DANMARKS OG GRØNLANDS GEOLOGISKE UNDERSØGELSE RAPPORT 2004/18

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Eastern Storøen seen from the east in early morning light. Note the extensive gossans.

Gossan landscape in eastern Storøen. In the background the south side of Qaqullussuit, with Akulleq supracrustals outcropping at the top of the cliff. Dark horizons in the Archaean gneiss below are amphibolite. Photo: E. Schou Jensen.

Contents

Abstract

Conspicuous gossans occur in the eastern part of Storøen, on the island Akulleq just to the east, and on the peninsula Qaqullussuit, Uummannaq district, central West Greenland. The gossans have developed by weathering of sulphide-rich layers within supracrustal rocks known collectively as the Akulleq supracrustals.

The Akkuleq supracrustals form layers up to 375 m thick. The dominant rocks in the Akulleq supracrustals are biotite-garnet gneiss/schist and biotite-sillimanite-garnet schist. In addition there are thin layers rich in pyrrhotite and graphite and also Fe silicate-rich layers (iron-formation). The latter contain grunerite and orthopyroxene (eulite), both together and as thin layers of one or other of these minerals interlayered with quartzite. On Qaqullussuit there is also banded amphibolite in close association with the Akulleq metasediments; this amphibolite most likely belongs to the supracrustals, although it could have been brought alongside the metasediments by thrusting.

The biotite-garnet gneiss/schist shows layering on a centimetre to 40 cm scale, granular layers alternating with foliated layers with a higher biotite content. This recalls layering in the metaturbidites of the Palaeoproterozoic Nûkavsak Formation farther north, and it is believed that the Akulleq metasediments are correlatives of the Nûkavsak Formation. Intense thrusting of the type demonstrated elsewhere in the region is thought to have given rise to interlayering of the Palaeoproterozoic supracrustals with gneisses of the Archaean basement.

Leucrocratic lenses and stringers in the metasediments provide evidence that a small degree of partial melting of the pelitic and semipelitic material has taken place. Further indications of the conditions of metamorphism are provided by the occurrence of sillimanite in conjunction with K-feldspar in pelitic layers in the Akulleq supracrustals, and of grunerite and eulite in the metamorphosed iron-formation. Together these pointers indicate that temperatures reached over 600°C during peak metamorphism.

24 hand samples from the Akulleq supracrustals, including 5 from small ultramafic pods in the metasediments, were analysed for As, Au, Co, Cr, Cu, Mo, Ni, Pb, V and Zn. In one sample, a garnet-rich gneiss, 0.1 ppm Au was recorded, but this determination has not been confirmed. The deviations from crustal averages for all the other elements were not large enough to suggest that economic concentrations of any of these elements exist under the ochre-rusty-coloured gossan. Five stream sediment samples, one from Storøen and four from Qaqullussuit, have been analysed for the same elements. The most significant anomalies were in two samples from Qaqullussuit which showed 10 and 18 ppb Au respectively. Since Archaean rocks greatly exceed Akulleq supracrustals in the catchment areas of the streams in question, this Au may have been derived from Archaean rocks.

Introduction

Travellers in the Uummannaq district in central West Greenland can hardly have failed to notice the striking yellow-ochre-weathering gossan zones that occur at the eastern end of the island Storøen (Greenlandic: Salliaruseq) and on the smaller island Akulleq (former spelling Akugdleq) to the east. Not only tourists but also professional geologists and prospectors have often wondered if mineral riches are concealed under the highly coloured rubble. In a short report on their field work in 1949 Rosenkrantz & Noe-Nygaard (1949) wrote (in translation from the Danish): "Samples of pyrrhotite were brought home from Karrats Isfjord and Storøen; in particular the occurrence on Storøen is thought to be of considerable size and will be examined more closely in the coming summer". No such closer examination appears to have been carried out in the following summer or at all until a party from the Canadian mining company Cominco Ltd visited the locality in 1968 (Paton 1968). Paton's description of the locality is however limited to one page and a sketch map. There is certainly no published description of the geology of eastern Storøen. For this reason the present report is timely, if not overdue. It is based on work carried out in the early 1980s as part of a master's project at the University of Copenhagen. The relevant sections of the thesis (Jørgensen 1983) have been abbreviated and edited, but only minor additions and changes have been made in the light of newer ideas and interpretations. The field work on which the thesis is based was carried out under the auspices of the Geological Survey of Greenland (GGU) with a view to preparation of the 1:100 000 map sheet 70 V.2 Agpat, published in 1987.

Setting

The Uummannaq area lies within the *c.* 1850 Ma old Rinkian orogenic belt in West Greenland (Henderson & Pulvertaft 1967, 1987; Grocott & Pulvertaft 1990). This belt can be divided into two areas based upon which rock units dominate at outcrop level. The northern area, which stretches from Maarmorilik at 71°07´N to Red Head at 75°04´N, is dominated by a several kilometres thick succession of Palaeoproterozoic supracrustal rocks, the Karrat Group (Henderson & Pulvertaft 1967). The southern area, extending southwards from Maarmorilik to Nuussuaq and beyond, is dominated by gneisses belonging to the basement complex that underlies the Karrat Group (Fig. 1).

In the northern area where the lithostratigraphy can be established, the Karrat Group has been divided into three formations, the Qeqertarssuaq and Mârmorilik Formations (lowest) and the Nûkavsak Formation. The Qeqertarssuaq and Mârmorilik Formations are regarded as correlatives deposited in different depositional sub-basins, the Qeqertarssuaq Formation north of Alfred Wegener Halvø (71°10´N) and the Mârmorilik Formation south of this.

The Qeqertarssuaq Formation varies in thickness from a few metres to more than 2 km. It consists mainly of siliciclastic rocks: quartzite, pelitic and semi-pelitic schist, and rare thin marbles. At the top of the formation, directly underlying the Nûkavsak Formation, there is an almost ubiquitous horizon of amphibolite and hornblende schist. A few ultrabasic lenses occur within the formation. In contrast, the Mârmorilik Formation consists almost entirely of dolomite and calcite marbles, with a basal unit of orthoquartzite and semipelitic schist (Garde 1978). The formation is *c.* 1.6 km thick in the Maarmorilik area, but elsewhere has been tectonically thinned, often to only a few metres. The Black Angel Zn-Pb deposit is hosted in the marbles near the top of the formation (Garde 1978; Pedersen 1980). In the Maarmorilik area the carbonates are capped by fine-grained semipelite and quartz-graphite-pyrrhotite-rich rocks which include a distictive ‘conglomerate’ consisting of rounded fragments of glassy quartz in a pyrrhotite-graphite matrix. This rock type has also been observed near the base of the Nûkavsak Formation (Pedersen & Gannicott 1980), which indicates that the Mârmorilik and Qeqertarssuaq Formations occupy the same stratigraphic position, i.e. immediately underlying the Nûkavsak Formation.

The upper and also most extensive formation of the Karrat Group is the Nûkavsak Formation, a ≥4 km thick unit built up of metagreywacke-pelitic schist couplets interpreted as turbidites (Henderson & Pulvertaft 1987). Occasional rusty-weathering graphite-pyrrhotite-rich layers provide local relief to the monotony of the formation.

Between 72°13´ and 73°11´N the Karrat Group has been intruded by the Prøven igneous complex. This consists of charnockite (the dominant component) and garnet leucogranite (Escher & Pulvertaft 1968). Hypersthene granite from the complex has yielded a Rb-Sr whole rock isochron age of 1860 ± 25 Ma (Kalsbeek 1981), providing a minimum age for the Karrat Group.

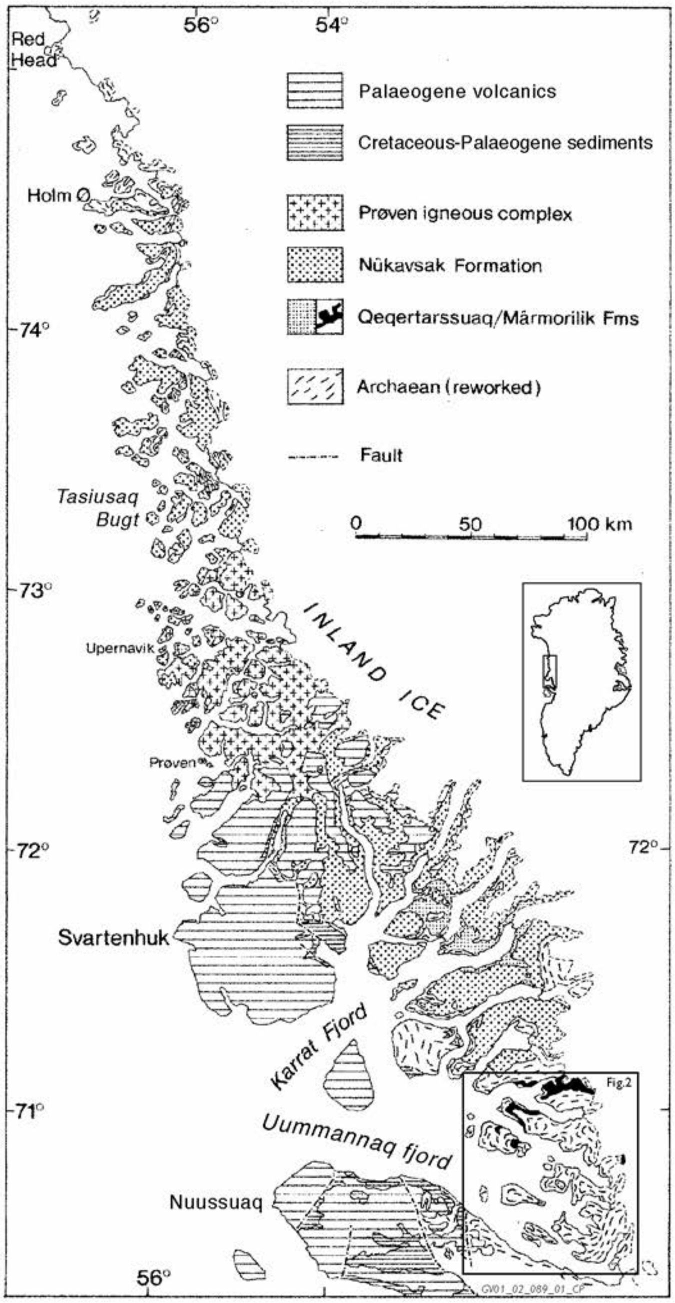

Figure 1. *Index maps showing the position of the area in relation to West Greenland and the Rinkian belt.*

The metasediments of the Karrat Group overly a gneiss-dominated basement. Locally an angular discordance is preserved between the basal quartzite of the Mârmorilik Formation and the underlying basement (Garde & Pulvertaft 1976). No obvious unconformity has yet been observed at the base of the Qeqertarssuaq Formation; instead, there is often tectonic interleaving and imposed conformity between quartzites and semipelites of this formation and basement gneiss.

In the southern area tectonic interleaving has completely taken over as the characteristic structural feature (Pulvertaft 1986; Grocott & Pulvertaft 1990), and metasediments of the Karrat Group are only present as relatively thin layers in thrust sheets and isoclinal folds in the gneiss-granite basement (Fig. 2). This is particularly well demonstrated in the case of the Mârmorilik Formation, which passing south from the type area becomes increasingly pinched and overridden by gneisses which form a major thrust nappe with a lateral extent of at least 700 km^2 (Pulvertaft 1986). Other thrust sheets can be only 170–400 m thick while having a comparable aerial extent. Similar thrusting of basement over rocks of the Karrat Group is seen in the western part of Alfred Wegener Halvø, where the Nûkavsak Formation has been reduced to horizons of biotite augen schist a few tens of metres thick below the overriding basement gneisses (Henderson & Pulvertaft 1987; Grocott & Pulvertaft 1990).

The basement in this part of the Rinkian complex is dominated by biotite gneisses of mainly tonalitic–granodioritic composition. Within these there are distinctive and often thick units of both coarse porphyritic granodiorite/augen gneiss and rather finer-grained ‘small augen’ tonalitic gneiss. Rb-Sr isochron ages obtained on these rocks range from 3087 ± 139 Ma to 2570 ± 90 Ma (Kalsbeek 1981; Andersen 1981; Andersen & Pulvertaft 1985). Although there is a discrepancy between the relative age of the dated units interpreted in the field and the relative isotopic ages (Henderson & Pulvertaft 1987), the isotopic ages are sufficient to establish the Archaean age of the gneiss basement in the Uummannaq district.

Just as in the Archaean craton to the south, anorthositic rocks occur in the basement in the Uummannaq district. These form horizons consisting of disoriented yet closely packed lenses of anorthosite and gabbro anorthosite in a granitic gneiss matrix (Andersen & Pulvertaft 1986).

Within the basement gneisses there are horizons of supracrustal rocks which can be divided into three groups: 1) horizons dominated by amphibolite; 2) horizons consisting of marble, often separated from the enclosing gneisses by a thin layer of quartzite and semipelitic schist; 3) horizons dominated by pelitic and semipelitic schist.

The amphibolite-dominated horizons often contain thin layers of garnet(-sillimanite) schist and also cordierite-anthophyllite rock, glassy quartzite containing a little green mica and sillimanite, and ultrabasic lenses occasionally with secondary spinifex-like textures (Knudsen 1980). These horizons are very like horizons known from the Archaean craton to the south, and for this reason and also from field relations are regarded as of Archaean age.

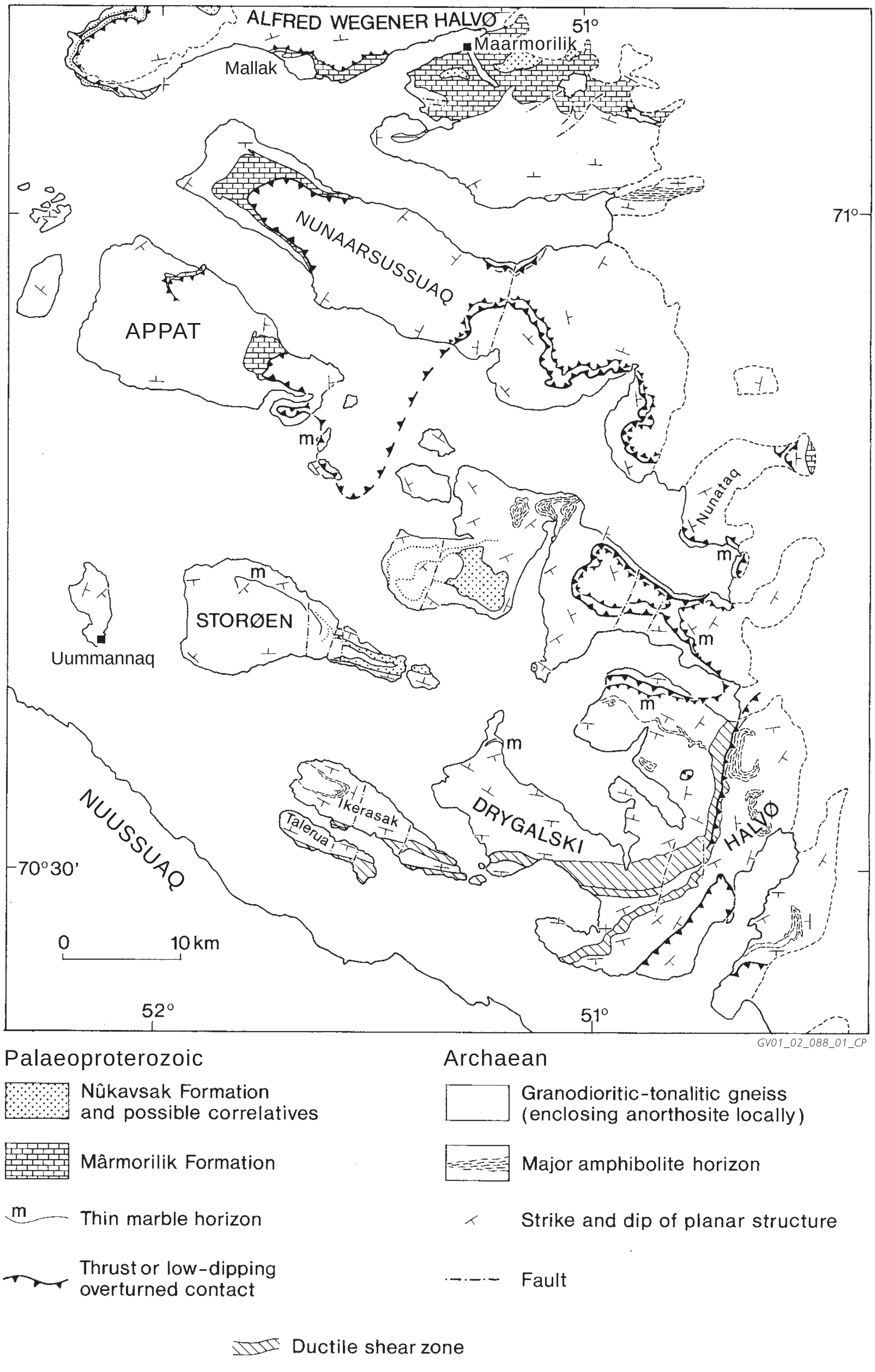

Figure 2. *Geological sketch map of the southern part of the Uummannaq area, central West Greenland. Slightly modified from Grocott & Pulvertaft (1990, fig. 8).*

The marble-dominated horizons can be correlated with the Mârmorilik Formation, and provide excellent evidence of tectonic interleaving of basement and cover in this region (Pulvertaft 1986). One such horizon occurs on the north side of Storøen (Fig. 2, Plate 1).

The horizons dominated by pelitic and semipelitic schist are the main subject of this report, since they host the precursors of the spectacular gossans mentioned earlier. For convenience they are termed the Akulleq supracrustals and are described fully in the next section.

Two groups of basic intrusions of Palaeoproterozoic age occur in the Uummannaq region. The earlier group comprises tabular, pod-shaped and irregularly shaped metabasite bodies (Schiøtte 1988). These often show primary discordance to structures in the gneisses but have also undergone disruption and deformation due to Rinkian movements. Small bodies belonging to this group occur on Storøen, Akulleq and Qaqullussuit (former spelling Qaqugdlugssuit) but are not shown in Plate 1 for reasons of scale.

The later group comprises completely undeformed dolerite dykes trending NNW–SSE (the dominant trend) and WNW–ESE. A >100 thick member of the NNW–SSE swarm has given a Rb-Sr isochron age of 1645 ± 30 Ma (Kalsbeek & Taylor 1986), which firmly establishes the minimum age of the Rinkian orogeny.

The Akulleq supracrustals and associated rocks

The informal term “Akulleq supracrustals” has been adopted as a designation for four spatially separated horizons consisting of biotite-garnet (-sillimanite) schist/paragneiss, Fe-rich metasedimentary rocks and occasional amphibolite layers. These outcrop 1) in eastern Storøen and on Akulleq; 2) in the cliffs and on the summit plateau of Qaqullussuit; 3) on the west side of Aappilattoq (former spelling Augpilagtoq; Fig. 2; Plate 1). The horizons vary in thickness from a few metres to 375 m. The horizons appear now as concordant layers in the country gneisses, except perhaps in eastern Storøen. Here the Akulleq supracrustals define an ESE–plunging syncline, with gneisses above and below. There is a certain amount of internal folding within the supracrustal unit that is not reflected in the gneisses above. This disharmony is thought to be due to differences in the ductility of the gneisses and supracrustals. It may explain why Kreuger (1928) thought that there was an unconformity on Storøen. The syncline has been displaced *c.* 650 m by a NNE–trending left-lateral fault running through the sound between Storøen and Akulleq.

South of Eqalussualik (former spelling Eqalugssualik; Plate 1) the western extension of the supracrustal layer on the northern flank of the syncline is involved in a tight–isoclinal recumbent fold closing to the north. The supracrustals in the upper limb of this fold are pinched out to the south (Plate 1). The outer closure of the fold as defined by the supracrustals lies about 1.3 km west of Eqalussualik, but is concealed by scree.

There are two horizons of Akulleq supracrustals on Qaqullussuit. The upper horizon occupies a wide area on the plateau, although extensively concealed below a boulder field, as well as being exposed in the higher parts of the cliffs all around the peninsula. The ≥350 m thick, eastern part of this horizon is the core of a recumbent isoclinal fold closing to the east, the gneiss core of which can be seen in the central part of the plateau. Neither the outer closure of the supracrustal layer nor the upper supracrustal-gneiss contact of the upper limb is exposed. Lineations and minor fold axes in the core of the major fold plunge at low angles to SSE or S. To the west the upper limb disappears above the plateau surface, while the lower limb extends as a *c.* 20 m thick layer to the top of the cliff on the south-west corner of the peninsula.

The lower horizon can be seen in the westwards- and northwards-facing cliffs. In the northwards-facing cliff it can be seen that this horizon is the core of a reclined isoclinal fold closing to the east; the orientation of the axis could not be ascertained because of inaccessibility.

The Akulleq supracrustals at the west corner of Aappilattoq form the core of a reclined tight fold closing to the east and plunging 160/30.

Largely because the Akulleq supracrustal horizons occur in three areas separated by wide fjords, it has not been possible to decide whether all the occurrences belong to a single layer repeated by tight to isoclinal folding or they occur in separate thrust sheets. Jørgensen (1983) has shown alternative models of structures that link the horizons in different ways, but was unable to choose decisively between the models.

The rock types that make up the Akulleq supracrustals can be divided into the following groups:

1) Biotite-garnet gneiss/schist
2) Biotite-sillimanite-garnet schist
3) Graphite schist
4) Massive sulphide
5) Quartz-Fe silicate rocks
6) Ultramafic rocks
7) Amphibolite

Where outcrop and scale allow, these groups have been mapped separately. On the southern flank of the syncline in eastern Storøen it seems that the graphite schist, massive sulphide, and quartz-Fe silicate rocks are grouped together at three levels, situated at the top, base and roughly the middle of the Akulleq supracrustal horizon. However, on the northern flank of the syncline, folding of the interlayered sulphide-graphite-rich rocks and biotite-garnet gneiss/schist has complicated the outcrop pattern. Added to this complication is the general problem that weathering of the sulphide-graphite-rich rocks has given rise to gravel which has crept down-slope and obscured contacts in many places (see frontispiece, p.2). For this reason an accurate map of the distribution of pyrrhotite-graphite-rich rocks could not be made in the time available, and these rocks are not distinguished everywhere in Plate 1.

The different rock types will now be described in turn before metamorphism and possible origins are discussed.

Biotite-garnet gneiss/schist

Biotite-garnet gneiss/schist is the dominant lithology in the Akulleq supracrustals. The texture of the rock varies from granular and rather fine-grained to coarser and foliated. It is often layered on a centimetre to 40 cm scale, granular layers alternating with foliated layers with a higher biotite content. Biotite, garnet, quartz and feldspar can be distinguished in hand sample, garnets up to a centimetre or more in diameter being particularly conspicuous. Within the biotite-garnet gneiss there are leucocratic layers up to 10 cm thick that show pinch-and swell structure and also concordant pegmatites. The latter contain fluorapatite, tourmaline and biotite, in addition to quartz and microcline.

In thin section the main minerals in the garnet-biotite gneiss/schist are quartz, plagioclase An_{c30}, red-brown biotite, garnet and microcline. Minor constituents are graphite, chlorite, hornblende, sericite, apatite, tourmaline, zircon, opaque phases and rare muscovite. Garnet forms porphyroblasts. Two types can be distinguished according to the size and distribution of inclusions in the grains. In the one type the inclusions – chiefly quartz, biotite and opaque phases – are about the same size as grains in the enclosing matrix. In the other type the inclusions are very small and confined to the centre of the porphyroblast, the rim being almost inclusion-free. In the first type the inclusions define a weak fabric parallel to the foliation in the enclosing matrix, while in the latter type the fabric defined by the inclusions is often oblique to that in the matrix. Fractures in both types of garnet are normal to the foliation in the enclosing rock.

Biotite-sillimanite-garnet schist

This rock type is seen particularly in horizons separating biotite-garnet gneiss/schist from graphite-sulphide-rich rocks. Sillimanite is conspicuous, both as fibrolite sheaves and as needles up to a millimetre thick and 5 cm long. In many places the parallel orientation of the larger prisms defines a lineation, in particular in eastern Storøen and on Akulleq where the lineation is oriented WNW–ESE, parallel to the axis of folding here. Within the schist there are often coarse leucocratic lenses and layers up to a few centimetres thick as well as pegmatite veins like those in the biotite-garnet gneiss/schist. Sillimanite occurs locally in these pegmatites.

The component minerals of this schist are quartz, biotite, sillimanite, garnet and microcline, together with small amounts of plagioclase and opaque material including graphite.

The biotite is a red-brown variety that has small pleocroic haloes. It shows some degree of orientation parallel to the foliation, but is not deformed.

Sillimanite occurs in all sizes from fine fibrolite up to robust needles. The fibrolite and smaller needles form sheaves and mats parallel to the foliation and running through quartz and microcline or wrapping around garnets. There are also seams of fine-grained, granulated quartz with embedded sillimanite needles. Some of the sillimanite is intergrown with or adjacent to biotite, and tiny sillimanite needles can also occur embedded in quartz, biotite or microcline, completely isolated from one another.

Microcline is often slightly perthitic, and is also sericitised in patches.

The leucocratic lenses and layers are coarse-grained and dominated by microcline and quartz. Plagioclase is present, but only in very small amounts, and there are also rare, small, ragged flakes of muscovite. Some sericitisation of both microcline and plagioclase has taken place.

Graphite schist

Graphite schist occurs in layers up to 2 m thick interlayered with biotite-sillimanite-garnet schist, quartzitic rocks and massive sulphide. It is a dark foliated rock, often containing much biotite. In thin section the main minerals are seen to be (approximate percentages in brackets): quartz (45%), brown biotite, partially altered to chlorite (20%), graphite (10%), feldspar and sericite (20%), accessories, mainly opaque phases (5%). In polished section examined in reflected light the following minerals have been observed: graphite (10%), pyrrhotite (3%), chalcopyrite (+), marcasite (+), rutile (+) and sphalerite (+). Pyrrhotite forms single crystals or composite grains <0.5 mm in size that are often oriented parallel to the foliation. Chalcopyrite is mostly found in contact with pyrrhotite. Marcasite occurs along fractures in pyrrhotite and seems to have formed by alteration of pyrrhotite.

Massive sulphide

Massive sulphide, mainly pyrrhotite, occurs in layers and pods up to 0.5 m in thickness and in veinlets in graphite schist. Often the mode of occurrence is impossible to determine because the rock weathers readily to a thick cover of gossan. The host rocks can be graphite schist, quartzitic rocks or biotite-sillimanite-garnet schist. In hand sample the massive sulphide is a dense, dirty (due to graphite) rock with lustrous surfaces where the sulphide is fresh.

Samples of the massive sulphide have been examined in polished sections and the following minerals have been observed: pyrrhotite (60–90%), pyrite (0–5%), marcasite (2–10%), graphite (5–30%), chalcopyrite (+), rutile (+), sphalerite (+), molybdenite (+).

The pyrrhotite forms grains up to 2 cm in size, with smooth grain boundaries and some development of subgrains. Locally smaller grain size is accompanied by the development of foam texture. Along grain boundaries pyrrhotite has often been altered to marcasite.

Pyrite occurs as idiomorphic porphyroblasts in contact with or as inclusions in pyrrhotite. Inclusions, mainly of graphite, define an internal fabric indicating post-kinematic growth of pyrite.

Graphite occurs in flakes from 1 mm to 1 cm in size. The flakes are frequently bent or kinked.

Rutile, sphalerite and chalcopyrite occur in grains less than 3 mm across. Chalcopyrite is also found as inclusions in sphalerite.

Within the massive sulphide there are sometimes minor domains <3 cm in size that are rich in graphite and silicates. Quartz and grunerite are the main silicates in these domains, while sercite, chlorite and apatite are present in smaller amounts.

Quartz-Fe silicate rocks

A spectrum of rocks ranging from almost pure quartzite to grunerite-layered quartzite or quartz-layered orthopyroxenite occurs in association with biotite-sillimanite-garnet schist and graphite schist.

Quartzite forms layers up to a metre in thickness. These are coarse-grained, although fine-grained quartz is found along boundaries between large grains. Apart from quartz, any of the following can occur in very minor quantities in the rock: biotite, muscovite, grunerite, graphite and other opaque minerals including pyrite.

Grunerite-(Fe sulphide)-layered quartzite occurs in layers up to a metre thick. These are associated with both quartzite and quartz-layered orthopyroxenite, biotite-sillimanite-garnet schist, graphite schist and massive sulphide.

The rock is medium-grained and shows layering on a centimetre scale expressed by paler and darker colours that reflect different concentrations of minerals. In thin section the minerals that have been observed are: quartz, grunerite, opaque phases, orthopyroxene, apatite, together with very small amounts of biotite, chlorite and sericite. Chlorite and sericite have formed by replacement of biotite. The pale layers consist of quartz (*c.* 65%) and grunerite (*c.* 35%). Most dark layers are made up of *c.* 85% quartz, Fe sulphides (*c.* 10%) and apatite (*c.* 5%). However, some dark layers are composed largely of orthopyroxene (up to 80%), with quartz, grunerite and pyrrhotite making up the remainder of the rock.

Quartz forms grains <0.5 cm thick that are elongated parallel to the layering. The large grains are often free of undulose extinction. Smaller grains are also commonly elongated. Grain boundaries can be both smooth and serrated. The serrated boundaries are the sites of smaller grains.

Grunerite (2V=90°; c^z=14°; Fe/Fe+Mg *c.* 72%) forms elongated, often twinned, grains with a weakly developed preferred orientation parallel to the layering.

Orthopyroxene (eulite) occurs in layers *c.* 0.5 cm thick. It contains inclusions of apatite, quartz and opaque minerals that define an internal fabric parallel to the layering. Apatite is also found as inclusions in quartz and grunerite.

The grunerite-(Fe sulphide)-layered quartzite has been examined in polished section. Pyrrhotite and pyrite are the main opaque phases. Pyrrhotite forms single grains up to a millimetre in size that are only rarely altered to marcasite. Pyrite occurs associated with pyrrhotite against which it shows well developed crystal faces, while it often pseudomorphs silicates. The other opaque minerals that have been observed in very small quantities are chalcopyrite, arsenopyrite (rare), sphalerite and graphite.

Quartz-layered orthopyroxenite forms layers or lenses up to a few metres thick that occur together with biotite-sillimanite-garnet schist/gneiss, graphite schist, quartzite and grunerite-layered quartzite. It is a coarse-grained, almost black rock with a layering defined by layers and lenses of quartz. Pyroxene, quartz, graphite and pyrrhotite can be identified in hand sample. The rock has a high density.

In thin section the following minerals have been identified, the volume percentages varying greatly from section to section: orthopyroxene (40–75%), grunerite (2–30%), quartz (2–20%), opaque phases (2–5%), biotite (2%), apatite (2%), garnet (0–2%), graphite (≤1%), hornblende (0%–+), sericite (+) and chlorite (+).

Orthopyroxene has $2V\alpha=70–74°$, $\Delta n=0.017$ and D=3.7, indicating a composition Fe/Fe+Mg *c.* 75% (eulite). Two types of eulite are found. One type forms large (≤3 cm), often elongate grains containing numerous inclusions and showing undulose extinction and deformation bands. The other type shows no undulose extinction and is often free of inclusions. Where inclusions do occur, they are very small and often define a ghost structure.

Grunerite replaces or is intergrown with eulite. It forms mm-long grains with twinning, 2V=90° and c^z=14°, indicating composition Fe/Fe+Mg *c.* 72%.

Quartz occurs mainly in layers where the grain size is up to 2 cm. It also forms smaller grains along the boundaries of eulite grains and inclusions in eulite. In quartz-rich layers quartz shows foam texture, inclusion inhibition and secondary grain migration.

Garnet occurs in small elongate grains that tend to be clustered together in domains.

Biotite occurs as undeformed grains up to 3 mm in size. Where biotite occurs in clusters it can exhibit decussate texture.

Graphite occurs as <1 cm long flakes penetrating quartz layers and eulite grains. It also occurs along boundaries between other grains.

Apatite is found as strings of inclusions in eulite.

In polished sections the following opaque minerals have been identified:
Pyrrhotite: grains up to a centimetre in size, often concentrated in zones parallel to the layering in the rock. The pyrrhotite is often altered to marcasite along fractures.
Magnetite: grains ≤ 0.2 mm, in some samples rimming pyrrhotite grains.
Chalcopyrite: grains ≤ 0.2 mm, commonly associated with pyrrhotite.
Arsenopyrite: found in only one sample as < 5 mm idiomorphic grains.
Graphite: flakes up a centimetre long, often concentrated in layers parallel to the layering in the rock.

Another Fe silicate-bearing rock type that rarely occurs consists of plagioclase (strongly altered to muscovite, sericite, and small patches of carbonate), quartz, orthopyroxene, clinopyroxene, pale brownish-green clinoamphibole, sphene and apatite. The pyroxenes are intergrown with or rimmed by amphibole in many grains. Sphene is often partially replaced by opaque material.

Ultramafic rocks

Ultramafic rocks occur as lenticular pods in sizes up to 20 x 80 m. These are particularly numerous in the Akulleq supracrustals on the southern flank of the syncline in eastern Storøen and have also been observed on Akulleq and Qaqullussuit; they have not been shown on Plate 1 for reasons of scale. The pods tend to be arranged in rows parallel to the overall layering in the supracrustals. This arrangement and the shapes of the pods suggest that they represent boudinaged, formerly planar, bodies.

The ultramafic rocks are coarse-grained and structureless. In hand sample pyroxenes, tremolite and pyrrhotite can be seen, while in thin section the minerals recorded and their approximate volume percentages are: clinopyroxene (+–10%), orthopyroxene (5–30%), olivine (5–10%), tremolite (50–80%) and opaque phases (5%).

Both clinopyroxene and orthopyroxene occur in grains up to 2 cm in size with inclusions of olivine and tremolite. The smaller inclusions (≤ 0.2 mm) are elongated and define a planar internal fabric in the host pyroxene grains. The grain boundaries of the pyroxenes are smooth and the grains are often rimmed by small tremolite grains.

Olivine is found as up to 5 mm long grains of irregular shape.

Tremolite occurs in grains up to 5 mm long that contain opaque inclusions and show undulose extinction. Tremolite is also found as fine-grained aggregates (grain size ≤ 0.3 mm) with foam texture; opaque inclusions are rare in these aggregates.

In polished section the opaque minerals have been identified are chromite, pyrrhotite, chalcopyrite, pentlandite and graphite. Of these chalcopyrite and pentlandite occur as inclusions in pyrrhotite.

Amphibolite

The upper of the two horizons of Akulleq supracrustals on Qaqullussuit is remarkable in that it is associated with horizons of layered amphibolite up to tens of metres thick. The amphibolite occurs both within the supracrustals (Fig. 3) and separating metasedimentary Akulleq supracrustals from orthogneiss horizons that are interleaved with the supracrustals in this area. The contacts with both the biotite-garnet gneiss and the mineralised rocks of the Akulleq supracrustals are often folded (Fig. 3).

The amphibolite horizons associated with the Akulleq supracrustals on Qaqullussuit are for the most part layered rocks. The layering is due to variations in mineral composition, both with regard to component minerals and the proportions in which these minerals occur. The grain size is medium to coarse. Parallel orientation of biotite flakes and hornblende grains impart a foliation on the rocks.

Figure 3. *Part of the cliff on the south side of Qaqullussuit, showing the relation of amphibolite (darkest rocks) to the Akulleq metasediments. The axes of the folds run at a narrow angle to the cliff face, so that the folds appear more extreme than they are in true profile. The pale rocks in the lower part of the photo are reworked Archaean gneisses. The base of the supracrustal unit on the right of the photo is at about 700 m a.s.l, and the top of the cliff at c. 880 m a.s.l.*

The chief component minerals in the amphibolite are blue-green hornblende and plagioclase An_{45-50}. These can be accompanied by small amounts of any of the following: brown biotite, diopside, garnet, quartz, sericite replacing plagioclase, chlorite replacing biotite, and sphene; chalcopyrite, pyrite, ilmenite, malachite and rutile also occur sporadically in very small amounts.

Diopside tends to be restricted to domains rich in plagioclase. The diopside grains contain small inclusions of hornblende with crystallographic axes parallel to the host diopside.

Garnet occurs in porphyroblasts that have an inclusion-rich core rimmed by more massive garnet. The internal fabric defined by inclusions is not always parallel to the fabric of the host rock.

Petrographically the amphibolite horizons associated with the Akulleq supracrustals are very similar to amphibolite occuring in horizons within the regional orthogneisses. In the field, however, there is one obvious difference, and that is in the amount of veining by quartzo-feldspathic material as both layers, schlieren and cross-cutting veins. This leucocratic veining is conspicuous in the amphibolite horizons within the regional orthogneisses but poorly developed in the amphibolites associated with the Akulleq supracrustals on Qaqullussuit.

Metamorphism

An immediate impression of the grade of metamorphism is provided by the presence of substantial amounts of sillimanite together with biotite, garnet and microcline in many of the pelitic schist layers in the Akulleq supracrustals. Sillimanite also occurs in pegmatites. The questions that arise are these: to just how high a grade of metamorphism the rocks have been subjected, and what reactions have given rise to the occurrence of sillimanite?

According to the most widely favoured P-T diagram for the aluminium silicate system, that of Holdaway (1971; see Spear 1993, p. 293), the minimum temperature for the first appearance of sillimanite in regionally metamorphosed rocks is 500°C at a pressure of 3.76 kb. This is the threshold for the first sillimanite zone in the Barrovian series. However, where sillimanite is associated with K-feldspar without the presence of muscovite or relics of other aluminium silicates or of staurolite, as is the case in the Akulleq supracrustals, the rocks have entered the second, higher-temperature sillimanite zone (Chinner 1966a, b; Spear 1993, p. 303). In the second sillimanite zone there is frequently evidence of partial melting of the rocks (Chinner *op. cit.*), just as there is in the Akulleq supracrustals in which there are lenses and stringers consisting mainly of microcline and quartz. This evidence of partial melting in the Akulleq supracrustals is a further indication that the rocks have been metamorphosed under second sillimanite zone conditions and also a further indication of the physical conditions governing this metamorphism. The minimum temperature of partial melting of pelitic rocks depends on both the precise composition of the rock and the degree of water saturation in the system. In the intermediate facies series melting can be expected to begin at temperatures between 600° and 650°C, while dry melting will not begin until the temperature is about 100C° higher (Thompson 1982).

The association of sillimanite with K-feldspar in pelitic schists is generally taken as being the result of the breakdown of muscovite (Chinner 1966b; Yardley 1989), thus:

$$\text{muscovite + quartz} \rightarrow \text{sillimanite + K-feldspar + } H_2O \text{ (or melt)}$$

In the Akulleq rocks in which partial melting has taken place, metamorphism has proceeded belond the point when muscovite could have survived; at least no muscovite relics have been observed in association with sillimanite, and the amount of muscovite that has been observed is negligible. Instead it is with biotite that the sillimanite is often associated. This relationship is common in metamorphic rocks (Chinner 1966b; Evans & Guidotti 1966, p.32; Shelley 1968). The replacement of biotite by sillimanite requires the exclusion of iron and magnesium; these elements could have been taken up by garnet and melt in a continuous prograde reaction.

An indication of PT conditions at peak metamorphism is also provided by mineral assemblages in the quartz-Fe silicate rocks. Evidence that these rocks are of sedimentary origin is provided by their layered structure, close association with typical pelitic metasediments and graphite-rich schists, and the common occurrence of graphite and apatite in the rocks themselves.

The following stable mineral parageneses have been observed in the quartz-Fe silicate rocks:

1) *Grunerite-quartz*-apatite-pyrrhotite
2) *Grunerite-quartz*-apatite-hornblende
3) *Eulite-quartz*-biotite-hornblende-grunerite-apatite-graphite
4) *Eulite-quartz*-biotite-hornblende-garnet-pyrrhotite-graphite
5) *Eulite-quartz*-hornblende-apatite-graphite.

As indicated, the most important constituents of the rocks are grunerite, eulite and quartz. The mineral parageneses indicate a high content of Si and Fe in the parent rocks, and low contents of K, Na and Al, i.e. that we are dealing with an iron-formation. Although no primary minerals or textures have been observed in the quartz-Fe silicate rocks, it seems most likely that the components of the parent sediments were similar to those found in unmetamorphosed or almost unmetamorphosed iron-formations in Precambrian terrains elsewhere in the world. Apart from Fe oxides, the minerals most common in these formations are quartz as chert, siderite, greenalite, minnesotaite and stilpnomelane. Of these, minnesotaite is usually interpreted as a metamorphic mineral, but its formation can also take place during late diagenesis (Klein 1978, 1983). Reactions that can form minnesotaite include (Klein 1983):

$$\underset{\text{greenalite}}{Fe_6Si_4O_{10}(OH)_8} + \underset{\text{chert}}{4\,SiO_2} \rightarrow \underset{\text{minnesotaite}}{2\,Fe_3Si_4O_{10}(OH)_2} + 2\,H_2O$$

$$\underset{\text{siderite}}{FeCO_3} + \underset{\text{chert}}{4\,SiO_2} + H_2O \rightarrow \underset{\text{minnesotaite}}{2\,Fe_3Si_4O_{10}(OH)_2} + 3\,CO_2$$

$$..\ \underset{\text{stilpnomelane-like}}{Fe_{2.7}(Si,Al)_4(O,OH)_{12}\bullet x\,H_2O} + 0.33\,Fe^{2+} \rightarrow \underset{\text{minnesotaite}}{Fe_3Si_4O_{10}(OH)_2} + H_2O + Al + \text{minor (Na,K)}$$

The formation of grunerite and eulite can take place via the following reactions (Klein 1983):

$$\underset{\text{siderite}}{8\,(Fe,Mg)CO_3} + \underset{\text{chert}}{8\,SiO_2} + H_2O \rightarrow \underset{\text{grunerite}}{(Fe,Mg)_7Si_8O_{22}(OH)_2} + 7\,CO_2$$

$$\underset{\text{minnesotaite}}{7\,Fe_3Si_4O_{10}(OH)_2} \rightarrow \underset{\text{grunerite}}{3\,(Fe,Mg)_7Si_8O_{22}(OH)_2} + 4\,SiO_2 + 4\,H_2O$$

$$\underset{\text{siderite}}{(Fe,Mg)CO_3} + \underset{\text{chert}}{SiO_2} \rightarrow \underset{\text{orthopyroxene}}{(Fe,Mg)SiO_3} + CO_2$$

$$\underset{\text{grunerite}}{(Fe,Mg)_7Si_8O_{22}(OH)_2} \rightarrow \underset{\text{orthopyroxene}}{7\,(Fe,Mg)SiO_3} + SiO_2 + H_2O$$

The first occurrences of grunerite and orthopyroxene have been used to characterise zones in metamorphosed iron-formations. The appearance of grunerite takes place under conditions corresponding to the upper biotite zone (Klein 1978, 1983) or garnet zone (James 1955) in pelitic schists, corresponding to upper greenschist/lower amphibolite facies, while the occurrence of orthopyroxene indicates a higher grade of metamorphism (e.g. James 1955, table 6; Kranck 1961; Butler 1969; Floran & Papike 1978; Klein 1983, fig. 11-16; Miyano 1987). Grunerite can however persist into the orthopyroxene zone and the two minerals can coexist in this zone.

Although studies of metamorphism in iron-formations are mostly of assemblages that include magnetite, there is sufficient agreement that the occurrence of orthopyroxene in these rocks requires temperatures at peak metamorphism in excess of 600°C (Miyano & Klein 1983, fig. 6, reproduced here as Fig. 4; note that, according to this figure, the formation of orthopyroxene + grunerite assemblages can place under a wide range of pressures). Thus there is consistency between the metamorphic conditions suggested by mineral parageneses in the sillimanite-bearing pelitic schist and by parageneses in the quartz-Fe silicate rocks, both of which indicate second sillimanite zone conditions. There are however two features that require comment.

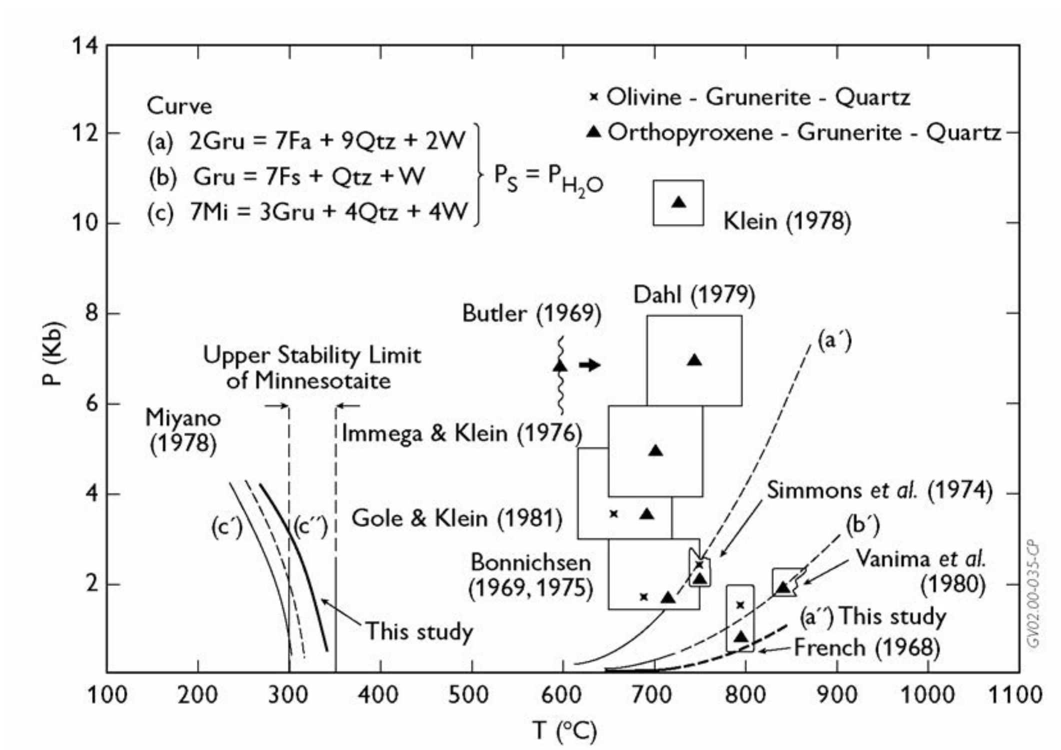

Figure 4. *Published metamorphic P-T conditions for various high-grade metamorphosed iron-formations (from Miyano & Klein 1983). Curves a´, a´´, b´, c´ and c´´ are equivalent to reactions (a), (b) and (c) respectively. Curve a´´ represents the apparent upper stability limit for grunerite, whereas curve c´´ is the adjusted upper stability limit for minnesotaite.*

Firstly, the Fe/Fe + Mg ratios of the grunerite and eulite occurring in the Akulleq supracrustals that have been determined from optical properties are approximately the same. This contrasts with the findings of Kranck (1961) and Butler (1969), and the plot shown in Miyano & Klein (1983, fig. 2C), all of whom indicate that orthopyroxene has a higher Fe/Fe + Mg ratio that coexisting grunerite. However, the optical data are only accurate to within a few percent Fe/ Fe + Mg, so no further comment is justified.

Secondly, in the Akulleq supracrustals, grunerite-layered quartzite and eulite-layered quartzite occur within layers up to a few metres thick that can be adjacent to one another. The temperature and pressure governing metamorphism in these layers cannot have been different, so some other factor(s) must have determined which Fe-rich silicate was formed. These can be 1) compositional difference between the layers in the parent rocks, and 2) differences in volatile activity in the different layers. These factors may have been interdependent.

Regarding the first factor, it is noticable that the grunerite-rich layers do not contain more than about 35% of the Fe-rich silicate, while eulite-rich layers contain up to 80% eulite. This suggests that the eulite-rich layers formed from a sediment rich in Fe and poor in silica (chert). An obvious possibility is that these layers were rich in siderite, which would mean that during metamorphism the partial pressure of CO_2 in the rock was higher than in the parent of the grunerite-rich layers. In the latter the grunerite could have been derived from hydrous minerals such as greenalite and minnesotaite; in the reactions leading to the formation of grunerite from these minerals water is released, so that the partial pressure of H_2O in the grunerite-rich layers was much higher than in the eulite-rich layers where CO_2 was the dominant volatile. According to Bonnichsen (1975, p. 334): “If the total pressure was approximately equal to the sum of the partial pressures of H_2O and CO_2 during metamorphism, it follows that the formation of pyroxene, rather than amphibole, was favored by the presence of a vapour phase relatively enriched in CO_2.”

Mobilisation of sulphides

As already described, massive sulphide, mainly pyrrhotite, occurs in layers and pods up to 0.5 m in thickness and in veinlets in graphite schist. This can hardly have been the mode of occurrence of the sulphide in the original sediments, implying that the sulphide was mobilised and migrated during metamorphism and deformation. Furthermore, the original, premetamorphic sulphide was probably pyrite, since pyrite gives way to pyrrhotite at higher grades of metamorphism (Thompson 1972; Spear 1993, p. 506).

Three mechanisms have been suggested for the mobilisation and migration of sulphides: 1) partial melting, 2) solid state flow, and 3) fluid state mobilisation.

Partial melting: Lawrence (1967) proposed that this process could take place during high grade metamorphism (temperature above 700°C). Vokes (1971) suggested that differential melting of sulphides may be initiated at lower temperatures, melting of copper sulphides taking place as soon as the temperature exceeds 500°C, while the melting of Cu-poor, Fe-rich sulphides requires rather higher temperatures.

Solid state flow: This mechanism has been proposed by Solomon (1965), McDonald (1967) and Pedersen (1980). It involves the separation of minerals due to differences in plasticity and mechanical mobility. Thus under high strain conditions minerals with a low strain resistance will migrate to more favourable sites, leaving minerals of high strain resistance behind.

Fluid state mobilisation: This is a process by which migrating pore fluids differentially take up minerals in solution in one area and redeposit them in another. The driving force is generally believed to be disequilibrium in stress or pressure (pressure solution) (Durney 1972). In non-porous rocks transfer is believed to have been effected chiefly by diffusion of dissolved ions in an aqueous intergranular ‘dispersed phase’ or ‘solution film’. In porous rocks ions in solution can be transported in a migrating intergranular fluid.

The massive, pyrrhotite-rich sulphide in the Akulleq supracrustals occurs in highly deformed rocks that have undergone high grade regional metamorphism at temperatures in excess of 600°C. Under these conditions any of the suggested mechanisms for the mobilisation of sulphides could have been operative. If partial melting has taken place, the pyrrhotite could be the residue left after the sulphides of e.g. Pb, Zn and Cu have been removed in the melt. However, apart from the Black Angel Zn-Pb deposit and other, smaller Zn-Pb concentrations in the Mârmorilik Formation, no base metal sulphide deposits have yet been found in the Uummannaq Fjord area.

Origin and possible correlatives of the Akulleq supracrustals

It is evident from the foregoing descriptions that the Akulleq supracrustals are totally recrystallised, highly deformed metamorphic rocks, so that the character of their parent rocks is a matter of conjecture. No isotopic studies have been carried out on the rocks, so nothing certain can be said about their age other than that they are Palaeoproterozoic or older.

The mineralogy of the pelitic gneiss and schist, in particular the occurrence of sillimanite, indicates that their parent rocks were greywackes and mudstones. The rocks are layered on a scale that is suggestive of turbidites.

The origin of the quartz-Fe silicate rocks has already been discussed, and it has been concluded that these were originally iron-rich sediments composed mainly of chert, silicates such as greenalite and minnesotaite, and siderite. The graphite schist and sulphide-rich rocks were deposited in a dysoxic environment.

The amphibolite layers resemble layered amphibolites in other areas that have been proved to be strongly deformed, regionally metamorphosed, basic pillow lavas and pyroclastics (e.g. Ehlers 1976; Myers 1978). The question here is whether the amphibolites associated with the Akulleq metasediments on Qaqullussuit belong to the Akulleq supracrustals or have been brought alongside and interleaved with the metasediments by thrusting. Before pursuing this further, the possible origin and age of the Akulleq supracrustals has to be discussed.

The questions to be addressed are these: are the Akulleq supracrustals Archaean or Palaeoproterozoic, and if Palaeoproterozoic, what are their most likely correlatives?

Although there is as yet no proven answer to these questions, the favoured answers are that the Akulleq supracrustals are Palaeoproterozoic in age, and that they can be correlated with rocks in the Karrat Group which are extensive and well preserved in the northern part of the Rinkian orogenic belt (Henderson & Pulvertaft 1987). The reasons for favouring these interpretations are set out in the following.

The fact that the Akulleq supracrustals occur as essentially concordant layers in an Archaean gneiss complex does not constitute evidence either way. It has been amply demonstrated that there was extensive thrusting in the southern part of the Uummannaq area during the Rinkian orogeny. This led to the stacking-up of thrust sheets consisting of thin layers of Karrat Group metasediments overlain by much thicker sheets of basement gneiss. The most conspicuous example of this thrusting involves Mârmorilik Formation marble which can be traced southwards from the type locality into concordant layers no more than a few metres thick in thrust sheets in the basement gneiss (Pulvertaft 1986). One such marble layer occurs on the north-east facing slopes of Storøen. Another has been involved in a spectacular recumbent isoclinal fold seen on a cliff on the north-west side of Dryglaski Halvø (see frontispiece in *Tectonics*, vol. 6, no. 4, August 1987). Farther north, on Alfred Wegener Halvø, basement gneiss has been thrust over thin layers of biotite schist that were derived from the flysch turbidites of the Nûkavsak Formation (Henderson & Pulvertaft 1987; Grocott & Pulvertaft 1990).

As already described, the interlayering of semipelitic and pelitic gneiss/schist on <1 m scale in the Akulleq supracrustals recalls layering in turbidites. Metaturbidites consisting of metagreywacke interlayered with pelitic schist completely dominate the Nûkavsak Formation in the Palaeoproterozoic Karrat Group farther north, providing an obvious candidate for a correlative of the Akulleq supracrustals.

Another similarity between the Akulleq supracrustals and the Nûkavsak Formation is the occurrence in both of rusty-ochre-weathering horizons rich in graphite and pyrrhotite. In the Nûkavsak Formation between 71° and 72°30´N Thomassen (1992) has reported both rusty-weathering horizons containing a few percent disseminated pyrrhotite and cherty layers rich in pyrrhotite and graphite, the latter often containing semi-massive pyrrhotite-graphite breccias.

Stream sediment analyses (see later section) also provide a hint that the Akulleq supracrustals are correlatives of the Nûkavsak Formation, in that, in the Uummannaq area, the only two samples outside the outcrop area of the Nûkavsak Formation that showed As values over 16 ppm were 510330 and 501334. These were collected at the foot of the two streams with proportionally the most extensive outcrop of Akulleq supracrustals in their catchment areas (Steenfelt *et al.* 1998, map 12).

Regarding the banded amphibolites occurring within and alongside Akulleq supracrustals on Qaqullussuit, the fact that it has been demonstrated in the Uummannaq area that lithologies of very different ages and origins have been brought together in concordant layers raises the possibility that these amphibolite layers are in fact of Archaean age and that the interleaving of these with the Akulleq metasediments is the result of thrusting. There is at present no good evidence either way concerning the age of these amphibolites. As already mentioned, the reason that they have been described together with Akulleq supracrustals per definition is that, unlike typical Archaean amphibolite horizons in the area, the amphibolites associated with Akulleq metasediments are never agmatitic and contain far less leucocratic material in the form of layers and schlieren than the Archaean amphibolites.

If one accepts that the Akulleq biotite gneiss/schist is a correlative of the Nûkavsak Formation in the Karrat Group and that the amphibolites associated with Akulleq metasediments belong to the Akulleq supracrustals, the latter must also belong to the Karrat Group and in that case are most likely correlatives of the amphibolites and hornblende schists at the top of the Qeqertarssuaq Formation, immediately underlying the Nûkavsak Formation.

At this point one can go even further, drop all inhibitions and begin to fantasise on the depositional setting of the quartz-Fe silicate rocks and sulphide-graphite-rich layers.

If the banded amphibolite in the Akulleq supracrustals on Qaqullussuit can be correlated with the amphibolite + hornblende schist horizon at the top of the Qeqertarssuaq Formation, then the depositional environment of the iron-rich rocks that lie alongside the amphibolite on Qaqullussuit resembles closely that of the iron-formations along the north-west flank of the Palaeoproterozoic Animikie Basin in Minnesota, USA (Morey 1983). These iron-formations occur at the transition between shelf quartzites and deeper water greywackes and slates, which matches the position of the starved basin deposits between the shelf and rift Qeqertarssuaq/Mârmorilik Formations and the deeper water flysch turbidites of the Nûkavsak Formation.

As for the source of the iron in the Akulleq iron-formation, this could be either the volcanic rocks at the top of the Qeqertarssuaq Formation or the hinterland of the sedimentary basin. No meaningful guess can be made on this question.

Economic geology

Gossan

The sulphide-rich rocks in the Akulleq supracrustals have undergone weathering, which has lead to the development of a thick crust of reddish-brown or ochre-coloured material. This crust has in many places disintegrated to similarly coloured gravel. The term used in mining circles for such reddish-brown or ochre-coloured weathering zones is *gossan*.

Gossan develops when sulphide deposits with a large content of pyrite or pyrrhotite are subjected to oxidising conditions in the presence of water. The depth to which gossans form is dependent on many factors such as climate, fracturing of the rocks, and depth to the water table.

Gossan consists mainly of hydrous iron oxides often collectively referred to as limonite. The commonest minerals constituting limonite are geothite (αFeO.OH), lepidocrocite (γFeO.OH) and turgite (haematite (αFe_2O_3) with adsorbed water) (Deer *et al.* 1963).

The gossans seen on eastern Storøen and on Akulleq are the most conspicuous in the entire Uummannaq region. They have formed by weathering of the pyrrhotite-graphite-rich rocks described in a foregoing section. When pyrrhotite- and pyrite-rich rocks are exposed to the atmosphere, the iron in these minerals combines with oxygen and water to form limonite, while the sulphur combines with oxygen and enters into solution as SO_4^{2-} (sulphate) ions.

The occurrence of gossan has led to the discovery of many important ore bodies. This is because pyrite and pyrrhotite can be accompanied by sulphides of many other metals such as Cu, Zn, Pb and Ag (see e.g. Edwards & Atkinson (1986) for examples). In Greenland it was the observation of a large gossan that lead to the discovery of the large Zn-Pb deposit at Citronen Fjord in North Greenland (van der Stijl & Mosher 1998). With this in mind, both rock samples and stream sediment samples have been analysed for selected elements. The results of analyses of elements considered most relevant in the present context are presented in Tables 1 and 2 and discussed in the following.

Geochemical studies

Nineteen rock samples from the Akulleq supracrustals and five samples from ultramafic pods in the supracrustals were analysed for the elements As, Co, Cr, Cu, Mo, Ni, Pb, V and Zn; the iron-rich metasediment samples were also analysed for Au. Furthermore, five stream sediment samples have been analysed for these and many other elements.

Hand samples

All samples were analysed by Atomic Absorption Spectrometry (AAS) using a Perkin Elmer 460 spectrometer. Two types of decomposition procedure were used. All samples were subjected to total decomposition, and three were subjected to a selective decomposition of sulphides. After decomposition the iron-rich metasediment samples were prepared for Au analysis by extraction of Au by dibutyl sulphide.

The results of the whole rock and sulphide analyses are shown in Table 1, and the values for the same group of elements in the upper continental crust and Archaean bulk crust are added for comparison (McLennan & Taylor 1999). Note that, because of differences in the dilution factors used in preparation of the samples, detection limits for As, Mo and Pb are not the same for all analyses. From an economic point-of-view, the results are not encouraging and call for very little comment.

Table 1. *Trace metal analyses of selected samples from the Akulleq supracrustals. All values in ppm. Crustal averages from McLennan & Taylor (1999).*

Sample number	Rock type	As	Au	Co in sulph.	Co in rock	Cr	Cu in sulph.	Cu in rock	Mo	Ni in sulph.	Ni in rock	Pb	V	Zn
249624	Amphibolite	<25		18	58	615	94	95	<5	14	175	<5	230	50
249715	"	<25		25		34	239	225	<5	8	98	<5	300	110
249610	Bi-gt gneiss/schist	<25			80	109		15	<5		113	15	100	60
249623	"	<25			63	97		35	<5		78	20	105	50
249625	"	<25			53	127		25	<5		63	25	100	90
249609	"	<25			178	7		30	<5		13	<5	15	6
249694	"	<25			55	191		10	<5		120	5	130	90
249693	Garnet-rich gneiss	<10	0.1	87	98	665	38	308	4	173	725	<2	215	100
249652	Massive sulphide	<10	<0.02		13	150		350	131		2500	8	655	490
249708	"	<10	<0.02		53	105		750	31		1200	4	330	700
249709	"	<10	<0.02		78	67		625	49		2300	<2	245	60
249679	Qz-layered opx'ite	<10	<0.02		15	16		230	2		1000	<2	55	610
249702	"	<10	<0.02		18	16		18	<2		1225	<2	40	53
249703	"	360	<0.02		25	23		35	2		1225	2	85	90
249690	Grun-layered qzt	<10	<0.02		88	97		400	7		550	<2	305	220
249691	"	<10	<0.02		160	33		165	4		220	4	75	180
249692	"	<10	<0.02		103	53		155	9		215	<2	160	120
249707	"	<10	<0.02		188	13		5	<2		25	2	25	33
249720	"	<10	<0.02		28	19		105	7		183	<2	40	140
249626	Ultramafic pod	<25			60	170		13	<5		215	30	90	50
249644	"	<25			65	253		43	10		180	<5	495	120
249645	"	<25			110	1345		25	5		725	<5	250	120
249678	"	<25			110	1985		200	<5		1725	<5	185	80
249688	"	<25			108	2120		118	<5		1425	<5	180	70
	Upper cont. crust	1.5	0.0018		10	35		25	1.5		20	20	60	71
	Archaean bulk crust				30	230		80			130		245	

As: Only one sample showed a detectable amount of As, even though arsenopyrite was observed in some of the polished sections.

Au: Only one sample showed Au above the detection level of 0.02 ppm; in this sample the value recorded was 0.1 ppm. No repeat run was carried out to confirm this analysis.

Co: Co values range from 13 to 188 ppm, highest in the grunerite-layered quartzite and ultramafic pods, but without any consistent relation to rock type.

Cr: As might be expected, Cr values are conspicuously higher in some of the ultramafic samples than in the metasediments or amphibolites, the highest value recorded being 2120 ppm.

Cu: Maximum Cu contents were recorded in the samples of massive sulphide. However, the maximum value of 750 ppm does not suggest the proximity of an economically viable Cu deposit. The Cu values in the amphibolite samples are disappointing considering that malachite stains have been observed on weathered outcrops of amphibolite.

Mo: Mo values are very low except in the massive sulphide samples from which a maximum value of 131 ppm was obtained. This is consistent with the observation of molybdenite in a polished section, but not enough to warrant any further exploration for this metal. Since Mo tends to be concentrated in mudstones deposited in oxygen-poor environments (Krejci-Graf 1972; Buchardt *et al.* 1997), higher values might has been expected in graphite-bearing samples.

Ni: Ni concentrations are highest in massive sulphide and in the ultramafic samples that also show the highest Cr values. The quartz-layered orthopyroxenites also have a relatively high Ni content; in contrast, these have the lowest Cr values. In the amphibolites and garnet-rich gneiss, only a small part of the Ni occurs in the sulphides.

Pb: Pb values are very low throughout.

V: The highest vanadium contents are found in massive sulphide and ultramafic pods, a maximum values of 655 ppm having been recorded in a massive sulphide sample. However, high values have also been recorded in the garnet-rich gneiss and in one sample of grunerite-layered quartzite. Vanadium is often concentrated in mudstones deposited in anoxic or dysoxic environments, so higher values might have been expected in graphite-bearing samples. In the stream sediments samples the highest vanadium values were recorded in samples showing the greatest influence of the Akulleq supracrustals (see later).

Zn: Zn contents tend to be highest in the massive sulphide and quartz-Fe silicate rocks, but a maximum value of 700 ppm is probably not enough to attract the mining industry.

In addition to the analyses carried out by the first-named author, four samples of massive sulphide from eastern Storøen were analysed by Cominco Ltd. The results indicated Ni values from 370 to 500 ppm, and 140 to 500 ppm Cu (Paton 1968).

Stream sediments

Samples of stream sediments were collected at the foot of five streams that cross outcrops of the Akulleq supracrustals. The samples were collected during a regional geochemical study of the region (Thomassen 1992; Steenfelt *et al.* 1998). The procedure adopted in collecting the samples and the methods used in subsequent analysis are described in Steenfelt *et al.* (1998), and the locations of the sampling sites are shown in Plate 1. The results for a selection of the elements determined are shown in Table 2.

Table 2. *Contents of selected metals in stream sediments from eastern Storøen and Qaqullussuit. Values in ppm where nothing else indicated. Sample locations shown on Plate 1. Data from Steenfelt et al. (1998).*

Stream sed. sample nr	As	Au ppb	Ba	Co	Cr	Cu	Mo	Ni	Pb	U	V	Zn	Fe2O3 %
501330	18	0	500	6	180	97	12	22	24	7.5	217	125	21.12
501331	10	10	420	13	210	109	6	60	19	4.8	186	140	16.38
501332	0	4	570	17	140	35	0	52	15	5.8	95	66	6.62
501333	4	18	410	17	130	63	5	57	22	5.0	126	107	8.47
501334	43	0		9	150	149	21	34	20	6.6	191	153	19.45
Regional mean	2.64	10.68	436.8	32.63	439.5	114.5	1.68	144.7	20.14	6.13	187.6	117.6	9.76

From the geological map Plate 1 it can be seen that the stream sediment samples 501330 and 501334 were collected at the foot of the streams that cross the widest area of outcrop of Akulleq supracrustals compared to reworked Archaean basement. These samples also show the highest Fe_2O_3 values. Sample 501331 can be expected to be influenced by material from the Archaean amphibolite horizon outcropping in the cliff above the sampling site as well as from the Akulleq metasediment horizon higher up. Ultramafic pods cannot have had any influence on 501330, and are not thought to be present in the catchment areas of the streams draining Qaqullussuit even though it cannot be excluded that there are small pods concealed under the boulder field on the plateau.

The As values greater than 16 ppm in the stream sediments have already been commented on. These are much higher than in average upper continental crust and in stream samples collected in areas of Archaean gneiss outcrop, but are in keeping with values recorded in the extensive areas of Nûkavsak Formation outcrops farther north.

Two stream samples show Au enrichment. It may be significant that Au values are highest in samples least expected to be influenced by Akulleq supracrustals, but two-digit ppb values can reflect no more than random single grains in the samples.

Co in the stream samples is lower than in most of the hand samples but the same as in continental crust. Cr is a little higher in the stream sediments than might be expected when compared to the hand samples, although no higher than in Archaean bulk crust. Cu, V and Zn are highest in samples with the greatest input from Akulleq supracrustals (501330 and

501334) or from Archaean amphibolite (501331). Mo is distinctly higher in samples influenced by Akulleq supracrustals, as might be expected, given that there is often Mo enrichment in ‘black shales’ (Krejci-Graf 1972; Buchardt *et al.* 1997). The same is true of Cu, Ni and V.

Evaluation of economic potential

Seen as a whole, the results of the geochemical investigations do not suggest that there are economically viable concentrations of base metals in the Akulleq supracrustals. Furthermore, the mineralogy of the associated quartz-Fe silicate rocks indicates that the original sediments constituted a normal iron-formation consisting of sulphide, carbonate and silicate facies deposits, in which there is not likely to have been any significant enrichment in base metals. The high graphite content in some layers indicates deposition in an environment with negative redox potential, but it does not seem that the concentration of any of the metals commonly enriched in this type of environment has been sufficient to warrant further attention.

There remains Au. The highest Au values were recorded in samples collected from steams draining Qaqullussuit but, as can be seen from a study of Plate 1, the Au in these samples is more likely to have been derived from Archaean gneiss or amphibolite than from the Akulleq supracrustals. Higher Au values were recorded in many samples collected at localities within the area of outcrop of the Karrat Goup farther north (Steenfelt *et al.*1998, map 13).

Dansk resumé

De fleste, der har sejlet i Uummannaq Fjord i det centrale Vestgrønland, har sikkert bemærket den kraftige rusten-gule farve, der præger skråningerne på øen Akulleq og den østlige del af Storøen. Såvel turister som professionelle minegeologer har spekuleret over, om der kunne være værdifulde mineralforekomster skjult under de stærkt farvede klippevægge og det løse grus. Området er blevet undersøgt, desværre med skuffende resultater. Nærværende rapport er en beskrivelse af lokaliteternes geologi og de lag, der er ophav til de stærke farver.

Rusten-gule bjergartsoverflader som dem, der ses på Storøen og Akulleq, betegnes *gossans* af fagfolk. Gossans opstår, når bjergarter med et stort indhold af jern-svovl mineraler (f.eks. pyrit, FeS_2, også kendt som svovlkis, og pyrrhotit, FeS, også kendt som magnetkis) udsættes for forvitring under tilstedeværelse af vand og ilt. I denne situation frigives svovl, der forbindes med ilt, og tilsammen fjernes som SO_4^{2-} (sulfat), som opløses i vand; vandet bliver faktisk til svag svovlsyre! Derimod indgår jernet i forbindelse med ilt og vand for at danne uopløselige hydrerede jern-oxider, som fagfolk kollektivt betegner limonit, i dagligsprog rust og okker. Det faktum, at forekomster af f.eks. kobberkis, zinkblende eller blyglans tit indeholder pyrit eller magnetkis udover de mere værdifulde mineraler, er årsagen til, at tilstedeværelsen af gossans ofte har ført geologer til opdagelsen af betydelige malmforekomster, senest i Grønlands sammenhæng zink-bly-forekomsten ved Citronen Fjord i Nordgrønland.

På Storøen og Akulleq er magnetkis det dominerende jern-svovl mineral. Det er koncentreret i lag indtil en meter tykke, hvor det typisk ledsages af grafit. Sideløbende med magnetkis-grafit-rige lag findes lag af kvartsit vekslende med tynde lag rige på jern-silikater (jern-kisel forbindelser). Meget små mængder af zinkblende og kobberkis findes spredt i de jernrige lag. De jern- og grafit-rige lag menes at være blevet aflejret på havbunden under iltfattige forhold.

Værtsbjergarten til disse jern-rige lag er en ca. 350 meter tyk biotit-granat skifer formation, som ligger som et tykt lag inden for de grå gnejser, der dominerer landskabet i det sydlige Uummannaq-område (biotit er en mørk-farvet jern-magnesium-holdig glimmer). Disse skifre var oprindeligt ler-holdige aflejringer, der blev afsat oven på gnejs-grundfjeldet. Tilsvarende bjergarter danner plateauet på toppen af Qaqullussuit og ses også ved Aappilattoq, øst for Akulleq. De anses som tilhørende den samme skifer-enhed, der dominerer den nordlige del af Uummannaq-området og ses især i de høje brunlige klinter, der flankerer de indre dele af Kangerluarsuk og Kangerlussuaq-fjordene. Man er nået frem til, at under den kraftige bjergkædedannelse, der fandt sted i den nordlige del af Vestgrønland for ca. 1.800 millioner år siden, blev store skiver af gnejserne skubbet oven på skifrene, således at relativt tynde flager af skifer nu forekommer som lag klemt mellem gnejserne. På tilsvarende vis er tynde lag af hvid Maarmorilik marmor, som oprindeligt lå ovenpå gnejserne, havnet som tynde lag klemt i gnejserne (ses f.eks. som et foldet lag på sydøstsiden af østenden af Appat).

Den kraftige deformation, der fandt sted under bjergkædedannelsen, førte også til dannelse af folder i alle størrelser, blandt andet den VNV–ØSØ-orienterede trug-formede struktur (synklinal), hvori biotit-granat skifer formationen er bøjet i den østlige ende af Storøen og også ombøjningszonen ved Aappilattoq. Samtidigt skete der en opvarmning af skifrene, således at helt nye mineraler blev dannet på bekostning af de oprindelige lermineraler. Mest synlig er biotit og granat, men også sillimanit blev dannet. Sillimanit er en aluminium-kisel forbindelse, der danner både en lysegrå "filt" af meget små tråde eller hvide nåle op til mere end en centimeter lange og en millimeter tykke. Laboratorieforsøg med bjergartsmaterialer har vist, at dannelsen af sillimanit kræver en temperatur på mindst 500°C og et tryk svarende til en dybde på mindst 10 km. Andre indicier tyder på, at temperaturen nåede over 600° under bjergkædedannelsens kulmination.

For at undersøge om der kunne være forekomster af værdifulde malmmineraler, er 24 håndstykker og 5 prøver af bæksedimenter blevet analyseret for en række grundstoffer (Tabel 1 og 2). Der blev ikke konstateret nogen markante afvigelser fra værdier i almindelige skorpebjergarter, med undtagelse af et håndstykke og en bæksediment prøve, der viste en lidt forhøjet guld-værdi. Dog er guld-indholdet i de to prøver ikke højt nok til at give anledning til yderligere efterforskning. Konklusionen er, at de jern-rige lag i biotit-granat skifrene på Storøen og Akulleq, er "ufrugtbare jernformationer" (engelsk: barren iron-formations). Dog skulle denne skuffende konklusion ikke holde folk fra at søge efter granat krystaller og magnetkis i de beskrevne lokaliteter.

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