orientation of the joints resulting from the fracture-cleavage in the area east of lake 430 m. B: Stereogram of smallfold-axes from the region around Nanortalik (Wulff net, lower hemisphere).

A multitude of small-scale structures with approximatively NW trending axes have been observed throughout the peninsula.

In the different places where first-phase isoclinal small-folds have been discerned, the second deformation with NW-axes can also be recognized. This second folding superimposes itself, on a small-scale, upon the first formed structure.

A striking example of such a small-scale refolding has been observed and reconstructed in the garnet-mica schist series on the eastern shore of lake 430 m (north of Quvnerssuaq, fig. 6). An isoclinal recumbent small-fold of a pyritic schist layer has here clearly been refolded by a NW small-fold system. This structure shows moreover the effects of a third deformation, which will be discussed in the next section. The reconstruction of the resulting small-scale structures, as represented on fig. 6, was made possible by a set of recent open fractures in the rock mass. These allowed the essential lines of the structure within the formation to be followed.

As during the preceding phase, it seems that the NW small-folds have affected the infrastructure to a higher degree than the suprastructure. The disharmonic character of these movements within the infra- and suprastructure commonly resulted in a tectonization of the border zone between the two levels. In consequence the original stratigraphic contact between the schists and the overlying quartzitic and volcanic rocks has often not been preserved. However, in this region the tectonic discordance between the two structural levels appears to exist only on a small scale. On a large scale there seems to be perfect concordance throughout the whole formation.

In some places the small-scale disharmony is compensated in the border-zone by the formation of "wildfolding". This can be clearly observed for instance in the quartzitic schist series which separates the volcanic unit from the schist unit near lake 560 m (north of Qagdlua mountain). On this spot, the small-folds show axes trending and plunging in all possible directions.

Otherwise the trend of the second-period large- and small-scale fold-axes is relatively constant within an arc between north and west (fig. 7). Their plunge varies considerably and is of course controlled by the dip of the layers prior to the second folding and by the effects of the subsequent third deformation. No preferential axial-plane inclination has been detected in the NW folds.

The third folding period

The third Ketilidian phase of folding is characterised in the area under consideration by the formation of very large-scale folds with faults and thrust-faults, and by the formation of very small-scale folds accompanied by a coarse fracture-cleavage. Intermediate-scale structures, due to this last phase, have not been found on the peninsula.

The trend of the fold-axes varies from ENE to NNE and is thus roughly the same as that of the first-period folds. This makes it of course often difficult to distinguish one from the other. The third-period large-scale folds possess very large wave-lengths (up to 10 km), and somewhat smaller amplitudes (up to 1.5 km).

These large structures have strongly influenced the present morphology of the peninsula. For instance, a very large syncline with a NNE axis has determined the present large depression which cuts diagonally across the peninsula across the lakes 210, 410, 670, and 480 m (see the detailed geological map on plate 6). This "synclinorium" affects the volcanic rocks and all the visible underlying series. To the SE a corresponding large anticline seems to have lifted the complex that was previously isoclinally folded during the first phase (NW of Quvnerssuaq mountain).

The non-synchronism of this large anticline with the equally NNE-directed first-phase folds may be assessed from the following facts:

1) The axial direction of the large anticline remains undisturbed by the second NW folding, whilst the first-formed structures possess strongly curved axes as a result of the secondary NW refolding (fig. 3).
2) The large anticline seems to have refolded the second-period NW folds an given them steep axial plunges.

3) There is a marked difference in the tectonic style between the large wide open anticline and the strongly folded, isoclinal first-period structures.

As is illustrated on the map fig. 3, all the important second-period NW structures possess plunging axes due to refolding by large-scale deformation of the third-period.

Several large NNE to NE oriented faults and thrust-faults seem to have been formed during the third and last Ketilidian deformation phase. One of these faults can be followed across the Quvnerssuaq mountain range (F 3 on figs. 3, 4 and 5), where it seems at first sight to date from the first deformation phase. However, when traced to the NE it does not follow the bend of the first-period axes, but cuts them at right angles (fig. 3). Moreover, it is cut in the SW by late-Ketilidian pegmatites. These facts enable the fault to be attributed to the third deformation period, and they, moreover, present additional proof of the existence of such a third phase in the examined area.

On a small scale the third Ketilidian folding period resulted in the formation of a multitude of relatively open small-folds with axial directions varying from ENE in the southern part, to NNE in the central and northern part of the peninsula (fig. 7). In many parts a coarse fracture-cleavage can be observed which appears to have been formed also during this deformation phase. Where fracture-cleavage and small-folds are combined, as in the area around lake 430 m (north of Quvnerssuaq), the resulting cleavage can be determined as a coarse strain-slip cleavage (Wilson 1961).

These small-scale deformations affected during the third period all the strata in the suprastructure as well as in the infrastructure. In the area east and south of lake 430 m the effects of the last phase have been examined in detail. There, the cleavage planes clearly cut all the older NNE and NW folds, and are closely connected with the third-period small-folds (fig. 6). The orientation of the cleavage planes is roughly the same as that of the simultaneously formed small-folds (fig. 7).

Discussion and conclusions concerning the Ketilidian structural evolution

It has been shown that the rock series on the peninsula situated between Sarqâ and Tasermiut fjords has been subjected to three successive deformations, with respectively NNE, NW and NNE to NE axial directions.

These three deformation phases present a striking difference in tectonic style:

I) The first phase (with NNE axes) is characterised by the formation of strongly isoclinal and often recumbent folds accompanied by thrust-faulting. These characteristics indicate a rather violent type of deformation.

II) The second phase (with NW trending axes) resulted in the formation of mostly open large- and small-scale structures that often strongly refold and twist the previously formed isoclixal structures. This deformation seems to have acted upon an easily deformable mass. A confirmation of this assumption is given by the fact that during the second phase the migmatisation reached a maximum of intensity (see chapter VI). The type of folding during this second phase appears thus to have been essentially plastic.

III) The third phase (with NNE to NE trending axes) is characterised by the formation of very large structures accompanied by large scale faulting. On a small scale, this phase resulted in the formation of fracture-cleavage, strain-slip-cleavage and smallfolding. These facts indicate that the deformation acted during a relatively long period upon a more rigid rock-mass with a higher elasticity limit than during the preceding phase (De Sitter 1956).

There appears to exist a great similarity between the above described succession of folding periods and that of other areas further to the north in Greenland. For instance, in the Ivigtut region Berthelsen (1960) also found three deformation phases with more or less the same successive axial directions (NE, NW and NE successively). Moreover, in the same region Bondesen (1960) and Oen Ing Soen (1962) also came to the conclusion that the second deformation was essentially of a plastic type, while the third folding acted upon a mass which had already gained a certain degree of rigidity.

However, several important differences between the structural evolution in the Ivigtut and in the Nanortalik area appear to exist. The third deformation seems to have been much less pronounced in the Nanortalik district than farther north, where, according to Berthelsen, NW folds were strongly refolded and twisted by this deformation. This could explain the relative simplicity of the resulting structures in the Nanortalik area, and the fact that in this region the difference between supra- and infrastructure is also less pronounced.

IV. PETROGRAPHY OF THE KETILIDIAN SUPRACRUSTALS

General

It has been shown in chapter III that the supracrustal rocks in the Nanortalik area can be divided into five lithologically different units. In the following pages it will be demonstrated how the petrographical composition varies from one unit to the other. In each group only the most important formation will be described. Every description is based on the microscopic examination of several rock slides. These descriptions correspond only to the apparently unmigmatised parts of the different rocks. All the clearly migmatic parts, such as aplites and pegmatites, will be described separately in chapter VI. The content of An in plagioclase was determined by means of extinction angles in sections normal to (010) or (001), and by the measurement of refractive indices. In the absence of a favourable section the universal stage was used.

The pelitic gneiss unit (Unit I, forming the base of the supracrustal rocks in the area).

a) The basal garnet-sillimanite gneiss:

Structure: Gneissic (banded, smallfolded and veined), medium- to fine-grained.

Texture: Grano-lepidoblastic, locally porphyroblastic with garnet porphyroblasts.

Principal constituents: Quartz, microcline, oligoclase (ca. An 20 $^0/_0$), biotite, sillimanite, garnet.

Accessory minerals: Cordierite, hypersthene, muscovite, ore and rutile, zircon and tourmaline as inclusions in quartz and biotite.

Alteration minerals: Chlorite, green biotite, sericite.

Microcline is often microperthitic and has obviously replaced the plagioclase. Oligoclase and microcline form locally coarse diablastic intergrowths. The garnet is strictly not birefringent, it is pink coloured and often possesses coronas of greenish biotite, which occasionally contains slender needles of sillimanite. The sillimanite has a prismatic habit and occurs generally as aggregates of relatively long needles (1–2 mm length). It has clearly developed from the biotite and relics of biotite are often present in the sillimanite aggregates. In some places, bundles of sillimanite needles have been deformed by the growth of garnet porphyroblasts. Oligoclase shows very rarely a mesoperthitic structure. The biotite is strongly pleochroic from yellowish green to very dark grey-brown. Hypersthene occurs as xenoblastic grains; it is altered in many places to a green biotite.

b) The gneissic garnet schist: (the upper part of unit I).

Structure: Gneissic beds alternating with schistose horizons; medium-grained, with fine-grained shear zones.

Texture: Grano-lepidoblastic, locally porphyroblastic with garnet metablasts.

Principal minerals: Quartz, microcline, albite (An 5 %), oligoclase (An 15–20 %), biotite, garnet.

Accessory minerals: Sillimanite, cordierite, muscovite, diopside, ore and zircon, rutile, allanite as inclusions in biotite and quartz.

Alteration minerals: Chlorite, epidote, sericite, green variety of biotite, muscovite.

The quartz occurs in xenoblastic aggregates; together with plagioclase it forms in many places graphic intergrowths. Quartz is locally found as plate-like grains forming fine-grained and almost monomineralic zones, which are probably due to movements causing both a comminution of the original coarser grain and a metamorphic differentiation.

The microcline is generally interstitial; it appears to have partially replaced most of the other minerals. Microperthite is in most cases developed within the microcline.

Late albite is often crystallised as large grains of uniform extinction, enclosing and partially replacing the garnets. Sillimanite is sparsely present in isolated aggregates. It developed mainly from biotite and is in many cases enclosed by late albite (fig. 8).

Biotite occurs in large quantities in the more schistose zones. Two generations can be recognized: a first generation showing deformed crystals being locally pushed aside by growing garnets; a second generation of undeformed crystals, partially replacing the garnets.

The gneissic garnet schist formation contains many interstratified amphibolite layers. These layers are essentially composed of nematoblastic hornblende aggregates. Locally biotite or diopside has developed from the hornblende (fig. 9). Oligoclase and quartz form a minor part of the amphibolite; they often corrode the hornblende crystals.

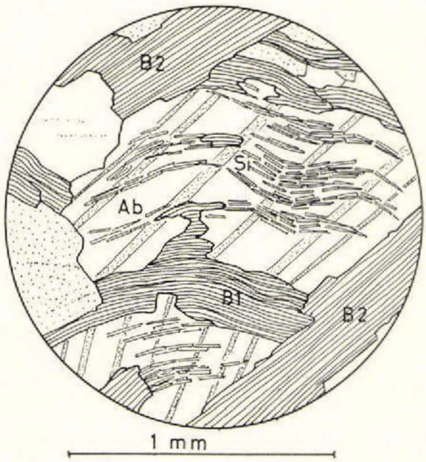


Fig. 8. Aggregates of sillimanite needles (Si) enclosed in late albite (Ab). The sillimanite needles themselves are not bent, but are arranged along a pre-existing curvature in the biotite flake which they partially replaced. Two biotite generations can be seen (B 1 and B 2). Slide No. 51625.

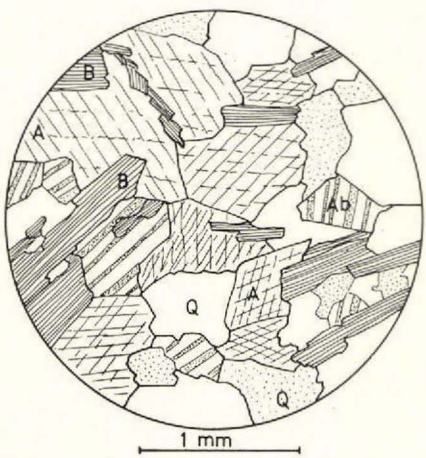


Fig. 9. Amphibolite. The amphibole (A) is corroded by quartz (Q) and albite-(Ab). Biotite (B) is developed interstitially between two amphibole crystals. Slide No. 63647.

The hornblende is pleochroic with X greenish yellow, Y deep green and Z deep bluish green; $2Vx = 75°$; $Z \wedge c = 22°$.

Garnet poikiloblasts, containing inclusions of hornblende and quartz, are not uncommon in amphibolite layers.

The semi-pelite unit (Unit II).

a) The black biotite schist: (Forms the base of unit II).

Structure: Schistose, fine- to medium-grained.
Texture: Predominantly lepidoblastic.
Principal minerals: Biotite, oligoclase (An $15\,^0/_0$), quartz.
Accessory minerals: Microcline, muscovite, zircon, allanite, ore, graphite.
Alteration minerals: Chlorite, green biotite.

Biotite is present in large quantities. Two generations can be recognized:
1) Plastically deformed flakes, often corroded by quartz.
2) Undeformed biotite cutting the first generation.

The biotites of the second generation are locally also deformed by shearing or by small-scale strain-slip cleavage. Quartz and oligoclase form small granoblastic aggregates. These aggregates occur as strings or lenses enclosed by lepidoblastic biotite masses. Oligoclase forms here and there bypautomorphous crystals, which often corrode and replace the biotite.

b) The quartz-biotite schist: (the middle and upper part of unit II).

Structure: Near Nanortalik harbour the structure is generally gneissic with intercalations of schistose zones; augen structures are common here. More to the NE the structure is essentially schistose. The grainsize is mostly fine.
Texture: Grano-lepidoblastic, locally porphyroblastic.
Principal minerals: Quartz, albite (An $7\,^0/_0$), oligoclase (An $20\,^0/_0$), biotite, microcline.
Accessory minerals: Muscovite, garnet, rutile, apatite, zircon, ore.
Alteration minerals: Chlorite, sericite, actinolite, saussurite, ore.

The feldspars (oligoclase and microcline) are locally developed as poikiloblastic rounded crystals. They seem to have grown during a period of deformation, preferentially along shear zones due to fracture cleavage.

Two generations of biotite can be recognized, the earlier one shows the effect of a strong plastic deformation, while the second-generation biotites are only locally deformed in shear zones. These zones affect also many other minerals. First-generation biotite is often partially replaced by oligoclase and quartz.

Quartzitic layers, composed essentially of quartz, plagioclase and minor amounts of biotite, are frequently found interbedded in the schists.

The pelitic schist unit (Unit III).

a) The biotite schist: (lower and middle part of unit III).

Structure: Schistose, fine-grained.
Texture: predominantly lepidoblastic.
Principal minerals: Biotite, oligoclase (An 15–20 $^0/_0$), albite (An 6 $^0/_0$), quartz.
Accessory minerals: Microcline, muscovite, rutile, allanite, tourmaline, apatite, graphite, ore.
Alteration minerals: Chlorite, green biotite, actinolite, zoisite, sericite, goethite.

Biotite occurs in two distinct generations. It is generally concentrated in lepidoblastic aggregates forming elongate zones probably concordant with the original bedding. These mafic zones are separated by thin beds composed essentially of a quartz-oligoclase assemblage. The quartz crystals contain many liquid inclusions, concentrated generally in the central part of the grains.

Albite replaces biotite and oligoclase and is itself replaced by late microcline. Oligoclase and microcline form locally diablastic intergrowths.

Many graphite- and pyrite-rich layers are found in the biotite schists. The pyrite is for the major part decomposed into goethite and other limonite-minerals. These alteration minerals account for the red-brown colour of the pyritic schist layers.

b) The biotite-garnet schist: (the upper part of unit III).

Structure: Schistose, generally fine-grained.
Texture: lepidoblastic, locally granoblastic or porphyroblastic, with garnet porphyroblasts.
Principal minerals: Biotite, oligoclase (15–20 $^0/_0$ An), albite (An 5 $^0/_0$), quartz, garnet, andesine (An 35 $^0/_0$).
Accessory minerals: Muscovite, microcline, epidote (clinozoisite), diopside, hornblende, zircon, rutile, graphite, ore, sphene.
Alteration minerals: Sericite, saussurite, epidote, chlorite, green biotite.

Albite developed interstitially from oligoclase and biotite. It is locally accompanied by minor amounts of epidote. Garnet occurs mostly as poikiloblasts, containing inclusions of quartz and biotite. It is often partially replaced by a fine-grained aggregate of granular plagioclase. Andesine xenoblasts have in some places partly replaced oligoclase. Quartz and oligoclase form interlobate granoblastic aggregates enclosed by biotite-rich zones. Diopside has rarely been observed forming diablastic aggregates with quartz.

The quartzite unit (Unit IV).

a) The quartz schist- (forming the whole of unit IV in the SW part of the area).

Structure: Massive granular zones alternating with more schistose layers. Fine- to very fine-grained.

Texture: Grano-lepidoblastic, generally isogranular.

Principal minerals: Quartz (35 to 50 % of the rock), oligoclase (10–20 % An), biotite.

Accessory minerals: Microcline, albite, muscovite, rutile, zircon, allanite, ore.

Alteration minerals: Chlorite, sericite.

Quartz, oligoclase and microcline form interlobate granoblastic aggregates in elongate zones and layers of variable thickness (2 mm to 3 cm) These acid zones alternate with more basic ones in which biotite is a major constituent. The biotite-rich layers are usually very thin. Near the top of the unit, diopside, epidote and hornblende appear together with the biotite in the basic zones.

b) The quartzite: (forming the whole of unit IV in the NE part of the area).

Structure: Massive, generally banded with darker horizons; fine- to very fine-grained.

Texture: Granoblastic isogranular.

Principal minerals: Quartz (50 to 70 % of the rock), biotite, oligoclase-andesine (An 15–35 %), microcline.

Accessory minerals: Muscovite, ore (predominantly hematite), allanite, albite, rutile, zircon, tourmaline. Moreover, at the top of the unit, the quartzite contains diopside, epidote, garnet, hornblende and sphene, while at the base sillimanite, cordierite and hypersthene have been observed.

Alteration minerals: Chlorite, epidote, sericite, saussurite.

Quartz forms with plagioclase and microcline xenoblastic interlobate masses in which biotite grew as isolated grains with random orientation. Mafic bands occur in the quartzites and are essentially composed of biotite with minor amounts of quartz and feldspar. Near the top of the unit the mafic bands contain diopside, hornblende and epidote. Diopside has here also been observed as isolated crystals in a matrix of finely crystallised quartz and feldspar. It appears to be pseudomorphous after chlorite and to follow an arrangement outlining the probable original borders of the clastic grains (fig. 10).

In some places near basic sills, the quartzite is impregnated with interstitial hornblende. Hematite-rich zones have frequently been observed, in which the hematite can amount to 20 % of the rock.

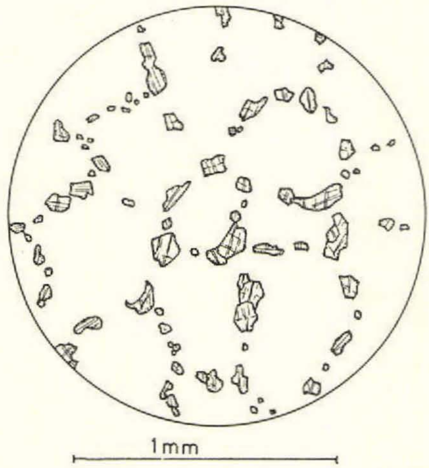


Fig. 10. Diopside crystal arrangement outlining the probable original borders of the clastic grains. The matrix is formed of finely crystallised quartz and feldspar. Quartzite near Tasermiut fjord. Slide No. 63607.

The base of the quartzite unit is characterised in the NE part of the area by the presence of sillimanite, cordierite and small amounts of hypersthene. The rock is here very fine-grained and could well be confused with an homogeneous gneiss.

The volcanic unit (Unit V forming the upper part of the supracrustal rocks of the area).

It has been demonstrated in chapter II that the volcanic unit consists of a great variety of rocks: pillow lavas, tuffites, lapilli, agglomerates, green schists, quartzites, quartzitic slabs, basic and ultrabasic sills, and mica shists. As the petrographic investigation is not extensive enough to give a detailed description of each of these rock species the descriptions will be limited to the three principal rock-types; the pillow lavas, the green schists and the meta-gabbro sills.

a) The pillow lava: (forming layers throughout the whole unit).

Structure: Real pillow structures can be observed in two localities in the investigated area: north of lake 410 m and on the western flank of Quagdlua mountain. Otherwise the pillows are always quite flattened and difficult to identify. Usually the borders of the pillows are dark green (amphibole), while the central part is more yellowish (epidote). This zoning is characteristic even in the places where the pillows are completely crushed, forming thin discontinuous layers of alternately amphibole-rich and epidote-rich beds.

Texture: Nematoblastic to grano-nematoblastic in the crushed and flattened parts, and heteroblastic in the more undisturbed parts.

Principal minerals: In the border zones of the pillows: amphibole (common green hornblende and actinolite), biotite, albite (An 5 %), quartz. In the central parts: epidote (clinozoisite and pistacite), zoisite, quartz, amphibole.

Accessory minerals: Oligoclase (An 25 %), calcite, muscovite, chlorite, sphene, apatite, zircon, microcline and ore.

Alteration minerals: Sericite, saussurite, chlorite, vermiculite.

Most common is dark green hornblende, forming together with albite-oligoclase, biotite and quartz nematoblastic or heteroblastic aggregates in the usually wide border zones of the pillows. Albite appears to expand in most cases by replacement of hornblende and biotite (fig. 11).

Fig. 11. Border zone of a pillow-lava near lake 410 m. Amphibole (A) is partly replaced by albite (Ab). Biotite (B) is developed in cracks and fissures in the amphibole. Epidote (e) is present in small isolated grains surrounded by quartz. Slide No. 51663.

It sometimes forms large poikiloblasts containing many remnants of amphibole, biotite and oligoclase. The central parts of the pillows are predominantly formed of heteroblastic associations of epidote, zoisite and hornblende with granular quartz or calcite (fig. 12). The crystal habit of epidote and zoisite suggests in some places that they filled original vesicles.

Moreover, in other places, cavities between pillows are filled with carbonates and quartz. However, in most cases the pillows are compressed and flattened in such a way that all original cavities or infilled materials have disappeared.

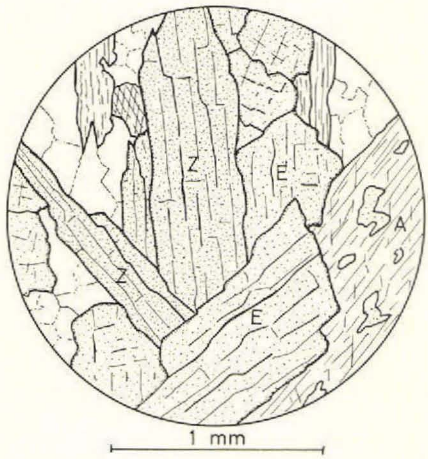


Fig. 12. Central part of a pillow in the pillow-lavas near lake 410 m. Epidote (E) and zoisite (Z) form aggregates of large crystals, possibly by infilling of original vesicles. Slide No. 51662.

b) The basic schists: (Green schists, interbedded in the volcanic rocks throughout the unit).

Structure: Schistose, medium- to fine-grained.

Texture: Predominatly lepidoblastic.

Principal minerals: Actinolite, albite-oligoclase (An 5–15%), chlorite, epidote, quartz, biotite. In the lowest part of unit V the green schists contain, moreover, appreciable amounts of common green hornblende, whilst in the upper part muscovite is an important constituent.

Accessory minerals: Sphene, apatite, zircon, tourmaline, allanite and ore.

Alteration minerals: Sericite, saussurite, chlorite, vermiculite, leucoxene.

Biotite presents here and there a strong pleochroism with X clear yellow and may therefore belong to the stilpnomelane variety.

Late albite replaces most of the other principal minerals. Allanite occurs within epidote crystals as small orange grains. Sphene forms often kelyphitic associations around ore grains.

Ultrabasic schists are sparsely interbedded in the green schists, particularly in the basal part of unit V. These schists are principally composed of talc, actinolite, chlorite and minor amounts of hornblende.

c) The gabbro: (Forming many sills, particularly in the lower part of the volcanic unit).

Structure: Massive, often fine-grained in the borders and coarser-grained in the centre. Some sills are entirely fine- to very fine-grained and may originally have been micro-gabbros.

Texture: Granoblastic to grano-nematoblastic. Nowhere can residual igneous textures be observed. In many places cataclastic textures due to intensive shearing can be seen.

Principal minerals: Amphibole (common green hornblende and actinolite), albite (An 5 %), biotite (particularly in the border zones of the sills).

Accessory minerals: Quartz, chlorite, epidote, microcline, sphene, apatite, ore (magnetite), oligoclase (An 25 %).

Alteration minerals: Chlorite, saussurite, sericite.

The border zones of the sills are generally characterised by the presence of large quantities of biotite forming elongate lepidoblastic aggregates parallel to the contacts. Intensive albitisation has in many places been observed. Close to pegmatites microcline sometimes replaces the "late" albite.

V. PETROGENESIS AND METAMORPHISM OF THE KETILIDIAN SUPRACRUSTALS

The metamorphic facies of the supracrustal rocks

In the preceding chapter it has been shown how the mineralogical composition varies from one unit to the other. It will now be established to which metamorphic facies the different supracrustal rocks belong. The classification used here is based on the subdivision of metamorphic rocks into characteristic mineral assemblages by ESKOLA (1939) and, with modifications and expansions, by TURNER (1948).

The highest degree of metamorphism is found in the basal gneisses of unit I in the coastal area E of Tusardluarnâq near Tasermiut fjord. The characteristic mineral assemblage, sillimanite-garnet-hypersthene, indicates that these rocks probably belong to the *granulite facies*. This is confirmed by the presence of mesoperthitic oligoclase and strongly microperthitic potash feldspar. However, the fact that biotite is still present and that hypersthene is only present in minor quantities indicates that the basal gneisses only belong to the lowest granulite facies. Another place, where probably low-granulite facies conditions were reached, is situated at the base of the quartzites in the NE part of the area. There, besides sillimanite and cordierite, small quantities of hypersthene were also found.

Most of the quartzites, gneissic schists and schists of the Nanortalik area belong however to the *amphibolite facies*. This is proved by the typical mineral assemblages, hornblende-biotite-plagioclase and biotite-plagioclase-microcline-quartz, which characterise the rocks of the semi-pelite, pelitic schist and quartzite units (units II, III and IV). The gneisses and gneissic schists of the upper pelitic gneiss unit (unit I), overlying the granulite-facies gneisses, belong also for the major part to the amphibolite facies of metamorphism. The presence here of accessory sillimanite and cordierite together with garnet shows, moreover, that these gneissic rocks reached the highest degree of amphibolite-facies metamorphism and that they belong to the *sillimanite-almandine subfacies*.

The lower limit of the amphibolite-facies metamorphism, where oligoclase becomes unstable and gives way to the assemblage albite-epidote, was reached within the upper quartzitic schists in the central part of the Nanortalik peninsula. This is clearly shown by the appearance in the upper part of these quartzitic schists of epidote together with hornblende and albite and by the disappearance of oligoclase. Therefore it can be said that the *albite-epidote-amphibolite facies* of metamorphism starts in the upper levels of the quartzitic schists of unit IV. The overlying volcanic rocks and green schists also reached to a small degree the albite-epidote-amphibolite facies of metamorphism. This is the case in particular with the lower green schists, in which the typical mineral assemblage, hornblende-epidote-albite–(biotite-quartz), can be observed.

The major part of the volcanic unit in the area under discussion, however, reached only the *greenschist facies* of metamorphism. This is shown by the assemblage, actinolite-chlorite-epidote-albite-quartz-biotite, which characterises most of the green schists, pillow-lavas and gabbro sills. In the upper part of the volcanic unit, and particularly in

Table 2. The principal supracrustal rock types and the corresponding metamorphic facies in the central and SW part of the Nanortalik area. (Classification according to Eskola modified and completed by Turner).

Unit	Rock type	Typical minerals	Metamorphic facies
V	Upper green schist	Muscovite-albite-chlorite epidote-quartz	Muscovite-chlorite subfacies
	Middle green schist, gabbro and pillow-lavas	Actinolite-chlorite-epidote-albite-quartz-biotite.	GREENSCHIST FACIES Biotite-chlorite subfacies
	Lower greenschist	Hornblende-epidote-albite-biotite-quartz	ALBITE-EPIDOTE-AMPHIBOLITE FACIES
IV	Upper quartzitic schist		
	Middle and lower quartzitic schist	Quartz-plagioclase-biotite-microcline	AMPHIBOLITE FACIES
III	Biotite-garnet schist	Biotite-plagioclase-quartz-garnet	
	Biotite schist		
II	Quartz-biotite schist	Biotite-hornblende-plagioclase-quartz	
	Black biotite schist		
I	Gneissic garnet schist	Sillimanite-cordierite-almandine-plagioclase	Sillimanite-almandine subfacies
	Garnet-sillimanite gneiss	Sillimanite-cordierite-hypersthene (-quartz-microcline-plagioclase-biotite)	LOW GRANULITE FACIES

the upper green schists, biotite has a tendency to disappear and to make way for the muscovite. This indicates that these schists belong to the lower greenschist facies (the *muscovite-chlorite subfacies*).

The central part of the gabbro sills and usually also the central zones of the groups of pillows seem, on the whole, to possess a slightly higher degree of metamorphism than the surrounding greenschists. For instance, the pillows often contain, beside the typical greenschist minerals, a basic oligoclase (An 25 %). This mineral is theoretically not stable in the greenschist facies (Turner and Verhoogen 1960).

In table 2 a summary is given of the different metamorphic facies and the corresponding rock-types which can be found in the Nanortalik peninsula.

Petrogenetic considerations; relations between deformation and metamorphism

It is not in the scope of the present investigation to give a detailed account of all the different transformation-phases of the rocks. However, the petrographic observations mentioned in the preceding chapter allow a few petrogenetic conclusions to be drawn.

Recapitulating, the principal observed facts are as follows:

1) In most pelitic rocks of the Nanortalik area at least two generations of biotite can be distinguished. The first of these generations usually shows micro-plications and appears to be the only important mineral which still presents the effects of a strong plastic deformation. This early biotite is partially replaced by most of the other minerals.
2) In the basal gneisses sillimanite seems to be undeformed in most cases. The apparent curvature of the sillimanite aggregates is generally only due to the fact that they replaced curved biotites. Each sillimanite needle taken separately is quite straight (fig. 8).
3) Garnet occurs mainly as poikiloblasts containing inclusions of quartz, oligoclase and early biotite. Late biotite often partially replaces the garnet. When in contact with sillimanite, garnet usually shows that it has grown by pushing aside the sillimanite bundles.
4) Hornblende, when present, is often corroded by late biotite, plagioclase, quartz and garnet.
5) Albite is in most cases developed as large interstitial crystals replacing all the other minerals except late microcline and quartz.
6) Microcline appears in two generations: A first generation of microcline probably formed together with quartz and oligoclase in the lower rock-units. This early microcline was formed locally as augen in shear-zones due to fracture-cleavage. A second generation of

microcline was developed, generally together with quartz, as a partial replacement of most of the other minerals, including late albite.

7) With the exception of the first-generation biotite, the rock-forming minerals appear in most cases to be undeformed. Locally however, small-scale fracture-cleavage and shearing can be observed which affects most minerals except late microcline, albite and quartz.

From these facts the following conclusions can be drawn:

a) The early biotite has been formed before the other actually observable minerals. It seems to be a remnant of a first synkinematic phase of metamorphism which probably took place during the second and principal Ketilidian folding period. This may be deduced from the fact that the early biotite is mostly deformed in a plastic manner and that the second folding period was characterised by a plastic deformation.

b) The first phase of metamorphism was followed by a second in which quartz, oligoclase, hornblende, biotite and in the basal formations sillimanite and cordierite were formed. All these minerals are deformed by small-scale shearing and fracture-cleavage which probably took place during the third folding period. Therefore it can be assumed that this second phase of metamorphism was effective after the second and probably at the beginning of the third deformation phase. As mentioned earlier in this paper, the third Ketilidian deformation period acted probably upon a less plastic and more rigid mass. This could explain the fact that the minerals formed during the second phase of metamorphism are generally undeformed except by shearing or fracture-cleavage.

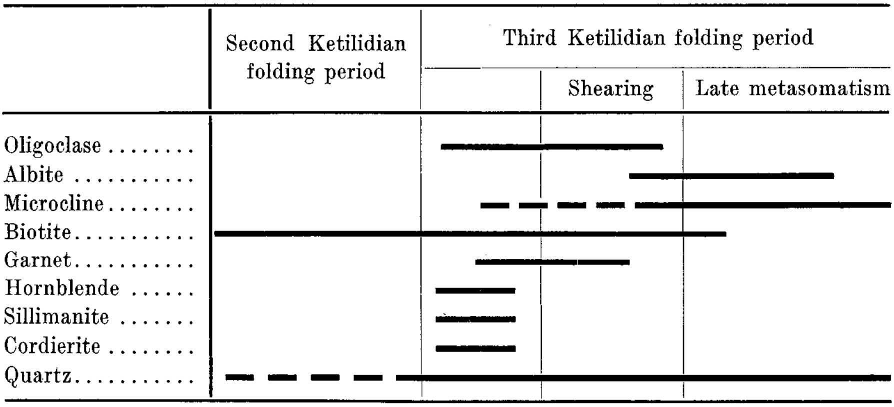
c) A third phase of metamorphism accompanied by metasomatism and addition of Si, Na and K resulted in the crystallisation of quartz, oligoclase, microcline, garnet and biotite. The fact that microcline and oligoclase developed as augen in shear zones, formed during the third Ketilidian folding period, indicates that the third phase of metamorphism was connected with the third folding period.

d) Considerable albitisation occurred at the end of the third folding period and after the third phase of metamorphism.

e) The albitisation was accompanied and followed by a large-scale microclinisation. It seems quite probable that the microclinisation continued even after the Ketilidian period.

All these conclusions concern only the most important minerals which were found in the middle and lower units. The introduction of microcline even after the albitisation is in accordance with the views of several authors (DRESCHER-KADEN 1948).

Table 3. Hypothetical representation of the different times at which the principal minerals of the pelitic schist unit were formed.



It seems likely that the first and second phases of metamorphism were progressive and acted under the influence of increasing temperature, as shown by the formation of sillimanite from biotite. On the other hand, the third phase of metamorphism appears to have been active during a decrease in temperature (obliteration of sillimanite and crystallisation of plagioclase and garnet). This third phase must thus have been slightly retrogressive.

Table 3 gives a hypothetical representation of the different periods at which the principal minerals of the pelitic gneiss unit were formed.

Retrograde metamorphism acted locally in fractures and shear zones and resulted in the chloritisation of biotite and amphibole, in the formation of epidote, and in the saussuritisation and sericitisation of feldspars. These alterations may be partly recent and may have been caused partly by the hydrating influence of metasomatism and pegmatitisation during late-Ketilidian times. Similar conclusions were reached by Oen Ing Soen (1962) in the Ivigtut district.

The present spilitic composition of the pillow-lavas might be a result of automorphism of the original basaltic lavas, followed by a further adaptation of the mineral assemblage to the greenschist facies of metamorphism (Turner and Verhoogen 1960).

The fact that the central zones of the groups of pillows and gabbro sills often possess a higher degree of metamorphism than the surrounding green schists can be explained by assuming that these zones were more isolated and thus kept longer their original composition.

VI. THE KETILIDIAN MIGMATITES AND ASSOCIATED PEGMATITES

Generalities, definitions

The term migmatite is used here according to the definition of TURNER and VERHOOGEN (1960): "A mixed rock in which a granitic component (granite, aplite, pegmatite, or the like) and a metamorphic host rock are intimately admixed on a scale sufficiently coarse for the mixed condition of the rock to be megascopally recognizable". In the area under examination most of the schists correspond to this definition and should be called migmatites. The underlying gneisses and the overlying quartzites and volcanic rocks show only local, minor migmatitic features. This migmatisation seems here to have been produced essentially by a metasomatic enrichment in quartzo-feldspathic material.

In the field the migmatitic rocks of the Nanortalik area are characterised by the presence of a multitude of small and large veins composed predominantly of quartz and feldspar. These veins correspond to the definition of *pegmatites* as used by RAMBERG (1956) in other, more northern regions. This definition comprises all kinds of essentially quartzo-feldspathic vein-, dyke-, or sill-shaped bodies which are coarser-grained and apparently younger than their host rocks, even if their structure is fine-grained and aplitic.

In the Nanortalik area the pegmatitisation is the most striking effect of the migmatisation. Therefore it is clear that by following the genesis and development of the former a fair idea about the evolution of the migmatisation process can be obtained. This is the reason why the Ketilidian pegmatites have been specially investigated in the Nanortalik area. Four different generation-groups of Ketilidian pegmatites have been distinguished, each successive generation cutting and partly replacing the former.

The first group consists of generally very thin (0.1–2.0 cm) and continuous veins, which are mostly parallel to the schistosity and which might be a result of pure gneissification due to early regional metamor-

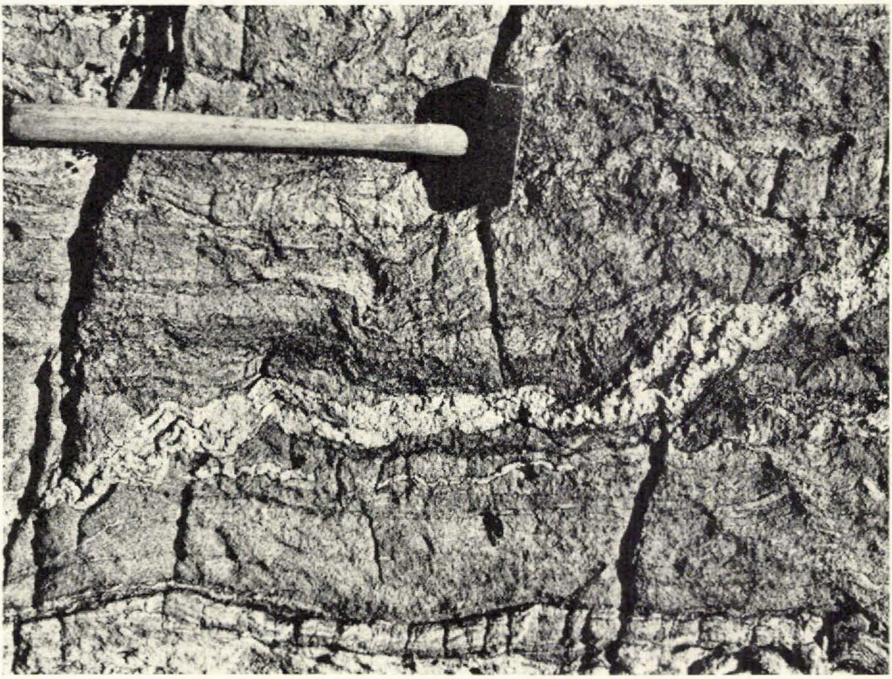


Fig. 13. Second-generation pegmatite in biotite schist near lake 430 m. The subconcordant vein cuts and partially replaces the thin first-phase pegmatites.

phism. Their structure is fine-grained aplitic, but the grain-size is usually coarser than that of the host rock.

The second group appears to be more important than the first. The pegmatites cut the earlier ones and commonly replace them along the schistosity planes (fig. 13).

The second-generation veins are generally discontinuous and can be concordant, subconcordant or discordant with respect to the foliation or bedding. Their thickness varies mostly between 0.5 and 10 cm and their length may reach several meters, but usually does not exceed 50 cm.

They commonly impregnate completely the schists and can amount to 60 vol.% of the rock. The veins are generally folded and locally present ptygmatic structures. The minerals are fine- to medium-grained.

The third group comprises medium- to coarse-grained, relatively thick (2–50 cm) pegmatites which seem to have formed preferentially along joints and fracture planes. Their length varies from 50 cm to 10 m. They are generally discordant and in many places show fold structures. When cutting earlier pegmatites, the third-generation veins often form cloudy aggregates of quartzo-feldspathic material, replacing the older pegmatites and parts of the host rock (fig. 21).

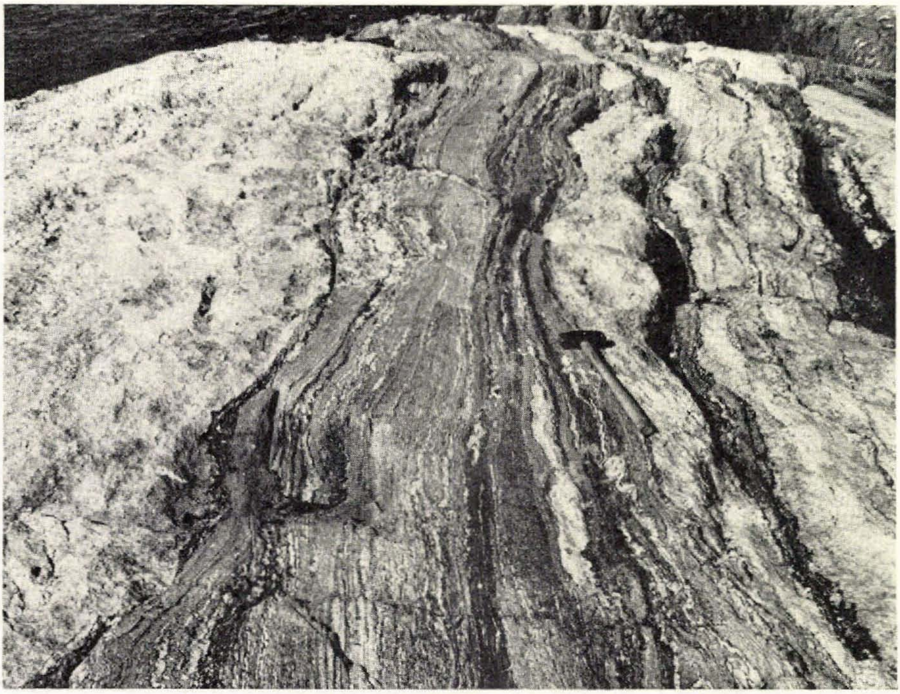


Fig. 14. Fourth-generation, large pegmatites concordant with the foliation. Gneissic schist formation on a small island S of Nanortalik harbour.

The fourth group is represented by very thick pegmatites (50 cm to 60 m). These in places assume the shape of huge lenses or undulating sills. They are generally concordant or subconcordant with the schistosity and can often be traced for several hundred meters. Characteristic is the presence of pinch-and-swell structures in which the pegmatitic material has partly pushed aside and has partly replaced the surrounding rock. Deformation seems to have acted only to a minor degree upon the Ketilidian fourth generation pegmatites. Frequently large pegmatites of this generation are found along concordant planes separating zones of very schistose rocks from more compact ones. The composing minerals are generally coarse- to very coarse-grained. However, medium- to fine-grained zones can in many places be observed within or on the borders of the pegmatite body.

The large fourth-group pegmatites are most conspicuous throughout the whole peninsula in the schists and gneissic schists (fig. 14). They were mentioned by WEGMANN (1938) in the Nanortalik island area.

Similar pegmatites with typical pinch-and-swell structures have been observed in the area NW of Ivigtut by OEN ING SOEN (1962).

Composition and grain size of the Ketilidian pegmatites

All the Ketilidian pegmatites are essentially composed of quartz and feldspar with minor amounts of biotite, muscovite, garnet and epidote. Locally, very small amounts of diopside, hornblende, chlorite, sillimanite, cordierite, sphene, apatite, graphite or zircon can be found. Moreover, ore minerals are present as accessories in most pegmatites. Tourmaline and beryl are only a minor constituent of the large fourth-generation pegmatites.

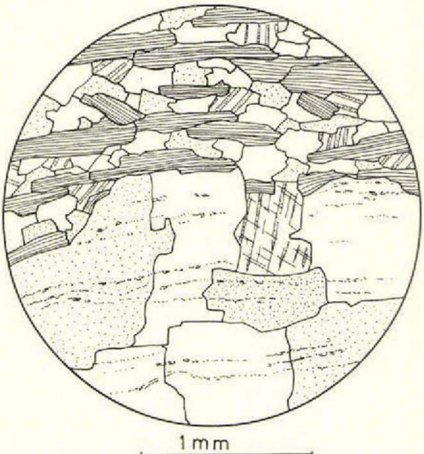


Fig. 15. Strings of inclusions in quartz in a small pegmatite vein near lake 430 m. The direction of the strings is clearly parallel to the elongation of the biotites in the schist. Slide No. 63631.

Quartz generally occurs as granoblastic aggregates and rarely as porphyroblasts. It often displays a strong undulatory extinction and it usually constitutes 30 to 60 vol. % of the pegmatite rock. Graphic intergrowth of quartz and potash feldspar occurs locally. With acidic plagioclase, quartz commonly forms irregular macromyrmekitic intergrowths. Many inclusions are found within the quartz crystals. They are generally very small and form strings parallel to the schistosity of the host rock. The direction of the strings is independent of the crystal structure and often cuts differently oriented quartz grains (fig. 15). The material of the inclusions is mostly formed of biotite, graphite and chlorite fragments.

The feldspar generally constitutes 40 to 70 vol. % of the pegmatite rock. In the first- and second-generation veins the feldspar is predominantly represented by an acidic plagioclase (oligoclase 18 % An or albite-oligoclase 5 % An). In the third- and fourth-generation pegmatites the most common feldspar is microcline. Orthoclase is only a minor constituent in all the pegmatites. The presence of microcline is characteristic of the very coarse-grained zones in the latest pegmatites. Antiperthitic plagioclase and mesoperthitic alkali feldspar only occur in pegmatites of the small gneiss area E of Tusardluarnâq. In all the other regions they appear to be absent. The texture of the feldspar grains is generally xenoblastic. The grains are moreover often strongly poikiloblastic and contain numerous relics of earlier formed minerals. This is particularly the case with the large microcline crystals which generally contain many relics of replaced plagioclases.

Biotite and muscovite occur as lepidoblastic aggregates and streaks in the pegmatites. They mostly seem to be relics of mica schists. The direction of the streaks and strings is almost always parallel with the schistosity of the surrounding rock. In some places in the fourth-generation pegmatites, aggregates of very large recrystallised biotites and muscovites are found. Biotite is generally more abundant than muscovite, but never amounts to more than 10 vol. $^0/_0$ of the pegmatite material.

The presence of the accessory minerals mainly depends on the composition of the host rock:

In garnet-rich schists and gneissic schists the pegmatites contain appreciable amounts of garnet crystals. This is particularly the case in the garnet-rich schist horizons of the area around lake 430 m. In these horizons the various generations of pegmatites also appear to contain different generations of garnets. The garnet crystals are generally recrystallised within the pegmatite as poikilitic porphyroblasts. Pushed-aside borders of small residual biotite flakes are commonly seen around the large garnets. The youngest garnets are always larger than the older ones. All the examined garnet crystals in the pegmatites probably belong to the manganese-rich variety, as shown by their low refractive index (1.74) and by the absence of birefringence.

Hornblende (common green variety), epidote (zoisite) and diopside occur in the pegmatite veins which transect the amphibole- and epidote-rich volcanic rocks.

Graphite is common in pegmatites cutting the graphite bearing schists, while sillimanite and cordierite have been observed only in the gneiss- and gneissic schist-pegmatites.

Allanite is sparsely found as inclusions in epidote or in biotite, showing a pleochroic halo in the latter case.

Sphene forms pink grains showing weak pleochroism and is often altered to a mass of leucoxene surrounding grains of ilmenite.

The grain size of the pegmatites is larger, as mentioned before, in the younger veins than in the older ones. The first- and second-generation veins are generally aplitic with grains varying in size between 0.6 and 1 mm. They are nevertheless coarser grained than the host rock. The third- and fourth-generation veins are generally coarse- to very coarse-grained with crystals varying in size from 0.5 to 5 cm. Exceptionally, perthitic microcline crystals with a diameter up to 40 cm were observed within the very large late-Ketilidian pegmatites.

Regular zoning, so common in later post-Ketilidian veins, does not seem to exist in the older pegmatites. However, an irregular zoning due to a difference in grain-size is common within the third- and fourth-generation veins, where irregular elongate aplitic zones can be observed forming the centre or the borders of the pegmatite. In every case observed, the large crystals appear to have grown later and partly at the expense of the small ones. Moreover, microscopic examination shows that the aplitic parts are essentially composed of quartz and albitic plagioclase (5–8 $^0/_0$ An), whilst the coarse-grained zones are characterised by the presence, besides quartz, of large quantities of microcline.

Other irregular zoning, due to differences in composition, is rare among the Ketilidian pegmatites. In the few known cases it is due to selective concentrations in marginal or central zones.

Composition of the Ketilidian migmatites

Besides the pegmatite veins, there are several other phenomena which characterise the Ketilidian migmatites. The most important are:

1) Cloudy aggregates of granitic composition replacing the schists.
2) Impregnation of the rock by numerous isolated feldspars (mostly acidic plagioclase) leading to a partial albitisation of the rock.
3) Concentration of calcic and ferromagnesian material within the schist in zones bordering the pegmatites.
4) Formation of calcite veins, probably due to expulsion of Ca from the replaced rock.
5) Formation of pure quartz veins.

However, all these characteristic mineral associations only form a very small part of the migmatite volume in comparison with the amount of pegmatite veins.

In three different places in the migmatitic schist area around lake 430 m, samples were taken of schist and pegmatite. In addition, in each locality the volume-proportion of the pegmatite to the schist was evaluated with help of a network superposed on the rock or on close-up photographs. Each rock sample was examined under the microscope and

Table 4. Estimated mineral compositions of three migmatite samples taken in the area near lake 430 m.

No. sample	63609 A, B Vol. %	63611 A, B Vol. %	63641 A, B Vol. %
Quartz	25	32	27
Microcline	12	15	18
Plagioclase (An 5–10 %)	28	20	16
Biotite	18	18	22
Muscovite and Chlorite	6	5	8
Garnet	7	6	4
Access. minerals	4	4	5

the proportions of the various predominant minerals were estimated by point-counting. The final result, giving the approximate composition of the complete migmatite (schist and pegmatite), is given in table 4, which shows that the migmatites in the vicinity of lake 430 m are relatively rich in quartz and that their composition is granodioritic for the first two samples and granitic for the last.

Localisation and genesis of the Ketilidian pegmatites

Most Ketilidian pegmatites are found in the schists and gneissic schists, while in the more compact underlying gneisses and overlying quartzites and volcanics these veins are unusual. Most of the tension-joint pegmatites in the volcanic series have been proved to be post-Ketilidian.

The localisation of Ketilidian pegmatites seems principally controlled by pre-existing planes such as fractures, faults, schistosity or bedding. These surfaces probably acting as "privileged paths", as defined by READ (1957), directed the migration and concentration of quartzo-feldspathic material.

Some Ketilidian pegmatites are localised in tension joints, developed in relatively compact rocks intercalated with more schistose layers. For instance this can be observed in some amphibolitic layers surrounded by gneissic schists SW of Tusardluarnâq. These basic layers contain in some places large masses of quartzo-feldspatic material, which replaces the amphibolite over considerable distances.

RAMBERG (1956) observed similar phenomena further to the north and explained them by a replacement of the ferromagnesian minerals by quartz and feldspar in local spots of maximum tensile stress within the originally homogeneous and continuous basic rock layers. The gneissic-schist and schist formations commonly contain such pegmatitic concentration formed in pressure shadows of more rigid layers.

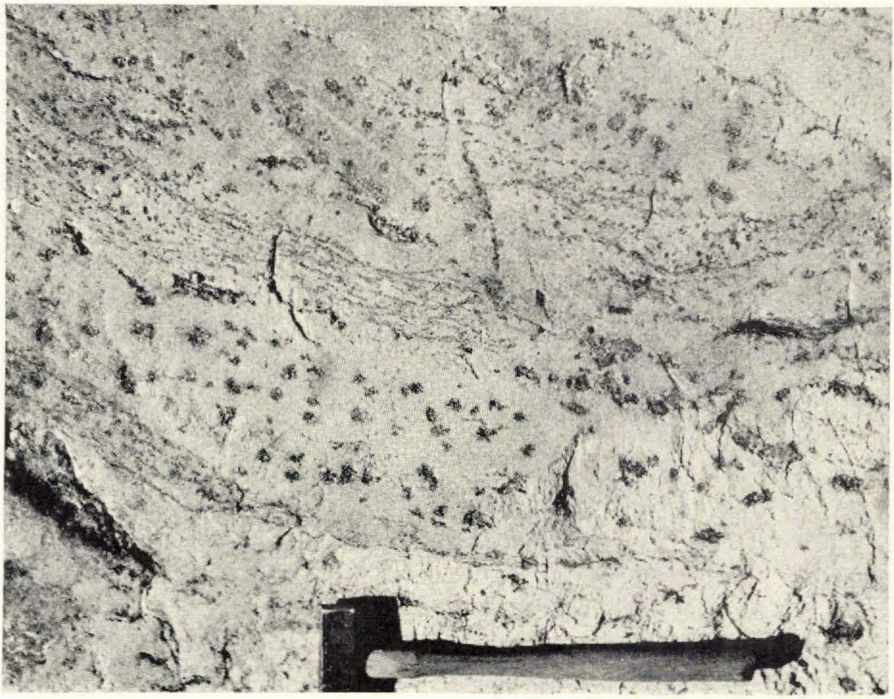


Fig. 16. Nebulitic structures within a large pegmatite on the S coast of Sarqâ fjord. The structures trend in the same direction as that of the enclosing schists.

It is clear that the emplacement of the pegmatites cannot be explained by one single mechanism and that it must be the result of several combined processes. However, field and laboratory observations show that all the Ketilidian pegmatites have been formed principally by *metasomatic replacement* and not by injection and crystallisation of a melt. This is demonstrated by following observations:

1) The pegmatites often contain practically undisturbed remnants of the partially replaced rock, forming nebulitic structures (fig. 16). The foliation observed in the inclusions is always concordant with the foliation of the host rock. This proves that the relics of the partially replaced rock have not been rotated or displaced. On a microscopic scale, similar phenomena can be observed particularly in quartz grains, which often contain inclusion-strings of the same orientation as the schistosity of the host-rock (fig. 15).

2) Often the geometry of a cross-cutting pegmatite and that of the enclosing rock shows that there has been no dilation of the original diaclase. This is clearly shown by the continuation in the same plane of distinctive layers on either side of the cross-cutting pegmatite (fig. 17).

Fig. 17. Example of nondilation replacement pegmatite. Typical layers in the schist continue in the same plane on both sides of the vein. Third-generation pegmatite near lake 260 m. Width of the vein 20 cm.

3) In some places it can be seen that Ketilidian pegmatites expanded by growth of more or less isolated quartz and feldspar porphyroblasts within the country rock. This is particularly clear around the large third- and fourth-generation pegmatites.

4) The fact, as mentioned in the preceding pages, that the composition of the pegmatites depends greatly on the composition of the host rock also indicates an probable origin by replacement.

5) The mineral associations in the pegmatite veins indicate that they belong in most cases to the amphibolite facies of metamorphism. This means that the pegmatites were formed at a temperature below 600° and consequently below magmatic temperature.

Several other, more theoretical arguments for an origin by replacement in the solid state are mentioned by RAMBERG (1956) and are also applicable to the Ketilidian pegmatites under discussion. It is unnecessary to repeat them here. From all the observations it follows that most of the Ketilidian pegmatites were formed by nondilation replacement (GOODSPEED 1940). A small number of them, however, appear to have

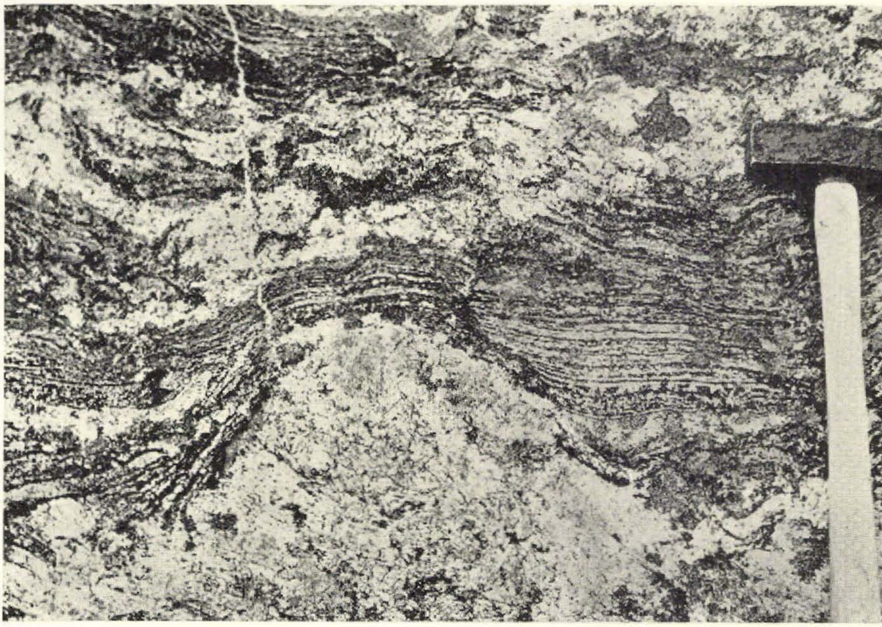


Fig. 18. Swell-structure on the border of a large fourth-generation pegmatite. It is obvious that the schist has been pushed aside plastically. Gneissic schist formation near Nanortalik harbour.

been formed in slowly expanding joints and were thus only partly formed by replacement.

Moreover, the large, generally concordant, pegmatites belonging to the fourth generation, often indicate that they have partly grown by pushing aside the host rock. These typical swell structures show that the schist has been pushed apart in a plastic way (fig. 18). As, according to Ramberg, no silicate melt can push aside the surrounding schist plastically, it seems likely that the expanding pegmatite grew in the solid state. The presence of the swell structures indicates that locally there has been more volume added than replaced. However, as the swells are generally accompanied by pinches in the same pegmatite, the structure could be the result of a tectonic stretching of the pegmatite, in which case the more mobile quartzo-feldspathic material must have moved by recrystallisation from high- to low-pressure points.

It has been shown previously that in the late-Ketilidian third- and fourth-generation pegmatites a zoning commonly occurs. This zoning is formed by plagioclase-rich fine-grained to medium-grained lenses generally enclosed in microcline-rich coarse-grained masses. It has also been demonstrated that the coarse-grained part always appears to have expanded partially by replacing and partly by pushing aside the finer-grained part of the pegmatite. These facts indicate that the large late-Ketilidian pegmatites were probably formed in two successive stages:

1) Formation of relatively thin and fine-grained veins of predominantly quartz and plagioclase (albite-oligoclase).

2) Partial microclinisation, and further expansion of the first-formed vein essentially by growth of coarse-grained quartz and microcline.

The first stage may represent the final part of the period during which albitic plagioclase was formed as a principal mineral in the migmatitic veins. For, as we have seen, albite-oligoclase is an important constituent in all the first- and second-generation pegmatites. Then the second formation-stage of the large third- and fourth-generation pegmatites corresponds probably to the beginning of a large-scale microclinisation. The fact that the microclinisation comes later than the "albitisation" can be explained by the greater mobility of Na compared to K (LAPADU-HARGUES 1945).

The localisation of the very large fourth-generation pegmatites is in some places determined by anticlinal structures in which the pegmatitic material became locked. This was observed in the more or less flat-lying schists NW and W of the Perserajuk mountain, where weak anticlinal arches localise very large late-Ketilidian pegmatites. Similar pegmatite occurrences have been described by NOE-NYGAARD and BERTHELSEN (1952) in the area near Nagssugtôq.

Recapitulating, it can be said that the Ketilidian pegmatites were formed essentially by migration and/or secretion of quartzo-feldspathic material in slip-planes and joints and by growth from these planes by replacement, partially by pushing aside the host-rock and earlier formed veins. The feldspar of the early-Ketilidian pegmatites is principally represented by an acidic plagioclase, while in the later-formed veins microcline is more abundant.

Relations between deformation and pegmatitisation during the Ketilidian period

Field and laboratory evidence show that in the Ketilidian a distinct relationship exists between the different deformation phases and the different pegmatites. Each folding period acted in a different way upon each generation of pegmatites. The area near lake 430 m (N of Quvnerssuaq) was chosen to examine this problem, because in this area the effects of the successive deformations are easy to recognize.

As already mentioned in chapter III, the rocks of the region were deformed during successive folding periods. The effects appear on a large and small scale.

I) The first deformation mainly produced isoclinal folds with axes trending NE to NNE.

II) The second deformation gave birth generally to open structures with axes trending NW.

III) The third deformation produced very large and open folds with ENE to NNE trending axes. This last Ketilidian deformation was accompanied by a coarse fracture- or strain-slip-cleavage.

The following results were obtained by the examination of large-scale rock sections perpendicular to the various foldaxes:

a) The thin first-generation pegmatite veins (P1) appear in most cases to be concordant with the bedding. Only when the layers are strongly isoclinally folded by the first deformation, do the P1 veins not follow the bedding, but continue along an axial-plane cleavage. On the other hand, they conform exactly to the younger folds. These facts seem to indicate that the first-generation pegmatites were formed along bedding- or schistosity-planes, and that they were formed after the first and before the second folding-period.

b) The second-generation pegmatites (P2) are the most abundant in the region. In the schist formations around lake 430 m, they generally form 20 to 40 % of the total rock volume. Numerous macroscopic sections perpendicular to the second fold-axis (NW), show:

1) That the P2 veins are mostly slightly discordant and rarely strongly discordant to the bedding.
2) That the P2 veins are folded in the same direction as but always to a lesser extent than the schist layers (fig. 19).

This last point permits one theoretically, by unfolding a P2 pegmatite, to ascertain at what stage of the second folding the pegmatite was emplaced. Most of the strongly discordant P2 veins show ptygmatic structures. This seems logical, for they have been more compressed longitudinally by the second deformation than have the more concordant veins (fig. 23). Natural rock sections parallel to the NW axes show in most cases a concordance in this direction between bedding and P2 veins (fig. 21). All these observations lead to the following conclusion: The P2 pegmatites were at one time formed *during* the second folding period along discordant slip-planes which were parallel to the NW axes.

c) The third-generation pegmatites (P3) have mainly developed along cleavage-planes and fracture-planes discordant to the bedding

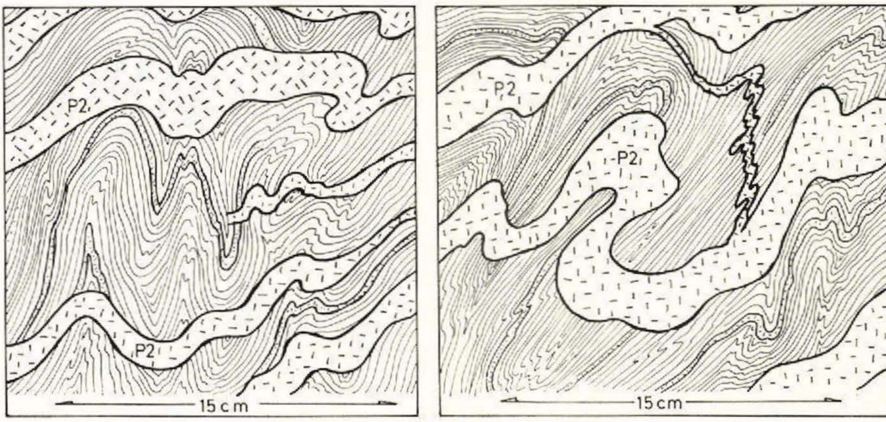


Fig. 19. Rock sections perpendicular to the second NW fold-axes. The P 2 pegmatites are clearly folded in the same direction as, but to a lesser degree than the schist layers. Schist formations around lake 430 m.

(figs. 21 and 23). These slip-planes are the result of a coarse fracture- or strain-slip-cleavage formed during the third Ketilidian deformation period (see chapter III). P3 pegmatites developed also partly as replacement of P2 veins, in which case they follow the pre-existing P2 fold-structures. Anyway, they never appear to be deformed by the second folding. On the other hand, the P3 veins often show fold-structures which must be a result of the third deformation period. This is clearly demonstrated in sections perpendicular to the third NNE fold-axes.

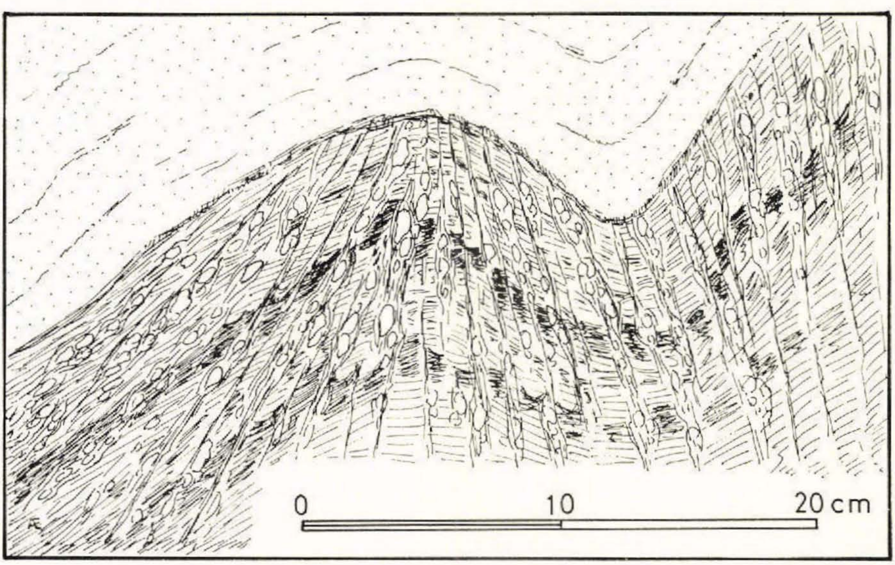


Fig. 20. Feldspar porphyroblasts growing as "augen" in fracture-cleavage planes cutting gneissic schists. Tusardluarnâq peninsula.

Fig. 21. Section normal to the third NNE fold-axes, showing the four generations of pegmatites. The P 3 veins follow the coarse strain-slip-cleavage planes formed during the third deformation period. Schist formation near lake 430 m. Drawn after a colour photograph.

From these facts we may conclude: The P3 pegmatites were formed after the second and during the third folding-period. They grew mainly along discordant fracture planes.

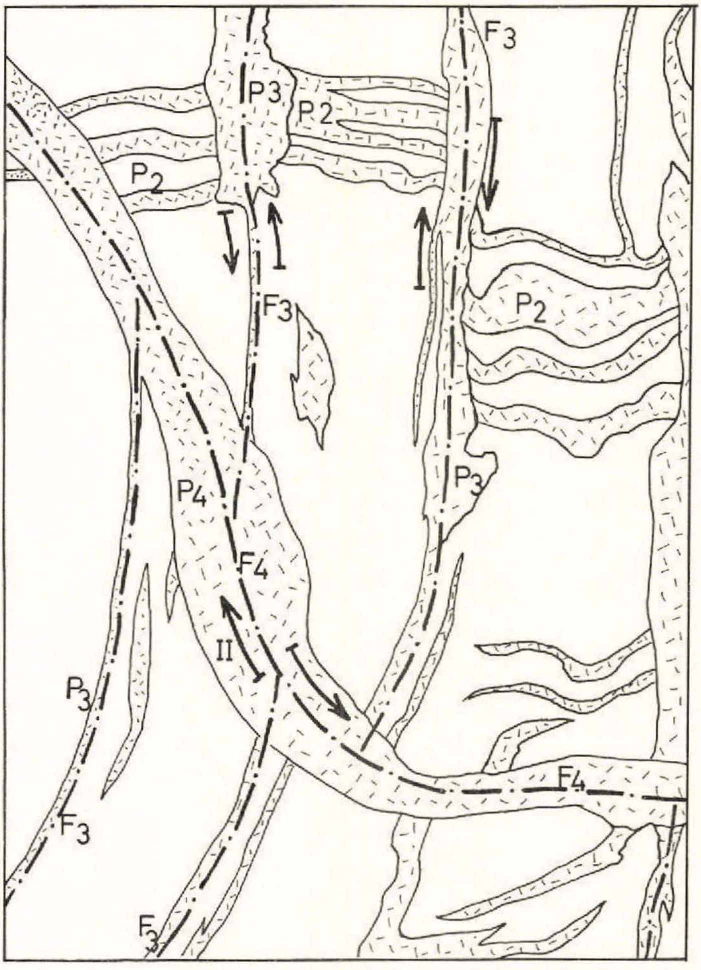


Fig. 22. Schematic interpretation of fig. 21, giving the probable directions of movement along the different fracture-planes (F 3, F 4) which gave birth to the pegmatites P 3 and P 4.

A phenomenon which must be more or less contemporaneous with the formation of the P3 pegmatites has been observed on the peninsula of Tusardluarnâq. In several places a type of augen-gneiss has been formed by the individual growth of large feldspar crystals in fracture-cleavage planes. This fracture-cleavage was formed during the third folding period as is indicated by its direction parallel to that of the third fold axes (NNE). It is particularly developed in the areas where zones of schists alternate with quartzites (fig. 20). The rounded form of the feldspar porphyroblasts shows that they were formed during the deformation.

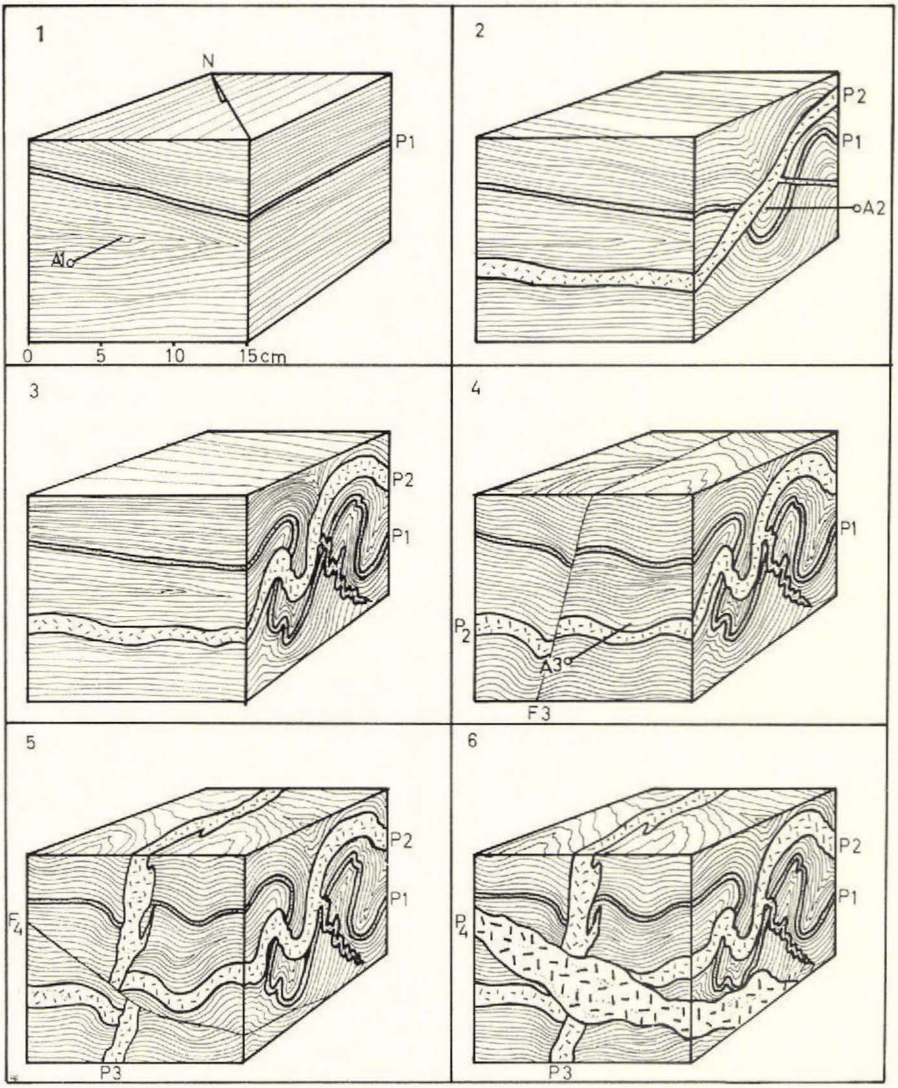


Fig. 23. Six theoretical stages during the Ketilidian deformations and pegmatitisations: 1) End of isoclinal folding I with axis A 1; emplacement of pegmatite P 1 along bedding or axial-plane cleavage. 2) Beginning of folding II with axis A 2; emplacement during this folding of pegmatite P 2 along slip planes parallel to the axis A 2. 3) End of folding II. 4) Beginning of folding III with axis A 3; formation of strain-slip cleavage plane F 3. 5) Continuation of folding III: emplacement of pegmatite P 3 along plane F 3, formation of fault F 4. 6) Formation of large pegmatite P 4 along plane F 4; end of folding III.

d) The very large fourth-generation pegmatites (P4) cut all the earlier ones, and often replace them partially or completely. They are usually roughly concordant or subconcordant with the layering, but transect all the minor structures. In some places the P4 pegmatites follow pre-existing cross-cutting fractures or faults (figs. 20 and 21).

Generally, the fourth-generation pegmatites do not seem to be deformed. Exceptionally, however, they can present fold structures which must be due to the last Ketilidian deformation period. Moreover they often present pinch-and-swell structures. From the above mentioned facts the following conclusions can be drawn: The P4 pegmatites were formed at the end of the third and last Ketilidian folding period. They were emplaced concordantly with the bedding and, now and then, also along pre-existing cross-cutting planes. These planes may be the result of fracturing and faulting generated by the third deformation. It is likely that the concordant planes in which the P4 pegmatites were mostly formed are the result of shearing between the schist- and gneissic-schist layers. The concordant shear-planes were probably formed during the third folding-period, when the deformed rocks were already slightly more rigid than during the preceding folding phases (see chapter III). They must have been important for they permitted the migration and fixation of large quantities of quartzo-feldspathic material.

It is clear that the described division into four pegmatite groups is artificial, and that in fact the genesis of pegmatites during the Ketilidian period was a more or less continuous phenomenon.

Discussion and conclusions concerning the Ketilidian migmatisation and pegmatitisation

It has been shown that the rock series on the peninsula situated between Sarqâ and Tasermiut fjords was subjected to a migmatisation which acted differently upon the various formations. The schists contain the largest proportion of pegmatite veins and are thus the most strongly migmatised, while the overlying quartzites and volcanics and the basal gneisses contain only a minor quantity of quartzo-feldspathic veins.

The fact that the schists belonging to the low-amphibolite facies of metamorphism are more migmatised than the low-granulite facies gneisses can be partly explained by a difference in compactness between the rocks. The more homogeneous and massive gneiss contains fewer fracture- and shear-planes able to give birth to the acid veins.

However, an additional explanatory theory has been given by Ramberg (1951) based on numerous observations and rock-analyses in other parts of western Greenland. The granulite facies gneisses probably belong to the top of the zone of degranitisation which must exist as compensation for the granitisation zone. This might mean that part of the pegmatite material (especially the elements Si, K and Na) forming the migmatite veins in the schists was "squeezed" out of the underlying gneisses. In the Nanortalik region this theory is confirmed by the fact that all the replacement granites are partially formed in amphibolite and epidote-amphibolite facies rocks (see following chapters), and not from the underlying gneisses or gneissic schists.

From the relations between pegmatitisation and folding during the Ketilidian period, the following conclusion can be drawn concerning the times of maximum migmatisation: the migmatisation started after the first folding period and reached a *maximum during the second Ketilidian folding period.* This fact explains partly the more plastic style of deformation of the second folding.

A second, but minor maximum of migmatisation was reached in the area towards the end of the third Ketilidian folding period, when numerous large pegmatites were emplaced along shear-planes and fracture- or fault-planes.

It has been established that at least four different generations of pegmatite veins grew during the Ketilidian period. The various compositions of these veins indicate that the migmatisation started with an enrichment of the rocks in quartz and acid plagioclase, and that it continued, during the third folding-period with a large-scale microclinisation. As stated previously, this succession can be explained by the fact that Na is more mobile and migrates faster than K.

Throughout the whole of the Nanortalik area the migmatite front reached the top of the schist formation and went locally even higher within the quartzites and volcanic rocks. It was however generally stopped by the quartzitic rocks overlying the schist unit.

VII. THE KETILIDIAN GRANODIORITES

Localisation and structural relation to the host-rock

Several relatively small granodiorite outcrops of Ketilidian age have been observed in the central and SW part of the area (see geological map on plate 5). Only the principal occurrences were mapped.

The Ketilidian granodiorites are characterised by a gneissic structure and by the presence of numerous inclusions of schist and gneiss. The granodiorites are commonly lens-shaped, the contacts on a large scale always being concordant with the foliation of the enclosing rocks. The same is true of the inclusions. The contact between granodiorite and country-rock is in most cases very sharp, without gradation. On a large and small scale the granodiorites appear to have been folded to a certain extent together with the host-rock. This proves their Ketilidian age.

The most important occurrence of synkinematic granodiorite was noted S of Qagdlua mountain, where it fills the core of a large anticline (Plate 4A). It is limited on all sides by a concordant bed of dark biotite schists belonging to the base of stratigraphic unit II. The general shape of the gneissic granodiorite body suggests that it was folded together with the enclosing schists during the second and third Ketilidian folding periods (see geological map on plate 6). This is confirmed on a small scale by the structural concordance between folded inclusions and similarly folded granodioritic material. The Qagdlua core-granodiorite contains many amphibolitic inclusions often forming layers which can be traced for several hundred meters. These layers lie parallel to the contacts and accurately follow the shape of the anticline. Near the top of Qagdlua mountain, the granodiorite also contains several large inclusions of garnet gneiss.

Composition, structure and texture

Structure: Generally fine- to medium-grained; often gneissic.

Texture: Predominantly granoblastic, with plagioclase porphyroblasts; locally grano-lepidoblastic. Interlobate textures are common.

Table 5. Chemical composition in weight $^0/_0$ of one sample of granodiorite and one sample of microcline-granite A.

	Granodiorite	Microcline-granite A
G. G. U. Nos.	51640	51689
SiO_2	70,88	68,84
TiO_2	0,13	0,75
Al_2O_3	16,15	13,47
Fe_2O_3	0,28	0,54
FeO	1,42	3,62
MnO	0,06	0,03
MgO	0,29	0,62
CaO	3,12	1,88
Na_2O	5,0	3,0
K_2O	2,2	6,0
P_2O_5	0,07	0,24
CO_2	0,00	0,00
H_2O+	0,28	0,58
Sum	99,88	99,57

Principal minerals: Quartz (15–30 vol. $^0/_0$); albite-oligoclase (An 4–15 $^0/_0$), forms 30 to 35 vol. $^0/_0$ of the rock; microcline (10 to 15 vol. $^0/_0$ of the rock); biotite.

Accessory minerals: Hornblende, muscovite, apatite, zircon, sphene, ore, allanite.

Alteration minerals: Sericite, vermiculite, epidote, chlorite.

Albite-oligoclase and microcline locally form coarse diablastic intergrowths. Mesoperthite was noted in a few places. The microcline generally forms interstitial crystals corroding most of the other minerals. Here and there lepidoblastic aggregates of biotite alternate with granoblastic or interlobate oligoclase-quartz masses. The biotite is pleochroic with X = pure yellow and Y = Z = dark brown to opaque.

One sample of the granodiorite situated S of Qagdlua mountain was chemically analysed in the G.G.U. laboratories by B. I. Borgen. This analysis, compared with those of post-Ketilidian granites, shows clearly the more sodic and calcic composition of the granodiorite (see table 5).

Genetic considerations and conclusions

It has been shown that the granodiorites are characterised by the following properties:

1) The contacts are always concordant.
2) The inclusions are undisturbed compared to the host-rock.

3) The structure is often gneissic.
4) The composition is more sodic and calcic than that of later granites.
5) The granodiorite bodies were folded together with the enclosing rocks during the second and third Ketilidian folding periods.

These observations lead to the conclusion that the granodiorites were formed essentially by replacement of gneisses and schists during the second Ketilidian period of deformation. The composition shows that they are mainly a result of quartzo-albitisation and that they were only slightly affected by the late-Ketilidian microclinisation.

The assumption that the granodiorites were formed essentially by replacement is confirmed when studying in detail the Qagdlua core-granodiorite. As has been shown, this granodiorite body is surrounded concordantly by dark schists belonging to the base of unit II. Moreover, it contains inclusions of amphibolite and garnet gneiss which are both characteristic of the upper part of unit I (see chapter II). It appears therefore logical to assume that the granodiorite core was formed by replacement of the rocks belonging to the upper part of unit I and that the granodioritisation was stopped by the anticlinical arch formed by the overlying biotite schists. As mentioned in chapter III, this anticline was formed during the first Ketilidian folding period, and was refolded later, after or during the granodioritisation.

It seems probable that the quartzo-albitisation which was the cause of the Ketilidian granodiorites was also responsible for the numerous second-generation pegmatites (chapter VI).

VIII. THE FIRST SANERUTIAN GRANITE (GRANITE A)

General character

The first Sanerutian granite presents itself in fresh outcrops as a light grey to grey-blue rock. It is medium- to coarse-grained and slightly porphyroblastic with microcline metacrysts. The length of these metacrysts varies from 0.2 to 1 cm.

Granite A is moreover characterised by the presence of a multitude of xenoliths which are in most cases between 5 and 50 cm long. However, very large "inclusions", up to several hundred meters long, can also be observed in places. All the inclusions have in most cases a common orientation.

Localisation

Granite A forms most of the higher parts of the Nanortalik peninsula. The Arpatsivîp-qáqâ, Pisigsip-qáqâ, Aniggup qáqâ and Quvnerssuaq mountains are all made of this granite (see geological map on plate V). In the NE of the area granite A also forms the highest part of a mountain chain. It is here underlain by younger granites. N of Aniggup qáqâ it occurs as relatively small isolated outcrops in the schists and quartzites.

Relative age determination

The post-Ketilidian and probably early-Sanerutian age of granite A is established by the following facts:

1) The granite cuts and replaces all the Ketilidian pegmatites, including the large late-Ketilidian P4 pegmatites. This can for instance be seen in the area around lake 430 m, where large fourth-generation pegmatites have recrystallised as granite.

2) The first generation of metadolerite dykes is partly granitised in the border zone of granite A near lake 560 m.

3) A second generation of metadolerites transects the granite (see details in chapter IX).

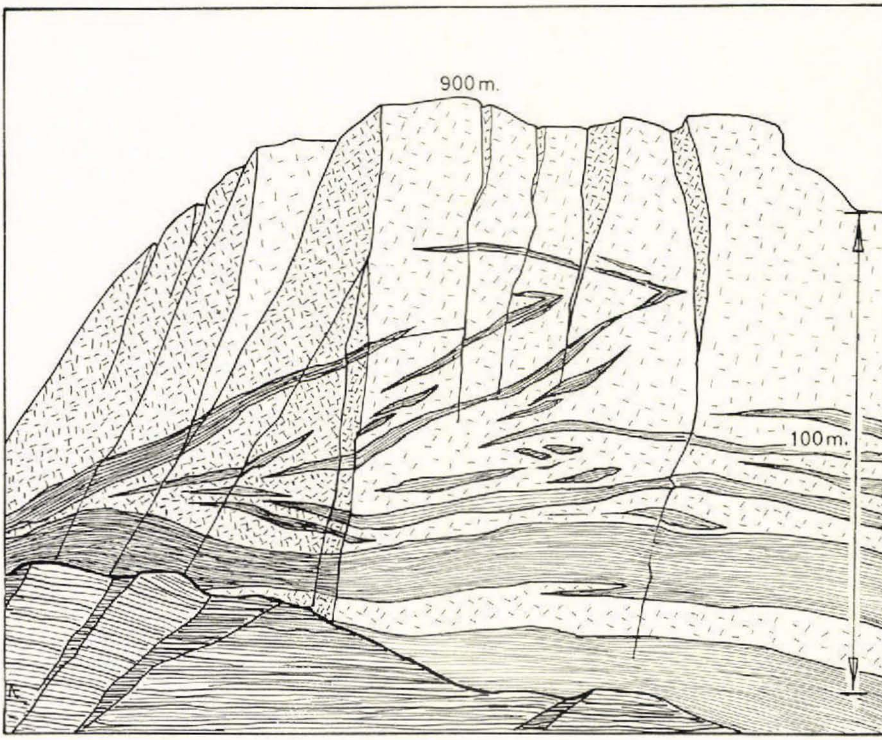


Fig. 24. View of the SW wall of the ridge below the summit of Quvnerssuaq mountain. The granite A, forming the top of the ridge, contains relicts of schists bent in a recumbent fold.

4) The granite does not appear to have been deformed during the Ketilidian orogeny. It locally contains relicts of folded schist layers, showing that the formation of granite A took place after the deformation of the schists (fig. 24).

Mineral composition, structure, texture

Structure: Generally granular, medium- to coarse-grained; in the contact zones and in agmatites often fine-grained. Locally, gneissic or nebulitic structures can be observed. Agmatites are common in the borders between granite and volcanics. A multitude of inclusions often give the granite, at first sight, a foliated aspect. This is in fact not true. The foliation exists only in the xenoliths and is totally absent in the granite material. Only a slight lineation, due to a preferred orientation of feldspar crystals, exists in some places in the granite. The foliation of the large inclusions corresponds in most cases to the bedding or stratification. Most of the xenoliths occur as elongated bodies varying in size between a few centimeters and several meters. The elongation is mostly parallel with the foliation.

Texture: Generally granoblastic to slightly porphyroblastic with microcline metacrysts varying in diameter from a few millimeters to one centimeter. Near the contact with the host-rock grano-lepidoblastic aggregates of biotite with quartz and feldspar can be seen. The texture of the inclusions is lepidoblastic in the schists and nematoblastic in the basic rocks.

Principal minerals: Quartz (15–30 vol. $^0/_0$), microcline (30–50 vol. $^0/_0$), albite-oligoclase (An 5–12 $^0/_0$; 10 to 15 vol. $^0/_0$), biotite (18 to 25 vol. $^0/_0$).

Accessory minerals: Andesine, hornblende, zoisite, epidote, sphene, apatite, rutile, allanite, zircon, muscovite, diopside, ore.

Alteration minerals: Sericite, chlorite, epidote, vermiculite.

Microcline is generally strongly microperthitic (vein perthite). It forms most of the porphyroblasts. However, porphyroblasts of quartz and albite also occur locally. The difference in size between the large and the smaller crystals is generally slight. It is not strictly true to speak of a ground-mass. Micrographic intergrowths of quartz and microcline are frequent. Myrmekite assemblages of quartz and plagioclase are common along the contacts between microcline and albite-oligoclase. The large microcline porphyroblasts are in most cases strongly poikilitic, containing a multitude of inclusions of plagioclase, quartz and biotite. Biotite is pleochroic with X = pure yellow and $Y = Z$ = very dark brown.

The composition of the granite is remarkable constant. Only slight variations have been observed. Near the contact with basic volcanic rocks, the content in quartz and microcline is a little lower than the average value, while near quartzitic rocks the granite is, to a small extent, richer in quartz. Andesine is present near most contacts.

In the NE part of the area, near the summit of Arpatsivîp, granite A seems to be more homogeneous and contains less inclusions than elsewhere. It is here also slightly richer in quartz and microcline (microcline can form up to 55 vol. $^0/_0$ of the rock). The granitisation has reached its maximum in this small area.

In several places large basic to ultra-basic masses are found within the granite, for instance in the region N of lake 560 m. These basic bodies differ from normal volcanic rock inclusions by their absence of foliation or banding and by the fact that their contacts with the granite are transitional. Moreover, characteristic volcanic structures are absent in these bodies. Their composition is rather homogeneous. The principal constituents are: hornblende, epidote, zoisite and oligoclase. Microcline and quartz occur locally as isolated poikilitic porphyroblasts surrounded by the ultrabasic material. Iron oxides are abundant in these rocks. In some places (N of lake 560 m), the basic bodies possess a broad marginal zone formed by an epidote-rich granite. This granite is characterised by its red and green colour due to iron oxides and epidote and it forms a transition between the normal granite and the basic mass.

It seems possible that these basic to ultrabasic bodies represent a "basic behind", in which calcic and ferric minerals were concentrated during the granitisation of the volcanic rocks and schists. Another possibility is that they may be metamorphosed intrusive ultrabasic plugs. This last hypothesis is however questionable, as nowhere in the area, outside the granite, were such large ultrabasic bodies found.

Chemical composition

Six representative samples of granite A were chemically analysed by B. I. Borgen in the laboratories of G. G. U. in Copenhagen. Table 6 gives the results of these analyses, arranged in order of increasing silica content.

Nos. 4 and 10 represent granite samples which were taken in the neighbourhood of volcanic rocks, while sample No. 9 was taken near

Table 6. Chemical compositions in weight % of six samples from granite A.

Nos. on map fig. 28	4	10	5	6	9	7
G. G. U. Nos.	51669	51677	51686	51688	51678	51689
SiO_2	67,53	67,63	67,67	67,87	68,82	68,84
TiO_2	0,84	0,79	0,78	0,74	0,70	0,75
Al_2O_3	14,42	14,45	13,79	13,92	14,14	13,47
Fe_2O_3	0,52	0,49	0,73	0,70	0,55	0,54
FeO	4,54	4,28	4,53	3,78	3,81	3,62
MnO	0,06	0,05	0,05	0,03	0,06	0,03
MgO	1,16	1,11	1,20	1,01	1,02	0,62
CaO	2,47	2,56	2,52	2,56	2,46	1,88
Na_2O	2,90	3,3	3,2	3,3	3,3	3,0
K_2O	4,8	4,2	4,5	4,3	4,0	6,0
P_2O_5	0,30	0,26	0,26	0,24	0,24	0,24
CO_2	0,00	0,00	0,00	0,00	0,00	0,00
H_2O+	0,24	0,48	0,58	1,27	0,50	0,58
Sum	99,78	99,60	99,81	99,72	99,60	99,57

the contact with quartzitic schists. Samples No. 5 and 6 represent a granite far from any contact. Finally, No. 7 gives the composition of the granite where it appears most homogeneous and contains least inclusions (see map fig. 28).

It is remarkable to see how little the composition varies from one point to another in granite A. This seems particularly curious, as the granite has been formed by metasomatic replacement of rocks of very different composition. Only slight variations in chemical composition appear in table 6:

1) The granite probably formed by replacement of basic volcanic rocks (Nos. 4, 10, 5 and 6) contains less SiO_2 and more $FeO + Fe_2O_3$, Al_2O_3 and TiO_2 than that resulting from granitisation of quartzitic schists (sample No. 9).

2) The samples taken near the contact with volcanic rocks (Nos. 4 and 10) are a little poorer in SiO_2 and slightly richer in Al_2O_3 and TiO_2 than those taken farther away from the contact (Nos. 5 and 6).

3) Sample No. 7, representing probably a maximum of granitisation, contains more SiO_2 and K_2O and less CaO, Al_2O_3, TiO_2, MgO and FeO than the other rock-samples.

These differences, though slight, show clearly that there exists a relation between the composition of the granite and that of the replaced rocks. The replacement of basic volcanic rocks gives a slightly more basic granite than the replacement of quartzitic schists. It is possible that the different parts of the granite tended towards the "ideal" composition formed in sample No. 7.

Structural relations between granite and country-rock

On a large scale, the first post-Ketilidian granite appears to be in some parts concordant with and in others discordant to the country-rock (see geological map on plate 5). However, when the contacts are carefully examined, frankly discordant borders can nowhere be seen. The apparent large-scale discordances are made up of wedge-like, alternating layers of granite and host-rock, whilst the borders of the granite "wedges" are concordant or subconcordant with the stratification of the enclosing rock (plate 4 B). The "wedge" contacts can be observed particularly in the central part of the area, in the border zone between granite A and the volcanic and quartzitic rocks.

The first post-Ketilidian granite appears to form a huge concordant sheet resting upon the pelitic schists (plate 4 B). The contacts between granite and schists dip slightly inwards on the NW and SE borders.

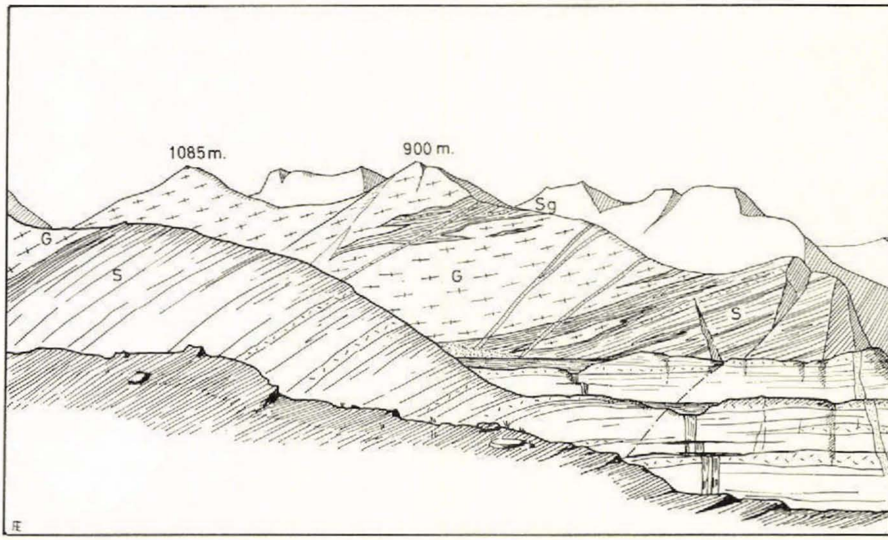


Fig. 25. View from the S of the mountains bordering lake 430 m. Granite A (G) lies conformably upon pelitic schists (S), belonging to the base of unit III. It contains a large undisturbed inclusion of garnet schist (Sg).

This is probably due to the dip of the schists, which form a broad NE syncline in this part of the peninsula. The granite sheet is overlain in the SW and NE by volcanic rocks. It contains a multitude of inclusions varying in size from a few cm to several hundred m. The material of which the inclusions are made is volcanic, quartzitic or schistose. Near lake 430 m, several very large inclusions of garnet schist were observed within the granite (fig. 25).

The inclusions lie perfectly parallel with each other and with the stratification of the country-rock. As the granite-sheet lies here concordantly on pelitic schists which probably belong to the base of unit III, it seems that the granite was formed by partial replacement of the garnet schists that normally form the upper part of unit III. The granitisation was incomplete and has left the inclusions as unmoved remnants of the original rock.

Similar observations were made more to the N in the area situated between the Arpatsivîp-qáqâ and Pisigsip-qáqâ mountains, where very large outcrops of volcanic rocks occur surrounded by granite A. By measuring the strike and dip of foliation-planes in the different large and small inclusions of volcanic rock, the structures, which must have existed before the partial granitisation took place, have been reconstructed. These structures show a syncline and an anticline with NW trending axes (figs. 26, 27, 28). They were probably formed during the second (NW) Ketilidian folding period (chapter III).

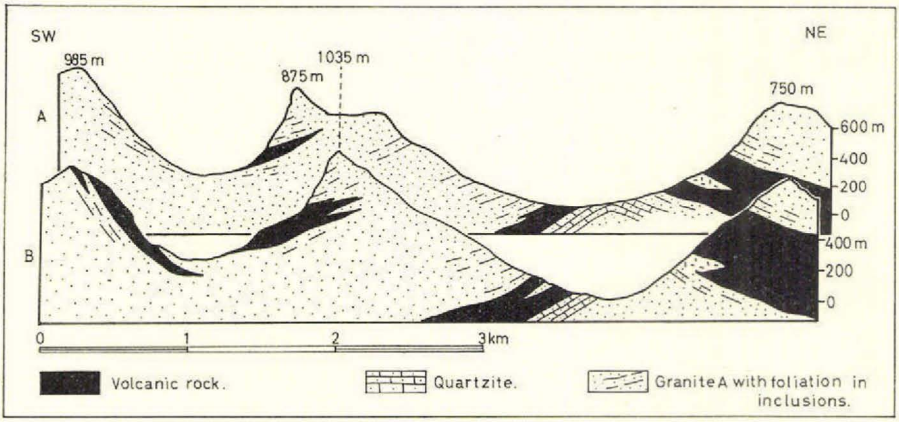


Fig. 26. Two schematic sections across granite A in the area N of Pisigsip-qáqâ mountain. The included relicts of volcanics and quartzites show the original structure before the granitisation. The trace of the sections is given on fig. 28.

The fact that all the various xenoliths observed form together a well defined structure (fig. 27) indicates that the granitisation took place after the folding and practically without mechanical movement of the rocks. It also excludes any magmatic origin for the granite. The only

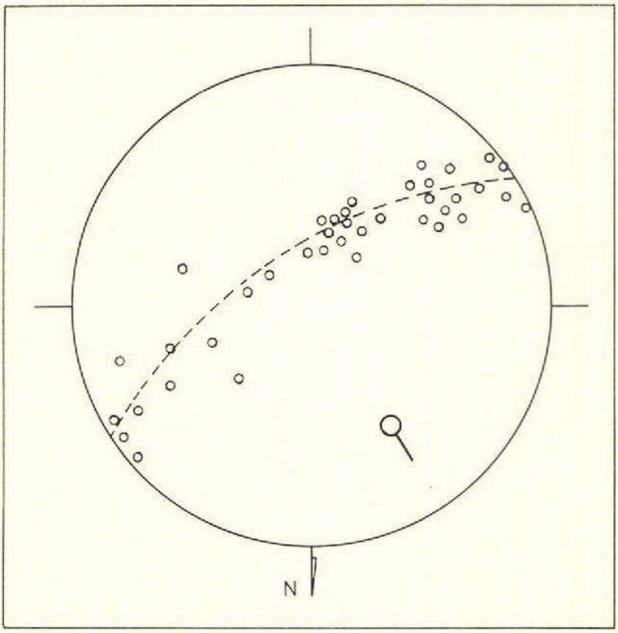


Fig. 27. Stereogram of the foliation poles of 39 large and small inclusions of volcanic rock in granite in the upper part of the synclinal valley between the summits 985 and 875 m (see fig. 26). The lower hemisphere of the Wulff net is used.

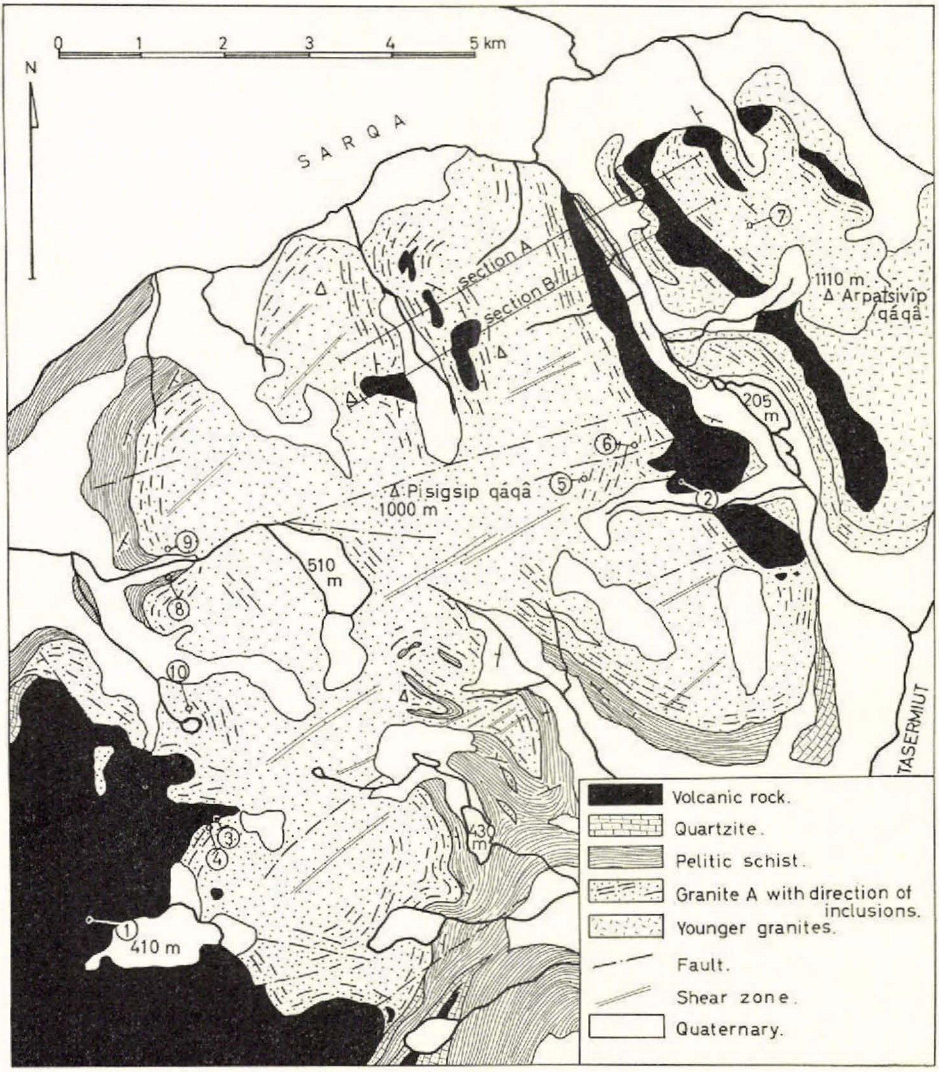


Fig. 28. Geological map of the central part of the Nanortalik peninsula, showing granite A and the directions of the foliation found in the inclusions. Nos. 1 to 10 correspond to samples taken for chemical analysis. Sections A and B correspond to those on fig. 26.

place where some movement occurred during the replacement is situated in a small area directly SW of lake 205 m, where the random orientation of the inclusions show that the granite was mobile at a certain period. Otherwise, all the measured directions of xenolith-foliations follow a definite pattern, concordant with the structure of the countryrock. This "foliation" of the granite was mapped in most of the zones lying near the contacts with supracrustal rocks. It is represented on the map of fig. 28 and clearly shows the concordance which exists between the foliation of the inclusions in the granite and that of the country-rock.

All the observations indicate that granite A was formed by metasomatic replacement of large quantities of volcanic rocks and of minor amounts of quartzitic and pelitic schists. It seems strange that the granitisation was stopped downwards, in most parts of the area, by the lower pelitic shists. The quartzites and volcanics, which generally are more resistent to the granitisation, are replaced, while the underlying schists remain for the greater part unchanged. Of course, it is not impossible that in its central part the granite massif invaded the schists to a much greater depth.

Contact relations between schists and granite

It becomes even more evident that the granite A was formed by metasomatic replacement when its contacts with the different country-rocks are examined in detail.

In the vicinity of lake 430 m the borders between garnet schist and granite are particularly well exposed. In a zone varying in width between 20 and 400 m, it can be observed how the schists, which are strongly migmatitic, gradually recrystallise and get a more granitic aspect, and finally pass into a granite.

The transition is gradual in places but also zones of practically unchanged schists alternate with concordant zones in which the schist

Table 7. Chemical composition in weight % of one sample of granite A and one sample of quartzitic schist. The samples were taken near their common contact.

Nos. on map fig. 28 G. G. U. Nos.	8 (schist) 51681	9 (granite) 51678
SiO_2	65,51	68,82
TiO_2	0,73	0,70
Al_2O_3	14,70	14,14
Fe_2O_3	0,63	0,55
FeO	5,21	3,81
MnO	0,09	0,06
MgO	1,88	1,02
CaO	3,31	2,46
Na_2O	3,2	3,3
K_2O	2,9	4,0
P_2O_5	0,21	0,24
CO_2	0,00	0,00
H_2O+	0,72	0,50
Sum	99,09	99,60

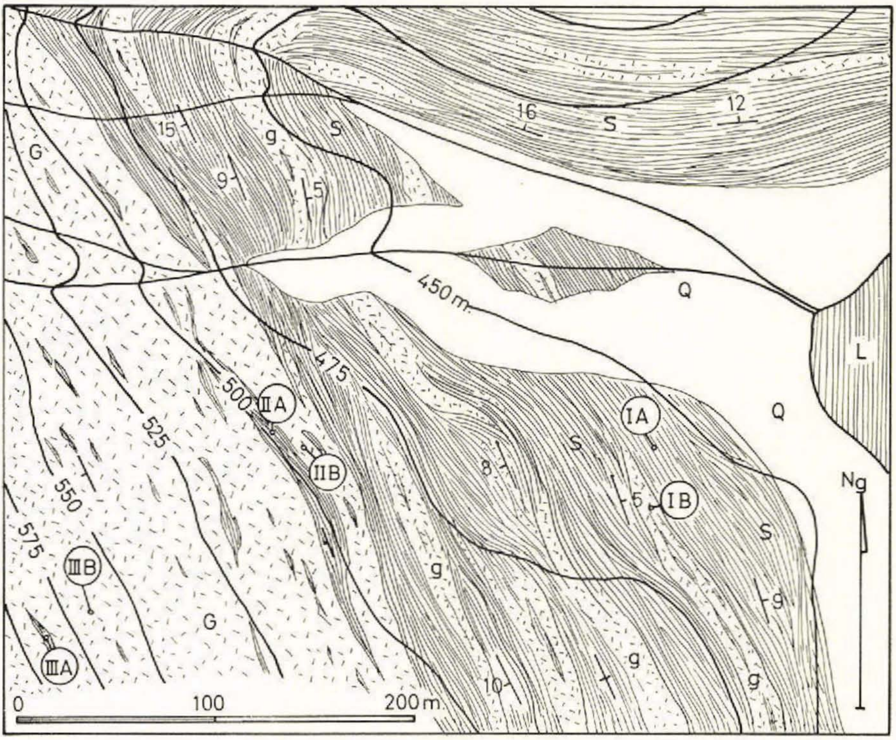


Fig. 29. Detailed geological map showing the contact zone between granite A (G) and garnet schist (S) W of lake 430 m. g = slightly granitised schist.

is more or less granitised. The contact between these zones is generally sharp. When the granite is approached, it can be seen how the granitisation increases in intensity and how the granitised zones become larger and more numerous, and finally give way to a more or less homogeneous granite in which only small inclusions of schist occur (fig. 29).

Looking in detail at the more and less granitised parts of the migmatitic schists, it can be seen that the principal effect of the granitisation is a rather coarse recrystallisation tending to homogenise the migmatitic schist: feldspar and quartz metacrysts grow within the dark biotite-rich parts of the schist, while biotite develops in the leucocratic veins (plate 1). On the other hand, the schistose, ungranitised remains, which form concordant lenses, become smaller nearer the granite and appear to be enriched in basic minerals.

All the Ketilidian pegmatites situated in the garnet schists are gradually recrystallised and changed into granite A, which clearly shows the post-Ketilidian age of the granite. These observations also indicate that granite A is not a direct result of Ketilidian migmatisation, but is due to a later reactivation of the migmatitic schists. The reactivation

Table 8. Average values of the mineral compositions in vol. % of three garnet schist samples and three granite A samples. (The samples were taken near one another in the area W of lake 430 m).

G. G. U. Nos.	63611, 63609, 63641. (Schist)	63614, 63613, 63643. (Granite)
Quartz	28	23
Microcline	15	41
Plagioclase (An 5–15 %)	21	12
Biotite	19	20
Garnet	6	—
Muscovite/chlorite	6	—
Access. minerals	5	4

was accompanied by a slight enrichment in Si and K and a loss of Ca, Fe and Mg, as proved by the following chemical analyses and evaluations of mineral contents:

1) B. I. Borgen has analysed representative samples of the rocks across the contact between granite A and quartzitic schist, N of lake 480 m. The results of these analyses are given in table 7. The corresponding numbers on the map of fig. 28 show where the samples were taken.

2) Three samples of granite A and three samples of garnet schist, taken near each other W of lake 430 m, were examined under the microscope. Their mineral composition was evaluated by point-counting and with the help of transparent networks superposed on detailed photographs of the rocks. In the composition of the schist are included the numerous pegmatite veins. The average composition of the three granite-samples is compared in table 8 with that of three schist samples.

Both tables, although representing different contacts, show that the granitisation was accompanied by an enrichment in K and, to a less degree, in Si. This was compensated by a loss in Ca, Mg and Fe. The content of Na seems to have remained more or less the same, because the microcline of the granite is strongly microperthitic and contains a relatively high amount of Na. From a mineralogical point of view, the granitisation appears to have been essentially a microclinisation.

The details of replacement and recrystallisation have been studied in 6 samples, which were taken close to the contact between granite and garnet-schist near lake 430 m. These samples represent three different stages of granitisation. The detailed map on fig. 29 shows their location.

The following pages describe successively each sample: (each stage of granitisation is illustrated separately in fig. 30).

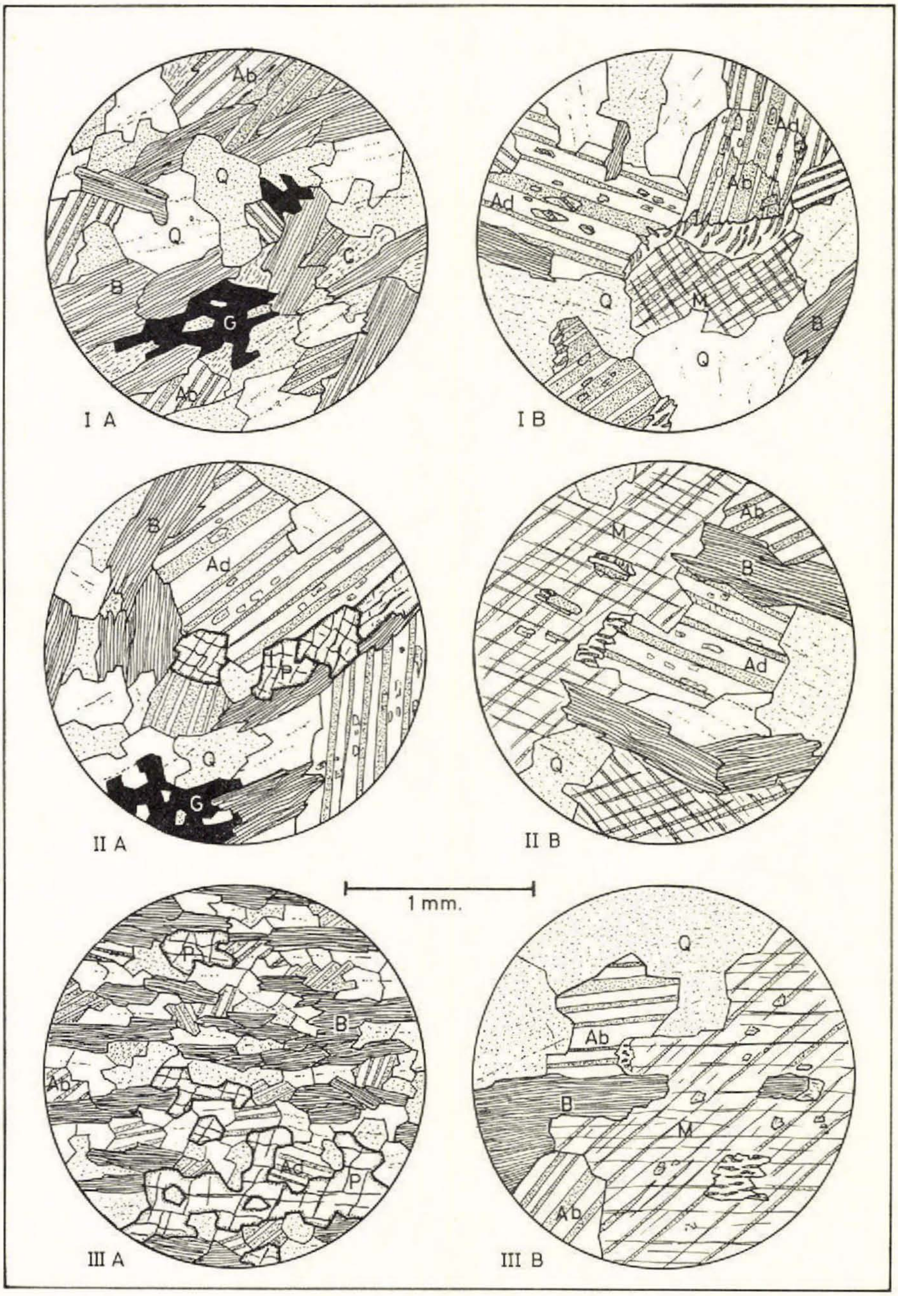


Fig. 30. Three stages of granitisation of garnet schists. The map on fig. 29 shows where the samples were taken. Q = quartz, M = microcline, Ab = albite-oligoclase, Ad = andesine, B = biotite, P = pyroxene, G = garnet, C = chlorite. For further explanations see text.

Sample No. I A (G. G. U. No. 63651): This sample represents the relatively unchanged garnet schist at a considerable distance from the contact. It is principally composed of grano-lepidoblastic aggregates of quartz, albite-oligoclase and biotite. Garnet forms interstitial poikiloblastic crystals.

Sample No. I B (G. G. U. No. 63652) represents the first zone of granitisation, relatively far from the granite. The composition is characterised by the appearance of andesine (An 35 $^0/_0$) and microcline as partial replacement products of albite-oligoclase, biotite and garnet. Quartz has recrystallised into large metacrysts. Andesine contains many inclusions of albite-oligoclase. Locally it appears to be pseudomorphous after the latter. Micrographic intergrowths between microcline and plagioclase are common.

Sample No. II A (G. G. U. No. 63639): This sample was taken near the contact in a schist zone surrounded by granitic material. The mineral composition shows that a slight basification of the rock took place. Andesine appears in large poikilitic crystals corroding and replacing the albite-oligoclase and the quartz. Pyroxene (diopside) has probably grown as pseudomorphs after garnet.

Sample No. II B (G. G. U. No. 63654): This represents a zone of already homogeneous granite, close to the contact. The mineral assemblage indicates a further microclinisation and acidification of the rock, compared with sample I B. Microcline forms large phenoblasts invading the rock and replacing most of the other minerals; it contains numerous, often large inclusions, principally of plagioclase. A new albite, slightly more acid than the albite-oligoclase of the schist (An 5–8 $^0/_0$) is present and has partially to completely replaced the andesine.

Sample No. III A (G. G. U. No. 63653): This sample was taken from a schist xenolith in the granite, relatively far from the contact. The composition is still more basic than that of the schist zones near the contact (Sample No. II A). Diopside is abundantly developed while garnet has disappeared completely. A new generation of small biotite flakes has grown, generally parallel to the pre-existing schistosity. Quartz and plagioclase have also recrystallised to fine-grained granoblastic aggregates. They form only a small percentage of the rock.

Sample No. III B (G. G. U. No. 63613): This represents granite A, relatively far from the contact. Microcline forms here large crystals. Andesine has completely disappeared, making room for albite. Biotite has recrystallised to large flakes and quartz has recrystallised to large xenoblasts.

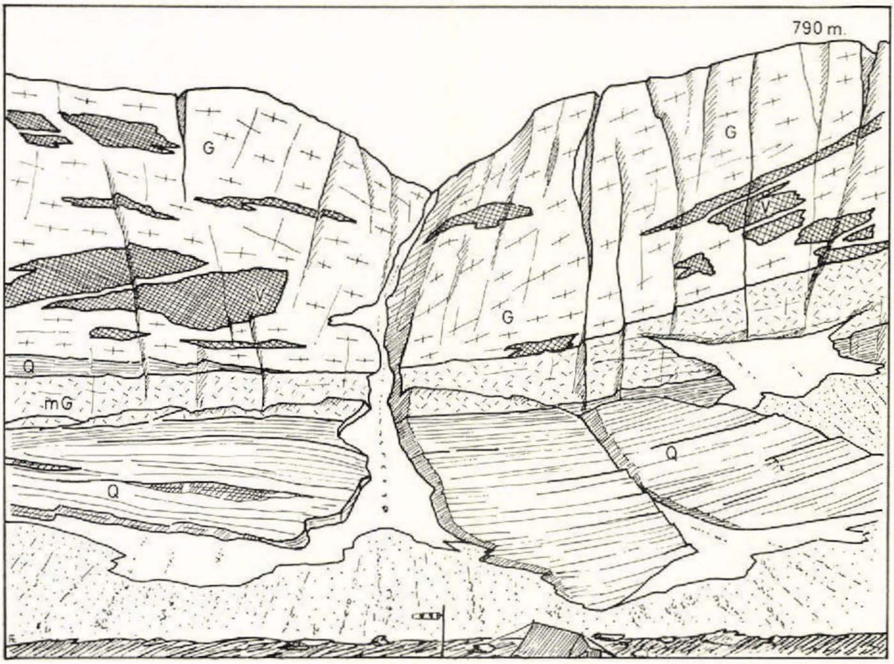


Fig. 31. NNW wall of Quvnerssuaq mountain. Granite A (G), containing large inclusions of volcanic rock (V), overlies conformably the quartzitic schists (Q). A later microgranite sheet (mG) follows more or less the contact.

Summarising these observations, the granitisation of the garnet schists can be divided into three main stages:

1) Disappearance of garnet; growth of andesine partially at the expense of albite-oligoclase. Formation of microcline.

2) Expansion of microcline, partially replacing the other minerals; appearance of new albite, partially replacing the andesine.

3) Further expansion of microcline. Disappearance of andesine. Formation of albite; recrystallisation of biotite and quartz into large flakes.

Each of these stages is accompanied by a higher degree of basifiction of the schist remnants.

The birth of andesine during stage 1 is probably due to the liberation of Ca by destruction of garnet.

Contact relations between volcanic rocks and granite A

The contact relations between volcanic rocks and granite A also show that the granite was formed by replacement without important mechanical movement. This can for instance be seen in the NNW wall

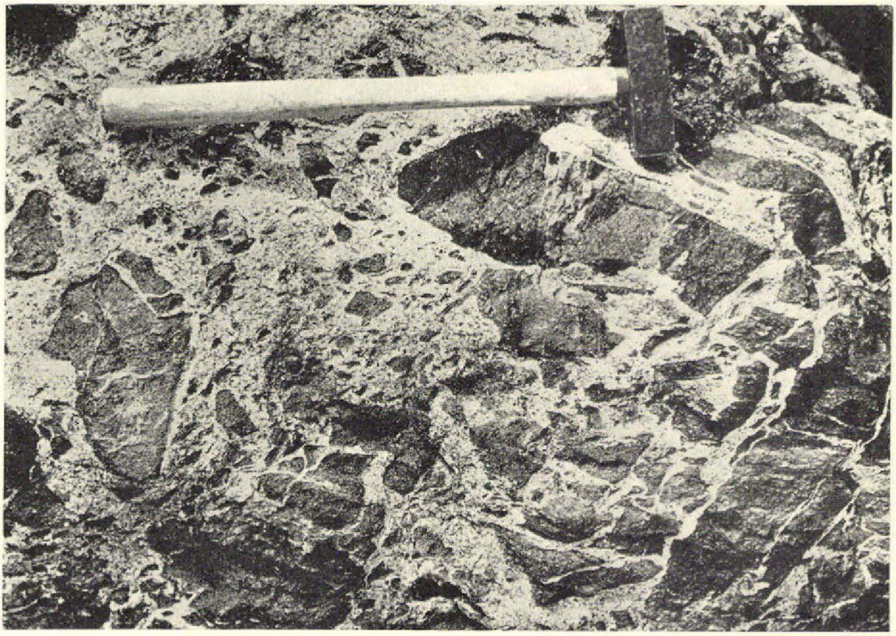


Fig. 32. Small-scale agmatite on the contact between granite A and volcanic rocks. Contact zone SW of lake 560 m (see geological map on plate 6).

of the Quvnerssuaq mountain ridge, where the granite lies concordantly on quartzitic schists and contains many large inclusions of volcanic rocks (fig. 31). These inclusions have apparently not been rotated and may be remnants of the volcanic series which originally overlaid the quartzitic schists.

The way in which the volcanics are replaced is different from the rather gradual granitisation observed in the schists. The contacts between granite and volcanic rock are always sharp and the granitisation progresses by small- and large-scale agmatitisation (fig. 32).

Petrographic investigation of samples, found in the contactzone N of lake 410 m, show that the granitic material in the agmatites is essentially composed of fine-grained aggregates of quartz, microcline and plagioclase. The plagioclase is principally an acidic andesine (An 30–35 $^0/_0$).

The volcanic rocks near this contact are predominantly composed of pillow-lavas. A striking change in the aspect and composition of the pillows can be observed when approaching the contact with granite A; the pillow structures gradually disappear and the rock becomes finer-grained and compact. The colour changes from dark green to brownish-yellow to light grey. The composition also varies: most of the hornblende is replaced by biotite, while epidote gives way to diopside and andesine. Locally small crystals of microcline grow between the other minerals.

Table 9. Chemical composition in weight % of three samples of pillow lava taken near the contact with granite A. The analyses are numbered according to decreasing distance from the contact.

Nos. on map fig. 28	1	2	3
G. G. U. nos.	51658	51691	51667
SiO_2	48,18	53,16	57,83
TiO_2	1,13	0,91	1,04
Al_2O_3	15,95	18,12	16,30
Fe_2O_3	0,83	0,89	0,41
FeO	9,93	7,50	7,57
MnO	0,22	0,13	0,04
MgO	7,56	4,51	3,40
CaO	11,31	6,91	4,93
Na_2O	1,5	3,0	2,7
K_2O	0,4	2,0	3,8
P_2O_5	0,14	0,17	0,26
H_2O+	1,36	2,14	0,78
Sum	98,51	99,44	99,06

In order to investigate how the chemical composition changes from pillow-lavas to granite, 7 samples of lava and granite, taken at different distances from the contact, were analysed in the G. G. U. laboratories. The results of these analyses are partially given in tabel 9 (for the lava samples). Analyses of the granite have already been given in table 6.

The analyses shown in table 9 give an idea about the chemical variations affecting the basic rocks near their contact with granite A. To compare these data with the analyses of granite samples, 7 chemical compositions of lavas and granites are represented on a Harker variation diagram (fig. 33). The map on fig. 28 shows the different localities where the samples nos. 1–7 were taken.

The rock specimens nos. 1, 2 and 3 are pillow lavas sampled at decreasing distances from the granite. No. 3 was taken at the contact in an agmatite. Specimens nos. 4 to 7 represent granite A at increasing distances from the contact. No. 4 was taken only some meters from no. 3, while a maximum of granitisation is reached in no. 7. All the granite samples were probably formed by granitisation of lavas or other volcanic rocks.

The diagram on fig. 33 shows that:

1) The chemical change in the lavas started already at some distance from the granite contact by an enrichment in Si, K, Na and Al, and by an impoverishment in most of the other composing elements, particularly Ca and Mg.

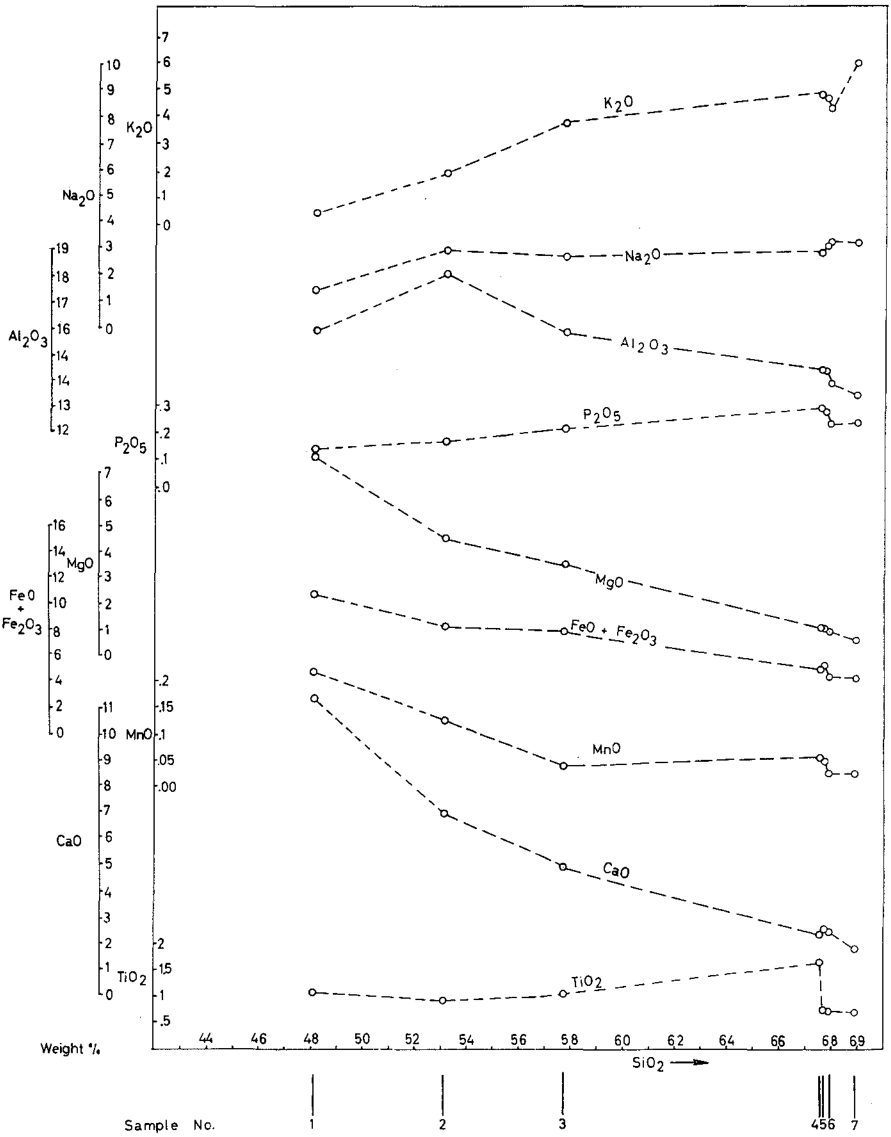


Fig. 33. A Harker variation diagram showing the differences in composition between pillow-lava and granite A in the contact zone between the two rocks. Nos. 1–3 represent lavas, nos. 4–7 were sampled in granites. The succession 1–7 corresponds to an increasing degree of granitisation.

2) The principal difference in composition between "unchanged" and completely granitised lavas is the increased content in K and Si, and the loss of Ca, Fe and Mg.

3) The variations are gradual.

4) Once the granite was formed, the composition remained remarkably constant. Only sample no. 7, representing a maximum of granitisation, shows an important enrichment in K and Si, and a loss in Ca. This means that most of the chemical variations occur within the lavas.
5) The granitisation of large quantities of pillow-lavas and other volcanic rocks involves an important supply of Si, K and Na, and the equally important removal of Ca, Fe, Mg, Mn and Al.

Genetic considerations and conclusions

All the observations mentioned in the preceding pages show that the first post-Ketilidian granite was formed mainly by metasomatic replacement, without important mechanical movement. The large-scale residual structures found within the granite suggest, moreover, that no important variations in volume occurred during the replacement. The granitisation progressed in a selective way, preferentially replacing volcanic rocks and quartzitic schists and leaving most of the underlying pelitic schists unchanged. This preference explains the sheet-like shape of the granite mass. No satisfactory explanation can be given for the fact that the more basic and compact volcanics were more easily granitised than the strongly migmatitic pelitic schists.

It is possible that the migrating granitising elements were more readily concentrated in the disturbed zone located between supra- and infra-structure. The granitisation probably started there and progressed further into the volcanic rocks following tension joints and cracks which must have existed in these brittle rocks in early Sanerutian times. This hypothesis could explain the preference shown by the granitisation for the volcanic rocks. For it is known that the mica schists have a pronounced tendency to yield to tensional stress by plastic flowage, while the basic rocks are considerably more brittle and compensate the tension by jointing (Ramberg 1956).

Petrographic observations show that the granitisation was essentially a microclinisation, often accompanied by an enrichment in quartz and, to a slight degree, in albite. A first stage of granitisation appears to be characterised in most contact-zones by the crystallisation of andesine. But the andesine disappears when the granitisation is more pronounced. Chemical analyses indicate that the replacement of large quantities of basic rocks involves an important introduction of Si, K and Na and an equally important removal of Ca, Fe, Mg, Mn and Al.

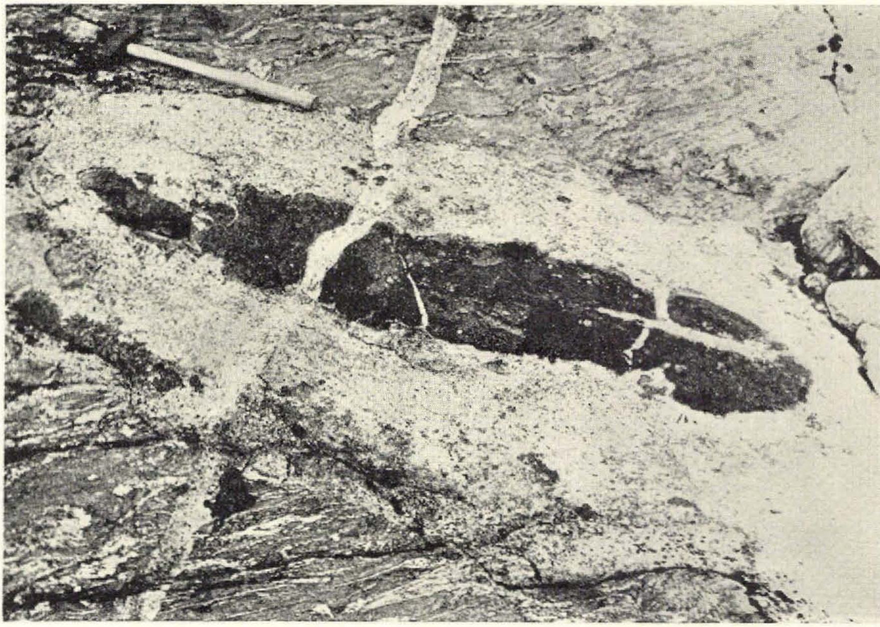


Fig. 34. Relicts of metadolerite enclosed in fine-grained microcline granite. Garnet schist formation near lake 430 m and near the contact with granite A.

IX. THE METADOLERITE DYKES

Localisation and relation to the host-rock

Metadolerite dykes are very scarce in the Nanortalik area. They were only observed in five places on the peninsula. The best metadolerite exposures can be seen in the area between lake 430 m and lake 560 m (geological map on plate 6).

The relations between dykes and adjacent rocks suggest that there are two generations, one intruded before and one after the formation of granite A.

The following observations characterise the first generation:

1) The direction of the dykes varies from E to NE.
2) They are often intersected and partially replaced by late-Ketilidian pegmatites.
3) The dykes are slightly folded in places.
4) In the neighbourhood of granite A the basic dyke material is for a large part replaced by a fine-grained microcline-andesine granite,

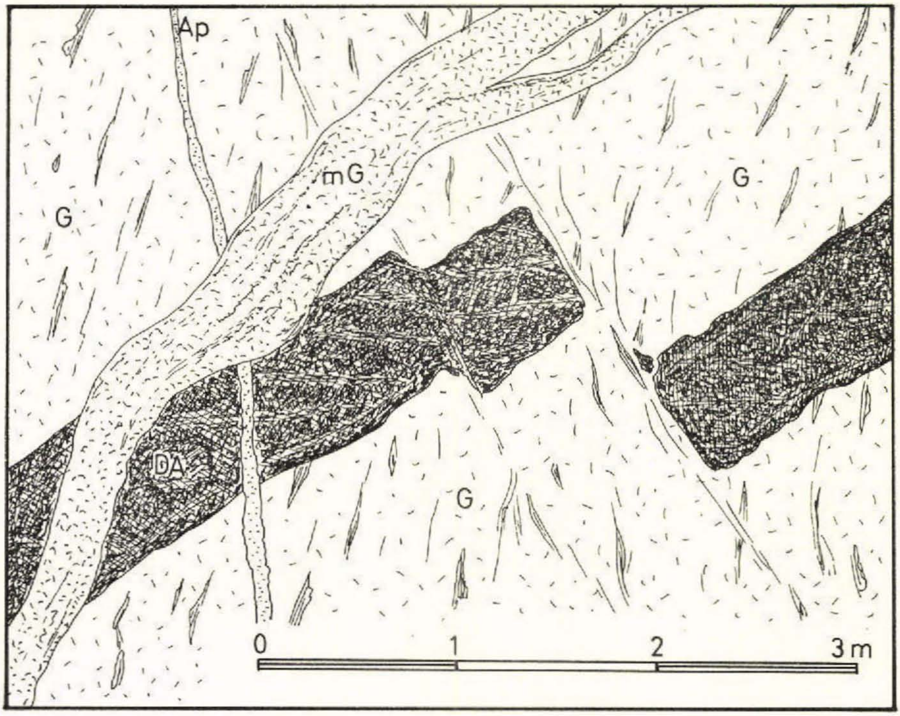


Fig. 35. Second generation metadolerite in granite A (G) near lake 560 m. The dyke contact transects inclusions. Dyke and granite are deformed by Sanerutian shearing and are cut by later microgranite (mG).

identical to that formed in agmatitic contacts between granite A and volcanic rocks. The granitisation was selective; it only replaced the dykes, leaving the adjoining schist unchanged (fig. 34).

5) Near the contact with granite A the dyke is completely granitised and disappears.

6) The metadoleritic material contains many microcline metacrysts.

The second generation of metadolerites is only represented by two dykes found in granite A on the N shore of lake 560 m. They show the following properties:

1) The dykes are oriented NW–SE, which is normal to the direction of the first-generation metadolerites.

2) The dyke contacts transect the feldspar porphyroblasts and inclusions of the granite (fig. 35).

3) The metadolerites are ungranitised and unfolded. Only locally do

shear planes, due to Sanerutian deformations, cut and slightly deform the dykes together with the granite.

4) The dyke material is relatively fresh. Microcline metacrysts are scarce.

Unfortunately, no intersections between first- and second-generation dykes were observed. However, even without this confirmation, the differences between the two groups are so obvious that their simultaneous intrusion seems impossible.

Composition

The metadolerites are essentially composed of andesine (An 40 to 55 $^0/_0$), hornblende and biotite. Relict ophitic texture is generally clearly defined by subhedral-euhedral laths of plagioclase. Pyroxene is scarce, only appearing as relic grains surrounded by hornblende. Microcline and quartz occur accessorily and form small granoblastic aggregates corroding the plagioclase and the hornblende. In the first-generation dykes microcline often forms poikiloblasts containing relicts of andesine. Late albite-oligoclase locally replaces the andesine, apparently without change of the crystal shape. Hornblende is partially altered to biotite, epidote and ore. Sphene, leucoxene and apatite form a minor part of the rock.

X. THE SECOND SANERUTIAN GRANITE (GRANITE B)

General character

The most noticeable feature of the second Sanerutian granite is the common occurrence of feldspar metacrysts. These crystals are irregularly distributed but generally form between 40 and 65 vol. $^0/_0$ of the rock and give the granite a very coarse-grained aspect. The length of the feldspar crystals varies from 5 mm to 15 cm. Many of the large megacrysts possess a coarse perthitic texture which gives schiller effects in hand-specimen. The weathered surface of granite B displays a typical dark reddish-brown colour.

These main features are very similar to those of the Frederiksdal and Sydprøven "New granites", as described by WEGMANN (1938) and BRIDGWATER (1963).

Localisation

Granite B occurs in two areas along Tasermiut fjord: In the region of Aniggoq mountain and, more to the NE, in the Arpatsivîp-qáqâ mountain range (geological map on plate 5). It seems likely that both occurrences are connected with the large "new granite" massif, found on the SE side of Tasermiut fjord (WALLIS 1962).

In the Arpatsivîp-qáqâ mountains granite B locally forms large concordant sheets between older granite and volcanic rocks.

Relative age determination

The lack in basic dykes makes it necessary to establish the age of granite B by means of its contact relations with the other granites.

N of Aniggoq mountain granite B replaces in parts granite A; large megacrysts grow separately or in groups within the older granite. Moreover, several inclusions of granite A are situated in the coarsely porphyroblastic granite. On the other hand, granite B is intruded and replaced by the fine-grained granite C, as can be observed in the NE part of our area.

These facts show that granite B must have been formed during the Sanerutian, after the emplacement of granite A and before the intrusion of granite C.

Mineral composition, structure, texture

Structure: Generally inequigranular, coarse- to very coarse-grained.

Texture: Predominantly porphyroblastic. The microcline phenoblasts are often elliptically shaped. Exceptionally, they may be surrounded by a mantle of plagioclase, giving the rock a rapakivi texture (SEDERHOLM 1928, ESKOLA 1928). Most of the microcline megacrysts are strongly poikilitic and often show concentric zones of inclusions.

Principal minerals: Microcline (35 to 65 vol.$^0/_0$ of the rock), albite-oligoclase (10 to 15 $^0/_0$ An), quartz, biotite, green hornblende.

Accessory minerals: Apatite, allanite, zircon, ore.

Alteration minerals: Chlorite, sericite, epidote.

Microcline forms most of the megacrysts. Its crystal shape can be tabular, subrounded or ellipsoidal. Microcline is strongly perthitic, often coarsely so; it appears to be considerably more perthitic than the microcline of the earlier granite A. According to MARMO (1962), the higher content in albite-perthite in younger microcline-granites is a general rule in Finland. Several of the large microcline crystals display irregular concentric zones of inclusions of biotite, plagioclase and quartz. These zones may correspond to succesive growth periods.

Many microclines possess lighter coloured margins due to a higher content in albite-perthite. Typical rapakivi texture, characterised by zones of plagioclase mantling the porphyroblasts, was seen in two places in the Aniggoq granite. WEGMANN's remark that the rapakivi structure is restricted to "those parts of the granite massifs in which inclusions are present as also where they are about to disappear", could not be confirmed in the Nanortalik area. Both occurrences of rapakivi texture were found in inclusion-poor parts of the granite.

Albite-oligoclase occasionally forms small porphyroblasts. The matrix of the rock is formed of granoblastic aggregates of plagioclase, quartz, biotite and hornblende. Quartz occurs mainly as anhedral interstitial grains. Hornblende and plagioclase are mostly poikilitic.

The general aspect of the different minerals suggests, in the majority of cases, a formation by recrystallisation and partial replacement of earlier minerals. This is proved for instance by the common orientation of small grouped inclusions, showing that they were not moved and are all relics of the same pre-exiting crystal. This phenomenon is particularly evident in the large microcline poikiloblasts (fig. 36). Plagioclase inclusions in microcline generally possess an albitic rim.

The composition of the granite is not constant, the main variable being the amount of potash feldspar; near the contacts with basic volcanic rocks the granite contains less microcline than near the borders with other granites or quartzites.

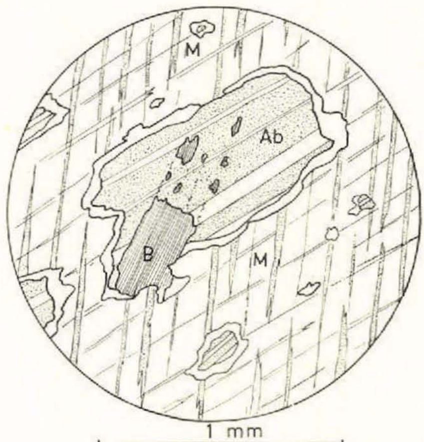


Fig. 36. Inclusion of albite-oligoclase (Ab) and biotite (B) in a large microcline crystal (M). It is evident that the plagioclase partially replaced the biotite, whilst the late microcline grew at the expense of both the plagioclase and biotite. The clear rim around the inclusion is formed by albite. Granite B near Aniggoq mountain. Slide No. 63627.

Relations between granite and country-rock

In the Arpatsivìp-qáqâ mountain range granite B forms a huge concordant sheet between volcanic rocks and quartzites. Its contacts are mostly parallel to the foliation of the country rock. The coarsely porphyritic granite contains there several large inclusions of volcanic rocks (particularly in the small area N of lake 55 m). These are all similarly oriented concordantly with the general trend of the adjoining volcanic rock layers. The inclusions do not seem to have been rotated or moved. In the Arpatsivìp area most of the contacts between granite B and the volcanic or quartzitic rocks are transitional: near the contacts with the granite the supracrustal rocks show an increasing content in potash-feldspar porphyroblasts which grow separately or in groups (fig. 37). This progressive impregnation by feldspar megacrysts is often accompanied by small-scale agmatisation.

More to the S in the Aniggoq area, the relations between granite B and host-rock are for the major part quite different: the granite appears

Fig. 37. Replacement of volcanic rock by granite B: The feldspar metacrysts have grown separately or in groups within the volcanic material. Small-scale agmatisation is visible in the lower left-corner of the photograph. Granite contact near lake 555 m, Arpatsivîp-qâqâ area.

to have intruded the quartzites and schists partly along pre-existing joints and faults. Most of the contacts are sharp and discordant to the foliation of the enclosing rocks. Moreover, several xenoliths oriented at random in the granite suggest that the latter was mobile at a certain time.

In the Aniggoq area numerous small discordant veins of granite B can be seen within the quartzites (fig. 38). The presence of moved inclusions in these veins suggests an intrusive origin.

Occasionally, the sharp discordant granite contacts are masked by the development of microcline megacrysts in the country rock. As these megacrysts are identical to those present in the granite, it seems probable that this phenomenon is due to late potash metasomatism accompanying the recrystallisation of granite B.

Genetic considerations, conclusions

The observations related in this chapter suggest that granite B was originally intruded as a melt in a restricted area, and that it expanded

Fig. 38. Discordant vein of granite B in quartzite. Small island in Tasermiut fjord near Aniggoq mountain.

from there by metasomatic replacement of the country rock. This expansion must have been important and was accompanied by a complete recrystallisation or "reactivation" of the granite. The present coarsely porphyroblastic texture was probably a result of this late recrystallisation.

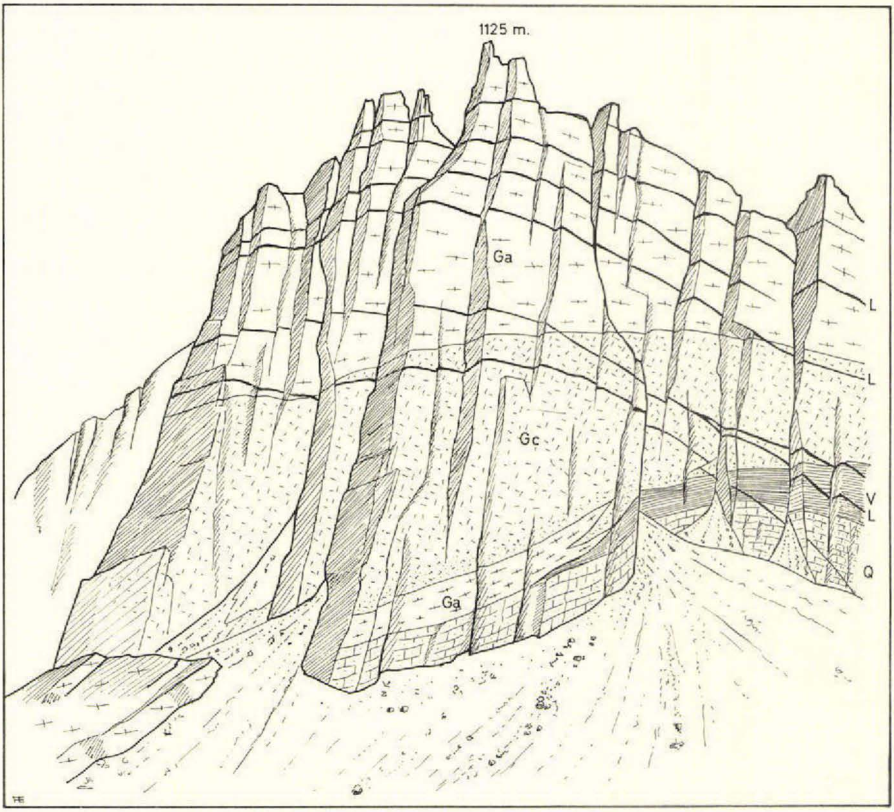


Fig. 39. View of the W wall of summit 1125 m in the Arpatsivîp-qáqâ mountain range, showing a huge sheet of granite C (Gc) lying between older granite A (Ga) and volcanic rocks (V). The whole structure is transected by numerous Gardar lamprophyre sheets (L).

XI. THE THIRD SANERUTIAN GRANITE (GRANITE C)

General character

Granite C appears in fresh outcrops as a very homogeneous light grey rock. It is distinctly more compact and less altered than the earlier Sanerutian granites. The main granite body presents a porphyroblastic texture with relatively small feldspar megacrysts (0.5–2.0 cm long). In dyke form granite C is characterised by a fine grain which gives it the aspect of a microgranite. Feldspar megacrysts occur occasionally in the fine-grained material of the dykes. The radioactivity of granite C is generally two to three times higher than that of the earlier granites.

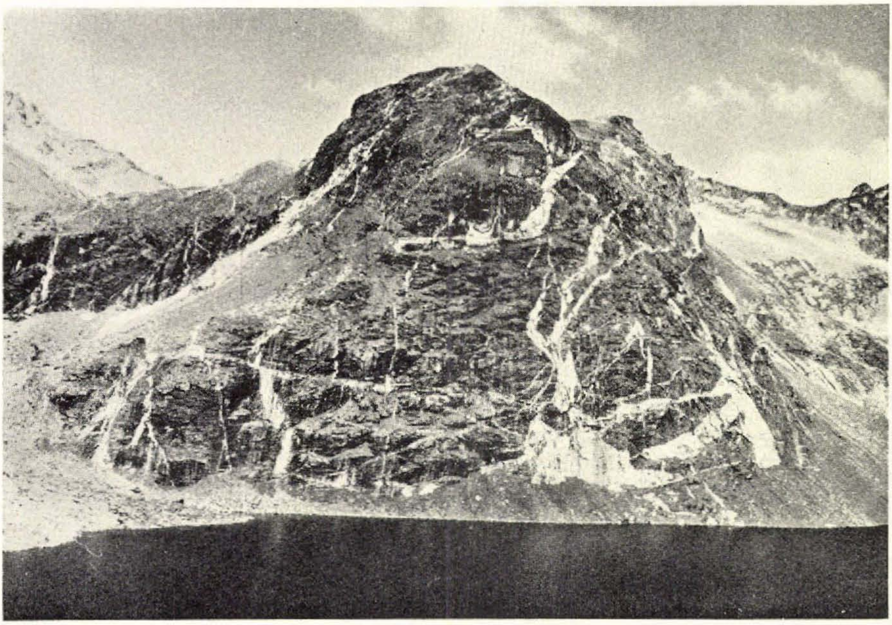


Fig. 40. Light-coloured veins of microgranite C in the volcanic rocks bordering the N side of lake 410 m.

Localisation

The third Sanerutian granite occurs mainly in the Arpatsivîp area, where it forms several mountains (see geological map on plate 5). It appears to have intruded the older granites, quartzites, and volcanic rocks. In the most NE part of the area granite C forms huge, apparently intrusive sheets between older granites and volcanic rocks. These sheets, often lying parallel with the foliation of the volcanic rocks, reach up to several hundred meters in width (fig. 39).

More to the S, in the central part of the Nanortalik peninsula, a multitude of dykes, sheets and veins of granite C transect the supracrustal series. They can often be traced for several kilometers. Their thickness varies from 5 cm to 80 m. The veins of granite C are particularly conspicuous in the dark volcanic rocks, in which they often appear as an intricate network following pre-existing shear planes and joints (fig. 40). They also often follow the limit between the earlier granite A and the overlying volcanic rocks.

Relative age determination

Granite C intrudes and transects most of the other rocks. The contact between this granite and granite B obviously cuts the feldspar megacrysts of the latter (fig. 41). Moreover, granite C contains several

Fig. 41. Contact between the light coloured granite C and the coarsely porphyroblastic granite B near Tasermiut fjord in the Arpatsivîp-qáqâ mountain area. The contact clearly transects the feldspar megacrysts.

inclusions of the coarsely porphyroblastic granite B in various stages of assimilation. As shown in fig. 39, granite C is transected by many early-Gardar lamprophyre dykes.

These observations give granite C a late Sanerutian age.

Mineral composition, structure, texture

Structure: In the main granite body, as in thick sheets and dykes, the structure is generally inequigranular, medium- to coarse-grained. In the borders, as in relatively thin veins and dykes, the granite is on the other side mostly equigranular fine-grained. An intense mafic layering can locally be observed in the sheets (fig. 42). This layering is in most cases quite discordant to the foliation of the country rock and can therefore not represent any ghost stratigraphy. It appears to be very similar to the rhytmic layering observed by Harry and Emeleus (1960) in the Tigssaluk granite sheets, SW Greenland. These

Fig. 42. Mafic layering in a microgranite sheet, in the Arpatsivîp-qàqâ mountain area.

authors ascribed the layering to the accumulation of dark minerals during the crystallisation of a granitic magma.

Texture: Although most of the macroscopic characters suggest that granite C was formed by intrusion and crystallisation of a granitic magma, nowhere have really igneous textures been seen: most of the crystals are more or less poikiloblastic and contain inclusions of one another. Apart from the microcline porphyroblasts, most of the minerals appear to have been formed more or less simultaneously by recrystallisation. The microcline megacrysts are probably the result of a still later recrystallisation. In the main granite body, as in the thick dykes and sheets, the texture is predominantly porphyroblastic. In the thinner veins it is generally fine-grained granoblastic, and often interlobate.

Principal minerals: Microcline (40 to 60 vol. $^0/_0$ of the rock), quartz, orthoclase (in the groundmass), albite-oligoclase, biotite.

Accessory minerals: Allanite, zircon, hornblende, muscovite, apatite and ore.

Alteration minerals: Sericite, chlorite.

All the large porphyroblasts are formed of a strongly perthitic microcline, which presents in many cases a simple Carlsbad twinning and euhedral crystal faces. It generally contains numerous inclusions, which are commonly oriented in the same direction and made of the same material, which shows that they are all relicts of the same mineral (fig. 43). The replacive character of the microcline is obvious in most parts of the granite.

Smaller porphyroblasts include quartz in rounded individuals, albite-oligoclase and microcline. The groundmass consists of quartz, microcline,

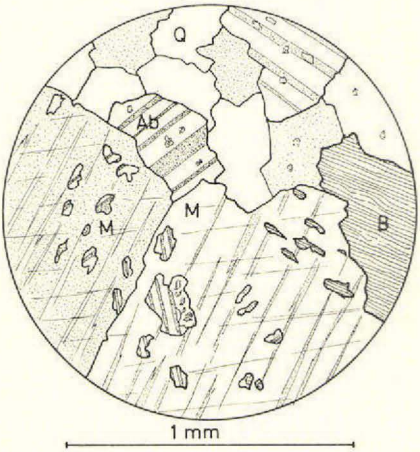


Fig. 43. Microcline (M) partially replacing biotite (B) and albite-oligoclase (Ab) in a vein of granite C. Slide No. 63642.

albite-oligoclase, orthoclase, biotite and muscovite. Micrographic intergrowths of quartz and microcline are common. A characteristic feature of granite C is the presence in quartz and feldspar of groups of similarly oriented biotite inclusions. This may mean that the granite was richer in biotite before the recrystallisation.

The composition of the granite varies only slightly from one place to another: near volcanic rocks and in dykes transecting the volcanic series, the granite appears poorer in microcline and quartz than in other places. Most large granite occurrences possess, moreover, aplitic margins in which biotite is almost absent.

Relations between granite and country-rock

Most of the granite contacts are sharp; only in the small area N of summit 1040 m, were gradational contacts between the granites A and C observed. Evidence of a slight contact metamorphism can be seen locally. The contacts are generally discordant to any foliation present

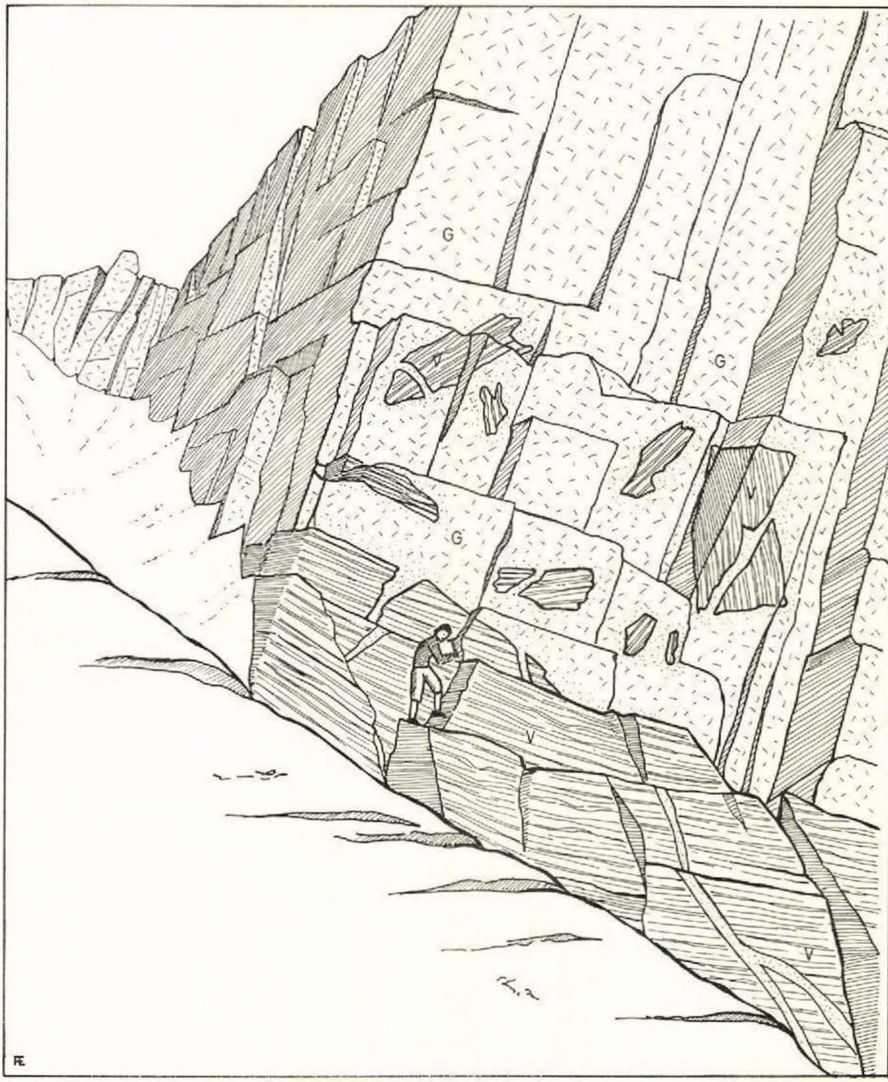


Fig. 44. North wall of mountain 1025 m in the Arpatsivîp-qâqâ area, showing moved and rotaded inclusions of volcanic rocks (V) in granite C (G) near the contact between granite and volcanics.

in the adjacent rock. The intrusive character of granite C is confirmed by the random orientation of most of the inclusions. This was for instance observed in the N wall of mountain 1025 m, where large xenoliths of volcanic rocks have been apparently moved and rotated in the granite quite near the contact with the volcanic series (fig. 44). The intrusive character of the granite-veins and dykes is also shown in several places by the presence of inclusions of material different from the host-rock, which proves that they were moved considerably.

Genesis and conclusions

The observations related in the preceding pages show that granite C was probably emplaced as an intrusive melt, and that it underwent in later times two successive recrystallisations, the last of these resulting in the formation of porphyroblasts. This seems to be the only way to explain the presence of a metamorphic texture in an intrusive granite. However, it is strange that much earlier metadolerite dykes still show clear remnants of igneous textures, whilst these are missing in all the later Sanerutian granites.

It seems likely that the relatively high radioactivity of granite C is due to a higher content in allanite and zircon.

XII. GENERAL PETROGENETIC CONSIDERATIONS AND CONCLUSIONS

The synkinematic migmatisation and granodioritisation

During the Ketilidian orogeny the rocks of the Nanortalik area were folded in three successive stages. The second and principal phase of deformation was accompanied by a strong migmatisation, giving birth to granodiorites and pegmatites. These rocks were mainly formed by replacement of pelitic schists and often the old stratification can still be detected.

The synkinematic granodiorites and pegmatites both contain much more acidic plagioclase (oligoclase) than K-feldspar, sodium being present in excess of potassium. The latter fact is confirmed by the chemical analyses of the granodiorite (table 5).

According to PETTIJOHN (1949), pelitic sediments contain as a rule more K_2O than Na_2O. One analyses of a quartzitic schist in the central part of the region also shows a higher content in potassium than sodium.

This means that during the granodioritisation of the schists Na_2O was added or K_2O was removed. MARMO (1962) and ENGEL and ENGEL (1953) consider the latter alternative more likely, and think that the extraction of potassium was responsible for the formation of synkinematic pegmatites which often occur together with the granodiotites. In the Nanortalik area all the observed synkinematic pegmatites possess, however, a characteristic granodioritic composition and could never have formed in that way. It seems therefore more likely that in the Nanortalik region the granodioritisation was mainly due to an important supply in Na_2O. It is possible that simultaneously small amounts of K_2O were removed into the country-schist, causing there the growth of late biotite.

According to many authors (TURNER 1948, MARMO 1958, ENGELHARDT 1961), the majority of the Precambrian synkinematic granitic rocks possess a granodioritic composition. This is the case in the Nanortalik area where the evolution of the schists and gneissic schists tended to such a composition during the Ketilidian orogeny. As shown by many observations (chapter V), the granodioritisation took place during an increasing regional metamorphism.

The late-kinematic pegmatites

At the end of the third and last Ketilidian folding period, many potassium-rich pegmatites were formed in the rocks of the Nanortalik peninsula. They are mainly composed of microcline and quartz and contain only small amounts of albite-oligoclase. They developed partially as replacement products of earlier sodium-rich pegmatites. The late-kinematic pegmatitisation was accompanied by a general microclinisation of the schists and gneissic schists. At the same time, a slight albitisation occurred, the oligoclase being replaced by almost pure albite.

The relations between the different minerals show that the regional metamorphism decreased in late-kinematic times. It seems thus that a lowering of the temperature accompanied the increase in the K/Na ratio of the feldspars. This may partly be explained by the experiments carried out by Orville (1961), who found that the K/Na ratio in alkali feldspars, which are in equilibrium with an NaCl vapour, increases with falling temperature.

In short, it can be said that during the end of the Ketilidian orogeny most of the rocks in the area were considerably enriched in K and Si, and, to a slight degree in Na.

The post-kinematic granites

Three post-kinematic granites occur in the Nanortalik region. Their K-feldspar is always a more or less perthitic microcline, specially so in younger granites.

The first post-kinematic granite (granite A) was mainly formed by replacement of large masses of volcanics and minor quantities of schists. This is evident by the presence in the granite of large-scale relict structures and by the nature of the contacts.

Except in a narrow border zone, the composition of granite A is remarkably constant. This shows that the homogeneity of a granite does not necessarily demand a magmatic origin. Near the impregnation contacts andesine is often present in granite A. It represents a transitory stage, in which parts of the calcium, removed from the replaced rock, were temporarily fixed. Obviously large amounts of K, Si and Na have been supplied to produce the replacement of the large masses of basic volcanic rocks by the granite. This is proved by the very different chemical compositions of the granite A and the volcanics.

The younger granites B and C appear to have been emplaced partially by the intrusion of a melt and partially by a subsequent replacement of the host-rock. This replacement is locally very important and must be responsable for approximatively 50 % of the granite B in the area.

The three post-kinematic granites all possess a more or less porphyroblastic texture with a granoblastic groundmass, even in the most obviously intrusive parts. This means that the intrusive parts recrystallised completely, probably in several successive stages, the last stage being the growth of the microcline porphyroblasts.

The fact that even the late-Sanerutian granite C recrystallised shows that its original mineral assemblage was unstable, compared to that of its host-rocks.

The microclinisation

Examination under the microscope shows that the microcline always developed directly from plagioclase, quartz or biotite within each of the three granites. It appears to be truly triclinic and never contains relicts of orthoclase. It can be assumed therefore that in the Nanortalik area the microcline was not derived from a pre-existing monocline sanidine or orthoclase. The latter origin was suggested by Guitard, Raguin and Sabatier (1960) for the microcline in the granites and gneisses of the eastern Pyrenees.

Heating experiments, carried out by Goldsmith and Laves (1954) indicate that microcline becomes unstable at higher temperatures (525° C under hydrothermal conditions) and is transformed into monocline sanidine. However, as Marmo (1958) points out, this does not prove that the inverse transition also takes place.

In the Nanortalik region all the microclines were probably formed by potassium metasomatism at a temperature below their instability point. Marmo and Permingeat (1957) suggest that the growth of most microclines was very slow an took a long time. This may also well have been the case in the Nanortalik region.

The origin of the granitising elements

According to Ramberg (1951) the granulite facies gneisses in western Greenland probably belong to the top of the degranitisation zone which must exist as compensation for the granitised upper levels. This means that part of the migmatising and granitising elements (Si, K, Na etc.) were "squeezed" out of the underlying gneisses and migrated to higher low-pressure areas. This theory is confirmed in the Nanortalik region by the following facts:

1) The granulite facies gneisses are only slightly migmatised compared with the amphibolite facies schists and gneissic schists.

2) All the replacement granodiorites and granites occur in the higher amphibolite-facies levels.

The first main element to migrate must have been sodium, which resulted in the synkinematic migmatisation and granodioritisation of the schists. This was followed in late- and post-kinematic times by considerable migrations of K, Si and Na which caused the formation of late-Ketilidian pegmatites and early-Sanerutian granites.

The evidence shows that the Sanerutian granites B and C were emplaced by intrusion. The local expansion of the "New granite" B by replacement of the surrounding rocks may be a result of the migration of granitising elements from the main "New granite" body situated on the other (SE) side of Tasermiut fjord.

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PLATES

Plate 1

Three stages of migmatisation and granitisation in the contact zone between granite A and migmatitic garnet schist near lake 430 m. A: migmatitic garnet schist. B: partly recrystallised and granitised schist. C: granite A with small schist-relicts.

Plate 2

Very simplified columnar sections at four places along Tasermiut fjord. The position of each section is given by the letters A, B, C and D on the geological map on plate 5.

Plate 3

Structural stereogram representing the lower surface of the volcanic formation (V), and two pyritic schist horizons (S), in the Quvnerssuaq and Qagdlua area. The situation of the stereogram is represented on the map on fig. 3.

Plate 4 A

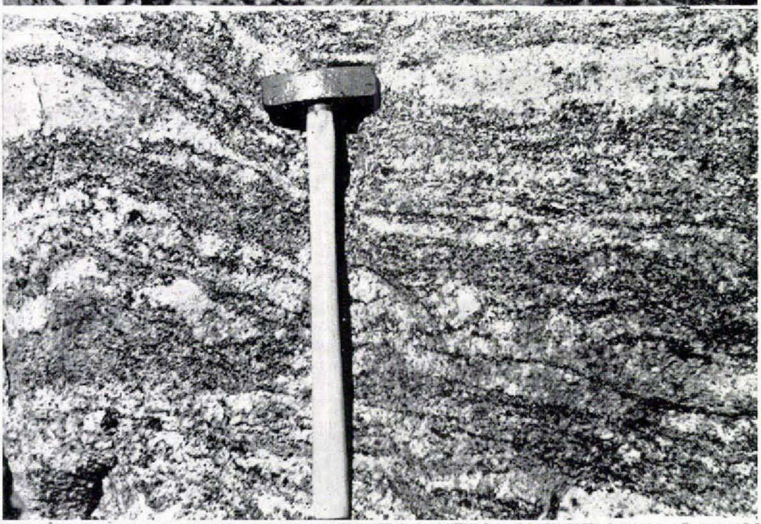
View from Tasermiut fjord of the S slope of Qagdlua mountain (1115 m). The gneissic granodiorite (G) forms the core of an anticline and is surrounded concordantly by dark biotite schist (S) containing several pyritic horizons (p). The granodiorite contains some amphibolite layers as inclusions (A). Volcanic rocks (V) and quartzitic schists (Q) overlie the pelitic schists.

Plate 4 B

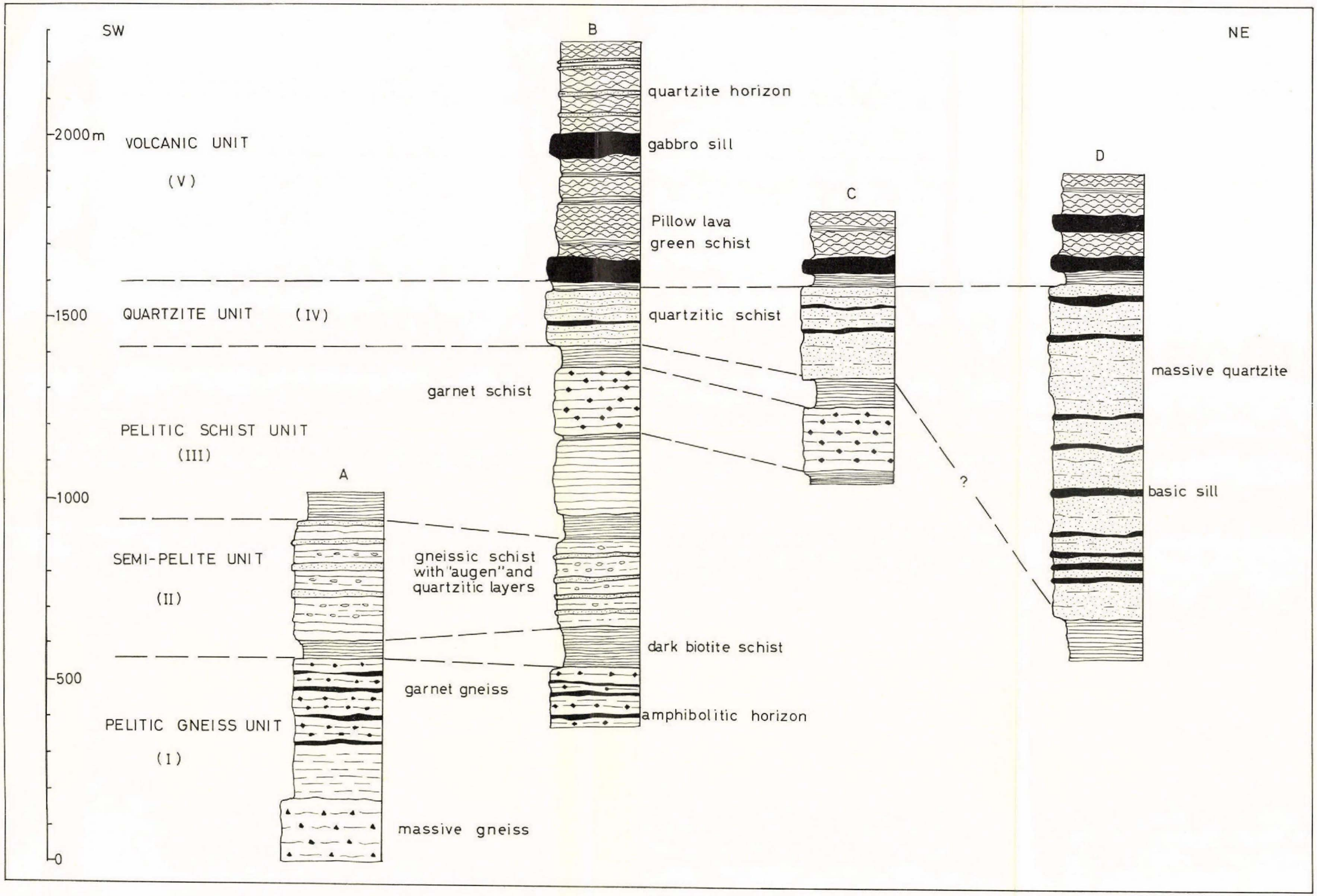
View from Sarqâ fjord of the mountains situated W of lake 510 m. The granite A (G) lies as a large concordant sheet between the schists (S) and the volcanic rocks (V). To the right, the granite forms wedges in the quartzitic schists (Q) and volcanics. Gardar dolerite dykes (D) cut the whole formation.

A

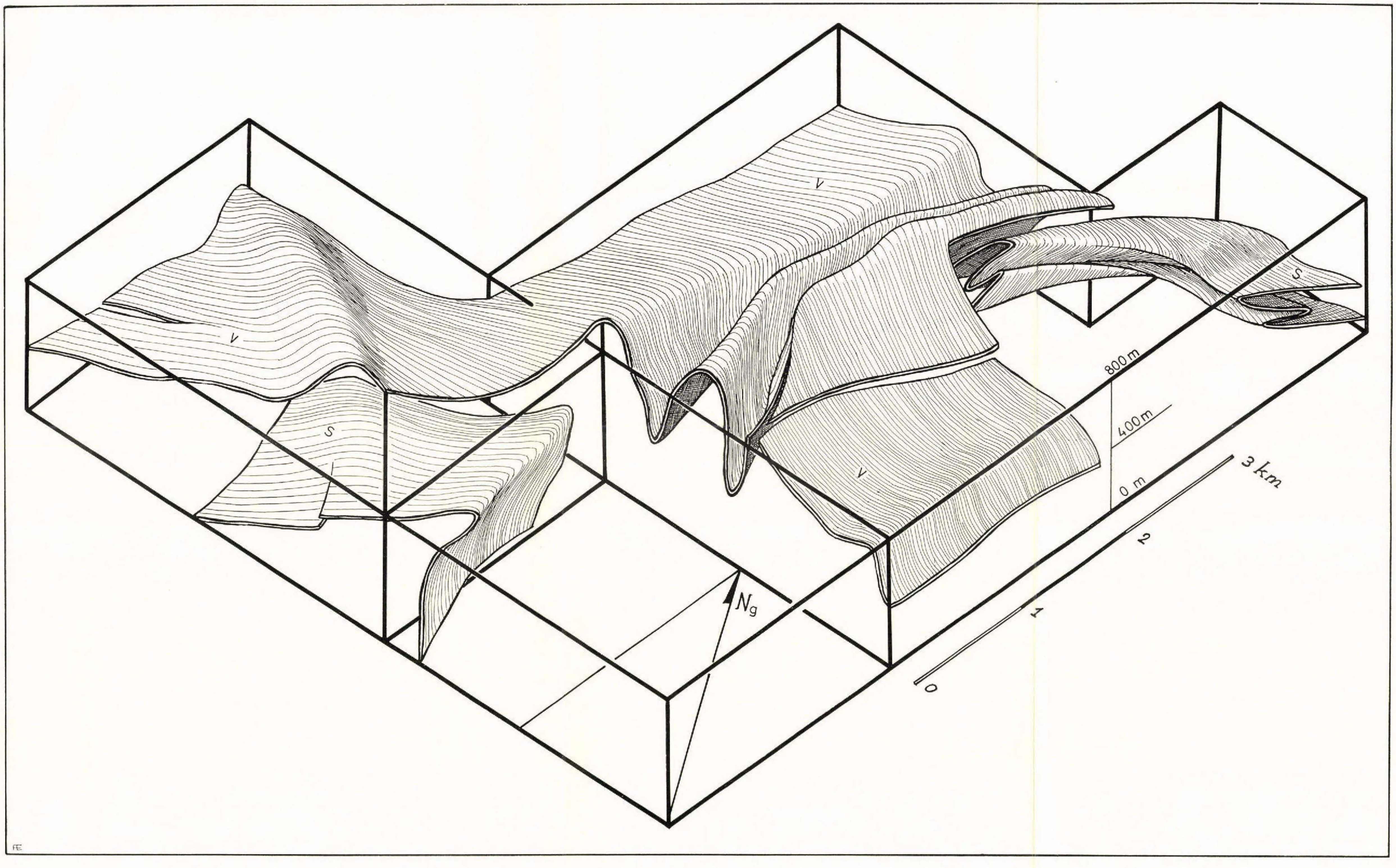
B



C



SW
NE
A
B
C
D
2000m
1500
1000
500
0
VOLCANIC UNIT
(V)
QUARTZITE UNIT (IV)
PELITIC SCHIST UNIT
(III)
SEMI-PELITE UNIT
(II)
PELITIC GNEISS UNIT
(I)
quartzite horizon
gabbro sill
Pillow lava
green schist
quartzitic schist
garnet schist
gneissic schist
with "augen" and
quartzitic layers
dark biotite schist
garnet gneiss
amphibolitic horizon
massive gneiss
massive quartzite
basic sill
?


V
V
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800 m
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1
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3 km
N_g
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PLATE 4 A

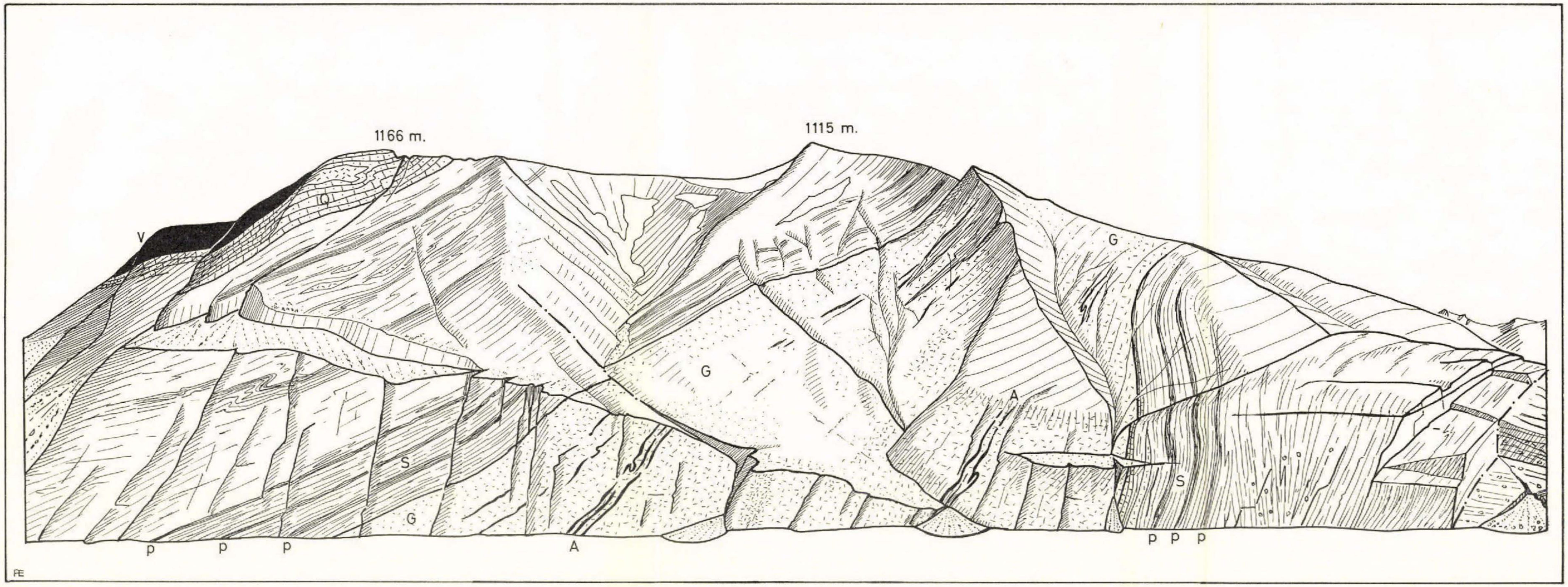
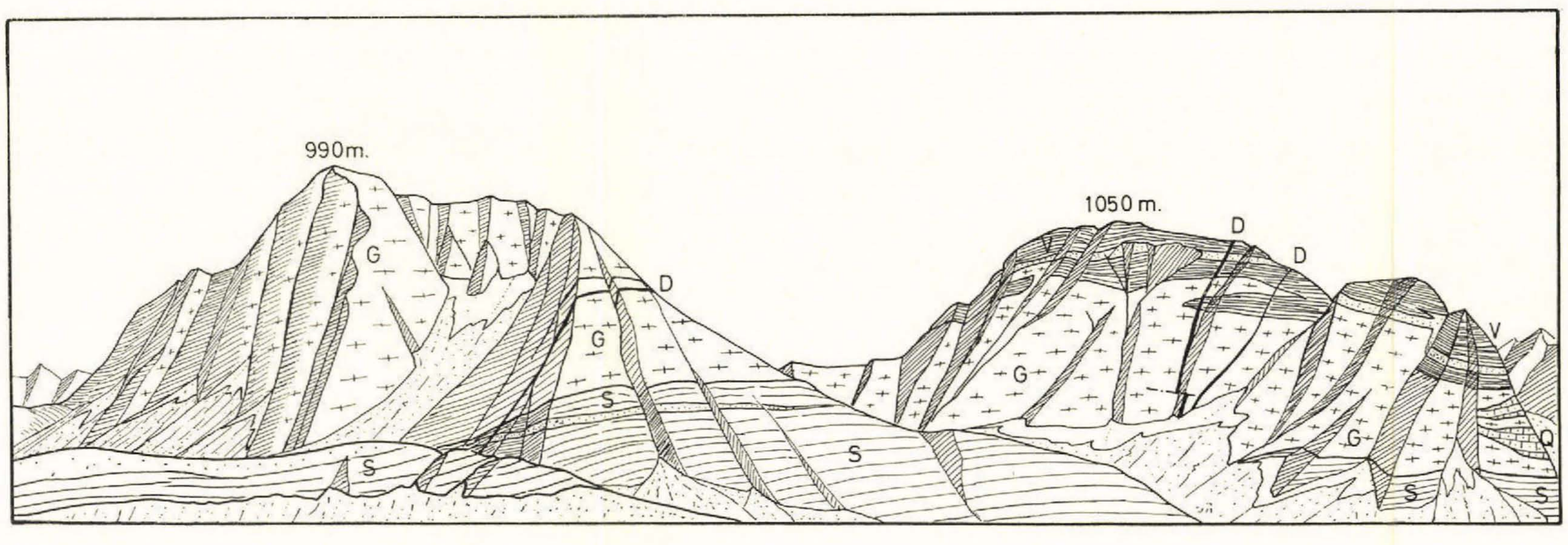


PLATE 4 B



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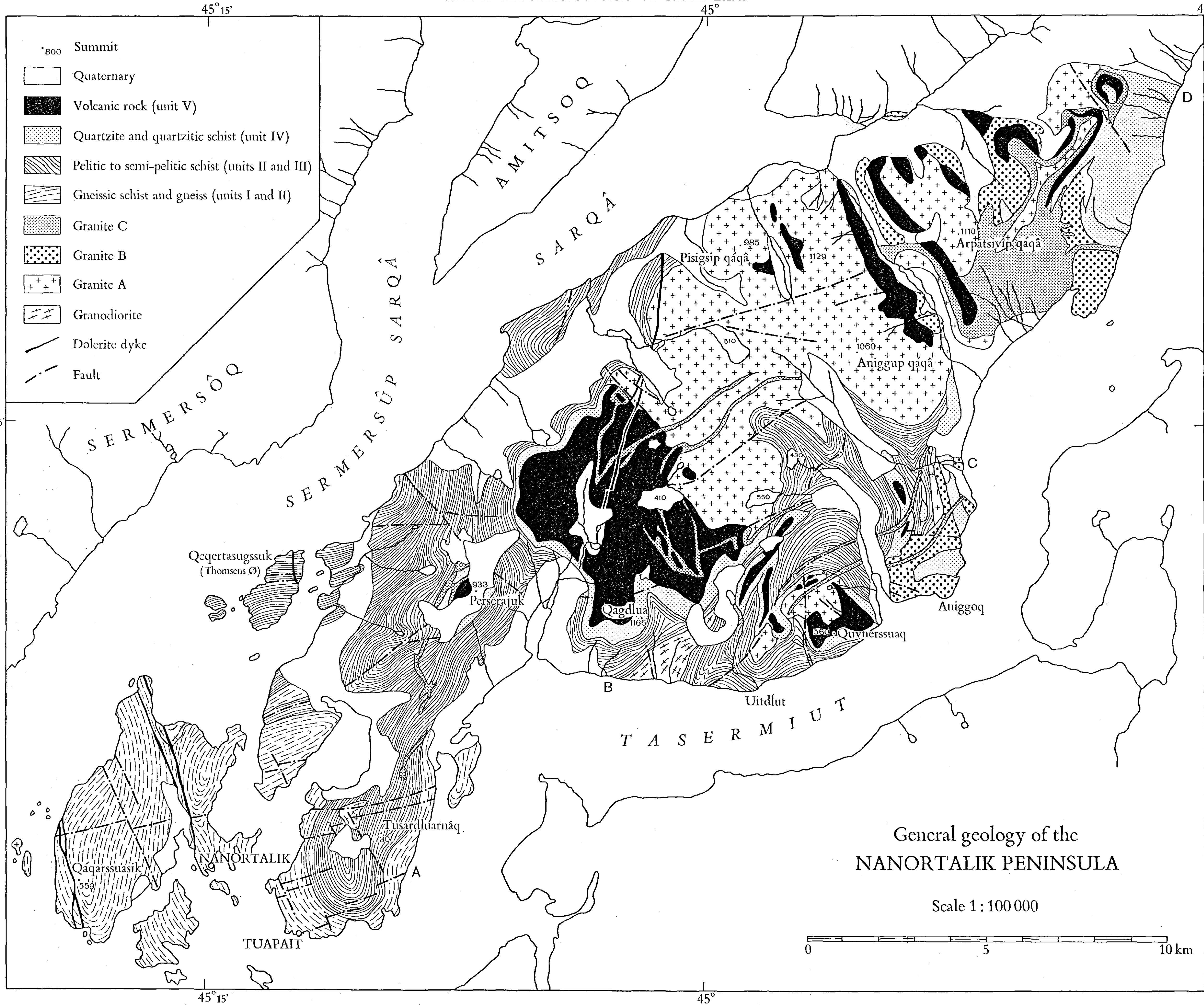
THE GEOLOGICAL SURVEY OF GREENLAND



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