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The early Archaean to Proterozoic history of the Isukasia area, southern West Greenland

by

Allen P. Nutman

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Abstract

The *c.* 3800 Ma Isua supracrustal belt and associated smaller bodies of supracrustal rocks are intruded by >3600 Ma orthogneisses. A coherent stratigraphic sequence is recognised consisting of interlayered metabasic rocks, metasediments derived from volcanic rocks, chemical sediments, and metabasic and ultramafic intrusions. Despite repeated deformation and high grade metamorphism sedimentary structures are locally preserved. The depositional environment was probably an immersed volcanic region remote from areas of significantly older crust. Conglomeratic structures in a metachert and banded iron formation unit suggest shoaling and shallow water conditions. Felsic sediments locally preserve evidence of deposition from turbidite flows. The Isua supracrustal rocks are regarded as thin fragments of a thicker, more extensive sequence.

The orthogneisses that intrude the supracrustal rocks consist of 3750–3700 Ma multiphase tonalites (the grey gneisses) which were first intruded by the basic Inaluk dykes, then by abundant shallow-dipping swarms of *c.* 3600 Ma granite sheets (the white gneisses) and finally by *c.* 3400 Ma pegmatitic gneiss sheets. These early Archaean rocks were metamorphosed under amphibolite facies conditions and repeatedly deformed prior to intrusion of the Tarssartôq basic dykes in the mid Archaean.

In the late Archaean (3100–2500 Ma) there was polyphase metamorphism up to amphibolite facies grade and two or more stages of deformation and local intrusion of granitic gneiss sheets and pegmatites. However, despite general strong deformation there is a large augen of low deformation preserved within the arc of the Isua supracrustal belt. During the Proterozoic there was intrusion of basic dykes, major faulting with associated recrystallisation under uppermost greenschist to lowermost amphibolite facies conditions, followed by heating and intrusion of acid dykes at *c.* 1600 Ma. No profitable mineralisations have been located.

Author's address:
Department of Earth Sciences
Memorial University of Newfoundland
St.John's
Newfoundland
Canada A1B 3X5

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Fig. 1. Oblique aerial photograph of the Isukasia area, looking southwest. Isukasia lies at the bottom right hand corner of the photograph. Area marked *Isb* between the dashed lines is the *Isua supracrustal belt.* Also shown are the central gneisses (Tarssartôq dykes are locally visible), the *Ataneq fault, lake 678 m* and *G,* northern Godthåbsfjord. Route 505 F2-V, no. 26672. Copyright Geodetic Institute, Denmark. Published with permission A.495/79.

INTRODUCTION

Geological importance of the Isukasia area

This bulletin describes the geology of the Isukasia area, part of the Archaean block of West Greenland (figs 1 & 2), and accompanies a 1:40 000 geological map of the same area (Plate 1). The area is geologically important because it contains the most extensive, best preserved *c.* 3800 to 3600 Ma old rocks known, and as such is the best sample available of the oldest-known crust. Prior to these rocks, the known terrestrial geological record is restricted to a few ion probe U-Pb ages of 4200 to 4100 Ma on zircons of detrital origin in 3600 to 3300 Ma old quartzites in the Yilgarn Block, Western Australia (Froude *et al.,* 1983). The early Archaean record of the area is of a supracrustal sequence that was repeatedly injected by granitic *(s.l.)* magmas and suffered high grade metamorphism and polyphase, heterogeneous deformation prior to cratonisation. Isotopic evidence suggests that these events occurred within the interval 3850–3350 Ma. The largest fragment in the area of the supracrustal sequence, the Isua supracrustal belt, has been the target of considerable geological investigation over the past decade.

The wide range of early Archaean lithologies together with the recognition of a well-defined sequence of events in the Isukasia area (Table 1) allows examination of early Archaean crustal evolution and crustal processes, such as sediment provenance and depositional environment, magma genesis, tectonic styles and thermal regimes. Although the Isukasia area is composed almost entirely of early Archaean rocks, it has not been immune to younger geological activity. Metamorphic and isotopic studies (Lambert & Simons, 1969; Rosing, 1983) and field observations (e.g. Bridgwater *et al.,* 1981; Nutman, 1982a, Nutman *et al.,* 1983, 1984a; Garde *et al.,* 1983) clearly show the importance of late Archaean and Proterozoic events in the geological evolution of the Isukasia area, whose effects hamper study of the area's early Archaean history.

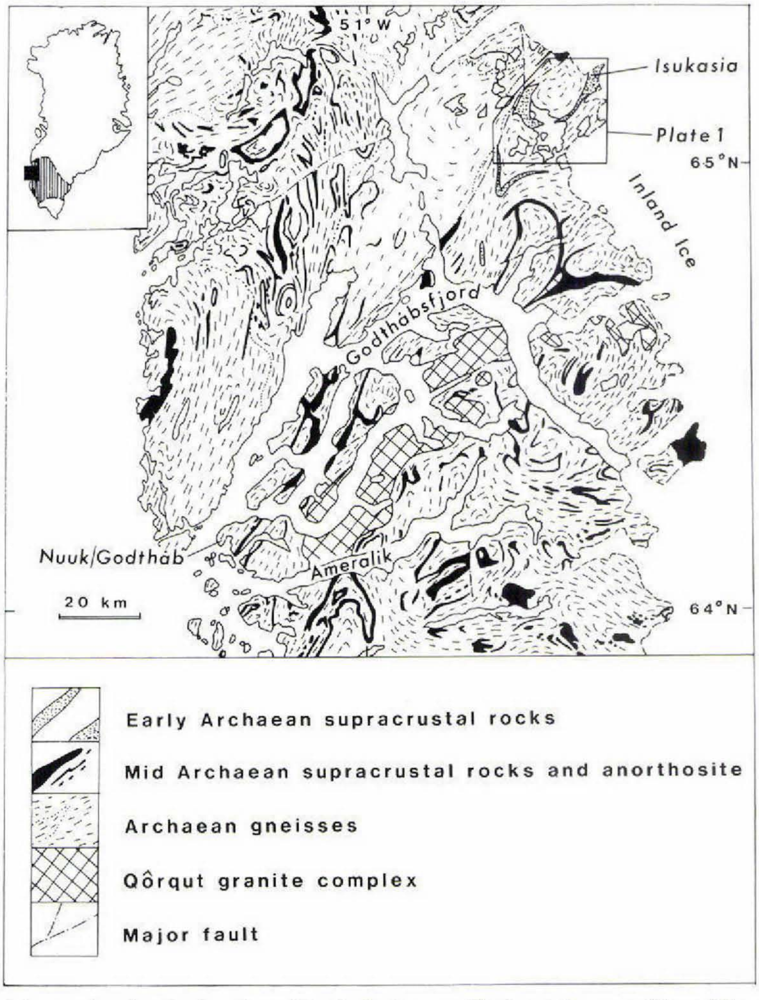

Fig. 2. Sketch map showing the location of the Isukasia area. The inset shows position of the main figure in the Archaean block of Greenland.

Character of the Isukasia area

The Isukasia area (Plate 1) is part of an extensive plateau dotted with lakes against the Inland Ice to the north of Godthåbsfjord (fig. 2). Sparsely vegetated glacial and fluvial/glacial deposits cover much of the bed rock. Some of these deposits form a thin or incomplete veneer, whilst elsewhere thick swathes of morainic deposits completely obscure the bed rocks over tracts of several square kilometres (inset on Plate 1). There are recent glacial deposits along the edge of and on the Inland Ice, and there are fluvial deposits along the major rivers of the area.

Isukasia, the peninsula of land projecting into the Inland Ice, was up to the end of the 1970s commonly referred to as Isua, which in fact lies 20 to 25 km to the northwest. Thus regrettably the main supracrustal unit of the area that disappears under the Inland Ice at Isukasia became known as the Isua supracrustal belt, a term that is now so entrenched in the geological literature that it is not practical to change it. Place names recognised by the Greenland Place Name Committee are scarce in the area. *Imarssuaq* and *Imarssuak* have been used to refer to the largest lake in the area. The origin of this name is not known, and at present it is not recognised by the Greenland Place Name Committee. Therefore it is referred to simply as 'lake 678 m' (Plate 1).

History of geological investigations

The Isukasia area was first explored by the commercial company Kryolitselskabet Øresund A/S, and academic interest in the area started almost a decade later when the antiquity of the rocks was established (Moorbath *et al.*, 1972). The first reconnaissance investigations of the area were made in 1962 by Kryolitselskabet Øresund A/S (Keto, 1970). These investigations identified a major arcuate 'schist formation' unit (now called the Isua supracrustal belt) and the dome-like structure in the adjacent gneisses, both of which are apparent on aerial photographs. In 1965 exploration activity by the company was concentrated in the region between Nuuk/Godthåb and Maniitsoq/Sukkertoppen, and the Isukasia area was revisited. On this occasion numerous erratic blocks of banded iron formation and 'schist formation' were found close to the edge of the Inland Ice, and the association suggested a relationship in the field. Aeromagnetic traverses, in progress at that time in the adjacent areas to the west, were extended to the Isukasia area. The first flights indicated significant magnetic anomalies associated with and conformable to the Isua supracrustal belt. The arc-shaped anomaly, 40 km long, disappears under the Inland Ice at Isukasia. By far the largest anomaly in the area (>12000 gammas) corresponds to large outcrops of magnetite-rich banded iron formation at Isukasia. These early geophysical observations led to more detailed studies in the following years. However in 1973, after completely assaying the volume and quality of the iron ore, the company ceased their operations. In the meantime, the Isukasia area was starting to attract scientific attention; at the time of writing there have been in the order of a hundred geological publications that deal directly with the area (bibliography available upon request from GGU). At first academic work concentrated largely on the Isua supracrustal belt, and it was not until 1980 that the adjacent gneisses were studied in detail.

A synopsis of publications covering the main fields of study in the area follows: Lambert & Simons (1969) presented a K-Ar date of 1940 ± 50 Ma from hornblendes in the Isua supracrustal belt, and Windley (1969) compared the belt to Proterozoic volcanosedimentary sequences. Further investigations suggest that this K-Ar date corresponds to a period of Proterozoic metamorphism and alteration (Rosing, 1983). V. R. McGregor visited the area in 1971 and suggested that the rocks there could be mostly of early Archaean age. Shortly afterwards, Rb-Sr and Pb-Pb isotopic studies by a group from Oxford University led by S. Moorbath showed the gneisses and the supracrustal rocks to be of early Archaean age (Moorbath *et al.*, 1972, 1973). In 1973 GGU geologists carried out 14 days of reconnaissance field work in the area and produced the framework of the current chronology (Bridgwater & McGregor, 1974).

Following on from these early studies, the late J. H. Allaart of GGU spent two field seasons investigating and mapping the Isua supracrustal belt and the contact gneisses at a scale of 1:20 000 (Allaart, 1975, 1976), and James (1975a, 1976) made reconnaissance structural studies. Some whole rock geochemical studies were carried out in this period by GGU and GGU-associated workers (Gill & Bridgwater, 1976; Moorbath *et al.*, 1975; Bridgwater *et al.*, 1976), and further isotopic investigations of the Isua supracrustal rocks and the contact gneisses were made (e.g. Moorbath *et al.*, 1975, 1977; Michard-Vitrac *et al.*, 1977; Baadsgaard, 1976; Hamilton *et al.*, 1978; Appel *et al.*, 1978). Meanwhile other groups undertook geochemical studies, mainly of the Isua supracrustal belt. P. W. U. Appel and co-workers from the Max Planck Institute for Cosmochemistry (Mainz) carried out diverse geochemical studies and carbon isotope studies (Appel, 1977, 1979a,b, 1980; Appel & Jagoutz, 1978, Schidlowski *et al.*, 1979). Oxygen and sulphur isotope analyses have been carried out on the supracrustal rocks and the gneisses by a group from Northern Illinois University (e.g. Oehler & Smith, 1977, Oskvarek & Perry, 1976; Perry & Ahmad, 1977), and organic chemistry studies have been performed by a group from Maryland University and by Nagy *et al.* (1975, 1977). These sophisticated geochemical and isotopic studies continued despite relatively poor knowledge of the tectonometamorphic history and lithological variation within the Isua supracrustal belt, whilst knowledge of the gneisses only a few hundred metres away from the belt was restricted to scattered spot visits for compilation of the 1:500 000 geological map (Allaart *et al.*, 1977). Pflug (1978) and Pflug & Jaeschke-Boyer (1979) interpreted microscopic artifacts in metacherts of the Isua supracrustal belt as early Archaean 'yeast-like' microfossils. These purported microfossils have been re-interpreted as fluid inclusions (Bridgwater *et al.*, 1981).

A visit by a NATO study group to the Isukasia area (Bridgwater *et al.*, 1979) led to a new wave of geological studies of the Isukasia area at GGU and the University of Copenhagen, with support from the Royal Society of London and the Carlsberg Foundation. These studies sought to redress the imbalance of work on the area by (1) more detailed mapping of the Isua supracrustal belt, (2) methodical mapping and detailed investigation of the gneiss-dominated parts of the area, (3) stratigraphic studies of the Isua supracrustal belt (some of this work was in conjunction with the late E. Dimroth, University of Quebec at Chicoutimi), (4) metamorphic studies (in cooperation with a group at Harvard University led by R. F. Dymek), (5) structural studies and (6) further geochemical and isotopic studies, in particular of the orthogneisses.

By 1983, there had been considerable progress made in further understanding of the geology of the area by the GGU and University of Copenhagen group. It was thus decided that the author should compile a GGU Bulletin with a 1:40 000 map, covering in some detail the geology of the Isukasia area. This project was made possible by funding from GGU, the Carlsberg Foundation and the Danish Natural Science Research Council. This Bulletin does not contain much in the way of new information – the aim has been to bring together data from many publications so that a systematic description of the geology of this classic area would be available under one cover. The author would particularly like to thank D. Bridgwater and M. Rosing for making this possible.

Table 1. Chronological sequence in the Isukasia area

(10)	(*youngest*) Injection of basic dykes and crustally derived granite sheets. Faulting under lowermost amphibolite facies to greenschist facies conditions and metasomatism. 1600 to 2100 Ma.
(9)	Intrusion of pegmatites. *c.* 2550 Ma.
(8)	*East of the Ataneq fault.* Development of a strong banding with local isoclinal folding (only recognised in the south of the area) followed by linear belts of strong deformation interspersed with southerly plunging folds with wavelengths of over 10 km. Amphibolite facies metamorphism. Local intrusion of granitic-granodioritic sheets. 2600 to 3100 Ma.
	West of the Ataneq fault. Development of banded gneiss complex with intrusion of the Taserssuaq tonalite and granitoid sheets. Metamorphism up to granulite facies followed by retrogression under amphibolite facies conditions. Isoclinal folding followed by formation of upright, irregular folds, with basin and dome forms.
(7)	Intrusion of Tarssartôq dykes, probably equivalent to the Ameralik dykes in Godthåbsfjord. *c.* 3200 Ma?
(6)	Deformation giving rise to upright folds.
(5) (*c*)	Intrusion of Amîtsoq pegmatitic gneisses (*c.* 3400 Ma) and thin tonalitic sheets.
(*b*)	Intercalation of supracrustal and gneiss units.
(*a*)	Deformation to produce a strong banding, most important in the south.
	It is possible that *a*, *b* & *c* were contemporaneous.
(4)	Intrusion of the Amîtsoq white gneisses. 3600 Ma.
(3)	Intrusion of the Inaluk dykes.
(2)	Intrusion of the Amîtsoq grey gneisses. This group probably includes early thin tonalitic sheets discussed in Nutman *et al.* (1983). Isoclinal folding of supracrustal rocks (1) probably occurred during intrusion of the grey gneisses. 3750 to 3700 Ma.
(1)	(*oldest*) Formation of Isua supracrustal rocks and Akilia association, and intrusion of basic and ultramafic rocks into them. *c.* 3810 Ma.

SUPRACRUSTAL ROCKS

Supracrustal rocks in the Isukasia area occur as strips and lenses included within the early Archaean Amîtsoq gneisses (McGregor, 1979). They are divided into two: (1) the Isua supracrustal belt and (2) the rest, that form part of the Akilia association (McGregor & Mason, 1977; see Table 1 for chronology and plate 1 for distribution of these rocks). As a visual estimate the Isua supracrustal belt comprises about 80 percent of the supracrustal rocks of the area. This division is rather arbitrary as these rocks were probably all derived from the same volcanosedimentary sequence (e.g. Baadsgaard *et al.*, 1984). It is adhered to here due to the shear size and amount of work done on the Isua supracrustal belt compared with the Akilia association units. Descriptions of these rocks here are biased towards their field characteristics. However, geochemical data are drawn upon in discussion of the protoliths of these rocks (representative analyses are given in Table 2). The composition of these rocks cannot be taken at face value to be original as there is considerable evidence of metasomatic alteration (e.g. Gill *et al.*, 1981; Rosing, 1983).

Isua supracrustal belt

Recognition of a stratigraphy

Major lithological divisions of the Isua supracrustal belt can be traced along strike, in spite of poor exposure and the disturbance of sequences by orthogneiss sheets and tectonic slides (in the sense of Hutton, 1979). These divisions, used in conjunction with locally preserved graded bedding from scattered localities throughout the Isua supracrustal belt, confirm the repetition of the succession by isoclinal folding suggested by Bridgwater & McGregor (1974). A detailed stratigraphy has been built up from sections of relatively low deformation. Using the stratigraphy and major structures as a control, individual lithological units can be traced through areas of varying strain, late retrogression and alteration, both along strike and on opposing limbs of folds. A coherent stratigraphy (Nutman *et al.*, 1983, 1984a) has been worked out along the whole length of the Isua supracrustal belt (sequence A). In the northeast of the Isua supracrustal belt a fault separates some units (sequence B) from sequence A (Plate 1 & fig. 3). Due to deformation the original thicknesses of the sequences are unknown. Excluding units of the garbenschiefer and ultramafic rocks (slightly discordant bodies of intrusive basic igneous origin) that occur in both sequences, they now have a total thickness of less than 1 km, depending on local strain.

The sedimentary and volcanic components of the belt are layered on a scale of a few millimetres to several metres. The main rock types are amphibolites (mostly layered), metacherts (stratified quartzites with no evidence of detrital origin), mica schists, felsic rocks (some demonstrably of detrital origin), banded carbonate and calc-silicate rocks, banded iron formation lithologies, and mixed rocks in which the lithologies listed are interlayered or grade into one another at a scale of a few centimetres. Different levels within the sequences are dominated by specific lithologies, allowing division of them into 'formations'. In sequence A these are in ascending order: (A1) amphibolite formation, (A2) lower banded iron formation, (A3) variegated schist formation, (A4) upper banded iron formation, (A5) calc-silicate formation and (A6) felsic formation. Sequence B comprises in ascending order: (B1) felsic formation and (B2) mica schist formation (fig. 3). Stratigraphic units of the Isua supracrustal belt are intruded by phases of the gneiss complex and are truncated at the edge of the Isua supracrustal belt (Plate 1).

Sequence A

A1 Amphibolite formation

The lowest part of this formation is well exposed at about 65° 6′N, 50° 8′W, west of lake 678 m. There are two layers up to several metres thick of clinopyroxene and hornblende-rich amphibolite with associated pale, homogeneous to faintly laminated hornblende-plagioclase amphibolite. Quartz-rich rocks form a unit up to 2 m thick. They have a weakly defined compositional layering of mafic layers (grunerite–hornblende–magnetite) less than 1 cm thick and quartz layers 1 to 10 cm thick. Upper parts of the

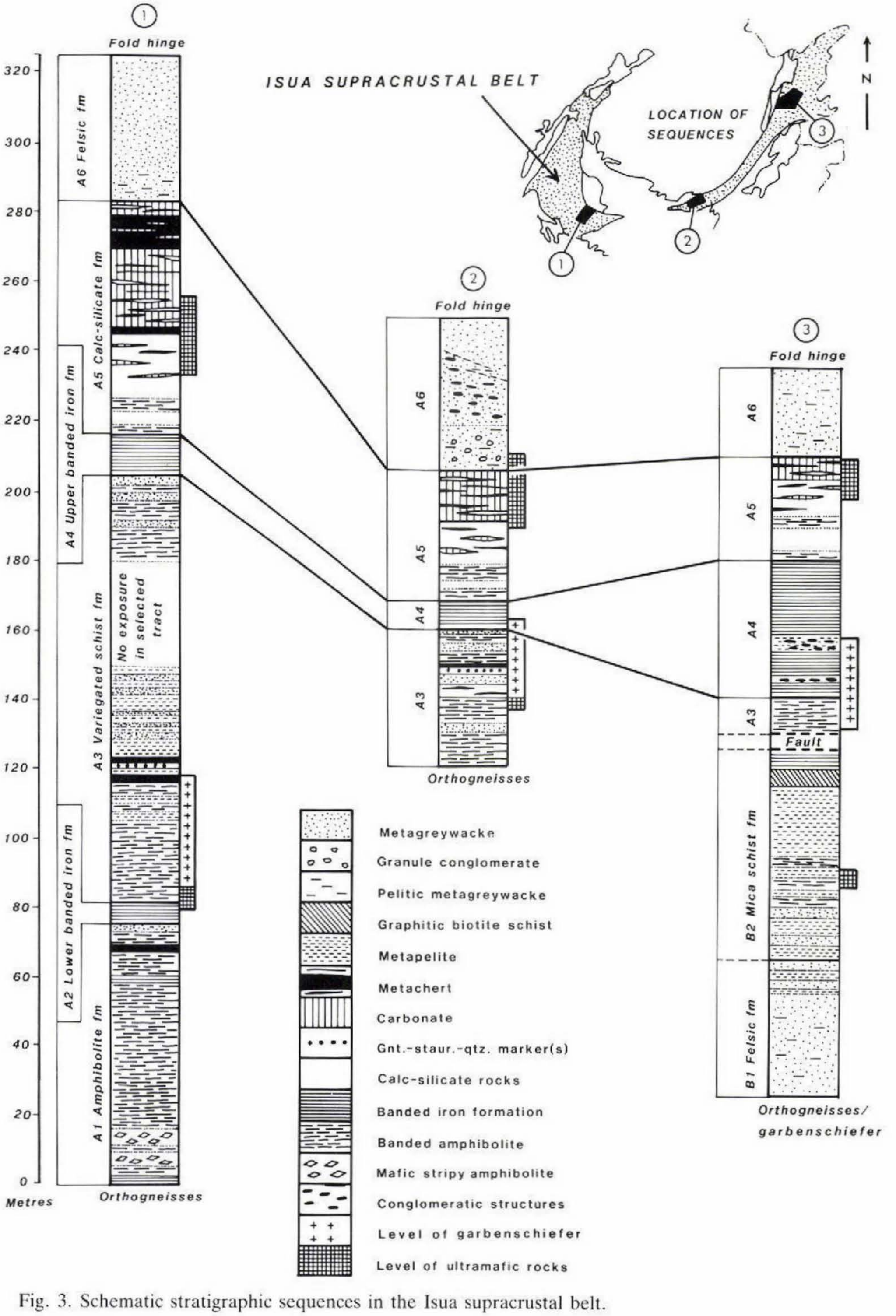

Fig. 3. Schematic stratigraphic sequences in the Isua supracrustal belt.

amphibolite formation are preserved over a wider area, mostly west of lake 678 m. Hornblende-plagioclase amphibolite layered at a 1 to 20 cm scale is the main lithology. Some layers of these amphibolites are garnetiferous and are interlayered with a few, rusty-weathering, quartz-rich bands less than 10 cm thick, probably of sedimentary origin. Some amphibolite layers are separated by seams up to 1 cm thick rich in hornblende or biotite. There are also a few units less than 2 m thick of rather homogeneous hornblende-plagioclase amphibolite with slightly coarser grained flecks of plagioclase. These might be gabbroic sheets (Rosing, 1983). Nodular amphibolites occur locally east of lake 678 m. They consist of pods of fine-grained, hornblende-plagioclase amphibolite set in an irregularly banded paler matrix. The pods are generally matrix supported and are up to 1 m long in the X direction of deformation. At the top of the formation there is a 10 m thick unit of pale banded amphibolite containing biotite and garnet.

Along the western edge of the Isua supracrustal belt there is a strip up to 100 m thick of predominantly basic rocks (Plate 1). These are in faulted contact with sequence A rocks to the east and bounded (and locally intruded) by orthogneisses to the west. This strip comprises predominantly banded amphibolites, probably correlated with the amphibolite formation. Locally exposed along its eastern side is a thin, lenticular, rusty, quartzitic unit (A2 lower banded iron formation?) and brown weathering, flaggy amphibolite (A3 variegated schist formation?). Along the northern side of the central gneisses there are two units of sheared greenschist facies mafic schist, devoid of lithological variation. Relations are not clear, but they are provisionally correlated with the amphibolite formation.

A2 Lower banded iron formation

This formation is best preserved on the eastern margin of the belt west of lake 678 m, where it can be followed for 15 km. On the southern side of the Isua supracrustal belt it is poorly exposed; there are sporadic outcrops of metachert and banded iron formation in a stratigraphic position implying that they belong to this formation. It is up to 15 m thick and is concordant with adjacent formations. It consists of laminated quartz–amphibole–magnetite rocks and metacherts interlayered in no apparent order. Layering is generally at a scale of 25 mm or less.

A3 Variegated schist formation

This formation is exposed throughout the Isua supracrustal belt, much of it as strips within the largest garbenschiefer unit. In the eastern and central parts of the Isua supracrustal belt the dominant lithologies are hornblende-plagioclase amphibolite layered at a 0.5 to 5 cm scale with some associated clinopyroxene and hornblende rich layers. There is also subordinate paler, biotite and garnet-bearing amphibolite. In the western part of the Isua supracrustal belt the latter increases in importance, and contains metapelitic biotite-rich interbeds and is locally associated with layered felsic rocks. In the northwest the scale of layering is greater, up to 1 m, probably because strain is less. Graded bedding is preserved in felsic units of this formation locally, notably in the northwest of the Isua supracrustal belt (V. R. McGregor, personal communication, 1981; Rosing, 1983; Nutman *et al.*, 1984a). The variegated schist formation contains rusty weathering metacherts up to 5 m thick with thin, magnetite-bearing banded iron formation interlayers. One metachert is garnetiferous and grades laterally into garnet-staurolite-graphite-quartz-amphibole rock, and locally contains Fe-rich orthopyroxene. These distinctive units are useful marker horizons.

A4 Upper banded iron formation

This formation consists of impure metacherts and diverse, thinly-layered quartz-magnetite-amphibole-carbonate rocks. The top and bottom are usually sharp and concordant to stratification in adjacent formations, although contacts gradational over 10 cm have been noted locally in the west of the Isua supracrustal belt (Rosing, 1983). There is marked lithological variation in this formation, and its outcrop thickness ranges from under 5 m to over 1 km reflecting changes in original thickness as well as regional variation in degree of strain. The extensive, thick outcrop of metachert and magnetite-bearing banded iron formation at Isukasia is interpreted as a local variation of this formation thickened by folding and

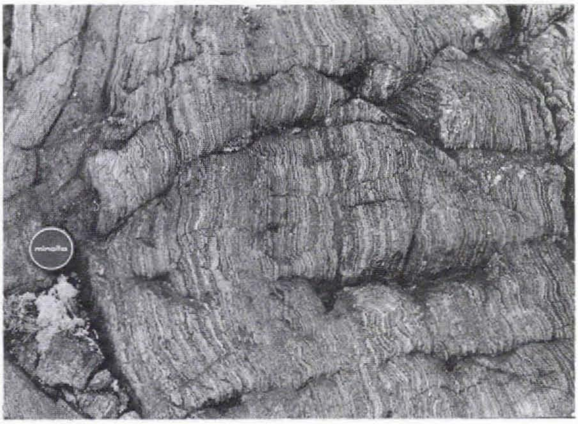

Fig. 4. Finely laminated quartz-magnetite banded iron formation, from the upper banded iron formation (A4). 65°5′52″N, 50°0′47″W.

probably by imbrication. Layering is normally at the millimetre scale (fig. 4) with quartz comprising over 50% modal of the rocks. In places at Isukasia, layering is thicker and the magnetite content is higher. Rocks described as metacherts from this formation in the northeastern end of the Isua supracrustal belt commonly contain thin layers of anthophyllite with subordinate amounts of talc or carbonate. These rocks are found interlayered with iron-rich metasediments, suggesting that the variation in iron and magnesium contents is diagenetic or depositional. In the northeast exposures of the formation there are a few thin layers of conglomeratic appearance. Whether these units are of sedimentary or tectonic origin is discussed below.

A5 Calc-silicate formation

This formation is divided into lower amphibolitic and upper carbonate-metachert units. The amphibolitic unit is characterised by melanocratic hornblende-rich rocks and amphibolites. The former have an irregular banding at a scale of less than 5 cm, accentuated by stringers and lenses of carbonate and quartz. The amphibolites are finely laminated to homogeneous hornblende-plagioclase rocks with some clinopyroxene-rich units. Felsic rocks are uncommon at this stratigraphic level. The upper carbonate-metachert unit has been severely disrupted by deformation and by reaction with adjacent ultramafic rocks. It ranges from ankeritic carbonate with widely spaced bands of metachert, calc-silicates (amphiboles or clinopyroxene) or magnetite, to metacherts with thin calc-silicate carbonate layers (fig. 5). Lo-

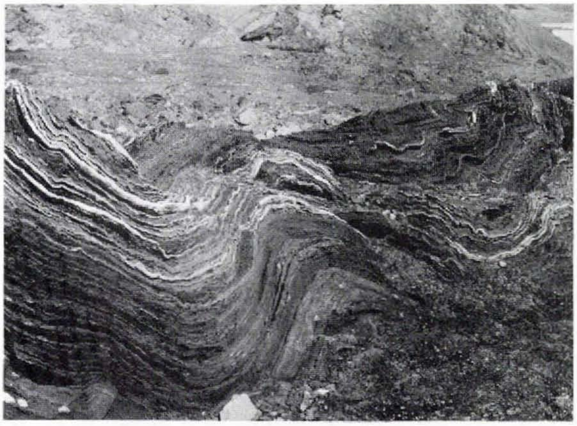

Fig. 5. Layered metachert (pale) and carbonate and calc-silicate (dark) rock, from upper part of the calc-silicate formation (A5). 65°7′4″N, 50°13′30″W.

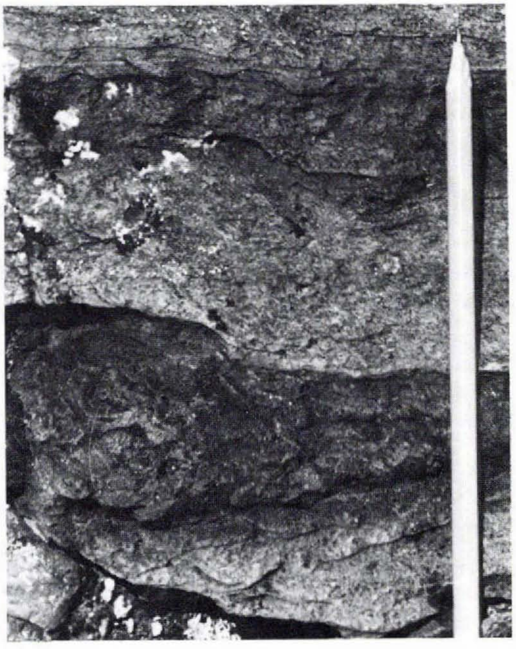

Fig. 6. Graded layer in felsic rocks of the felsic formation sequence A. Direction of younging is upwards in the figure. 65°5'52"N, 50°0'22"W.

cally there are calc-silicate units up to 2 m thick. Chrome muscovite is locally abundant in quartz-rich rocks close to ultramafic units. Layering in these rocks ranges from less than 1 cm to 2 m, and is commonly irregular or lenticular.

A6 Felsic formation

This formation is the uppermost division of sequence A. It has been referred to as an acid volcanic unit in previous accounts (Allaart, 1976; Bridgwater *et al.*, 1976). It outcrops in the core of an isoclinal fold and has been thickened by repetition in small tight folds. The formation is divided into a thin lower unit of 'granule conglomerate' and an overlying unit that varies laterally from a coarse 'conglomeratic' lithology to fine-grained, layered felsic rocks (fig. 6). The granule conglomerate unit is approximately 5 m thick on the eastern shore of lake 678 m, where it is overlain by the contrasting coarse conglomeratic rock. In other parts of the Isua supracrustal belt the granule conglomerate unit is overlain by the fine-grained variety of the upper unit, and they cannot be distinguished. The granule conglomerate unit consists of plagioclase-quartz-biotite-muscovite rocks with garnet, alkali feldspar and carbonate present locally. There are altered, detrital plagioclase grains up to 5 mm diameter preserved locally, the largest of these grains are commonly multigranular. Carbonate and coarse-grained quartz are secondary components, probably introduced from the adjacent calc-silicate formation. Coarse conglomeratic rock of the felsic formation is best developed on the eastern shore of lake 678 m. Along strike in both directions this lithology is largely replaced over 5 km by layered felsic rocks. It comprises prolate nodules of pale, fine-grained felsic material in a darker matrix that ranges from impure carbonate to felsic granule conglomerate. The same lithology as that of the nodules also occurs as layers 10 to 20 cm thick alternating with darker matrix. In the west of the belt some massive felsic units are interlayered with layered felsic rocks and are cut by gneiss sheets. Thus some of these massive units are likely to belong to the supracrustal sequence rather than being intrusive tonalite sheets.

Sequence B

Sequence B outcrops along the west side of the northeastern end of the belt (Plate 1). Its way-up is shown by the sense of graded bedding near its western margin. The sequence is poorly exposed, contains small isoclinal folds, and is estimated to be less than 150 m thick. To the east sequence B is in faulted contact with sequence A.

B1 Felsic formation

This formation comprises plagioclase–quartz–biotite–muscovite rocks layered at a 2 to 10 cm scale. Usually they are uniformly fine grained, locally with plagioclase grains possibly of detrital origin. Some graded beds occur near the base of the formation. Garnetiferous mica schist units occur sporadically throughout the formation, and increase in frequency and thickness upwards. These schists are similar to the dominant lithology of the overlying mica schist formation. Thus the boundary between these formations is placed arbitrarily in this packet of interlayered felsic rocks and mica schists that marks the transition between them.

B2 Mica schist formation

This consists predominantly of chlorite-biotite-garnet-quartz schists that locally contain staurolite, kyanite, carbonate and graphite. These schists have a faint, but regular layering at all scales. They are interlayered with thin metachert units and rarely with the magnetite banded iron formation. There are also some interbeds of hornblende-plagioclase layered amphibolite in the middle levels of the formation. Uppermost in the formation is a unit of calc-silicate banded iron formation with metachert and sporadically developed biotite-quartz-graphite schist.

Relation between sequences A and B

Three possible relations between the sequences are given here: (1) If the garbenschiefer unit at the base of sequence B is assumed equivalent to the one in sequence A (fig. 3), then sequence B could be a thicker lateral equivalent of the variegated schist formation of sequence A, which contains essentially the same lithologies but in different proportions. (2) If instead, the similarity between the felsic formations of the two sequences is given more weight in correlation, this would place the upper parts of sequence B stratigraphically above sequence A. (3) Alternatively, as the two sequences are in faulted contact, there may not necessarily be a correlation between them.

Mafic intrusions

Garbenschiefer units

These occur as units up to more than 1 km broad that form approximately 25 percent of the Isua supracrustal belt. They are a pale green basic rock that is massive or faintly banded at a scale of 1 m or less. It often shows well-developed garbenschiefer texture of amphiboles on its foliation surfaces, hence its name. Although this texture is not unique to this lithology, it is where it is most strikingly developed. Garbenschiefer units occur in both sequence A and B (Plate 1), and they are slightly discordant to lithological layering in adjacent rocks and contain inclusions of supracrustal lithologies. The garbenschiefer is interpreted as gabbroic, possibly sill-like intrusion(s).

The garbenschiefer unit is most commonly found as a green amphibole-chlorite schist, locally containing carbonate. Development of the hornblende garbenschiefer texture is controlled by water content of the rocks (Rosing, 1983). From field and isotopic evidence Rosing shows that hydration occurred in the late Archaean and Proterozoic (discussed below). Locally garbenschiefer has strongly reacted with iron-rich inclusions. For example it is commonly garnetiferous where in contact with biotite schist. The thickness of the garbenschiefer unit in sequence A varies considerably. On the southern limb of the early isoclinal fold in this sequence it is less than 20 m thick and discontinuous along strike. At both ends of the Isua supracrustal belt there are extensive outcrops of garbenschiefer, due to its repetition and thickening in folds. On the northern limb of the isoclinal fold the garbenschiefer is thinned tectonically and is faulted out at the margin of the Isua supracrustal belt and at the eastern margin of sequence B (Plate 1 & fig. 21). In sequence B garbenschiefer occurs at the edge of the Isua supracrustal belt at the margin of the Inland Ice and thins southwards (Plate 1 & fig. 21).

Ultramafic rocks

These occur at most stratigraphic levels in the Isua supracrustal belt (Plate 1 & fig. 3). They are repeated by isoclinal folds and some intrude garbenschiefer. Ultramafic units are up to 100 m wide, and some are slightly discordant to layering of the supracrustal sequences. They are homogeneous to slightly banded, a structure interpreted as tectonically modified igneous layering. Their volatile-free composition ranges from metadunite to pyroxenite. They are interpreted as intrusions into the supracrustal sequence.

No regular composition variation across units has been observed and different units seem to be dominated by one composition. Most of these rocks are rich in ortho-amphibole, and where best preserved they contain relict olivine and pyroxene set in fibrous amphibole matrix. Where most altered they are fibrous amphibole-talc or serpentine schists, locally rich in carbonate. Concentrations of chromite or other spinels have not been found. Ultramafic units within formations A1 to A3 can be traced for up to 15 km. In areas of lower deformation they are continuous sheets, but where more deformed they are disrupted into trains of pods. Ultramafic rocks that occur in the upper part of formation A5 and locally in formation A6 are generally strongly broken up and have reacted with adjacent carbonate and silica-rich rocks. They probably correlate with the better preserved ultramafic rocks of formations A1 to A3. A talc schist unit up to 5 m thick occurs in formation B2. This is interpreted as a strongly altered sheet of igneous origin rather than related to the mica schists and amphibolites with which it is found.

Structures in Isua metasediments

Planar layering in rocks of contrasting composition is ubiquitous throughout the Isua supracrustal belt. It is concordant to major lithological divisions and is interpreted as tectonometamorphically modified sedimentary layering and volcanic structures. A striking feature of much of the supracrustal sequence is the closeness of interlayering of contrasting lithologies, even allowing for (probably considerable) tectonic thinning. Occurrences of nodular, lensoidal and irregular layering are found sporadically throughout the Isua supracrustal belt. Most cases can be attributed to the way layers of different rheological properties and thickness have behaved during deformation. Locally in metachert and carbonate rocks there are lenticular or nodular units alternating with planar layered units, suggesting that some of these differences in layering could be pre-tectonic. In the northwest of the Isua supracrustal belt small scale lenticular structure is prevalent in most lithologies, due to intersection of compositional layering and foliation by a later schistosity.

Several types of conglomeratic structure are found in the metasedimentary rocks. Since these rocks have been strongly deformed several times, these structures cannot be taken at face value to be of sedimentary origin. The origin of the three main types of conglomeratic structure is discussed here.

Flat pebble conglomeratic structure

This structure occurs in the northeastern exposures of the upper banded iron formation (Dimroth, 1982; Nutman *et al.*, 1984a). Flat pebble conglomeratic units consist of tabular, angular metachert fragments set in a groundmass of quartz, magnetite and carbonate (fig. 7). The fragments are up to 15 cm long, 3 cm thick and some are layered. They are commonly arranged in an imbricate fashion, rather than parallel to stratification. The units are normally conformable, but some show

Fig. 7. Flat pebble conglomeratic unit in the upper banded iron formation (A4). 65°10′25″N, 49°49′10″W.

Fig. 8. Round pebble conglomeratic unit in the upper banded iron formation (A4). The largest quartz nodules are *c.* 3 cm in diameter. 65°10′18″N, 49°48′41″W.

possible erosional bases. In most cases there is little or no rheological contrast between the fragments and their matrix, reducing the possibility that they are of tectonic origin. These units resemble 'edgewise conglomerates' formed by break-up of desiccated laminated sediment in the lower supratidal zone (James, 1979).

Round pebble conglomeratic structure

This occurs in the northeastern exposures of the upper banded iron formation (Allaart, 1976; Dimroth, 1982; Nutman *et al.*, 1984a) as one unit that grades laterally into a mica schist horizon. It is oligomict with rounded quartz nodules and biotite-rich clasts with an open or closed fabric set in a matrix of biotite-garnet-quartz schist gradational into carbonate or calc-silicate rocks (fig. 8). Graded bedding has not been observed. The diameter of the quartz nodules in *Y-Z* sections is 0.5 to 7.5 cm. The nodules are prolate with their *X* direction parallel to the axes of steeply plunging small folds and a mineral lineation at the locality. Commonly the nodules are arranged in distinct layers. In the most closely packed layers the matrix is reduced to a film coating the nodules.

Dimroth (1982) concluded that nodular structure in this unit could be conglomerate of sedimentary origin. There is no doubt that some quartz nodules in the Isua supracrustal belt (including some in the discussed unit) are tectonically broken-up layers and veins. Field relations are equivocal and a tectonic origin for all the nodules cannot be ruled out. Whatever the origin of this unit it is noteworthy, because its mica schist matrix with biotite-rich clasts indicates a spell of terrigenous sedimentation within the development of a thick unit of chemical sediments. If the quartz nodules of the unit are sedimentary, they could be pebbles derived from adjacent metacherts. Lack of graded bedding in this unit would suggest it is a traction current deposit (Nutman *et al.*, 1984a). Such quartz pebble conglomerates generally form in shallow water environments.

Fig. 9. Felsic nodules (note occurrence as trains) in carbonate-rich matrix, felsic formation sequence A. 65°5′52″N, 50°0′22″W.

Conglomeratic structure in the felsic formation of sequence A

The most spectacular occurrence of this lithology is in the core of a Z-style fold (65° 6′N, 50° 1′W) on the east shore of lake 678 m (Allaart, 1976; Bridgwater *et al.*, 1976; Dimroth, 1982; Nutman *et al.*, 1983, 1984a). This locality is discussed in detail here, as it has been widely publicised as a sedimentary conglomerate. It consists of fine-grained felsic nodules set in a matrix ranging from felsic detrital sediment to carbonate and calc-silicate rocks (fig. 9). Felsic matrix is locally graded and commonly shows patchy enrichment in carbonate and calc-silicate minerals. Disruption

Fig. 10. Felsic nodules produced by disruption of continuous felsic layers between carbonate-rich matrix, felsic formation sequence A. 65°5′52″N, 50°0′22″W.

and alteration of the edge of a Tarssartôq dyke (Table 1) by carbonate-rich matrix suggests that at least some carbonate is secondary. On the limbs of the fold similar nodules are produced by disruption of continuous layers (fig. 10).

The nodules are normally matrix-supported and are commonly concentrated into horizons, with interlayers up to 2 m thick free of them. The size range of the nodules is large, and they are unsorted. The nodules are prolate, with X parallel to the steeply plunging mineral lineation and to Z-style fold axes. However, the nodules are locally deformed around a younger minor set of folds. They range from 2 cm to over 1 m in diameter in Y-Z section with X:Y commonly greater than 10:1 (James, 1976). Some of the largest nodules are penetrated locally by tongues of carbonate-rich matrix. Some nodules show partial fission into several smaller ones (Nutman *et al.*, 1983, fig. 3). The nodules consist of weakly foliated felsic rock with a fine-grained, mosaic-textured groundmass of alkali feldspar + biotite + muscovite + quartz + minor plagioclase, with quartz and feldspar megacrysts and carbonate in some cases. Tourmaline has been noted in several of the nodules. Within the groundmass are lenses of slightly coarser grained muscovite + alkali feldspar + biotite. These lenses have ragged terminations, and resemble fiamme (Nutman *et al.*, 1983, fig. 5). The texture and composition of the nodules suggest they could be derived from felsic volcanic rocks.

The protolith of the conglomeratic unit was probably a mixture of altered felsic volcanic rock and felsic detrital sediment. If the nodules are of sedimentary origin, then the massive, graded character of the matrix and the large size and nonsorted nature of the nodules would suggest that the conglomerate unit consists of high density mass flows, such as debris flows (e.g. Dimroth, 1982). However, away from the hinge of the fold at the specified locality there is a progression from nodules to tabular bodies and finally continuous layers (Rosing, 1983; Nutman *et al.*, 1984a), and in a large scale context this conglomeratic structure occurs in a part of the Isua supracrustal belt that contains prominent Z-style folds. Thus a tectonic origin for this unit is most likely. The conglomeratic structure was probably best developed at the specified locality because the rather unusual layering plus carbonate-rich nature of the matrix resulted in a greater ductility contrast than is normal between layers of the unit.

Graded bedding

Variation across layers in the abundance of feldspar granules in a finer-grained, relatively homogeneous matrix is preserved locally in felsic rocks. In the felsic formation of sequence A graded bedding occurs as packets of beds, all facing the same way. On the eastern shore of lake 678 m graded bedding occurs as units up to 50 cm thick, with sharp tops and bases (fig. 6; Rosing, 1983; Nutman *et al.*, 1983, 1984a). Eastwards, graded beds of the same type are preserved less commonly. Near the base of the felsic formation of sequence B graded beds less than 10 cm thick and

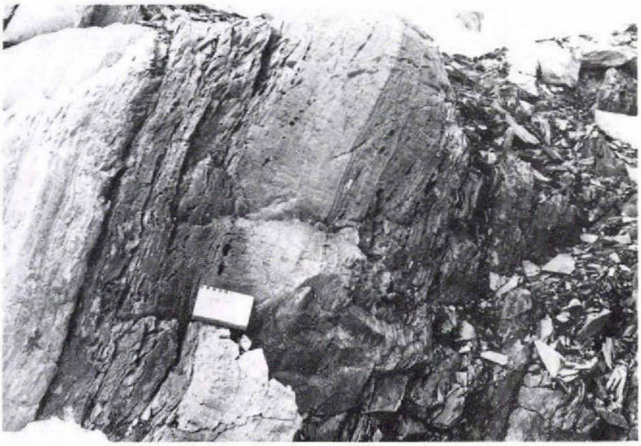

Fig. 11. Graded layering in felsic rocks of the variegated schist formation, in the northwest of the Isua supracrustal belt. 65°8′18″N, 50°11′0″W. Photograph courtesy of M. Rosing.

with sharp tops and bases are locally preserved (Nutman *et al.*, 1983). Graded beds also occur in the variegated schist formation. In the northwest of the Isua supracrustal belt these are well preserved in an augen of low deformation (V. R. McGregor, personal communication, 1981; Rosing, 1983; Nutman *et al.*, 1984a). Complete graded units show a transition from felsic to pelitic and are up to 50 cm thick (fig. 11). The bases of the graded units are sharp and locally erosional. The felsic parts appear to have been silicified and locally contain clasts of mafic or pelitic material (Nutman *et al.*, 1984a, fig. 4e), and upwards they become finer grained and darker in colour. Dark grey to black mafic metapelite overlies felsic parts of the units. Commonly there are packets of several graded felsic layers with little or no interlayered pelite. These graded units can be matched with Bouma sequences described by Walker (1984). As most units of felsic rocks contain locally preserved graded bedding, at least parts of them were deposited from turbidity currents. The source of the felsic sediment is discussed in a later section.

Akilia association

Akilia association units of the Isukasia area are considerably smaller than the Isua supracrustal belt and consist predominantly of amphibolite. Subordinate lithologies, listed in decreasing importance are: ultramafic rocks, quartz-rich metasediments and felsic, pelitic and calc-silicate rocks. Apart from the ultramafic rocks, most of these rocks have a strong compositional layering, interpreted as tectonically modified primary structures. Layering is commonly disrupted by coarse-grained segregations that probably formed under high-grade metamorphism.

Unlike the Isua supracrustal belt, a coherent stratigraphy is not apparent for these smaller units of supracrustal rocks. Apart from one unit in the southwest of the area, they all bear a strong lithological resemblance to formation A1 of the Isua supracrustal belt. The anomalous unit is dominated by well-layered brown-weathering biotite-amphibolite with interlayers of mica schist and metachert. Furthermore it is associated with a chloritic amphibolite that resembles the garbenschiefer. This supracrustal unit strongly resembles formation A3 of the Isua supracrustal belt.

Amphibolites

By far the most common variety is hornblende-plagioclase amphibolite layered at a 0.5 to 10 cm scale. Some layers of these contain garnet or clinopyroxene, and layering is locally interspersed with thin seams rich in either biotite or hornblende. Locally there are more homogeneous, concordant units of amphibolite up to 5 m thick that may be associated with them. A distinctive, but not very common, type of amphibolite is a clinopyroxene-rich, melanocratic variety that forms concordant units up to 5 m thick. It is irregularly striped, with clinopyroxene and hornblende lenses distributed in a finer-grained plagioclase and hornblende matrix. They grade into coarse-grained clinopyroxene-hornblende rocks, locally containing minor amounts of orthopyroxene. A third, not particularly common variety is homogeneous to faintly laminated, hornblende-plagioclase amphibolite, commonly found in association with the clinopyroxene-rich amphibolite. Like similar rocks of the Isua supracrustal belt, the clinopyroxene-rich rocks and homogeneous amphibolites might be derived from fractionated, layered igneous bodies (Nutman *et al.*, unpublished). Leucoamphibolites are rare. Where relatively undeformed they are homogeneous, medium grained plagioclase–hornblende rocks that lithologically resemble Akilia association leucogabbroic intrusive sheets south of Godthåb (Nutman, 1980; Nutman *et al.*, unpublished). Locally they are found in association with enclaves of metagabbroic amphibolite. A unit of faintly banded, chloritic leucoamphibolite less than 25 m thick that resembles the garbenschiefer occurs in a supracrustal unit in the southwest of the Isukasia area (Plate 1).

Quartz-rich metasediments

These form units generally less than 1 m thick. Two main types of lithology occur: quartz-magnetite banded iron formation and quartz-amphibole-clinopyroxene-garnet rock. Layering in the quartz-magnetite rocks is well developed at a scale of less than 1 cm, with the magnetite-rich layers subordinate to the quartz-rich ones. Some of the latter type also have a well developed planar layering, but irregular and lensoid layers of the ferromagnesian minerals in a quartz matrix are more common. Calc-silicate rocks, dominated by amphibole and clinopyroxene with subordinate quartz and carbonate occur locally.

Felsic and peltic metasediments

These are only of local importance. Their main occurrence is in a supracrustal unit in the southwest of the Isukasia area (Plate 1). In this unit there is a 50 m thick packet of mesocratic, brown-weathering biotite-garnet amphibolites layered at a 2 to 10 cm scale, which contain thin horizons of biotite-rich schist. Associated with them is a distinctive garnet-biotite-quartz-plagioclase schist unit only 1 m thick, which can be followed for almost 2 km. Elsewhere there are sporadic occurrences of faintly layered felsic gneisses, which form units up to 1 m thick within supracrustal units. Most contain hornblende and biotite, but some are garnetiferous.

Ultramafic rocks

In the larger supracrustal units in the southwestern corner of the Isukasia area ultramafic units are up to 50 m thick and can be followed for up to 2 km along strike (Plate 1). The relation between the supracrustal rocks and the ultramafic units is not clear in the field, but as they do not continue into the adjacent gneisses, they are grouped with the supracrustal rocks. They probably correlate with the ultramafic rocks within the Isua supracrustal belt. Isolated pods of ultramafic rocks occur within the gneisses. In areas of low deformation, where relations with the adjacent gneisses are preserved, they can be distinguished from ultramafic Tarssartôq dykes. Where strain is higher it is generally impossible to tell in the field whether ultramafic pods belong to the supracrustal sequence or are broken-up ultramafic Tarssartôq dykes. Ultramafic rocks are homogeneous or have faint compositional banding. These rocks commonly consist of ortho-amphibole with some relict olivine, orthopyroxene and clinopyroxene; and some contain carbonate, talc or serpentine locally. Chadwick & Crewe (1982) reported chromite-rich ultramafic inclusions in the gneisses no more than 10 km south of 65°N, but this lithology has not been found in the Isukasia area.

Protoliths of the supracrustal rocks

There are five main compositional categories of supracrustal rocks found in the area; amphibolites, felsic rocks, mica schists, chemical sediments and lithologies that are mixtures of the above types. In addition, these rocks are found in association with garbenschiefer and ultramafic rocks. The field and geochemical characters of these lithologies where least altered are reviewed here, with the aim of identifying their protoliths.

Amphibolites

The commonest amphibolites are laminated varieties. Most of them may have been formed under water because they are interbanded with metachert and banded iron formation. Appel (1977) reported thin amphibolitic layers within major banded iron formation units, and interpreted them as aeolian differentiates of basaltic ash. Banded amphibolites vary greatly in appearance in the field, but petrographically their variation is more limited.

Where higher-grade assemblages are preserved, they contain hornblende, plagioclase (An >50), garnet or clinopyroxene locally. Cummingtonite, quartz and biotite also occur. These assemblages are characteristic of upper to middle amphibolite facies metamorphism. Some amphibolites contain a large amount of biotite, which in some cases clearly overgrows amphibole, suggesting secondary K_2O-enrichment. Retrogression is common, with chloritisation of garnet, overgrowth of hornblende and clinopyroxene by pale amphibole, increase in the amount of clinozoisite and reduction of An content of plagioclase. In most extreme cases (in fault zones) amphibolites are converted to greenschists with no preservation of relics of higher metamorphic grade assemblages.

The amphibolites have a significant range in composition (Table 2), which to some extent may be due to variation in the proportion of mafic to felsic layers among the samples analysed. The composition of more homogeneous layers found within these amphibolites falls in the range of the banded types. All these amphibolites are probably of Archaean high-Fe tholeiite affinity (Gill *et al.,* 1981). The clinopyroxene-rich amphibolites that occur locally in the Isua supracrustal belt and

Table 2. Representative analyses of supracrustal rocks in the Isukasia area

Sample	225956	292258	225996	292107	158497	167690	158526c	167655	171756	175554	117988a	292162
SiO_2	47.79	50.86	50.01	87.19	13.74	57.34	66.56	60.92	45.58	49.85	38.84	75.98
TiO_2	0.33	1.10	1.17	0.04	nd	0.39	0.51	0.67	0.22	0.31	0.03	nd
Al_2O_3	8.43	13.70	13.76	nd	0.05	16.59	16.34	15.45	15.33	17.08	0.38	0.16
Fe_2O_3	3.32			10.87	1.18	10.97	0.77	2.19	2.01	0.16	1.73	7.34
FeO	8.43	12.83	15.34		4.07	7.53	3.16	4.29	7.31	9.47	6.00	11.01
MnO	0.34	0.31	0.26	0.11	0.59	0.40	0.07	0.14	0.18	0.25	0.12	0.39
MgO	9.67	5.43	5.52	0.76	16.80	2.64	1.21	2.46	16.30	10.40	48.90	2.59
CaO	17.20	9.23	10.21	0.34	28.22	0.85	2.96	6.08	7.83	7.71	0.09	2.68
Na_2O	0.39	2.61	1.94	0.01	nd	0.29	1.63	2.43	0.78	1.36	nd	0.02
K_2O	0.26	1.01	0.36	0.10	0.01	1.37	5.86	2.72	0.07	0.10	nd	nd
P_2O_5	0.02	0.12	0.07	0.03	0.17	0.10		0.09	0.05	0.02	0.09	0.03
loi	1.44	1.07	0.83	2.50	0.22	2.20	0.94	0.92	4.65	1.88	3.27	1.26
CO_2	1.40				35.06		0.65	1.16				
	99.23	99.87	99.50	99.40	100.11	100.67	100.84	99.52	100.30	98.59	99.45	101.46
ppm												
Cl		433										
Rb	7	34	1	2	nd	65	145	120	nd	4	nd	
Sr	70	106	90	nd	23	30	75	55	55	61	nd	10
Ba	15	133	34	6	60	84	430	182	21	nd	27	
Y	9	25	27	2	9		5		10	16	nd	
Zr	33	109			4	113	164	175	18	28	nd	
Pb	30	11			10		26		1	9	1	
Th	nd	9				1						
Cu	14				nd	38	nd	55	nd	16	nd	
V	140	197	274	9		98		73		137		
Ni	436	53	29	35	6	205	23	85	540	160	3975	
Cr	1035	119	55	28	20	500	65	155	1090	279	1040	
La	1.47					8.04						
Ce						16.40				2.56		
Nd	2.65					7.00						
Sm	0.87								0.33	0.31		
Eu	0.33					0.61			0.17	0.13		
Gd	1.13											
Tb									0.18			
Dy	1.35											
Er	0.85											
Yb	0.82					1.41			1.33	0.71		
Lu	0.12					0.19			0.39	0.19		
Hf						2.14						

225956 and 292258 associated clinopyroxene-rich amphibolite and homogeneous grey amphibolite respectively, lower part of the amphibolite formation.
225996 banded amphibolite, upper part of amphibolite formation.
292107 silica-rich banded iron formation, western part of the upper banded iron formation.
158497 carbonate rock with calc-silicate layers, upper part of the calc-silicate formation.
167690 garnet-mica schist, upper part of the mica schist formation.
158526c layered felsic rock, felsic formation of sequence A.
167655 layered felsic rock, felsic formation of sequence B.
171756 and 175554 mafic and felsic garbenschiefer unit amphibolites respectively.
117988a metadunite. All these rocks come from the Isua supracrustal belt.
292162 banded iron formation enclave, in the gneisses south of the Isua supracrustal belt.

nd not detected
loi loss on ignition
Analyst: XRF, Ib Sørensen, GGU

more frequently in the Akilia association units are interlayered with metasediments and other amphibolites, suggesting they are of volcanic or shallow intrusive origin. Their distinctive geochemical characters are high MgO, CaO, Cr and Ni and low

Al_2O_3 and TiO_2 contents. The grey, rather homogeneous amphibolites found in association with clinopyroxene-rich amphibolites are of Archaean high-Fe tholeiite affinity. Thus the clinopyroxenic amphibolites and the homogeneous amphibolites are interpreted as clinopyroxene rich and residual liquid parts respectively of layered picritic-tholeiitic sills and flows. The hornblende-rich amphibolites at the base of the calc-silicate formation are interpreted as derived from picritic units.

Felsic rocks

Locally preserved graded bedding shows that at least some of the felsic rocks are sediments deposited from turbidity currents (Rosing, 1983). The lack of coarse-grained quartz suggests that they do not contain significant input of material derived from emerging granitic plutons or gneissose basement. On the other hand, their fine-grained felsic nature and their close association with rocks of volcanic origin suggests they could have been derived from local felsic volcanic rocks (Bridgwater & McGregor, 1974; Nutman *et al.*, 1984a).

The felsic gneisses contain plagioclase–biotite–quartz–muscovite±K-feldspar±garnet, are commonly well-banded at a scale of 1 cm or less and are fine grained. In some cases larger clasts consisting of several plagioclase grains are present. From field evidence these could be of detrital origin. Carbonate is present locally, but seems to be due to secondary processes. Epidote overgrowths and some muscovite are secondary.

Geochemically the most outstanding characteristic of most of the felsic rocks is their high K_2O contents, particularly those of the felsic formation of sequence A (Table 2). In contrast Na_2O contents are low and show a moderately good negative correlation with K_2O (Nutman *et al.*, 1984a). For samples not strongly affected by post-depositional alteration, contents of SiO_2 of between 62 and 67 wt% and of Al_2O_3 of between 14 and 16 wt% are normal (Table 2). A few samples of the felsic formation of sequence B have rather higher SiO_2 contents, which is perhaps due to silicification during diagenesis. Ba abundances are commony high, whilst Rb contents are low, in view of the potassic nature of most of these rocks. Ni and Cr abundances are low in the felsic formation of sequence A, suggesting that input of material from a mafic source is not important. In the felsic formation of sequence B, Ni and Cr abundances are somewhat higher (Table 2), suggesting that there might have been a minor input from a mafic source for these rocks.

High K/Na values are a character of felsic volcanic rocks where K-enrichment and Na-depletion of sodic protoliths was due to alkali exchange between rocks and hot saline groundwaters causing growth of potassic authigenic minerals or to subaerial weathering (Radulescu, 1966; Krogh, 1977; B. Fryer, personal communication, 1982). Alkali exchange by either of the above processes provides an excellent explanation of Na/K variation, but fairly constant SiO_2 and Al_2O_3 contents of these felsic rocks (Nutman *et al.*, 1984a). Field evidence shows that at least some of these rocks were deposited from turbidity currents. These rocks could represent sediment formed from weathered or altered (emergent?) andesitic or dacitic volcanic rocks that were periodically dislodged, transported down the flanks of volcanic centres as turbidite flows and redeposited as graded beds to form prograding wedges of sediment. More massive, less potassic units, such as in the western part of the felsic formation of sequence A could be related felsic volcanic rocks.

Mica schists

The mica schists are faintly layered at all scales, but no structures are preserved to indicate how they were formed.

A mica schist from the mica schist formation of sequence B contains garnets up to 5 mm diameter in a groundmass of quartz, biotite, chlorite and muscovite. Minor amounts of staurolite coexisting with garnet also occur. The garnets have a core rich in inclusions (mostly quartz and Fe-Ti oxide) with a distinct rim with fewer inclusions. The inclusions form trains that are oblique to the foliation outside the garnets. Garnets are partially altered to chlorite. The biotite, chlorite and muscovite of the groundmass form a foliation that wraps around the garnets. Also present in the groundmass are blebs of Fe-Ti oxide and small blades of graphite.

Most of the mica schists are potassic, but are richer in FeO, MgO, Ni and Cr (Table 2) compared with most other pelites. Thus it is likely that they were derived from mafic rocks (Boak *et al.*, 1983; Nutman *et al.*, 1984a). Heavy REE abundances in the mica schists are similar to those of typical layered amphibolites in the Isua supracrustal belt; but light REE are enriched relative to the amphibolites, perhaps due to weathering during formation (Boak *et al.*, 1983; Nutman *et al.*, 1984a).

Chemical sediments

The lithological variation of these rocks is mirrored by a considerable range in their composition (Table 2). The chemical sediments comprise quartz±magnetite±clinopyroxene±amphibole(s)±garnet±biotite±carbonate± graphite±pyrite and in rare cases Fe-rich orthopyroxene and staurolite.

The petrography of these rocks varies greatly – three samples representative of banded iron formation, metachert and carbonate-rich rocks are described. In the banded iron formation quartz-rich layers alternate with mafic layers at a 3 mm or less scale. The mafic layers are rather thinner than the quartz-rich layers. The quartz-rich layers contain minor amounts of carbonate. The mafic bands consist predominantly of magnetite. Quartz and magnetite commonly coexist, but in some cases grunerite, partly altered to hornblende, mantles the magnetite. In the metacherts the scale of layering may be over 1 cm and is marked by flecks of magnetite, amphibole (hornblende or grunerite mantled in hornblende) or clinopyroxene (partly altered to amphibole). Some samples contain minor amounts of either carbonate or graphite. In slightly more mafic varieties, clinopyroxene and amphibole lenses up to 1 cm thick form discontinuous layers. These may represent continuous layers broken-apart during deformation. In the representative carbonate thin section (upper part of the calc-silicate formation) dolomite occurs with tremolitic amphibole, diopside and minor white mica. The grain size is up to several millimetres, but there are zones where the grain size is less, probably due to higher strain. The tremolite and the white mica occur as lenses orientated parallel to these higher strain zones, whilst clinopyroxene forms equidimensional, rather irregular grains. Twinning in the dolomite is commonly deformed.

The upper part of the calc-silicate formation is not only lithologically but also chemically distinct from other chemical sediments of the area, as shown by their higher CaO relative to FeO+MgO contents (Table 2). These differences may suggest that the protoliths of the calc-silicate formation rocks were dolomite/ankerite and chert mixtures, whilst the protoliths of other chemical sediments (apart from anthophyllite metacherts) were once authigenic silicate, siderite, Fe-oxide and chert with minor dolomite/ankerite mixtures. Such differences could be a function of depositional or diagenetic environment. Variations in abundance of Al_2O_3, Na_2O, K_2O, P_2O_5, MnO, TiO_2, most trace elements and REE (Appel, 1980) are erratic (Table 2). These variations are attributed to incorporation of small amounts of terrigenous material of volcanic origin, plus secondary chemical changes (Nutman *et al.*, 1984a). Appel (1979b) suggested that chromite grains in some of these rocks are of cosmic origin, although their present compositions are within the range of metamorphosed terrestrial chromites. The anthophyllite metacherts have higher MgO/FeO, Al_2O_3 and Zr abundances than banded iron formation rocks with which they are interlayered (Nutman *et al.*, 1984a), and perhaps their high Al_2O_3 and MgO content is due to palygorskite in their sedimentary protolith.

Review of models for the formation of Precambrian chemical sediments (e.g. James, 1966; Drever, 1974; Sighinolfi, 1974) suggests that the types of chemical sediments of the Isukasia area (magnetite banded iron formation and carbonates) could be mostly shallow water deposits. This is in agreement with the flat pebble conglomerates locally preserved in metacherts.

Mixed rocks

Well layered rocks containing layers of mixed parentage are common. Their well stratified nature suggests that their mixed parentage is a sedimentary phenomenon. The most common type is rocks of mixed terrigenous and chemical sedimentary parentage. For mixed rocks lying towards the chemical sediment end of the spectrum their mixed parentage may not be apparent in the field because they resemble silicate facies banded iron formation. However, they are readily distinguished on geochemical grounds as the former are richer in Al_2O_3, Ni and Cr and total REE relative to the pure chemical sediments (Nutman *et al.*, 1984a). Mixed rocks towards the felsic and basic end of the spectrum are generally detected by their higher SiO_2, CO_2, FeO or in some cases CaO contents compared with rocks of solely terrigenous parentage.

Garbenschiefer

Because of its MgO and Al_2O_3 rich composition the garbenschiefer is interesting in terms of its petrogenesis, and also because it is mineralogically sensitive to changes in metamorphic conditions (Rosing, 1983).

Rosing (1983) made detailed petrographic studies of the garbenschiefer. The least hydrated and altered garbenschiefers are pale brown to pale greenish grey amphibolites. This type comprises predominantly Fe-Mg amphiboles and An >70 plagioclase with some quartz, kyanite and rutile. Carbonate and biotite may be present in small amounts and are late alteration products. The type is commonly faintly layered at a 5 mm scale, and the Fe-Mg amphibole grains are up to 10 mm long, and commonly form a garbenschiefer texture. There is a progressive overgrowth of the Fe-Mg amphibole by hornblende, which is coupled with the growth of chlorite, disappearance of kyanite and decrease in plagioclase content. The final stage is a chlorite–hornblende–quartz–ilmenite±plagioclase rock. In areas of low deformation garbenschiefer texture survives. Rosing (1983) attributes these changes to progressive hydration during late Archaean and Proterozoic events.

Garbenschiefers have high Al_2O_3, (and like komatiites) high MgO, low TiO_2, P_2O_5, Zr and Ba contents (Table 2), and low FeO/MgO and Y/Zr ratios. Geochemical traverses have not revealed any regular geochemical trends across, or compositional contrast between units. Gill *et al.* (1981) suggested that the garbenschiefer's precursor was high-Al basic igneous rock that contained up to 25 normative percent entrained magnesian olivine and orthopyroxene. Thus less mafic (more common) compositions are regarded as a first approximation to represent liquid compositions. These possible liquid compositions have low FeO/MgO values and high Al_2O_3 content when compared with common basaltic liquids. Nutman *et al.* (unpublished) suggest that the primitive liquid that gave rise to the garbenschiefer may have formed from clinopyroxene-depleted mantle that had already undergone partial melting to supply basic magma, possibly the parental liquids of the tholeiitic rocks interlayered with metasediments (of the Isua supracrustal belt and the Akilia association). Olivine followed by orthopyroxene fractionation during ascent of these second generation melts gave rise to the garbenschiefer compositions.

Ultramafic rocks

All the ultramafic rocks that are not associated with or grade into amphibolite (or are strongly carbonated) are CaO poor. They have lower Al_2O_3 and higher MgO (*c.* 40 wt%) contents than primitive liquids such as peridotitic komatiites (Nutman *et al.*, unpublished). Recalculated as H_2O and CO_2 free norms the ultramafic rocks are predominantly olivine-bearing orthopyroxenites with minor clinopyroxene. Some samples are normatively quartz-bearing orthopyroxenite, suggesting the SiO_2 has been mobile. This in turn throws doubt on whether pyroxene/olivine ratios in olivine normative varieties are igneous, or whether they too have been affected by SiO_2 mobility. Ni/Cr ratios may not be original as Cr has probably been leached from the ultramafic rocks and fixed in adjacent lithologies in chrome muscovite (Dymek *et al.*, 1983). Variation of Rb, Sr and Ba is erratic, but abundances of these elements are generally low. Variation of Y, Zr, Nb, P_2O_5 and TiO_2 are somewhat irregular and do not show chondritic ratios. Chondrite normalised REE abundances are low, with moderate $(La/Yb)_N$ values. Variation in the light REE chemistry could be an alteration effect. It is clear that the igneous precursors of these rocks have been significantly disturbed by metasomatism. Their local layering and occurrence as locally discordant sheet-like bodies suggest they are intrusive, and their compositions suggest they are derived from basic liquid, olivine and pyroxene mixtures. Some might represent cumulate facies of basic intrusions from which the residual liquid had been expelled, whilst others could have been feeders to higher level basic volcanics and intrusions.

Depositional environment of the supracrustal sequences

The Isua supracrustal belt and Akilia association rocks are interlayered sediments and volcanic rocks, and show marked lateral variation. All terrigenous sediments are interpreted as derived from volcanic sources, as there is no evidence that older basement or emerging granitic plutons contributed to the terrigenous material (Nutman *et al.*, 1984a; Hamilton *et al.*, 1983). The amphibolites were probably derived from flows and tuffs of tholeiitic affinity. Felsic rocks were probably derived from andesitic to dacitic volcanics that underwent severe alteration during weathering and redeposition. Mica schists are predominantly pelitic rocks derived from altered mafic rocks. The presence of graded bedding and chemical sediments throughout the succession suggests that most of these rocks were deposited under water. The intimate association of chemical sediment and terrigenous material occurs at scales down to those of individual beds and shows that volcanic activity and deposition of chemical sediments were coeval. Also the intimate association of amphibolites, mica schists and felsic rocks suggests that basic and acid-intermediate volcanism overlapped. The Isua supracrustal rocks contain a considerable amount of felsic material and may thus be rather atypical of the supracrustal succession to which it belongs. On the other hand, the smaller Akilia association units in the Isukasia area are perhaps more representative of early Archaean supracrustal rocks elsewhere in the North Atlantic Craton (McGregor & Mason, 1977; Nutman, 1980, 1982b; Bridgwater *et al.*, 1978, in press; Chadwick & Crewe, 1982; Hall, 1978; Friend & Hall, 1977; Allaart *et al.*, 1977). However, on the whole these rocks are poorly preserved so that less information can be gleaned from them than from the rocks of the Isua supracrustal belt. They are predominantly basic volcanic rocks mostly of tholeiitic affinity with interlayers of chemical metasediment and minor amounts of felsic rocks, perhaps derived from intermediate volcanic parents. Ultramafic rocks are interpreted as parts of layered intrusions and feeders to intrusions and volcanic rocks. The garbenschiefer and Akilia association leucocratic amphibolites elsewhere in the area are interpreted as intrusions of strongly fractionated high-Mg-Al magma into the supracrustal rocks.

The present favoured interpretation of the Isua depositional environment (Nutman *et al.*, 1984a; Boak *et al.*, 1983; Rosing, 1983) is of an immersed volcanic region remote from exposed areas of significantly older crust. Water depths during at least some of the stratigraphic record were shallow. The only units that preserve evidence of deposition in an environment with appreciable topography are felsic rocks, and it is likely that these were sediments reworked off the flanks of felsic volcanic centres. The volcanic rocks and volcanic sediments were coeval with deposition of chemical sediments which locally formed thick units and show evidence of shoaling.

GNEISSES, INALUK DYKES AND PEGMATITES

Orthogneisses, the bulk of which are of early Archaean age, underlie approximately 80 percent of the Isukasia area. The rocks discussed here have been referred to collectively as the Isua gneisses in some publications. They form part of the Amîtsoq gneiss complex of West Greenland (McGregor, 1973, 1979). Although their polyphase nature was recognised in earlier studies (Moorbath *et al.*, 1973; Bridgwater *et al.*, 1976; McGregor, 1979) it is only since 1980 that their history has begun to be elucidated in detail (Table 1).

The least deformed gneisses occur in the core of the central gneisses (the area within the arc of the Isua supracrustal belt). These are slightly deformed, polyphase, sheeted meta-igneous complexes consisting of two main groups of gneisses: (1) older, tonalitic *grey gneisses* that are intruded by (2) younger granitic *white gneisses,* separated in time by the intrusion of the *Inaluk (basic) dykes* (figs 12 & 13). In addition, there are two generations of fine-grained tonalitic sheets and a generation of pegmatitic gneiss sheets (Table 1). These gneisses provide the best opportunity known for the study of the earliest-known sialic rocks. Under deformation these (intersheeted) gneiss phases are transformed into banded or schlieric gneisses in which individual components of the complex can no longer be identified with certainty. Within these banded gneisses that dominate the gneiss complex there are augen of lower deformation, where grey gneisses are cut by Inaluk dykes and then broken-up by white gneisses (Nutman, 1982a). Granitic and pegmatitic sheets that post-date the Tarssartôq dykes occur locally (Table 1). These are volumetrically insignificant compared with the pre-Tarssartôq dyke (early Archaean) components of the gneiss complex. The late Archaean Taserssuaq tonalite (Garde *et al.*, 1983, 1986) and associated banded gneisses and granitoid sheets and lenses occur west of the Ataneq fault (Plate 1).

Early isotopic studies of the gneisses from the Isukasia area were mostly based on samples of highly deformed, commonly strongly banded gneisses close to the Isua supracrustal belt. Using U-Pb, Pb-Pb and Rb-Sr isotopic systems, ages of *c.* 3800 to 3600 Ma have been obtained from these rocks (Moorbath *et al.*, 1972, 1975, 1977), with suggested crustal residence prior to 3600 Ma of less than 200 Ma (Moorbath *et al.*, 1973). More detailed interpretation of the data is not feasible because of the heterogeneous nature of these suites of samples. More recent Sm-Nd, U-Pb and Rb-Sr studies on suites of samples comprising individual phases of the gneiss complex have not revealed any component older than the Isua supracrustal belt, and have shown that the grey gneisses are 3750 to 3700 Ma old, the white gneisses *c.* 3600 Ma old and the pegmatitic gneisses are *c.* 3400 Ma old (Hamilton *et al.,* 1983; Baadsgaard, 1983; Baadsgaard *et al.,* in press a,b). A swarm of undeformed pegmatite dykes concentrated south of the Isua supracrustal belt give an Rb-Sr whole rock age of *c.* 2550 Ma (Baadsgaard *et al.,* 1985). An undeformed microgranite dyke that cuts the Isua supracrustal belt gives a Rb-Sr whole rock age of *c.* 1600 Ma (Kalsbeek *et al.,* 1980, Kalsbeek & Taylor, 1983).

Gneiss phases

Grey gneisses

These rocks are predominantly metatonalites, but also include some granodiorites and quartz-diorites. They form at least 70 percent of the central gneisses and a somewhat higher proportion of the gneisses to the south of the Isua supracrustal belt. In their least deformed state they have a weakly developed gneissic banding defined by biotite and are plagioclase-flecked. They were cut by sparse pegmatite veins prior to the intrusion of the Inaluk dykes. The grey gneisses are polyphase, with rafts of mafic tonalite occurring in more voluminous, leucocratic types. The mafic tonalite is most abundant along the inner western margin of the Isua supracrustal belt, but it also occurs in a mappable unit in banded gneisses approximately 5 km south of the Isua supracrustal belt. Locally the grey gneisses have coarser-grained patches and lenses, some of which contain hornblende, which developed prior to or concomitant with intrusion of the white gneisses.

The grey gneisses are essentially plagioclase (An 40–25) +quartz+biotite rocks with minor microcline. Where least deformed the grey gneisses contain plagioclase megacrysts up to 1 cm long. More mafic varieties of the grey gneisses also contain hornblende. Biotite forms mats that define the gneissosity and is normally coarse grained, but locally there are overgrowths of fine-grained biotite, commonly associated with colourless mica. Hornblende, when present, generally lies within the foliation, and is either replaced by or appears to coexist with biotite. Locally hornblende appears to be late, forming blocky grains that overgrow the gneissosity. These hornblendes are sieved, containing inclusions of quartz, apatite and magnetite, and have locally been altered around their edges to biotite. The abundance of allanite in the grey gneisses seems to correlate with the abundance of white gneiss in the vicinity. Allanite is commonly rimmed by epidote. Alteration, shown by partial replacement of biotite by colourless mica, and by replacement of plagioclase by albite, phengite and epidote is common, particularly along the northern and western edges of the area. This alteration is probably associated with the Ataneq fault zone.

Grey gneisses contain inclusions of supracrustal rocks, and (locally discordant) grey gneiss sheets intrude the Isua supracrustal belt (Moorbath *et al.,* 1977; Bridgwater *et al.,* 1978). Most of these sheets are finer grained than typical grey gneisses, probably due to strong deformation. Many of the sheets are severely epidotised and sericitised. However, the centres of some of the thicker sheets are coarser grained and resemble typical grey gneisses.

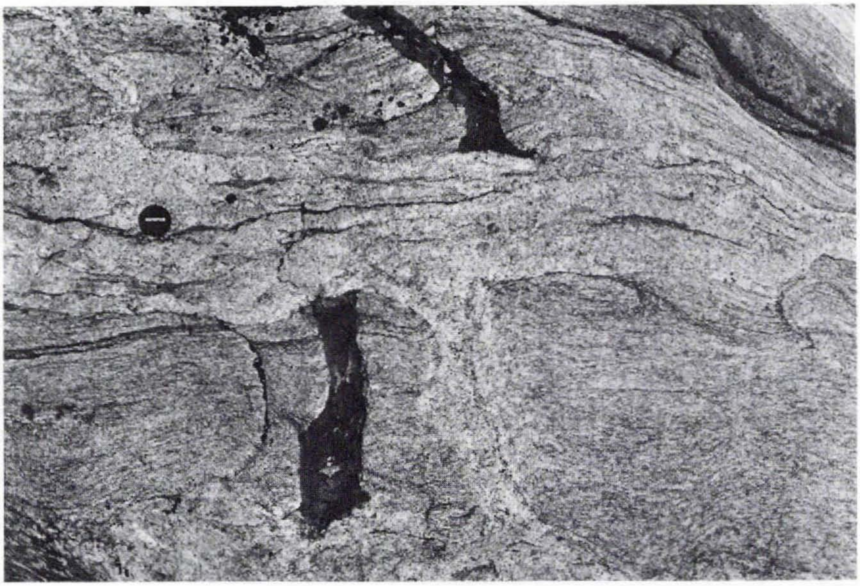

Fig. 12. Grey gneiss (pale variety) cut by Inaluk dyke and then by white gneiss sheets. Note the irregular margins of the Inaluk dyke and its offset across the white gneiss sheet. Note also the locally blurred contacts between the grey and the white gneisses and the biotite schlieren in the white gneiss. 65°11′10″N, 50°1′20″W.

Inaluk dykes

Sparse basic dykes, which intrude the grey gneisses and are cut by the white gneisses are known as *Inaluk dykes* (Nutman *et al.*, 1983). They have not been found in the Isua supracrustal belt, but at a few localities they cut amphibolite inclusions in the gneisses that are probably of supracrustal origin. Inaluk dykes are fine to medium grained, hornblende and biotite bearing intrusions normally less than 2 m wide that in the least deformed parts of the central gneisses are steeply inclined and highly discordant to the (younger) white gneiss sheets (fig. 12). Commonly the Inaluk dykes have a faint flow(?) banding and irregular, cuspate margins. Some contain small, altered plagioclase megacrysts. Locally wisps of basic material are found in white gneisses where they cut through Inaluk dykes, and in one place Inaluk dyke material is either cut by, or is intimately intermixed with granitic gneiss. This suggests that some Inaluk dykes were intruded after intrusion of the white gneisses had begun.

The Inaluk dykes consist essentially of hornblende, biotite and plagioclase with minor quartz and alkali feldspar. The proportions of hornblende and biotite vary greatly, perhaps due to biotite growth during potassium metasomatism when the white gneiss sheets were intruded. Most of the biotite seems secondary and forms a foliation concordant to that in the host gneisses. Accessory mineralogy is diverse; apatite, zircon, ilmenite/magnetite and sphene are abundant in most dykes, but in a few they are rare.

Locally hornblende contains blebs of magnetite/ilmenite, perhaps of exsolution origin. Further alteration marked by epidote overgrowths and replacement of plagioclase by albite, phengite and epidote is locally important.

Poorly exposed ovoid hornblendite bodies occur in the gneisses southwest of the Isua supracrustal belt. They are less than 100 m wide and up to 1 km long. They are generally strongly agmatised by pale gneisses and contain enclaves of grey gneisses. At the end of one of these bodies there is an igneous breccia of biotite-hornblende rich matrix that contains jumbled blocks of grey gneiss, resembling late Archaean igneous breccias from the Buksefjorden region further south described by Chadwick & Coe (1983). These hornblendite bodies are correlated with the Inaluk dykes.

White gneisses

These occur as swarms of anastomosing sheets generally less than 25 m thick (fig. 13). By far the most important white gneiss lithology is white, medium-grained, homogeneous granite. Subordinate phases are potassic granodiorite similar in appearance to the granites, and sheets of white and pale pink pegmatite of granitic composition. Pegmatite also occurs as selvages to granitic white gneiss sheets. The white gneisses are not distributed evenly in the central gneisses. In the northeast they coalesce to form a single lens of *c.* 10 square kilometres in extent, of which the centre is devoid of grey gneiss inclusions, whilst in the northwest white gneisses are generally scarce. Overall the white gneisses form up to 30 percent of the central gneisses and a lower proportion of the gneisses south of the Isua supracrustal belt.

The white gneisses consist essentially of microcline, plagioclase (An_5-An_{20}) and quartz, with minor biotite and muscovite. Where least deformed they are equigranular and medium grained. Grain boundaries are generally complex and amoeboid, with myrmekite commonly found between feldspar grains. Muscovite rims biotite and locally replaces it, with production of finely disseminated magnetite/ilme-

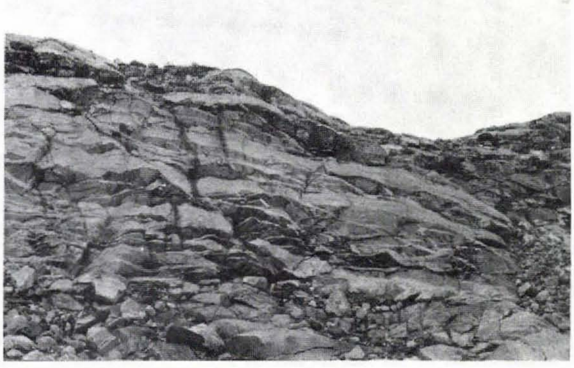

Fig. 13. Grey gneisses cut by Inaluk dykes (see right hand side of photograph), intruded by anastomosing swarms of white gneiss sheets. White gneiss sheets dip at less than 30°. Cliff is 50 m high. 65°10′25″N, 50°0′50″W.

nite. Subhedral epidotes occur that contain allanite cores. Alteration is common, with irregular overgrowths of epidote and breakdown of plagioclase to albite, phengite and epidote. Locally there are biotite-rich smears and clots up to 5 mm thick. These are interpreted as remains of 'digested' thin slivers of grey gneisses and other country rocks.

In the north of the core of the central gneisses, the white gneiss sheets are gently to moderately inclined and cut the Inaluk dykes with high angles of discordance, showing that post white gneiss ductile deformation there was minimal. These characters suggest that the bulk of the white gneisses were emplaced as gently inclined bodies. Inaluk dykes are locally displaced laterally across white gneiss sheets (fig. 12). Commonly contacts between grey gneiss and white gneiss are slightly blurred, and grey gneisses that occur as inclusions in, or close to, large units of white gneisses are coarsely recrystallised. In extreme cases such inclusions take on a ghost-like appearance. Thus there is field evidence that voluminous white gneiss units have metasomatised adjacent grey gneisses.

Late grey sheets

Rare, thin, fine-grained tonalitic sheets cut deformed grey and white gneisses.

Pegmatitic gneisses

These are quartzofeldspathic sheets with a gneissose fabric that range from granite to albitic trondhjemite in composition. They are a volumetrically insignificant component of the gneiss complex and mainly occur as sheets up to several metres thick in the Isua supracrustal belt and along lithological boundaries, such as the borders of the Isua supracrustal belt. The white gneisses were already deformed when the pegmatitic gneisses were intruded.

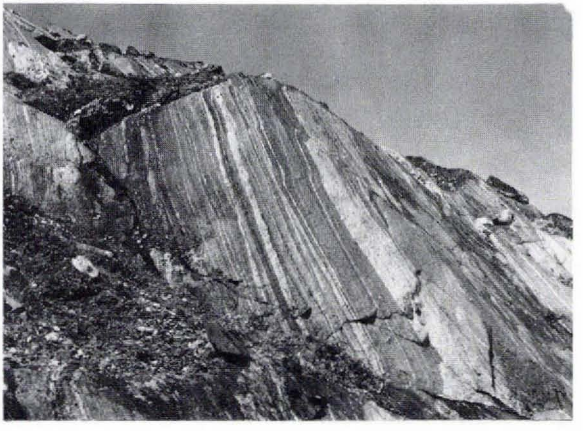

Fig. 14. Grey and white gneisses strongly deformed to give a banded gneiss. 65°6'42"N, 49°59'5"W.

Fig. 15. Pegmatitic lenses in grey gneisses cut by a thin Tarssartôq dyke, 5 km south of the Isua supracrustal belt. Note mafic selvages locally on the pegmatitic lenses. 65°63'58"N, 49°54'10"W.

Pegmatitic gneisses are coarse-grained, feldspar, quartz, mica rocks. Muscovite is the most important mica, biotite only occurring in some sheets. Alkali feldspars are generally heterogeneous with perthite stringers. Normal zoning is common in plagioclase, and myrmekite occurs between feldspar grains. Allanite is a common accessory mineral, and garnet and pyrrhotite occur in some sheets. Strongly developed flaser texture is seen in some units, particularly those apparently intruded along lithological boundaries (Nutman, 1984, fig. 7b). Alteration, shown by sericitisation and overgrowth of epidote is extensive.

Banded gneisses

These are the dominant gneisses south of the Isua supracrustal belt. They also occur in the west and the extreme north of the area. The banded gneisses are generally finer grained than their less deformed white and grey gneiss equivalents in the central gneisses. This is probably due to variation in the degree of deformation during recrystallisation under amphibolite facies metamorphism. The mineralogy of the banded gneisses is essentially the same as that of their grey and white gneiss equivalents in the central gneisses. Biotite forms a prominent foliation, corresponding with the foliation and compositional layering seen in the field.

The banding of these gneisses derives from two main sources. First, it is caused by intense deformation that has reduced the grey and white gneiss components of the migmatite precursor to more or less parallel bands that can be less than 5 cm thick, with obliteration of all intrusive relations (fig. 14). Secondly there is a smaller scale banding and lensoid structure in the grey gneiss components. This is a partial separation into biotite-rich melanosome and quartzofeldspathic leucosome (fig. 15). Such segregation is considered to be the result of amphibolite facies metamorphism and has been described in detail from elsewhere (e.g. Mehnert, 1968). Variation in the metamorphic history, degree of strain and different proportions of grey and white gneisses across the area gives the gneiss complex a greatly varied appearance from place to place. Banded gneisses in the central gneisses (mainly along the contacts of the Isua supracrustal belt) owe their character to strong deforma-

Fig. 16. Granitic gneiss sheet containing inclusion of deformed Tarssartôq dyke. 65°0′11″N, 50°19′55″W.

tion only. Southwards, the gneisses are banded due to combinations of deformation, metamorphic segregation and mobilisation.

Post Tarssartôq dyke pegmatites and gneisses

These occur in small volumes, intruded into the early Archaean gneisses. Two types are distinguished; pegmatite dykes (generally undeformed), and poorly banded granitic-granodioritic gneisses.

The granodioritic to granitic gneisses occur as subconcordant sheets that break up the early Archaean rocks in the extreme southwest of the Isukasia area (fig. 16). Although still commonly discordant to their host, they are markedly deformed and pre-date the Ataneq fault zone. Granodioritic to granitic gneisses also occur as discordant sheets cutting early Archaean rocks north of the faults that mark the northern edge of the central gneisses.

The (*c.* 2550 Ma) pegmatite dykes occur predominantly as a swarm that is concentrated around the southern part of lake 678 m (Nutman, 1982a; Baadsgaard *et al.*, 1985). A few members of this swarm occur in the Isua supracrustal belt and the southern edge of the central gneisses. The pegmatites are subvertical and have a dominant trend of *c.* 080° and are generally less than 5 m wide. However, on the east side of lake 678 m a spectacular member of the swarm occurs that is up to 100 m wide and contains microcline crystals up to 1 m across and large, angular, distinct-edged enclaves of host gneisses. The pegmatites consist of microcline, plagioclase, quartz and interlayered muscovite and biotite. Locally they contain garnet, and allanite is locally abundant in the wall rocks. Over the central and eastern parts of the area these pegmatite dykes are undeformed apart from sheared margins locally. Approaching the major faults in the west of the area similar pegmatites become first sinusoidally folded across, and then swing towards concordance with the gneissic layering. A thin greyish-brown microgranite dyke with a Rb-Sr age of *c.* 1600 Ma occurs in the northeast end of the Isua supracrustal belt (Kalsbeek *et al.*, 1980), and a few lithologically similar microgranite dykes trending east–west and less than 2 m thick cut the central gneisses.

Rocks in extreme west of the area

Multiphase tonalitic to granitic gneisses and the *Taserssuaq tonalite,* all of which contain scattered amphibolite inclusions (Garde *et al.*, 1983), occur west of the Ataneq fault and the zone of strong ductile deformation that runs from the southwest corner of Plate 1 to the northwestern end of the Isua supracrustal belt. Banded gneisses only occur locally as a thin unit against the Ataneq fault. They comprise pegmatite-banded tonalitic hornblende-biotite gneisses that have been invaded by sheets and irregular masses of pale, more homogeneous granitic gneisses. North of 65° remnants of Tarssartôq dykes have not been seen in these gneisses. However, south of 65° these banded gneisses outcrop more extensively and are better preserved, and it is possible that they contain an early Archaean component (Chadwick *et al.*, 1983). The Taserssuaq tonalite is a fairly homogeneous, but multiphase body of hornblende-biotite tonalites, with minor components of early, mafic diorite and late granodiorite. The Taserssuaq tonalite intrudes banded gneisses on its eastern side (Garde *et al.*, 1983), and samples of it yield a U-Pb zircon concordia intercept age of *c.* 3000 Ma (Garde *et al.*, 1986). The Taserssuaq tonalite is truncated by the Ataneq fault so that it is in tectonic contact with the early Archaean gneisses to the east. West of the Ataneq fault there are undeformed granitoid-pegmatite sheet swarms. The most prominent of these swarms trends approximately 020° and dips eastwards. They are affected by movement on the Ataneq fault zone.

Geochemistry and origin of the early Archaean gneisses and Inaluk dykes

Reconnaissance geochemical studies of the Amîtsoq gneisses throughout their known extent (O'Nions & Pankhurst, 1974; Bridgwater *et al.*, 1976; Lambert & Holland, 1976; McGregor, 1979) included samples of banded gneisses close to the Isua supracrustal belt. These studies showed that in their general character these gneisses resembled calc-alkaline gneiss complexes of younger age. Following the division of the gneiss complex of the Isukasia area into several components (Nutman, 1982a; Nutman *et al.*, 1983), Nutman & Bridgwater (in press) made further studies of the gneisses, using samples that are composed of only one phase (Table 3).

Metasomatic alteration of the grey gneisses associated with intrusion of the white gneisses is assessed by comparing samples of grey gneisses 'swamped' in white gneisses (for example from 65° 11′ 13″N, 49° 56′ 28″W) or taken close to white gneiss sheets, with samples of grey gneisses more remote from white gneiss sheets.

Where the grey gneisses have remained as distinct inclusions, major elements (with the possible exception of some K-enrichment) are unchanged, but they have above average LREE, Rb, Pb and Th contents. No significant changes in Zr, Y and Ba content are apparent. On the other hand, where grey gneiss inclusions are ghost-like, not only Rb, Pb, Th and LREE are enriched, but also K, Na and Ca abun-

Table 3. Representative analyses of gneisses in the Isukasia area

Sample	237000	236991	236969	236966	236955	263949	229472	229402	225859	225917	292202	292192
SiO_2	64.12	68.02	70.52	71.72	72.64	47.45	74.16	70.15	72.02	66.10	66.77	67.10
TiO_2	0.47	0.41	0.34	0.27	0.17	0.95	0.01	0.39	0.26	0.45	0.49	0.39
Al_2O_3	14.76	15.22	14.83	14.60	14.40	11.31	14.25	14.70	14.90	15.29	16.21	15.94
Fe_2O_3	2.22	1.29	1.23	0.40	0.25	3.42	nd	0.95	0.86	0.73	0.85	0.80
FeO	5.12	2.66	2.19	1.46	0.99	8.90	0.44	2.79	1.11	3.00	2.63	2.69
MnO	0.08	0.04	0.04	0.03	0.02	0.23	0.01	0.03	0.03	0.05	0.05	0.06
MgO	1.25	1.17	0.71	0.45	0.25	10.64	0.01	0.77	0.47	1.74	1.62	0.99
CaO	4.51	3.66	3.05	1.86	1.45	9.60	0.85	3.13	1.94	3.18	3.86	2.52
Na_2O	3.77	4.62	4.50	4.02	3.94	1.65	2.24	4.65	3.95	3.53	4.75	4.65
K_2O	1.92	1.41	1.14	3.74	4.73	2.77	6.30	1.42	1.87	1.63	1.51	2.88
P_2O_5	0.21	0.14	0.10	0.08	0.06	0.51	0.01	0.12	0.08	0.14	0.16	0.13
loi	0.92	0.71	0.49	0.71	0.54	1.26	0.15	0.54	0.98	2.22	0.83	0.86
	99.35	99.35	99.14	99.34	99.44	98 69	98.42	99.64	98.47	98.06	99.73	99.01
ppm												
Cl	1020	284	52	63	54		55			78		
Rb	64	87	67	141	116	69	225	77	67	46	51	154
Sr	267	263	226	224	118	189	70	259	195	259	510	340
Ba	334	199	178	626	567	2054	270	192	578	239	378	551
Y	26	20	14	6	7	28	nd	7	7	13	13	10
Zr	158	169	201	212	151	89	18	271	190	135	132	160
Pb	12	17	11	27	28	11	110	10	12	13	11	31
Th	1	4	4	16	18	6	4	10	11	8	5	5
Cu	29	16	1	6	2	26	nd	2	4	nd	29	8
V	46	35	26	20	12	223	1	18	13	48	45	33
Ni	6	2	nd	nd	nd	126	2	2	1	20	14	10
Cr	15	14	9	4	3	598	7	14	11	49	25	21
La	19.2	16.9		45.2	39.6	37.2		49.7		29.2		
Ce	40.0	38.6		91.7	56.7	83.7		96.5		67.4		
Nd				20.1	16.2			29.1		19.4		
Sm	4.92	4.24		3.64	3.12	10.20		3.16		4.6		
Eu	1.65	0.93		0.73	0.58	2.73		0.92		0.95		
Tb	0.75	0.44		0.25	0.18	0.84				0.29		
Yb	1.80	1.33		0.21	0.34	1.40		0.51		0.84		
Lu	0.24	0.19		0.09		0.20		0.12		0.12		
Hf	4.12	4.35		5.63	4.13	3.61		6.44		3.64		

237000 mafic grey gneiss, southern central gneisses.
236991 typical grey gneiss, southern central gneisses.
236969 pale grey gneiss, middle of the central gneisses.
236966 and 236955 typical white gneisses, northern central gneisses.
236949 Inaluk dyke, northern central gneisses.
229472 pegmatitic gneiss sheet, northern central gneisses.
229402 early tonalitic sheet, southern central gneisses.
225859 late tonalitic sheet, southeast central gneisses.
225917 grey gneiss sheet in the southwest of the Isua supracrustal belt.
292202 and 292192 banded gneisses *c.* 5 km south of the Isua supracrustal belt.

nd not detected
loi loss on ignition
Analyst: XRF, Ib Sørensen, GGU

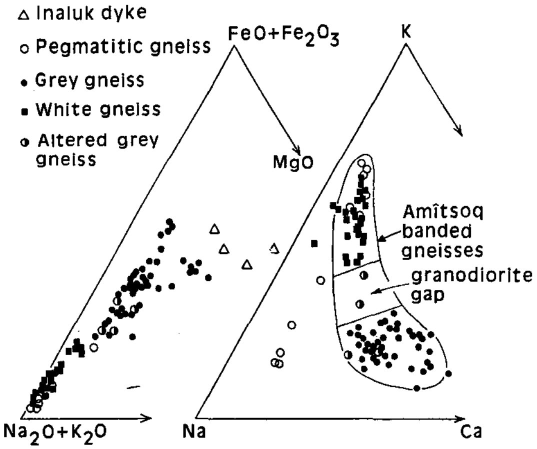

Fig. 17. AFM and Ca–Na–K diagrams for the early Archaean gneisses. *Granodiorite gap* is the compositional break between the grey gneisses and the white gneisses.

dances are strongly disturbed (fig. 17). Such changes, associated with intrusion of the white gneisses, explain the disturbance to the U-Pb zircon, Rb-Sr and Pb-Pb isotope systems at *c.* 3600 Ma (Baadsgaard *et al.,* in press a, and unpublished data). Metasomatic alteration of the grey gneisses in the late Archaean and mid Proterozoic has been detected by petrographic, geochemical and O and Pb isotopic studies (Baadsgaard *et al.,* in press b). However, by consideration of the grey gneisses that are best preserved on the basis of field, petrographic and isotopic criteria and the avoidance of elements such as Rb, Sr, Ba, and Pb it is possible to use the geochemistry of the grey gneisses to provide clues on their origin.

The grey gneisses have Ca:Na:K proportions similar to other Archaean tonalites, being more sodic than many Phanerozoic tonalite/dacite suites (McGregor, 1979). Most grey gneisses have over 15 wt% Al_2O_3, and SiO_2 contents range from 60 to 70 wt% (Table 3). Mafic grey gneisses have $(La/Yb)_N$ values as low as 5 and fairly straight REE patterns, whilst relatively unaltered, less mafic grey gneisses have somewhat higher $(La/Yb)_N$ values and slightly downward bowed patterns (Nutman & Bridgwater, in press). Eu anomalies are small and may be positive or negative. Such REE patterns are common in other Archaean tonalites (e.g. Arth & Hanson, 1975; Compton, 1978). The grey gneisses contain inclusions of a metasedimentary sequence, much of which is rich in amphibolite, and thus partial melting of crust dominated by basic rocks, equivalent to the exposed fragments of volcanosedimentary sequence in the Isukasia area, could have been the source of the grey gneiss protoliths (McGregor, 1979; Nutman & Bridgwater, in press).

The grey gneiss sheets within supracrustal units have compositions within the range of the grey gneisses (Table 3), apart from the more altered units. Some of these altered units contain carbonate, and some have higher K_2O and lower Na_2O contents, and abundances of light REE, Ba, Pb, Rb, Cr and Ni can be anomalously high. Samples with high Cr content come from tracts where there is sporadic development of secondary chrome muscovite, after the Tarssartôq dykes were intruded. Gold is present in very small amounts in some of these rocks (D. Bridgwater, personal communication, 1983).

Inaluk dykes

Inaluk dykes are basic rocks with high FeO/FeO+MgO values (fig. 17), but other aspects of their chemistry are variable (Table 3). High K_2O contents are found in more biotite-rich samples and Rb, Ba, Sr and Th contents are erratic, but commonly high. Such variations could be attributed to contamination of the melts during their ascent or to alteration, either during emplacement of the white and pegmatitic gneisses or during subsequent tectonometamorphic events. Abundances of P_2O_5, TiO_2, MnO, Y and LREE are generally high (Table 3). Basic rocks of this type are commonly found in association with granitic rocks (e.g. Emslie, 1978). This is in agreement with field observations from the Inaluk dykes which were intruded just prior to, and together with, the white gneisses. Models for the origin of such basic-granitic complexes consider the basic components to be strongly fractionated melts of mantle origin, contaminated to varying degrees by interaction with crustal material (e.g. Emslie, 1978). Nutman & Bridgwater (in press) argue that such an origin is the most feasible explanation for the chemistry of the Inaluk dykes.

White gneisses

These rocks contain 3 to 6 wt% K_2O, 70 to 75 wt% SiO_2 and 12 to 14 wt% Al_2O_3 and have a low content of ferromagnesian components (Table 3). The abundances of Rb, Sr and Ba in the white gneisses are high (Table 3) and typical for granites. REE patterns of white gneisses have $(La/Yb)_N$ values of between 40 and 140 and are highly enriched in light REE with respect to chondrites (Nutman & Bridgwater, in press). There are both small positive and negative Eu anomalies and the patterns are somewhat downward curved.

The majority have a granitic composition and span the Qtz-Or-Plag cotectic in the salic tetrahedron with PH_2O = 7 to 2 kb (Nutman & Bridgwater, in press). Granitic white gneisses locally grade into pegmatite, showing that final crystallisation was in a fluid-rich environment. On this evidence the white gneisses could be early, low temperature melts formed at depth of a quartzofeldspathic source, locally modified by hydrothermal processes during crystallisation. The most likely quartzofeldspathic source for the white gneisses is anatexis at the depth of the country rocks – grey gneisses with supracrustal inclusions. Geochemical studies of the white gneisses (Nutman & Bridgwater, in press) show that on the basis of their quartzofeldspathic components they could be interpreted as low temperature melts (in the sense of Winkler, 1974) at an undisclosed PH_2O. Using the arguments of Fyfe (1970), the white gneiss magmas must have originally been water undersaturated and originated at depth as they are clearly intrusive, with no evidence that they are formed *in situ*. As they crystallised at their level of intrusion they exsolved an H_2O-rich fluid, with resultant pegmatite formation and metasomatism of the country rocks.

Late grey sheets

Samples of these sheets (Table 3) have Na/K ratios between those of the white and the grey gneisses and plot in the 'granodiorite gap' of fig. 17. Sample 225859 has been analysed for REE, and shows considerably higher $(La/Yb)_N$ value than grey gneisses, and its chondrite-normalised pattern shows a slight positive Eu anomaly and is downward-curved. The late grey sheets are clearly geochemically different from the grey gneisses.

Pegmatitic gneisses

These are silica-rich and poor in ferromagnesian elements (Table 3), and have considerable range of Na/K values, giving rise to a general linear trend of granitic to albitic trondhjemitic compositions (fig. 17). Granitic compositions lie within the region of low temperature, early melts in the salic tetrahedron, but the more sodic ones plot outside this region (Nutman & Bridgwater, in press). Their spread within the system is best attributed to marked variation in Na/K ratio. Abundances of Rb, Ba and Sr are generally much higher in the K-rich than the Na-rich units, but variation of Pb, Th and Zr shows no apparent correlation to this factor. Ni and Cr abundances are generally low, and for samples of granitic composition light REE abundances are lower and heavy REE abundances are higher than for white gneisses with similar major element chemistry (Nutman & Bridgwater, in press). The field character of the pegmatitic gneisses suggests that they crystallised in a fluid-rich environment, and therefore not only igneous but also hydrothermal processes will have controlled their chemistry. The pegmatitic gneisses may have originated from granitic magma, which interacted with alkali and halogen-bearing fluids during crystallisation, giving rise to the Na enrichment coupled with K depletion observed in some samples (Nutman & Bridgwater, in press).

Banded gneisses

Samples of these mixed rocks from the Isukasia area produce a calc-alkali 'variation trend' spanning the range of compositions of the grey and white gneisses (fig. 17). There are now over 70 analyses of single-phase grey and white gneiss samples from the area. None of these samples fall in the 'granodiorite gap' of fig. 17. This suggests that the variation trend for the banded gneisses, with compositions falling in the 'granodiorite gap', is due to mechanical mixing of two distinct groups of gneisses (caused by strong deformation such that samples of a few kilograms may contain several bands of the grey and white gneiss), and by metamorphic conditions that result in formation of schlieric gneisses and pegmatitic neosome producing small scale compositional heterogeneity. The origin of granodioritic whole rock compositions by these methods does not preclude that granodiorites in other parts of the Amîtsoq gneiss complex could be derived from a single igneous protolith.

Significance of the early Archaean gneisses

The early Archaean gneiss complex of the Isukasia area demonstrates that the oldest known continental crust is of multiphase, polygenetic igneous origin. The grey gneisses are grossly geochemically similar to many younger Archaean tonalitic gneiss suites, and it is likely that they formed by partial melting of crust dominated by basic rocks. The tectonic setting for their generation is not known, but melting of underthrust slabs of the parent is possible because large isoclinal folds in the Isua supracrustal belt could be rotated nappe-like structures formed at approximately the same time as the grey gneisses were intruded (Nutman *et al.*, 1983; Nutman, 1984). The white gneisses are strong candidates for products of recycling of early Archaean sialic crust at *c.* 3600 Ma by partial melting. The *c.* 3400 Ma pegmatitic gneisses record further purging upwards of low temperature melting fractions of the crust. Their wide variation of composition is attributed to modification of igneous compositions by hydrothermal processes. Field evidence suggests that their emplacement may have been associated with tectonism. The diverse origins for the gneiss complex at Isukasia demonstrates that like the Amîtsoq gneisses south of Godthåb it is polygenetic in origin (e.g. Nutman *et al.*, 1984b).

The grey gneisses, white gneisses and the pegmatitic gneisses are between 3800 and 3400 Ma old; they could have formed in a 'super event' that took place within one tectonic setting. However, it is just as likely that the phases formed in discrete, smaller events with different tectonic settings. The implications of the sequence of gneiss intrusion on the metamorphic history of the area is discussed in a later section of this bulletin.

TARSSARTÔQ DYKES

The *Tarssartôq dykes* (Nutman *et al.*, 1983) are several sets of basic and ultramafic dykes found in the Isukasia area (fig. 18 & Plate 1). These dykes have been correlated with the mid Archaean *Ameralik dykes* (McGregor, 1973) by Allaart (1976), Bridgwater *et al.* (1976) and Gill & Bridgwater (1976, 1979) and in some accounts have been referred to as the Isua dykes. Throughout much of the central gneisses late Archaean deformation is so low that it can be difficult in the field to distinguish between dykes of Proterozoic and Archaean age (Table 1); it is only in

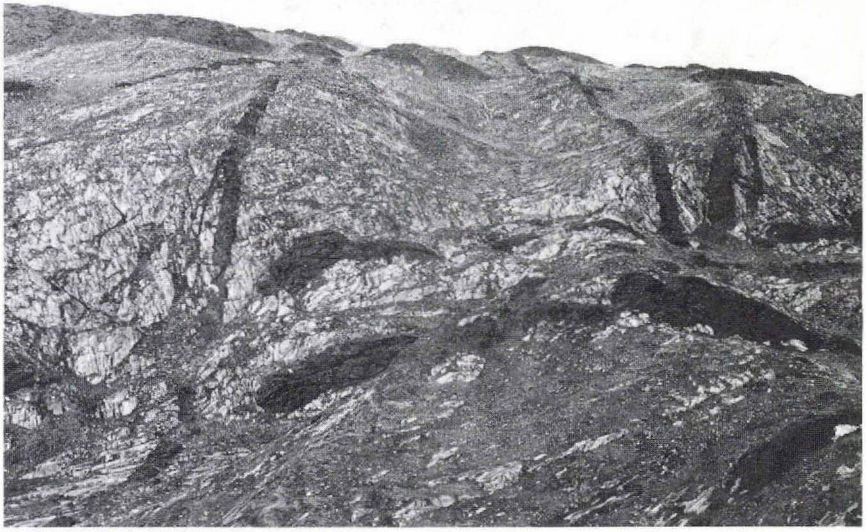

Fig. 18. Podded, discordant Tarssartôq dykes in the southern margin of the central gneisses. Tangential dykes run left to right and radial dykes run from top to bottom of the figure. Looking north from 65°6′22″N, 49°58′25″W. Hill is approximately 300 m high.

areas of stronger late Archaean deformation that the dykes of these different ages can be readily distinguished in the field. Thus the name Tarssartôq (*Greenlandic:* hillside with dark lines) was adopted for the dykes in the Isukasia area, rather than collectively correlating all of them with Archaean Ameralik dykes to the south (Nutman *et al.*, 1983). The description of the Tarssartôq dykes is based on dykes that, where followed out into areas of strong late Archaean deformation, are deformed and recrystallised into amphibolites, demonstrating that they are of Archaean age.

Throughout much of the central gneisses the Tarssartôq dykes are parallel-sided, straight or slightly warped, and although recrystallised under amphibolite facies conditions, they retain igneous textures, igneous plagioclase grains and in some dykes igneous pyroxene grains. At the edge of the central gneisses, the Isua supracrustal belt and further south, the Tarssartôq dykes are deformed and recrystallised into lineated hornblende amphibolites. By 65°N, the southern boundary of the Isukasia area, the Tarssartôq dykes resemble more closely the Ameralik dykes in Godthåbsfjord than their undeformed counterparts 15 km to the north (Nutman *et al.*, 1983). The deformation that has affected the Tarssartôq dykes is correlated with late Archaean structural events to the south (Chadwick & Nutman, 1979; Chadwick, 1985). The dykes can be used as a marker to distinguish between deformation in the early Archaean (3800 to 3400 Ma) and in the late Archaean (3100 to

Table 4. Intersection relations between Tarssartôq dykes

(a)	Ultramafic dykes are cut by tangential (E – W) basic dykes.
(b)	Tangential basic dykes and large mafic coarse-grained dykes are cut by radial (N – S) basic dykes.
This gives the following chronology:	
(3)	(*youngest*) Radial basic dykes.
(2)	Tangential basic dykes and large mafic coarse-grained dykes.
(1)	(*oldest*) Ultramafic dykes.

2500 Ma). Also, metamorphic assemblages in the dykes give some constraint on the age of metamorphic assemblages in the host early Archaean rocks.

Well-preserved Tarssartôq dykes

Four groups of Tarssartôq dykes are recognised in the field, which in decreasing volumetric importance are as follows: (1) so-called E–W or tangential basic dykes (more or less parallel or tangential to the Isua supracrustal belt), (2) so-called N–S or radial basic dykes (strongly discordant to the trend of the Isua supracrustal belt), (3) mafic, coarse-grained dykes and (4) ultramafic dykes. Descriptions of the first two categories of dykes have been given by Bridgwater & McGregor (1974), Allaart (1976), Bridgwater *et al.* (1976), James (1975a, 1976) and Gill & Bridgwater (1976, 1979). In the central gneisses intersections between the dykes indicate the age relations between some of the sets (Table 4). Where free of late Archaean regional ductile deformation the dykes locally have sheared margins. There is no clear evidence that this is an intrusive feature; sheared contacts are probably due to local movements during late Archaean and Proterozoic events.

Ultramafic dykes

These are rare and less than 10 m thick in the central gneisses. Thicker (up to 100 m wide), but more deformed ones occur south of the Isua supracrustal belt. The longest (>3 km) in the southern part of the central gneisses trends approximately north–south and is up to 10 m wide (Plate 1). It has sharp contacts with the host gneisses with no mobilisation of the country rocks observed. Over most of its length it is straight and essentially undeformed. However, within 2 km of the Isua supracrustal belt it is tectonically thinned, schistose, rotated towards the east and consists of orthoamphibole–serpentine±carbonate. In the north it is isotropic and small relict olivine grains occur.

Mafic coarse-grained dykes

The two most prominent of these dykes can be followed for 20 km across the middle part of the central gneisses (Plate 1). They are in places over 100 m wide and have the same orientations as the tangential basic dykes. Despite their width, in the field they appear to be homogeneous. In the middle of the central gneisses they are only slightly deformed, but to the east and west, as they approach the Isua supracrustal belt, they are deformed into arcuate trains of pods.

Their igneous assemblage is hypersthene–plagioclase–olivine. Hypersthene grains are up to 5 mm across and are commonly well preserved apart from replacement at their margins and along cracks by pale amphibole and chlorite. Olivine is found as inclusions in hypersthene grains. Plagioclase forms laths up to 5 mm long and exhibits ophitic and subophitic textures with hypersthene. There is some alteration of plagioclase to epidote, phengite and scapolite. In places, particularly near their contacts, these dykes contain biotite.

Tangential basic dykes

In the northern part of the central gneisses, where there is no appreciable late Archaean deformation, there are dykes which intersect at approximately 30° (Plate 1). These probably belong to the tangential generation of dykes. The relationship between the dykes that intersect at 30° is uncertain – there is no evidence that one cuts the other. Thus on the basis of their intersection angle they are provisionally regarded as conjugate dykes (Nutman *et al.*, 1983). As these dykes become progressively deformed, their angle of intersection decreases, so that near the Isua supracrustal belt and further south they give the appearance of dykes of a single orientation (Plate 1). The tangential dykes are commonly 20 to 50 m wide, black to deep brown in colour, fine to medium grained, with in the field a clear doleritic texture. Chilled margins are widely preserved. These dykes have a simple morphology, with only rare *en echelon* relations, apophyses or degeneration into numerous thin dykes. Some contain calcic plagioclase phenocrysts, like some of the Archaean basic dykes in Godthåbsfjord. The phenocrysts are up to 2 cm across and are usually concentrated at one side or in the centre of the dyke (fig. 19).

Fig. 19. Tarssartôq dyke cutting garbenschiefer in the western part of the Isua supracrustal belt. Plagioclase megacrysts are concentrated towards the eastern side of the dyke. 65°6′33″N, 50°10′20″W.

Where they have not been affected by late Archaean deformation, these dykes commonly retain ophitic textures, with laths of plagioclase in a groundmass of randomly orientated pale green amphibole or hornblende aggregates replacing pyroxene, which is not commonly preserved. Chilled contacts are preserved in some places and contain microlites of plagioclase. However, the margins of the dykes are commonly the site of strong alteration and recrystallisation. Skeletal Fe-Ti oxides form up to 2 percent of most sections and probably formed as a result of breakdown of pyroxene. Locally there is a thin rim of biotite on the Fe-Ti oxides. Plagioclase is slightly cloudy and shows oscillatory zoning and complex twinning that is probably of igneous origin (Gill & Bridgwater, 1979). Compositions range from An_{70} to An_{30}. Large plagioclase phenocrysts present in some dykes are more calcic than plagioclase in the groundmass. There is widespread alteration of plagioclase to epidote, phengite and scapolite. The scapolite shows considerable variation in Na/Ca and in some cases mantles clinozoisite (Wagner, 1982).

Radial basic dykes

Lithologically, morphologically and petrographically these dykes are similar to the tangential basic dykes that they cut. Where free of late Archaean regional ductile deformation they are straight, parallel-sided and subvertical.

Deformed Tarssartôq dykes

Close to the Isua supracrustal belt the tangential dykes are locally strongly podded (fig. 18 & Plate 1) and James (1976) and Gill & Bridgwater (1979) suggested that they could have been intruded into hot country rocks. Podding in the dykes is parallel to the mineral lineation in the host gneisses that is not present or very weak to the north, where the dykes are not podded. Although the centres of the dyke pods retain igneous textures, their margins commonly have a hornblende lineation. Furthermore, the gneisses between the dyke pods are lineated and the younger, radial basic dykes are slightly podded (Plate 1 & fig. 18). This suggests that the podding is tectonic.

In areas of strong post-dyke deformation, the form of the dykes depends very much on their original orientation and on the rheological properties of their country rock. Thus, in the south of the area where the dykes cut fairly uniformly banded gneisses, despite rotation and marked tectonic thinning, they remain coherent units that can be followed for up to several kilometres (Plate 1). Dykes that cut carbonates, calc-silicate rocks and ultramafic schists are broken-up and often have undergone marked reaction with their host. For example in ultramafic schists it is common that only the centres of dykes are well-preserved hornblende-plagioclase amphibolite, with zones of amphibole + serpentine or talc ± carbonate on either side, and it can be hard to locate where the original margin of the dyke lay. Dykes in supracrustal units that seem reasonably well preserved in the field are often found to be strongly altered in thin section, with development of micas, carbonate, epidote and scapolite.

Correlation of the Tarssartôq dykes

The Tarssartôq dykes are between 3400 Ma and 2550 Ma old because they cut 3400 Ma pegmatites, but are cut by 2550 Ma pegmatites (Table 1). South from the

Table 5. Representative analyses of dykes in the Isukasia area

Sample	117947	173278	158440	158401	173232	173236	236930$	242631	225964d	225944	248142	248102
SiO_2	47.9	48.1	49.0	49.5	46.7	48.3	49.14	53.1	51.93	43.10	53.94	55.66
TiO_2	0.79	0.78	1.05	0.88	0.88	0.94	0.95	0.2	0.34	0.39	0.39	0.56
Al_2O_3	15.43	15.15	14.69	14.35	14.50	13.61	13.63	11.9	9.47	4.54	11.89	13.64
Fe_2O_3	3.00	1.75	2.02	1.60	2.52	2.17	2.36			0.98	1.04	2.20
FeO	8.86	9.86	9.98	9.89	11.45	12.33	11.56	7.5*	9.82	5.92	6.94	6.81
MnO	0.17	0.22	0.24	0.24	0.20	0.26	0.22	0.15	0.17	0.12	0.12	0.12
MgO	7.95	7.94	7.25	7.05	6.83	6.33	6.50	18.5	17.44	19.90	12.22	7.26
CaO	10.50	11.31	11.80	12.24	9.50	11.05	11.20	5.2	4.21	3.39	8.09	8.54
Na_2O	2.73	1.84	1.15	1.15	1.50	1.67	1.50	2.1	1.10	nd	1.80	2.45
K_2O	0.07	0.17	0.13	0.13	1.60	0.35	0.36	0.44	0.97	0.03	0.55	0.91
P_2O_5	0.13	0.06	0.07	0.07	0.10	0.10	0.08	0.05	0.04	0.05	0.05	0.10
loi	1.70	1.47	1.60	1.59	2.36	1.56	1.74			11.89	2.62	1.26
CO_2	1.34	0.85	0.90	1.32	1.42	0.90						nd
	100.6	99.5	99.9	100.0	99.6	99.6	100.24	99.1	95.49	100.31	99.62	99.51
ppm												
Cl										135		
Rb	2	2	2	4	92	5			40	nd	17	25
Sr	100	78	50	76	67	93			51	38	193	311
Ba	36	37	14	33	169	33			58	31	237	435
Y									12	3	11	17
Zr	120	45	59	50	53	53			55	62	50	80
Pb										6	10	14
Th										5		
Cu										2	82	57
V								93	129	55	115	146
Ni	190	157	115	88	125	74		632	599	2293	202	179
Cr	290	310	210	310	221	77		2678	2669	534	1078	287
La							2.18					
Ce	24.0	6.76	8.4	10.1	6.9	10.5	6.30					
Nd	12.7	5.44	6.3	6.8	4.6	7.2	6.58					
Sm	3.5	1.78	2.1	2.5	1.7	2.4	2.33					
Eu	0.83	0.60	0.67	0.83	0.76	0.94	0.80					
Gd	4.1	2.38	2.9	3.3	2.5	3.0						
Tb							0.56					
Dy	4.4	2.94	3.6	3.9	3.6	3.9						
Er	2.8	1.91	2.2	2.2	2.2	2.3						
Yb	2.4	1.85	1.8	2.8	2.7	1.9	1.94					
Lu							0.34					

117947, 173278, 158440 and 158401 tangential (E – W) Tarssartôq dykes.
173232, 173236 and 236930$ radial (N – S) Tarssartôq dykes.
242631 and 225964d large, E – W mafic Tarssartôq dykes cutting the middle of the central gneisses.
225944 N – S ultramafic Tarssartôq dyke, southwestern central gneisses.
248142 and 248102 N – S Proterozoic dykes cutting the Isua supracrustal belt.

* all Fe as FeO
nd not detected
loi loss on ignition
Analyst: XRF, Ib Sørensen, GGU

Isukasia area the Tarssartôq dykes are progressively more deformed and resemble the Ameralik dykes, the most prominent, widespread generation of Archaean basic dykes in the Godthåbsfjord region, which are older than *c.* 3000 Ma Nûk gneisses (McGregor, 1973). Pb-Pb and Rb-Sr isotopic studies (Wagner, 1982) of Tarssartôq dykes that are less deformed than average suggest they are older than 3100 Ma. These isotopic data support the field evidence that Tarssartôq dykes most likely correlate with the (>3000 Ma) Ameralik dykes.

Composition of the Tarssartôq dykes

Ultramafic dykes

These dykes are MgO, Cr and Ni rich and CaO and Al_2O_3 poor (Table 5). They are interpreted as crystallised from basic liquids that contained an entrained mafic phase. The high Ni/Cr ratio and low SiO_2 contents of the dykes would suggest that olivine was the most important entrained phase.

Mafic coarse-grained dykes

These dykes are rich in MgO, SiO_2, Cr and Ni, and poor in CaO and Al_2O_3. Apart from somewhat lower CaO/Al_2O_3 ratios, their composition resembles that of parental compositions proposed for the Bushveld complex (Cawthorn & Davies, 1982). Some samples studied contain rather high contents of alkalis. It is not known whether this is due to post-intrusion alteration or to contamination from the host gneisses during intrusion.

Tangential and radial basic dykes

Gill & Bridgwater (1976, 1979) discussed the tangential and radial basic dykes in detail as a single group. They documented geochemical changes in the dykes due to events since their intrusion, and concluded that the dykes were of low-Ti, low-K Archaean tholeiite affinity. Covariation of MgO/(MgO+FeO), Al_2O_3, TiO_2, Ni and Cr can be attributed to fractionation of liquidus phases from a single parent magma whilst variation in alkalis could be attributed to alteration (Gill & Bridgwater, 1979). Gill & Bridgwater showed that there is no geochemical evidence that these dykes have strongly interacted with or assimilated significant amounts of their host gneisses during intrusion. Basic dykes with higher MgO/(MgO+FeO) are more Al_2O_3-rich than typical tholeiites and bear compositional resemblance to some basaltic komatiites.

Setting of the Tarssartôq dykes

The Tarssartôq dykes and Ameralik dykes together form major dyke swarms that cut the Amîtsoq gneisses. Their formation could be associated with the initial stages of rifting of the Amîtsoq continental crust (Gill & Bridgwater, 1979; Chadwick, 1981; Nutman & Bridgwater, 1983), perhaps heralding the formation of mid Archaean supracrustal sequences dominated by basic rocks, represented by the Malene supracrustal units found intercalated with Amîtsoq gneisses in the Godthåbsfjord region.

PROTEROZOIC DYKES

Late mafic dykes occur in the Isukasia area that post-date folding and formation of the regional mineral lineation under amphibolite facies conditions and cut *c.* 2550 Ma pegmatites (Table 1). They have been correlated with swarms of Proterozoic basic dykes that occur throughout the Archaean of West Greenland (Bridgwater *et al.*, 1976; Nutman *et al.*, 1983), that give Rb-Sr ages in the range 2150 to 1950 Ma (Kalsbeek *et al.*, 1978; Kalsbeek & Taylor, 1985; D. Bridgwater & B. J. Fryer, personal communication, 1983). The late mafic dykes of the Isukasia area yield a scatter about a Rb/Sr reference line of 2056 Ma (Wagner, 1982), supporting this field correlation. All the Proterozoic dykes of the Isukasia area have chill facies, and contacts are sharp, regular and approximately parallel. Locally the edges of dykes are recrystallised to mafic schist containing chlorite and actinolitic amphibole. This is interpreted to be due to slight shearing after the dykes were intruded. Although the dykes are unaffected by regional ductile deformation, they are offset and thoroughly recrystallised in most fault zones of the area. Assemblages in the fault zones are greenschist facies.

Two types of Proterozoic dykes are present; mafic, north–south dykes and 110° and 020–040° black basic dykes. The relative ages of these dykes are not seen in the Isukasia area. But elsewhere in the Archaean block of West Greenland, north–south trending dykes (MD1s) that are lithologically similar to the north–south dykes of the Isukasia area are early in the sequence of Proterozoic dyke intrusion (Bridgwater *et al.*, 1976).

The north–south dykes are the most prominent. The largest of them occurs in the east close to the Inland Ice, and can be followed for 15 km (Plate 1). It is in places over 100 m wide, and *en echelon* relations are common. The north–south dykes are coarse grained and brown weathering, and igneous layering has not been seen in them. Several of these dykes contain inclusions of gneisses up to 10 m long, with lobate and commonly indistinct margins. In the vicinity of such inclusions the dykes have pale blotches that are perhaps incompletely assimilated host gneisses. Except in fault zones these dykes have doleritic texture, with pyroxene grains in some cases completely surrounded by porphyritic andesine plagioclase. The plagioclase is locally epidotised. The pyroxene forms up to 40 percent of the dykes and is uralitised to varying extents. Wagner (1982) suggests from textural evidence that these dykes cooled slowly.

Dykes with a trend of 110° and less than 20 m thick occur in the west of the area, and a dyke with a trend of 020 to 040° cuts the southeastern part of the Isua supracrustal belt. These dykes are fine grained and dark grey or black, with plagioclase laths up to 1 mm long. As with the north–south dykes pyroxene in them is slightly uralitised.

The north–south trending dykes are rich in SiO_2, MgO, Ni and Cr, and poor in CaO (Table 6). They resemble in composition lithologically similar dykes of the same trend throughout the Archaean block of West Greenland (Rivalenti, 1975; Hall *et al.*, 1985; D. Bridgwater & B. J. Fryer, personal communication, 1983). The north–south dykes from Isukasia have Rb, Sr, Ba and volatile contents higher than in many suites of basic dykes, and have a large content of Pb derived from older continental crust (Wagner, 1982). This is probably due to contamination from continental crust during intrusion of the magma (Wagner, 1982; D. Bridgwater, personal communication, 1984). Evidence of this in the Isukasia area comes from partially assimilated gneiss inclusions in the dykes.

DEFORMATION

Throughout most of the Isukasia area the Tarssartôq dykes are discordant, but are recrystallised and deformed to lineated amphibolites that locally contain garnet. This shows the severity of middle to late Archaean tectonometamorphic events. In the core of the central gneisses there is widespread preservation of high angle dyke discordances (Plate 1) plus preservation of igneous plagioclase and pyroxene in the dykes. This shows that post-Tarssartôq dyke deformation was generally less severe there than elsewhere, hence this area provides a window for the examination of early Archaean crustal deformation.

Early Archaean structure

James (1975a, 1976) showed that the Isua supracrustal belt is strongly deformed and interpreted it as cover to the adjacent gneisses that was strongly sheared and infolded into its basement gneisses prior to intrusion of the Tarssartôq dykes. However, it is now known that the gneiss phases are younger than, and intrude, the Isua supracrustal belt. Thus this interpretation is incorrect.

James (1976) derived values for pre-Tarssartôq dyke strain in the Isua supracrustal belt from what he interpreted as flattened and then folded pebbles (fig. 20). However, these pebbles have been reinterpreted as boudinaged layering (Nutman *et al.*, 1984a), invalidating the strain values. Linear fabrics are generally parallel in the early Archaean rocks and the Tarssartôq dykes, suggesting that if any fabrics originated in the early Archaean they have been severely modified by subsequent tectonometamorphic events. Nutman (1984) studied the early Archaean

Fig. 20. Strongly rodded felsic rocks, giving rise to a conglomerate-like lithology, felsic formation sequence A of the Isua supracrustal belt. 65°5′52″N, 50°0′22″W.

tectonics of the area by examination of the relations between four *chronostratigraphic units* (Table 6). The characters of these units have been described above.

Isoclinal folds in the Isua supracrustal belt (Plate 1 & fig. 21) are not parallel to the trend of the Isua supracrustal belt and are not apparent in the adjacent areas of gneiss. Locally, only moderately deformed tonalitic sheets are slightly discordant to strong tectonometamorphic layering in the Isua supracrustal belt (Nutman *et al.*, 1983; Nutman, 1984) suggesting that the isoclines are older than or contemporaneous with intrusion of the *c.* 3750 Ma grey gneisses.

In areas where the white gneisses are virtually undeformed, white gneiss sheets are seen to cut a biotite fabric in the grey gneisses. The biotite fabric shows small

Table 6. Chronostratigraphic units and early Archaean history

(7)	(*youngest*) Deformation giving rise to upright folds.
(6c)	Intrusion of pegmatitic gneisses (chronostratigraphic unit).
(6b)	Intercalation of supracrustal and gneiss units.
(6a)	Deformation that caused development of planar fabrics, strongest in the south.
(5)	Intrusion of thin tonalitic sheets (volumetrically insignificant).
(4)	Intrusion of the white gneisses (chronostratigraphic unit).
(3)	Intrusion of Inaluk dykes.
(2)	Intrusion of the grey gneisses (chronostratigraphic unit).
(1)	(*oldest*) Deposition of the supracrustal rocks and intrusion of mafic and ultramafic rocks (chronostratigraphic unit).

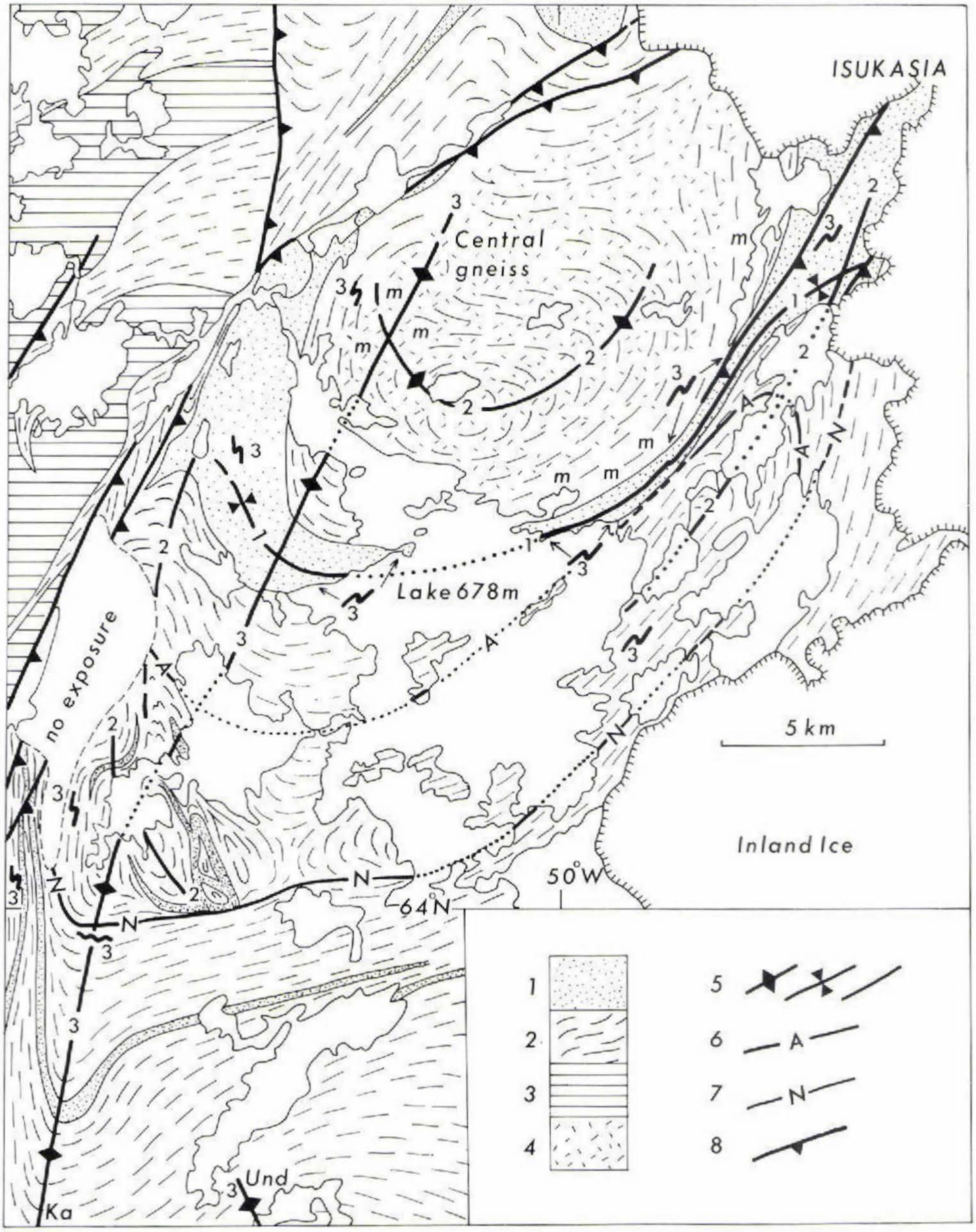

Fig. 21. Structural sketch map of the Isukasia area. *1* supracrustal rocks. *2* gneisses. *3* Taserssuaq tonalite. *4* area where late Archaean lineation is weak or absent (east of Ataneq fault only). *5* fold axial traces (*1* isoclinal fold in the Isua supracrustal belt, *2* post gneiss intrusion, pre-Tarssartôq dyke folds, *3* late Archaean upright folds with *m* prominent mullioning, *Ka* Kangerssuaq antiform and *Und* Ujuragssuit nunât dôme). For lines *A* and *N* (*6,7*) see the text. *8* Proterozoic fault.

isoclinal folds locally. This demonstrates that the grey gneisses had been deformed (and metamorphosed) prior to intrusion of the white gneisses.

Contacts between gneisses and large units of supracrustal rocks such as the Isua supracrustal belt are sharp and are locally cut by Tarssartôq dykes. Lithological layering of supracrustal units is locally discordant to, and truncated at, their contact with adjacent gneisses (Plate 1). Small enclaves of supracrustal rocks occur in the grey gneisses, and the southern part of the Isua supracrustal belt west of lake 678 m is heavily sheeted by tonalitic gneiss. But generally there are only sparse grey gneiss sheets in the major supracrustal units, although the adjacent gneiss areas are dominated by grey gneisses. Another striking feature of most of the major supracrustal units is that they are not bordered by migmatite zones rich in enclaves of the supracrustal rocks. Pegmatitic gneiss sheets of 3400 Ma age intruded along the margins of supracrustal units commonly have a flaser texture, suggesting these contacts are zones of displacement. Nutman (1984) suggested that the early Archaean crust of the area consisted of slices of supracrustal rocks cut by gneiss sheets, *intercalated* with slices of multiphase gneisses with supracrustal enclaves. As edges of these slices are cut by Tarssartôq dykes and are locally marked by 3400 Ma pegmatitic gneisses, it is suggested that intercalation occurred in the early Archaean (Table 1).

Subsequent to intrusion of the white gneisses there was early Archaean (pre-Tarssartôq dyke) deformation and metamorphism up to amphibolite facies (Nutman *et al.*, 1983) which in the south transformed the white and grey gneiss migmatites into strongly banded, locally schlieric gneisses, in which the relations between the component phases have been obliterated (Nutman, 1982a). Line A in fig. 21 shows the approximate southern limit of where intrusive relations between grey and white gneisses are frequently well-preserved. This development of a strong planar fabric in the gneisses of the south of the area might have occurred at

Fig. 22. Tarssartôq dyke cutting fold in banded gneisses. 65°1'31"N, 50°17'20"W.

the same time as the tectonic intercalation of gneiss and supracrustal units suggested above (Table 1). Pre-Tarssartôq dyke non-cylindrical folds (fig. 22) that post-date development of this banding are described by Nutman *et al.* (1983) and Nutman (1984) and are shown on fig. 21. Before late Archaean deformation these folds could have been upright, non-cylindrical structures, perhaps with a gentle basin and dome morphology. Southwards, early Archaean structures are increasingly modified and then generally obliterated by increase in late Archaean (post-Tarssartôq dyke) deformation, and south of line *N* (fig. 21) survival of early Archaean structures, apart from strongly modified compositional layering and small, tight, intrafolial folds, is rare.

Early Archaean crustal accretion in the area probably took place by magmatic injection of tonalite units into a supracrustal sequence, perhaps at the same time as isoclinal folding in the supracrustal rocks. It was followed by intrusion of granites and interdigitation of units with pegmatite injection, followed by further folding under high-grade metamorphic conditions. The early Archaean crustal accretion, reworking and deformation of the Isukasia area may have been in several distinct phases, which on the basis of present isotopic data occurred between *c.* 3800 and 3400 Ma. At *c.* 3600 Ma the thickness of the sialic crust was at least 25 km in places (Griffin *et al.*, 1980). Crustal thicknesses during formation of the white gneisses were probably at least 20 km. However, neither the original thickness of the grey gneiss accretionary pile nor crustal thicknesses between 3600 and 3400 Ma are known, but they are likely to have been substantial.

Late Archaean structure

The late Archaean structure of the Isukasia area is examined by comparison to that of areas to the south. This is followed by placing the Isukasia area into models of late Archaean crustal evolution for the Godthåbsfjord region.

On the basis of studies of late Archaean structural evolution in the Godthåbsfjord region and the coastal region south of Nuuk/Godthåb (McGregor, 1973; Bridgwater *et al.*, 1974; Chadwick & Nutman, 1979; Hall & Friend, 1979; Brewer *et al.*, 1984; Chadwick, 1985; Friend *et al.*, unpublished) late Archaean deformation is divided here into two types. In the aforementioned area, structures and fabrics of the earliest type are expression of sub-horizontal shear within the crust. Most important is intercalation (probably by thrusting) of units of different age and origin, formation of recumbent isoclinal folds of up to nappe-like dimensions and development of a strong sub-horizontal *S* fabric. In most places early Archaean fabrics and structures were strongly modified or obliterated during this type of deformation. Also, the Ameralik dykes (probably equivalent to the Tarssartôq dykes of the Isukasia area) were converted into amphibolite strips that are commonly broken up into trains of pods and are generally concordant with layering of their host rocks. This stage has generally been regarded as synchronous with intrusion of late Archaean gneisses of the region (Bridgwater *et al.*, 1974). However, structures of this type have recently been found which involve thrusting of 2800 Ma granulite facies rocks over amphibolite facies rocks, and therefore must be younger (*c.* 2700 Ma?) than the aforementioned intercalation event. The event(s) are referred to as the horizontal stage(s) in the following sections.

Later structures are upright, non-cylindrical folds, commonly with basin and dome form (Chadwick & Nutman, 1979), which are interspersed with linear zones of deformation, known as straight belts which normally trend between 170° and 020°. A mineral lineation developed widely throughout the Godthåbsfjord region is approximately coaxial to the upright, non-cylindrical folds. In the cores of some basin and dome folds the lineation is so strong that *S* fabrics have been obliterated, resulting in lenticular areas of *L* tectonites (Chadwick & Nutman, 1979). Lineation of the same type and direction is also found in most straight belts. The aligned minerals are parts of amphibolite facies assemblages giving ambient metamorphic grade during development of these structures (Chadwick & Nutman, 1979; Friend *et al.*, unpublished). The domal structures and straight belts were intruded by 2700–2600 Ma granitoid rocks during their formation (McGregor *et al.*, 1983; Chadwick, 1985), but are cut by the undeformed *c.* 2550 Ma old (Brown *et al.*, 1981; Moorbath *et al.*, 1981) Qôrqut granite complex. Structures of this type are designated as belonging to the upright stage. In the following sections this second stage is referred to as the upright stage.

In the southernmost part of the Isukasia area, Tarssartôq dykes swing into concordance with banding of the gneisses and are considerably thinner compared with those to the north. The transition takes place over a few hundred metres, and is shown on fig. 21 as line *N*. North of line *N* there are a few tracts in the gneisses no more than a few hundred metres wide where Tarssartôq dykes have been rotated into parallelism with banding of their host. The Isua supracrustal belt is a zone of high strain, and may also have been deformed during this event. Crossing to the south of line *N* the dykes show several forms, depending on their original orientation and thickness. Those dykes originally only slightly discordant to line *N* were thinned and rotated to be parallel with the banding of their host, and some of the thicker ones were also podded. Dykes originally highly discordant to line *N* reacted in several ways. Some simply swing into concordance with the banding of their host. Others develop sinusoidal folds across the layering or are mullioned and then with further strain are broken up into trains of amphibolite pods lying in the foliation. Dykes on both sides of line *N* have the mineral assemblage hornblende+plagioclase±garnet, but those south of line *N* contain coarser-grained hornblende than those to the north. The change of the dykes across line *N* is accompanied by the banding of the gneisses becoming somewhat closer.

These changes clearly show a marked increase in strain and degree of recrystallisation across line *N*. Modification of the dykes suggests that flattening, attributed to the horizontal stage(s) of deformation, was appreciable. It would appear that north of line *N* strain was generally low during the horizontal stage(s) of deformation.

Strong, generally southerly-plunging mineral lineations occur throughout the Isukasia area apart from most of the central gneisses. With the exception of those in

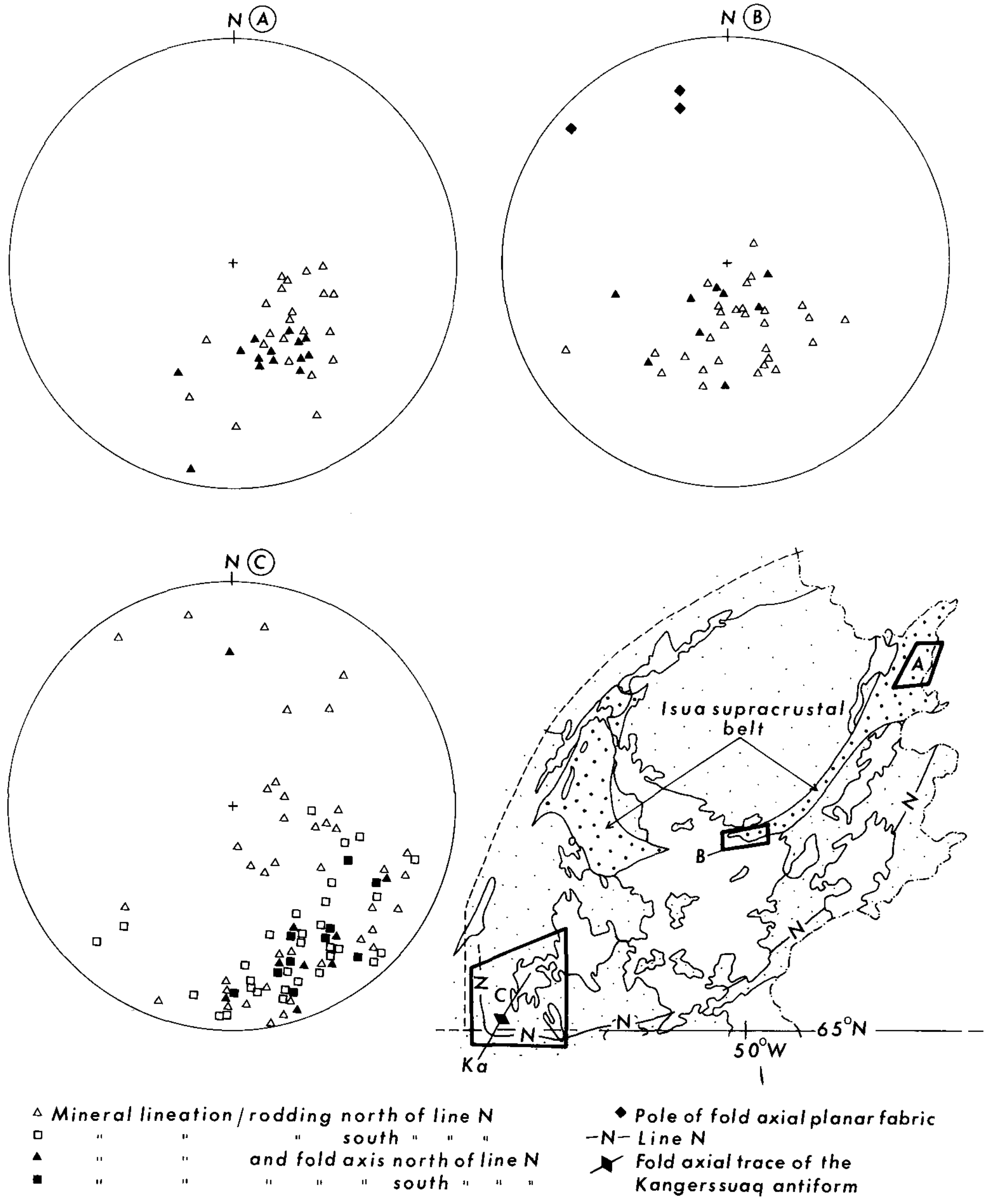

Fig. 23. Structural data from three parts of the Isukasia area.

the central gneisses, Tarssartôq dykes are recrystallised into lineated amphibolites with the assemblage hornblende–plagioclase±chlorite±epidote±garnet. Thus the lineation-forming deformation event is post-dyke, and has affected the whole of the area apart from much of the central gneisses and some other smaller augen of low deformation, where the dykes are statically recrystallised under amphibolite facies conditions. The strength of the lineation varies throughout the area, and is generally strongest in hinges of southerly plunging late folds. The rocks of the Isua supracrustal belt generally have a strong mineral lineation. The Isua supracrustal belt seems to have acted as a high strain zone during this event, which could explain why the central gneisses have escaped from high strain. Parallel to the mineral lineation are late fold axes, *X* axes of stretched plagioclase aggregates in Tarssartôq dykes, and axes of boudins and mullions (fig. 23). These fabrics were analysed by James (1976) who concluded that they are fabrics of extension related to the development of major late folds in the area. Competent layers in supracrustal rocks can break up into rods that at some localities resemble deformed conglomerates (fig. 20; Nutman *et al.,* 1984a). Quartz veins disrupted in the lineation-forming event can also give rise to rods that superficially resemble quartzite pebbles, particularly on *Y-Z* sections.

Small-scale *Z*-style late folds with a fold-axial mineral lineation are particularly abundant in the central and eastern parts of the Isua supracrustal belt (Plate 1). A fold axial planar fabric is present locally in calc-silicate rocks and ultramafic schists (fig. 23), and these folds are not so common and are smaller in the adjacent gneisses. The style of the folds in the Isua supracrustal belt and the trend of the dykes in the adjacent gneisses suggest that this part of the Isua supracrustal belt could have been a shear zone with dextral movement (Plate 1). Large Tarssartôq dykes of all directions at the edge of the central gneisses to the north are tectonically podded (Plate 1 & fig. 18). This style of podding is generally attributed to compression rather than stretching along the dyke (cf. Escher *et al.,* 1976, fig. 76). Modification of the dykes is perhaps a response to shortening along the adjacent parts of the Isua supracrustal belt. In the eastern end of the Isua supracrustal belt the steeply plunging *Z*-style folds are larger, not as tight (as shown by the form of the Tarssartôq dykes) and have refolded pre-Tarssartôq dyke folds (Plate 1). The lineation there is generally strong, with some areas of *L* tectonite. In such areas the Tarssartôq dykes are lineated amphibolites with fluting on their margins parallel to the mineral lineation. In the west of the Isua supracrustal belt rodding fabrics are locally strong, but small *Z*-style folds are not found.

In the western part of the area there is a major overturned antiformal closure, verging to the west (fig. 21). It is the northerly continuation of the Kangerssuaq antiform described by Chadwick *et al.* (1983). The axial trace of the Kangerssuaq antiform continues further north across the Isua supracrustal belt (with the westerly part of the belt lying in the westerly, overturned limb of the fold), and continues into the central gneisses, where large Tarssartôq dykes are mullioned in its

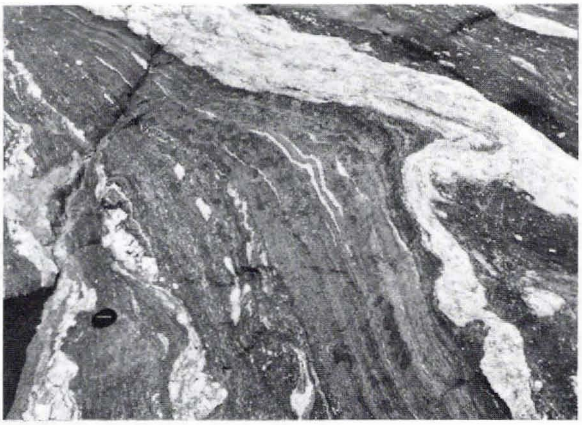

Fig. 24. Strongly sheared amphibolite with disrupted granitic veins on edge of a supracrustal unit on the western limb of the Kangerssuaq antiform. 65°0'32"N, 50°19'24"W

hinge, but the regional mineral lineation is rare. The occurrence of this fold in the central gneisses was noted by James (1975a, 1976). Parasitic folds on the antiform have a fold axial lineation, correlated with the lineation found throughout most of the area. On the overturned, easterly-dipping western limb of the fold deformation is very strong, and locally Tarssartôq dykes are only well preserved in cores of small *S*-style folds and are converted to mafic smears on the fold limbs. Post-Tarssartôq dyke (late Archaean) pegmatite and granitoid sheets are also folded round the *S* folds. Contacts between some supracrustal units and adjacent gneisses are marked by hornblende-biotite schists with disrupted pegmatite and granitoid veins adjacent to sheared gneiss (fig. 24). These schist zones are younger than the Tarssartôq dykes, contain the regional mineral lineation and are interpreted as syn-Kangerssuaq antiform tectonic slides. They are concordant to the banding of less deformed gneisses of the same tract.

Development of the Kangerssuaq antiform might have been coupled with the rising of the Ujaragssuit nunât dome (Chadwick *et al.*, 1983; Chadwick, 1985) to the south. The Kangerssuaq antiform is a structure so large that it encompasses most of the Isukasia area (fig. 21). In its southerly plunging mineral lineations, minor folds, rodding structures, replacement of some fold limbs by shear zones and locally developed basin and dome structures, but general lack of related planar fabrics, it resembles and is correlated with folds of the upright stage to the south.

The Isukasia area has clearly been affected by both styles of late Archaean tectonism that have been recognised in outer Godthåbsfjord and Buksefjorden. The area north of line *N* was on the whole a relatively low strain region during the horizontal stage(s) (*c.* 3000 or 2700 Ma), but its original extent and shape are not known. On the basis of pillow younging directions in units of Malene supracrustal rocks intercalated with early Archaean rocks in northern Godthåbsfjord, the Isukasia area might have been at a structurally higher level than the rocks to the south (Chadwick, 1985). The extent of the regional mineral lineation shows that with the

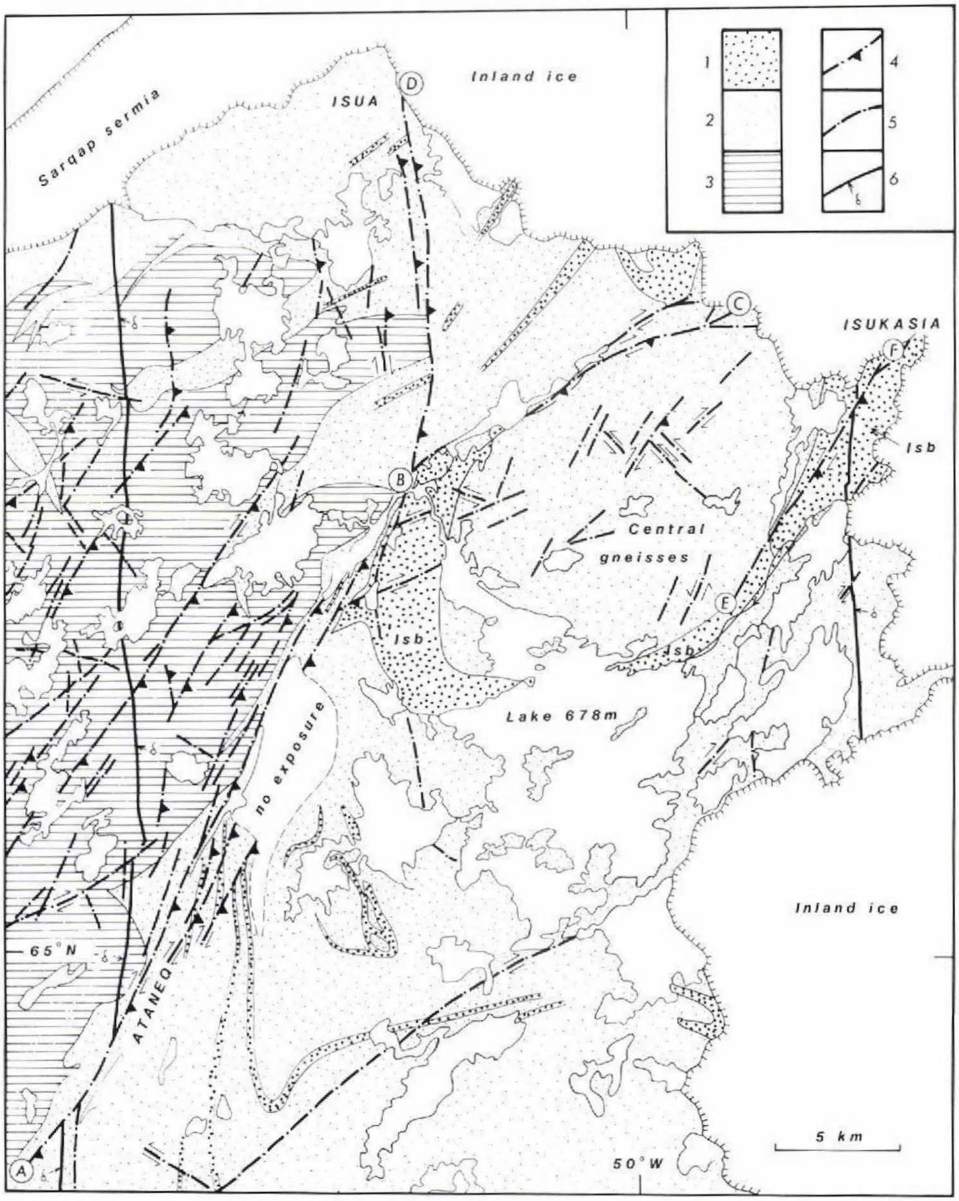

Fig. 25. Faults in the Isukasia and adjacent areas. *1* supracrustal rocks, *2* gneisses, *3* Taserssuaq tonalite, *4* Proterozoic fault dipping to south or east, *5* Proterozoic fault steeply dipping and *6* large Proterozoic dykes. *Isb* Isua supracrustal belt. *ABC* Ataneq fault, *BD* Isua fault and *EF* Isukasia fault, in the north-east of the Isua supracrustal belt. Mapping south of 65°N by geologists from Exeter University.

exception of parts of the central gneisses, the rocks of the Isukasia area were markedly deformed and recrystallised during the upright stage (2700–2600 Ma).

Faults

The Isukasia area is cut by fault zones, some of which have displacements of more than 1 km. Of greatest importance are (a) the Ataneq fault zone (McGregor, 1979), (b) the Isua fault zone, which runs from the northwest end of the Isua supracrustal belt northwards to Isua and (c) the Isukasia fault zone in the northeastern end of the Isua supracrustal belt (fig. 25). Marked chemical alteration is associated with the faulting.

Ataneq fault zone

This consists of a major arcuate fault with associated minor faults on either side (fig. 25). The southern part of the Ataneq fault zone (A to B, fig. 25) is 100 m wide, trends 020 to 030° and dips between 60 and 40° SE. The southern part of the Ataneq fault zone lies within, and perhaps exploited the western limb of the Kangerssuaq antiform, characterised by strongly banded or flaggy gneisses with concordant tectonic slides that dip to the east. The northern sector of the Ataneq fault zone (B to C, fig. 25) is poorly exposed, is about 50 m wide, and trends 050 to 060° and dips 40 to 60° SE over most of its length, but 2 km from the Inland Ice it divides into a splay system of steeply inclined faults. Small overturned folds with a fold axial planar crenulation cleavage with similar orientation to the fault zone are common in amphibolites adjacent to the fault.

The southern part of the Ataneq fault zone is dominated by mylonites that have completely obliterated earlier fabrics of the rocks. The mylonites have a planar fabric of 1 mm ribbon quartz and 1–5 mm thick bands of fine-grained recrystallised quartz, clouded feldspar, epidote, with mica flakes, predominantly muscovite, parallel to the foliation observed in the field. In some cases there are feldspar porphyroclasts set in a uniform fine-grained matrix. In zones of lower deformation the foliation may be undulating, with augen present in which pre-mylonite layering is preserved. The layering in the augen is thinner than layering outside the fault zone and forms irregular tight folds and sheath folds. Folds of this type probably form by simple shear (Cobbold & Quinquis, 1980), and it is likely that they formed during the development of the mylonites. In the northern sector of the fault zone strongly sheared rocks frequently rich in quartz with a layering at a 1 cm scale are common. Brittle deformation has occurred locally, as shown by bands of cataclasite and pseudotachylite concordant to the mylonite fabric. Close to the Ataneq fault zone a faint fault-parallel mica and epidote fabric in the gneisses cuts Archaean planar fabrics, whilst in basic rocks a distinct crenulation cleavage is locally developed. Adjacent to the fault zone basic rocks have commonly recrystallised to greenschists containing quartz and carbonate, whilst quartzofeldspathic rocks show breakdown of plagioclase to epidote and phengite, and replacement of biotite and hornblende by muscovite, epidote and chlorite.

North of 65°N faults west of and parallel to the Ataneq fault zone do not displace a large north–south trending Proterozoic dyke, with one possible exception (fig. 25), but further south this dyke is displaced 4 km dextrally across the Ataneq fault (Brewer *et al.*, 1984). Using Proterozoic dykes and Akilia association units as markers, dextral displacements of up to 1 km or more are observed on faults parallel with the Ataneq fault in the southwest of the area. Distortion of Tarssartôq dykes where they meet the northern part of the Ataneq fault zone also indicates dextral displacements. Thus there is dextral displacement of at least 5 km across the Ataneq fault zone.

Isua fault zone

Only the southern end of this fault zone (B to D, fig. 25) is within the area covered by Plate 1. The main fault zone is 50 to 100 m wide, trends 170 to 180° and dips between 70 and 40° eastwards. Some sections of the fault zone bifurcate, and contain lenses up to 100 m wide of less deformed rocks. It cuts earlier structures with a high discordance. Associated with the main fault are minor parallel ones up to 5 m wide on either side. Lithologically the fault rocks are similar to those of the Ataneq fault zone, suggesting predominantly ductile deformation with some brittle deformation. There is no correlation of marker horizons on either side of the fault, suggesting displacement of at least several kilometres. Rotation of fabrics close to the fault and small displacements over minor parallel faults indicate dextral movement.

Isukasia fault zone

Sequences A and B of the Isua supracrustal belt are separated by a fault running 030° dipping moderately eastwards (E to F, fig. 25). It dies out southwestwards, *c.* 15 km from where it emerges from under the Inland Ice. The fault zone is less than 20 m wide and comprises mica schists with quartz-rich lenses and is dominated by ductile deformation. Kyanite is found locally in the mica schists against the fault (sample 167679, Boak *et al.*, 1983). Locally there are isoclinal folds with amplitudes of 10 m or less adjacent to the fault that are thought to be related to the faulting. A large Proterozoic basic dyke in this area is not exposed where it meets the fault. The nearest dyke exposures to the fault suggest that it has either been displaced dextrally across the fault, or has followed the fault for less than 200 m.

Minor faults

Minor faults occur between the fault zones described above. Most of them have the same orientations as the faults described above. Their displacements are all less than 100 m and are mostly dextral. There are a few small faults trending between 120 and 140° with sinistral displacements. This suggests that nearly all the faults discussed belong to a conjugate system with approximately E–W compression.

Age and crustal setting of the faults

The faults cut folds and fabrics developed during late Archaean high-grade metamorphism and displace *c.* 2550 Ma granite and pegmatite dykes. The presence of mylonites, cataclasites and pseudotachylites suggests faulting with different strain rates and possibly under different *P-T* conditions. It is likely that the faults have had a protracted history of movement. The faulting clearly postdates *c.* 2550 Ma pegmatites and could have continued until after *c.* 1950 Ma when the youngest of the regional Proterozoic dyke swarms were intruded (Kalsbeek *et al.*, 1978). A similar time span is reported for faulting in the Nuuk/Godthåb area (Smith & Dymek, 1983), faulting in northern Godthåbsfjord (James, 1975b) and for movement on the southern end of the Ataneq fault zone (Brewer *et al.*, 1984) and on the Fiskefjord fault zone west of the Isukasia area (Garde *et al.*, 1983).

METAMORPHISM

In the Isukasia area early Archaean mineral assemblages recrystallised extensively in the late Archaean and in some parts of the area in the Proterozoic (Rosing, 1983). This successive metamorphic overprinting means that mineral-chemical studies to establish conditions of metamorphism can only be applied directly to the youngest events in the area. However, field relations, structural history, parageneses in the Tarssartôq dykes, likely thermal affects of intrusion of the gneiss phases and combined isotopic and petrographic studies (Rosing, 1983) can be used to give some constraints on early and late Archaean metamorphic conditions.

Early Archaean metamorphism

Intrusion of the grey gneisses and the white gneisses would have been accompanied by heating, and subsequent strong ductile deformation would have been accompanied by recrystallisation (Table 1). These events are likely to have left metamorphic imprints, evidence for which is discussed below.

Metamorphism associated with grey gneiss intrusion

The central gneisses are least affected by post-Tarssartôq dyke events. Supracrustal inclusions in the central gneisses are deformed, but are cut by virtually undeformed Tarssartôq dykes. Inclusions of supracrustal amphibolite have upper to middle amphibolite facies assemblages (partly retrogressed under either lower amphibolite facies or greenschist facies conditions) and contain coarse-grained, deformed metamorphic segregations that perhaps formed by incipient partial melting (fig. 27). Well-preserved, strongly discordant Inaluk dykes in the central gneisses are weakly foliated amphibolites, but lack the coarse-grained segregations ob-

Fig. 26. Coarse-grained, deformed segregations in amphibolite, from the northeast of the Isua supracrustal belt. 65°11′7″N, 49°49′10″W.

served in the supracrustal amphibolite inclusions. In contrast, many Tarssartôq dykes are fine grained, and although they have been affected by static recrystallisation with the growth of hornblende, actinolite, chlorite, epidote and scapolite, igneous textures and igneous laths of plagioclase are widely preserved, as is igneous pyroxene locally. This shows that the textures related to high-grade metamorphism in the amphibolite inclusions developed before the Tarssartôq dykes were intruded, and that static amphibolite facies metamorphism followed after the dykes were intruded. Similar textural contrasts are seen in the Isua supracrustal belt and the area to the south (fig. 26), but were generally strongly deformed during later events.

A realistic temperature for the (tonalitic) grey gneisses upon intrusion would be in the order of 900°C (Wyllie, 1977). With these rocks forming at least 70 percent of the early Archaean crust of West Greenland their intrusion would have been accompanied by a marked thermal input into the crust, sufficient to cause upper amphibolite facies metamorphism in the supracrustal rocks (see discussion of late Archaean crustal accretion by tonalite injection by Wells, 1980). Thus the early, coarse-grained segregations in the supracrustal rocks can be related to upper amphibolite facies metamorphism most likely associated with intrusion of the grey gneisses. In the best preserved of the grey gneisses there is no garnet, or pseudomorphs after garnet. This absence of garnet places a loose upper limit on the confining pressure during and after their intrusion (fig. 28). Coarse-grained segregations in upper amphibolite facies basic rocks are commonly attributed to incipient partial melting, because in many terrains they are often associated with and grade into trondhjemitic patches and mafic residual zones. Thus the lowest temperature limit of the metamorphism associated with intrusion of the grey gneisses can be assumed to be the amphibolite solidus with excess water present (field a, to the right of curve 5 fig. 27).

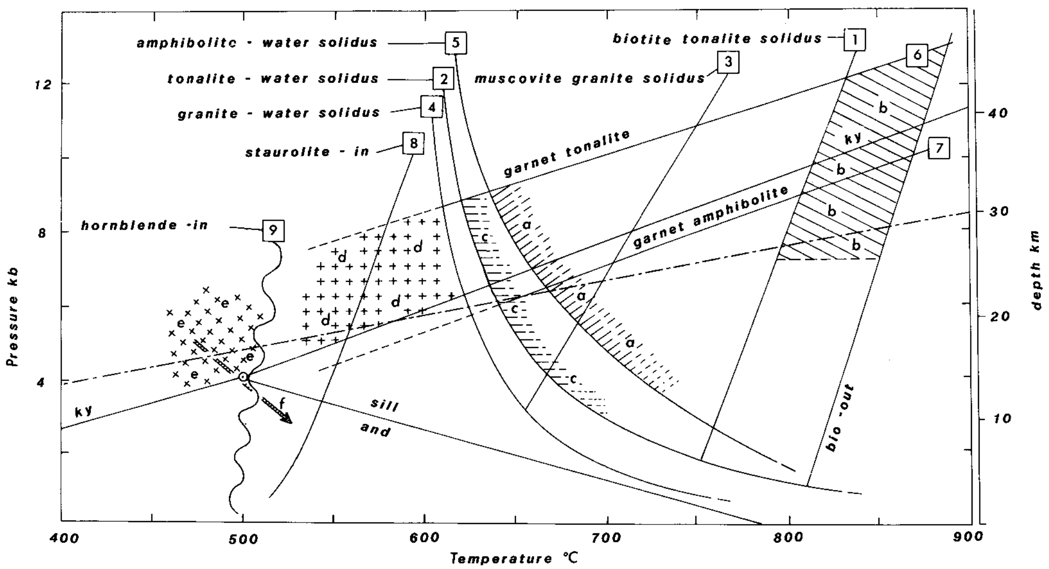

Fig. 27. P,T diagram; data from Wyllie (1977) and Winkler (1974). (a) Minimum P,T conditions when the grey gneisses were intruded. (b) Likely conditions for generation of the white gneisses. (c) Possible metamorphism during white gneiss intrusion – localised? (d) Conditions during early Archaean (*c.* 3600 – 3400 Ma) and late Archaean (3100 – 2600 Ma) reworking. (e) Mid Proterozoic conditions. (f) Formation of orthoamphibole+hercynite+cordierite hornfels. *Ky, Sill, And* are the stability field of kyanite, sillimanite and andalusite respectively.

Metamorphism associated with white gneiss intrusion

Grey gneisses least affected by late Archaean and younger deformation and cut by well-preserved white gneiss sheets have in some places coarse-grained patches and elongate lenses that locally contain hornblende. These clearly formed before or during intrusion of the white gneisses. This texture could be associated with continuing metamorphism in the terminal stages of grey gneiss (over)accretion (see Wells, 1980) or specifically with the intrusion of the white gneisses. The latter alternative is discussed below.

From experimental petrology, the likely temperature of the white gneiss magmas upon intrusion would be *c.* 650°C (e.g. Wyllie, 1977). In the central gneisses the white gneisses are most abundant, forming up to 30 percent of the area. Thus they are likely to have had a marked thermal impact upon their host. Pegmatite in the white gneiss group indicates that the final stages of crystallisation of these rocks took place in a hydrous environment. The solidus temperature for water-saturated tonalite lies within the solidus interval likely for the white gneisses (fig. 27). It is possible that water-rich fluid could have migrated into adjacent grey gneisses as the white gneisses crystallised, and was responsible for forming the coarse-grained segregations and general coarsening of grain size observed in some places. This sug-

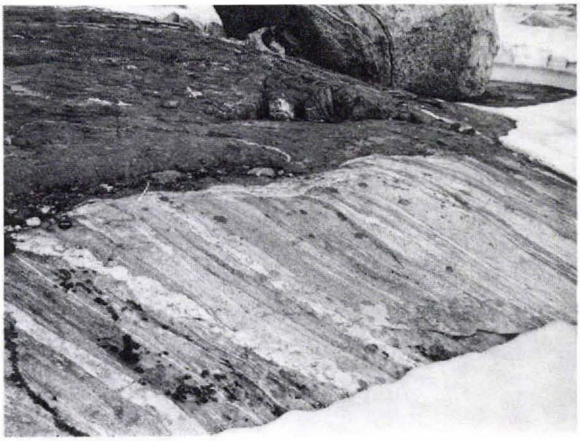

Fig. 28. Strongly banded gneisses cut at a high angle of discordance by a Tarssartôq dyke. 65°2'4"N, 50°17'30"W.

gests metamorphic conditions (at least locally) of middle amphibolite facies (fig. 27), associated with voluminous white gneiss intrusion.

Early Archaean metamorphism during post white gneiss deformation

In some places south of the Isua supracrustal belt Tarssartôq dykes cut (folded) banding in the early Archaean rocks at high angles of discordance (fig. 28). Both the banding and the folds developed after intrusion of the white gneisses (Nutman, 1984). This demonstrates that parts of the early Archaean complex were strongly deformed and recrystallised after injection of gneisses, but before the Tarssartôq dykes were intruded. The nature of the banding suggests deformation under medium to high grade conditions.

Rosing (1983) recognised progressive hydration in the garbenschiefer units of the Isua supracrustal belt, with the earliest (fairly rare) assemblage:

orthoamphibole or cummingtonite + plagioclase + quartz + rutile ± kyanite

being converted to:

hornblende + chlorite + quartz + ilmenite + plagioclase.

Rosing (1983) discussed Rb-Sr and Sm-Nd isotopic data of samples of garbenschiefer hydrated to varying degrees. The Rb-Sr results show that the hornblende-bearing varieties formed in the late Archaean, and that the earlier kyanite-bearing assemblages could be early Archaean. Kyanite in the early assemblage indicates a geothermal gradient of less than 30°C/km and the assemblage as a whole indicates metamorphism under amphibolite facies conditions. Boak & Dymek (1982) and Boak *et al.* (1983) favoured their Isua main stage metamorphism to be an early Archaean (perhaps >3600 Ma) event. However, it is argued below that it is more likely that they recorded a late Archaean event.

Fig. 29. Coarse-grained segregations in strongly deformed, garnetiferous Tarssartôq dyke. 65°5′42″N, 50°16′5″W.

Late Archaean metamorphism

The mid Archaean Tarssartôq dykes can be used as indicators of late Archaean metamorphism across the Isukasia area. In addition, detailed petrographic work combined with isotopic studies (Rosing, 1983) of the garbenschiefer units and re-interpretation of mineral-chemical studies of Boak *et al.* (1983) can give estimates of conditions during late Archaean metamorphism. Tarssartôq dykes south of line *N* (fig. 21) locally contain coarse-grained, strongly deformed metamorphic segregations that consist of hornblende–plagioclase–quartz±garnet (fig. 29). Tarssartôq dykes to the north have the same amphibolite facies assemblages, but they do not have coarse-grained segregations, and garnet is present only locally – mainly in thin dykes that cut Fe-rich, Ca-poor rocks in the Isua supracrustal belt. Although the presence of garnet in amphibolites is partly controlled by bulk composition of the rock, the absence of garnet and segregations in typical dykes in the north of the area and their presence in the south of the area suggests that post-dyke recrystallisation and metamorphism were stronger in the south. The Isukasia area is interpreted as higher in the crust relative to the rocks to the south in the middle to late Archaean (Chadwick, 1985). Southwards from the Isua supracrustal belt, across and beyond line *N*, deeper levels of the crust are exposed at the surface, and an increase in metamorphic grade might be expected.

The strong linear fabric present throughout most of the Isukasia area is due to mineral growth during amphibolite facies metamorphism. During this event Tarssartôq dykes were widely recrystallised to rodded, lineated amphibolites with An >17, indicating metamorphism under amphibolite facies conditions (fig. 30). Garnet in dykes in the south of the area is commonly recrystallised to fine-grained plagioclase-amphibole intergrowths stretched out parallel to the regional mineral lineation. It is possible that breakdown of garnet occurred with the formation of the

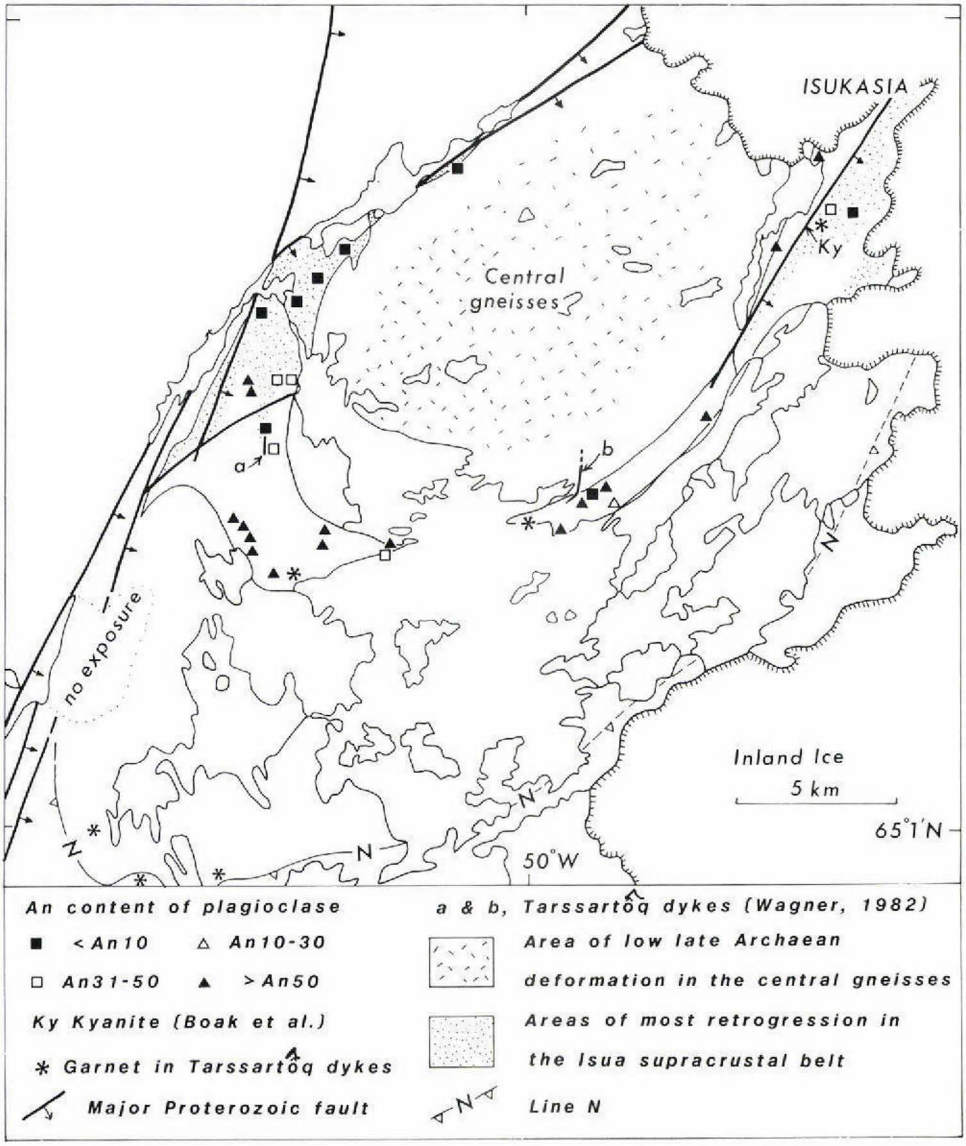

Fig. 30. Map showing miscellaneous metamorphic mineral data. Plagioclase determinations D. Bridgwater (personal communication, 1983).

lineation. This would indicate recrystallisation in the south under a somewhat lower metamorphic grade (still within amphibolite facies) during the lineation-forming event. The linear fabric has been correlated (see above) with a similar fabric widespread in amphibolite facies Archaean rocks of the Godthåbsfjord region. Some Isua metapelites with this linear fabric have the assemblage biotite–staurolite–garnet–muscovite–quartz, reflecting metamorphism up to middle amphibolite facies (fig. 27) in the late Archaean. Boak & Dymek (1982) and Boak *et al.* (1983) calculated garnet–biotite equilibrium temperatures of (541±43°C) for their Isua main stage metamorphism using Isua metapelitic rocks of middle amphibolite facies grade, which they considered to be an early Archaean event. In view of the widespread evidence of late Archaean amphibolite facies metamorphism in the area, Boak & Dymek (1982) and Boak *et al.* (1983) did not present compelling evidence to show why they considered their Isua main stage (amphibolite facies) metamorphism to be early Archaean, rather than late Archaean.

Areas of low deformation during the late Archaean show the development of garbenschiefer textures. The most spectacular of these is the growth of hornblende in the garbenschiefer units (fig. 31) and some calc-silicate rocks.

Proterozoic metamorphism

Major ductile fault zones that were active in the mid Proterozoic occur in the Isukasia area. Adjacent to and within these faults, basic rocks, including Proterozoic dykes, are recrystallised under greenschist facies conditions. Away from the faults the amphibolite facies assemblages with plagioclase (An ≥17) are widespread in basic basement rocks (fig. 30), whereas Proterozoic dykes widely retain their igneous mineralogy. Petrographic evidence shows that these assemblages are due to

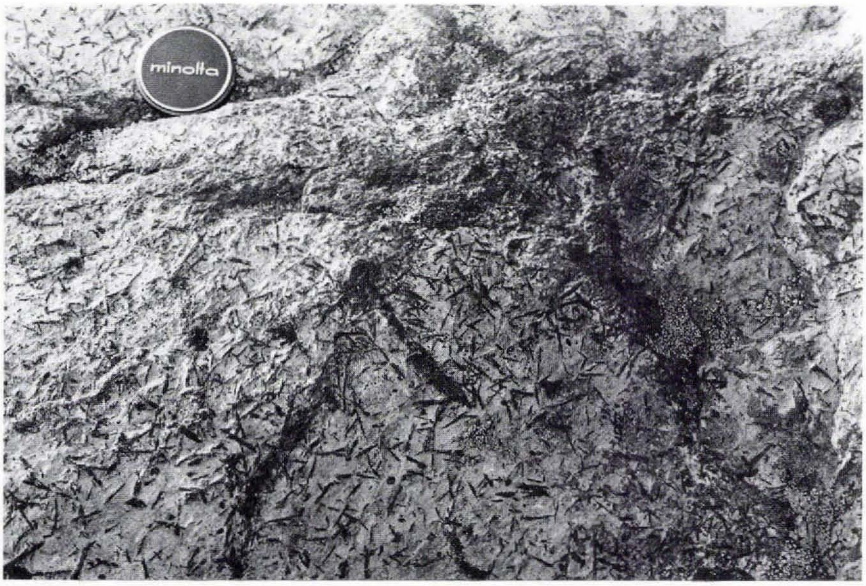

Fig. 31. Hornblende garbenschiefer texture, in the garbenschiefer unit of the Isua supracrustal belt. 65°7′57″N, 50°10′32″W.

retrogression, and they are clearly related in the field to Proterozoic faulting. Boak *et al.* (1983) report kyanite from sample 167679, a strongly altered metapelite in the Isua supracrustal belt. This sample comes from the edge of a Proterozoic fault with a marked displacement (fig. 30). Boak *et al.* (1983) used kyanite in this sample to attempt to place constraints on P_{load} of early Archaean metamorphism. Rosing (1983) reports kyanite in post-Tarssartôq dyke shear zones elsewhere in the Isua supracrustal belt. These kyanite occurrences close to and within faults are best related to metamorphic conditions during Proterozoic faulting. The presence of kyanite in faults and assemblages in basic rocks suggests that faulting and retrogression occurred at somewhat below 500°C and at >3kb (fig. 27). The conditions for this Proterozoic retrogressive metamorphism in and close to fault zones coincide with those given by Boak & Dymek (1982) and Boak *et al.* (1983) by garnet-biotite geothermometry for their *Isua 'retrogressive' metamorphism,* which they considered to be a late Archaean event. Perry & Ahmad (1977) studied the carbon isotope chemistry of mica schists containing graphite and siderite, from close to the Isukasia fault in the northeast of the Isua supracrustal belt. They concluded that the graphite formed by reduction of carbonate and the carbon isotope distribution indicated equilibration between 400 and 500°C. This distribution probably reflects conditions during the Proterozoic faulting.

K-Ar ages on garbenschiefer hornblendes (Lambert & Simons, 1969) and petrographic and mineral-chemical studies of the garbenschiefer combined with interpretation of Sm-Nd and Rb-Sr isotopic data (Rosing, 1983), indicate at maximum lowest amphibolite facies metamorphism (*c.* 500°C, calculated from garnet-biotite pairs) and hydration with marked alteration in places leading to development of biotite, white mica, epidote and carbonate in the garbenschiefer at 2000 to 1800 Ma. These events are correlated with the Proterozoic faulting. Major faults may experience significant shear heating during movement (Graham & England, 1976), and the retrogressive reactions are exothermic. Thus temperatures of up to 500°C for recrystallisation in and close to the faults may be appreciably above the ambient temperature of the crustal level the faults traversed.

Rosing (1983) reports a locality in the southeast of the Isua supracrustal belt where a crenulation cleavage related to the Proterozoic faulting is locally overgrown by the assemblage spinel–cordierite–orthoamphibole with a hornfels texture. This reflects a pressure decrease and temperature increase (to above 500°C) relative to the conditions during the retrogressive metamorphism associated with the Proterozoic faulting. It is possible that the hornfels could have grown during the 1600 Ma thermal event (Baadsgaard *et al.*, in press b) which is the youngest recognised in the area.

METASOMATISM

Rocks have interacted with fluids on many occasions throughout the history of the area. Examples of metasomatism/alteration in the early Archaean have already been discussed. This section will discuss the most obvious evidence of late Archaean and Proterozoic metasomatism. Representative analyses of altered and retrogressed rocks are given in Table 7.

Table 7. Representative analyses of altered rocks in the Isukasia area

Sample	175574	175570b	171755	171757	173164	292481	292480	292482	248109b	248107	236930Xa	Xb
SiO_2	32.68	50.70	48.67	51.11	52.68	70.72	72.93	73.19	45.14	48.33	47.34	48.93
TiO_2	0.20	0.19	0.29	0.43	1.22	0.12	0.10	0.15	0.77	1.25	0.86	0.93
Al_2O_3	1.19	4.68	18.82	16.32	13.02	13.47	13.29	13.46	13.95	13.65	13.49	13.63
Fe_2O_3	4.56	1.46	1.88	1.68	2.53		0.54	0.80	1.01	3.20	2.47	3.02
FeO	10.81	14.34	7.41	5.98	11.42	1.16*	0.48	1.22	9.73	7.12	11.87	10.93
MnO	0.31	0.34	0.26	0.09	0.23	0.02	0.01	0.07	0.11	0.36	0.19	0.22
MgO	28.82	22.88	7.59	10.34	5.35	0.27	0.15	0.40	7.68	5.63	6.19	6.23
CaO	1.54	0.74	10.30	10.31	4.94	1.28	0.85	1.37	7.80	15.64	5.51	10.96
Na_2O	0.01	0.03	1.06	2.63	1.06	4.45	5.19	3.96	2.17	0.13	nd	1.17
K_2O	0.01	0.93	1.64	0.53	3.89	4.17	1.85	2.90	0.05	1.93	5.40	0.75
P_2O_5	0.01	0.01	0.06	0.07	0.18	0.04	0.02	0.06	0.05	0.08	0.07	0.08
loi	19.95	3.83	1.92	1.29	1.08	0.97	0.95	2.13	10.76	1.82	4.37	1.69
CO_2					2.05		0.02					
	100.01	100.13	99.90	100.83	99.65	96.66	96.36	99.71	99.17	99.14	97 76	98.54
ppm												
Cl						23	nd					
Rb	1	197	76	41	78	134	64		19	34		
Sr	21	1	62	197	24	156	99	67	54	303		
Ba	nd	1025	134	80	27	405	394		525	3754		
Y	6	7	20	13		9	12		22	28		
Zr	28	15	43	31	110	134	112		46	64		
Pb	12	13	7	5		19	19		12	24		
Th						9	20					
Cu	74	19	55	nd	134	4	16		14	16		
V	38	108	188		220	100	6		168			
Ni	3414	1362	219	280	10	5	4		139	169		
Cr	2731	3665	240	270	31	10	6		367	214		
La			2.23	5.35							2.71	3.20
Ce			6.29	10.12							7.79	6.99
Nd											6.68	6.38
Sm			0.98	1.46							2.32	2.36
Eu			0.32	0.38							0.71	0.80
Tb			0.46	0.44							0.54	0.57
Yb			1.94	1.86							1.54	1.98
Lu			0.32	0.38							0.29	0.34

175574 and 175570b ultramafic schists, Isua supracrustal belt.
171755 and 171757 altered garbenschiefer unit amphibolites.
173164 banded amphibolite, Isua supracrustal belt.
292481, 292480 and 292482 white gneisses close to the northern end of the Ataneq fault (292482 is strongly sheared).
248109b and 248107 altered Tarssartôq dykes cutting the Isua supracrustal belt.
236930Xa,b margin and 10 cm from margin of *c.* 5 m wide Tarssartôq dyke, northern central gneisses.

nd not detected
loi loss on ignition
Analyst: XRF, Ib Sørensen, GGU

In the Isua supracrustal belt quartz–chrome muscovite–sulphide veining is most concentrated in zones of Z-style folding at contacts between ultramafic and chemical sediment units. Also in zones of pronounced Z-style folding, carbonatisation of lithologies adjacent to units of carbonate-bearing chemical sediments is common. This is seen for example in felsic metasediments and Tarssartôq dykes (248109b, Table 8) on the eastern shore of lake 678 m. These carbonatised felsic metasediments show disturbance of their Rb-Sr isotope systems since the early Archaean (data *in* Moorbath *et al.*, 1975). As discussed above, the garbenschiefer units show progressive development of hornblende and chlorite (Rosing, 1983). Rosing shows that some hornblende growth was due to hydration in the late Archaean, with the addition of Rb and K_2O.

In the mid Proterozoic major faulting in the north and west of the Isukasia area caused widespread recrystallisation under greenschist to epidote amphibolite facies conditions (fig. 30). In and near the fault zones there are quartz and carbonate veins, and basic rocks locally contain abundant mica. Changes are so extreme in the far northwestern end of the Isua supracrustal belt that it can be hard to differentiate between Tarssartôq dykes (fig. 32), banded amphibolite and garbenschiefer. Further away from the fault there is a zone where the rocks have not been strongly deformed during faulting, but have generally been retrogressed with earlier higher-grade assemblages partially preserved. Chemical variations are erratic, but the main changes are hydration, possibly leaching of SiO_2, sometimes addition of CO_2, disturbance of alkali contents, addition of Rb, Cl, Pb and N and probably redistribution of REE (Rosing, 1983; Nutman & Bridgwater, unpublished data). In the western part of the Isua supracrustal belt the garbenschiefer is locally strongly altered. Hamilton *et al.* (1983) suggested the garbenschiefer could have been metasomatised since intrusion, and further isotopic and geochemical studies by Rosing (1983) suggest that at 2000–1800 Ma (the time of faulting) there was a flux of Cl-

Fig. 32. Strongly altered Tarssartôq dyke near to the northern part of the Ataneq fault. Dark lenses are weathered-out carbonate. 65°9′30″N, 50°9′18″W.

bearing fluid through these rocks. The main Proterozoic faults dip to the east and south, so that they underlie the Isukasia area. Throughout much of the area compositional layering is steeply inclined, providing suitable pathways for fluids from the fault zones to react with the rocks overlying the faults, such as garbenschiefer in the west of the area as discussed above.

Susceptible to mobility during events between 3800 and *c.* 1800 Ma are K_2O, Rb, Sr, Pb, U, Th and in some cases REE. This is unfortunate as these elements involve all of the commonly-used isotopic dating systems. In other words, isotopic compositions of samples are a function of original composition and changes during metasomatic events, up to 2000 Ma later. This leads to difficulties in establishing the true age of units, their crustal residence (time elapsed since the protoliths were incorporated into sialic crust) and the isotopic character of their source.

MINERAL OCCURRENCES

Discussed here are mineral occurrences in the Isua supracrustal belt that have been investigated for their commercial interest. Beginning in the 1960s the commercial potential of the Isukasia area was first studied by the company Kryolitselskabet Øresund A/S. In particular the study concerned an iron ore deposit that outcrops at Isukasia, at the northeast end of the Isua supracrustal belt (fig. 30). The Isua supracrustal belt, being a fragment of an Archaean volcanosedimentary sequence is a possible unit to contain base metals, gold and nickel. There are some occurrences, for example disseminated sulphides and thin veins that contain massive chalcopyrite (Appel, 1979a). There has already been some use in jewellery of an attractive looking green chromium mica bearing quartzite named 'grønlandit' ('greenlandite') found in the area.

Iron ore

Banded iron formation lithologies outcrop throughout the Isua supracrustal belt and in the other smaller supracrustal units within the area. Apart from at Isukasia all these units are too small to be considered as commercial iron ore prospects. The banded iron formation at Isukasia is 1.3 km wide. Prevalent metamorphic grade is amphibolite facies (Bridgwater *et al.*, 1981) and the iron formation is markedly deformed. Where exposed it is predominantly oxide facies, consisting of banded quartz-magnetite rocks containing minor amounts of amphibole and disseminated sulphide. To the south and west are associated units of magnetite and amphibole banded metachert. Although loose blocks of quartz-hematite iron formation are common, only one small exposure of this lithology is known, on the eastern side of Isukasia (P. W. U. Appel, personal communication, 1983).

After the first airborne geophysical surveys of the area, the banded iron formation at Isukasia was concluded by Kryolitselskabet Øresund A/S to be the only one large enough to warrant more detailed investigation (fig. 30). The Kryolitselskabet Øresund A/S computed that there are 1900 million tons of ore containing 34% Fe (Keto, 1970). The plans for mining the ore went as far as outlining possible routes to transport the ore down to the coast. In 1973 Kryolitselskabet Øresund A/S decided not to mine the iron ore and their geological exploration at Isukasia ceased.

Gold

Gold determination has been carried out on the main units of the Isua supracrustal belt (D. Bridgwater, M. Rosing and P. W. U. Appel, personal communications, 1984). These studies showed that Au abundance was commonly below detection limits, but values up to 1.07 ppm are found in silicate facies of the iron formation.

Scheelite

Stream sediment samples on the Isua supracrustal belt contain up to 47 grains per 5 kg sample of scheelite (P. W. U. Appel, personal communication, 1984). Scheelite has not been detected in gneiss samples from the Isukasia area (Appel, in press) but only in rocks of the Isua supracrustal belt.

Sulphides

Appel (1979a) reports copper sulphide and pyrrhotite from layered amphibolites and banded iron formation lithologies of the Isua supracrustal belt. He suggested that much of the 'stratabound' mineralisation was coeval with deposition of the supracrustal rocks and is of submarine exhalative origin. Recent detailed reappraisal of some Archaean 'stratabound' gold and sulphide mineralisation of Canadian greenstone belts (Colvine *et al.*, 1984) shows that this type of mineralisation is preferentially developed in high strain zones by deposition of the metals from a metasomatic fluid in the sheared (i.e. banded) rocks. This occurs tens of millions of years after the formation of the supracrustal hosts. This thoroughly documented mechanism is a possible model for 'stratabound' mineralisations in the Isua supracrustal belt.

Chrome mica occurrences and 'grønlandit'

Chrome-bearing muscovite showings are largely restricted to the supracrustal units, in particular the Isua supracrustal belt. Generally there is not sufficient chromium in the muscovite to classify it as fuchsite (Dymek *et al.*, 1983). Concentrations of the Cr-muscovite show up readily in the field due to the distinctive green colour of this mineral. The Cr-bearing minerals are of sporadic, but widespread occurrence and are found in a wide variety of units: metaquartzites, quartz veins, felsic metasediments, intrusive granitic and tonalitic gneiss sheets, carbonate rocks and along shears. All the main Cr-muscovite occurrences are in contact with, or close to, ultramafic units, which are regarded as the source of the Cr via a metasomatic fluid (Dymek *et al.*, 1983).

Green quartzite from the Isukasia area that contains the chrome muscovite is called 'grønlandit' and has been used on a small scale to produce ornaments, pendants and brooches. Based on visual surveys of the outcrops there are perhaps in the order of several tons of high quality 'grønlandit' readily available.

Soapstone

Soapstone forms lenses within sheared and altered ultramafic rocks that form a subconcordant unit within the amphibolite formation (A1) in the northwestern part of the Isua supracrustal belt (M. Rosing and J. H. Allaart, personal communications, 1981). It has a very pale green colour, is rather coarse grained, with a contorted fabric. It has been found to be suitable for carving.

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This bulletin includes geological work in the Isukasia area by numerous workers since 1965. The account is mainly based on field and laboratory work since 1980 by H. Baadsgaard, D. Bridgwater, R. C. O. Gill, A. P. Nutman and M. Rosing, initiated as a result of a NATO scientific research project in 1978 (Grant 949), set up to provide a sound field geological basis for an understanding of the complex geology of the Isukasia area (Bridgwater *et al.*, 1979). Research has been supported by GGU, the Royal Society of London, the Danish Natural Science Research Council, the Carlsberg Foundation and the Canadian N.S.E.R.C. Compilation of the map and this Bulletin was financed by special grants to D. Bridgwater and K. Ellitsgaard-Rasmussen from the Carlsberg Foundation and the Danish Natural Science Research Council. The studies since 1980 were founded on earlier investigations, notably reconnaissance work in the area (Bridgwater & McGregor, 1974) and mapping of the Isua supracrustal belt in 1975 and 1976 by the late J. H. Allaart (Allaart, 1976). Although written and compiled by A. P. Nutman, this Bulletin (like other GGU regional descriptions) has been prepared in consultation with other geologists who have worked in the area. Early drafts of this account were circulated to people directly or indirectly involved in work on the Isukasia area for comment and criticism. In this respect I thank P. W. U. Appel, D. Bridgwater, B. Chadwick, A. A. Garde, R. C. O. Gill, F. Kalsbeek, V. R. McGregor, J. F. W. Park, M. Rosing and K. Secher for their help.

REFERENCES

Allaart, J. H. 1975: Field mapping of the pre-3760 m.y. old supracrustal rocks of the Isukasia area, southern West Greenland. *Rapp. Grønlands geol. Unders.* **80,** 53–56.

Allaart, J. H. 1976: The pre-3760 m.y. old supracrustal rocks of the Isua area, central West Greenland, and the associated occurrence of quartz-banded ironstone. *In* Windley, B. F. (edit.) *The early history of the Earth.* London: Wiley, 619 pp.

Allaart, J. H., Jensen, S. B., McGregor, V. R. & Walton, B. J. 1977: Reconnaissance mapping for the 1:500 000 map sheet in the Godthåb-Isua region, southern West Greenland. *Rapp. Grønlands geol. Unders.* **85,** 50–54.

Appel, P. W. U. 1977: Aeolial differentiation of basaltic tuffs in the early Precambrian Isua supracrustal belt, West Greenland. *Neues Jb. Miner. Mh.* **1977,** 521–528.

Appel, P. W. U. 1979a: Stratabound copper sulphides in the banded iron-formation and the basaltic tuffs in the early Precambrian Isua supracrustal belt, West Greenland. *Econ. Geol.* **74,** 45–52.

Appel, P. W. U. 1979b: Cosmic grains in an iron formation from the early Precambrian Isua supracrustal belt, West Greenland. *J. Geol.* **87,** 573–578.

Appel, P. W. U. 1980: On the Early Archaean Isua iron-formation, west Greenland. *Precambrian Res.* **11,** 73–87.

Appel, P. W. U. (in press): Stratabound scheelite in the Archaean Malene supracrustal belt, West Greenland. *Miner. Deposit.*

Appel, P. W. U. & Jagoutz, E. 1978: On the geochemistry of metabasalts and metarhyolites in the early Precambrian Isua supracrustal belt, West Greenland. *Proc. 11th int. Miner. Ass. Novosibirsk,* Abstr. Vol. **2,** 152 only.

Appel, P. W. U., Moorbath, S. & Taylor, P.N. 1978: Least radiogenic terrestrial lead from Isua, West Greenland. *Nature, Lond.* **272,** 524–526.

Arth, J. G. & Hanson, G. N. 1975: Geochemistry and origin of the early Precambrian crust of northeastern Minnesota. *Geochim. cosmochim. Acta* **39,** 325–362.

Baadsgaard, H. 1976: Further U-Pb dates on zircons from the early Precambrian rocks of the Godthåbsfjord area, West Greenland. *Earth planet. Sci. Lett.* **33,** 261–267.

Baadsgaard, H. 1983: U-Pb isotope systematics on minerals from the gneiss complex at Isukasia, West Greenland. *Rapp. Grønlands geol. Unders.* **112,** 35–42.

Baadsgaard, H., Nutman, A. P., Bridgwater, D., Rosing, M., McGregor, V. R. & Allaart, J. H., 1984: The zircon geochronology of the Akilia association and the Isua supracrustal belt, West Greenland. *Earth planet. Sci. Lett.* **68,** 221–228.

Baadsgaard, H., Nutman, A. P., Rosing, M. & Bridgwater, D. 1985: A late Archaean pegmatite dyke swarm from the Isukasia area, southern West Greenland. *Rapp. Grønlands geol. Unders.* **125,** 48–51.

Baadsgaard, H., Nutman, A. P. & Bridgwater, D. (in press a): Geochronology and isotopic variation of the early Archaean Amîtsoq gneisses of the Isukasia area, southern West Greenland. *Geochim. cosmochim. Acta.*

Baadsgaard, H., Nutman, A. P., Rosing, M., Bridgwater, D. & Longstaffe, F. (in press b): Mid-Proterozoic metasomatic alteration of Amîtsoq gneisses from the Isukasia area, southern West Greenland. *Geochim. cosmochim. Acta.*

Berthelsen, A. & Bridgwater, D. 1960: On the field occurrence and petrography of some basic dykes of supposed Pre-Cambrian age from the southern Sukkertoppen District, western Greenland. *Meddr Grønland* **123**(3), 43 pp.

Boak, J. L. & Dymek, R. F. 1982: Metamorphism of the *ca.* 3800 Ma supracrustal rocks at Isua, West Greenland: implications for early Archaean crustal evolution. *Earth planet. Sci. Lett.* **59,** 155–176.

Boak, J. L., Dymek, R. F. & Gromet, L. P. 1983: Petrology and rare earth element geochemistry of clastic metasedimentary rocks from the Isua supracrustal belt, West Greenland. *Rapp. Grønlands geol. Unders.* **112,** 23–33.

Brewer, M., Chadwick, B., Coe, K. & Park, J. W. F. 1984: Further field observations in the Ivisârtoq region of southern West Greenland. *Rapp. Grønlands geol. Unders.* **120,** 55–67.

Bridgwater, D. & McGregor, V. R. 1974: Field work on the very early Precambrian rocks of the Isua area, southern West Greenland. *Rapp. Grønlands geol. Unders.* **65,** 49–54.

Bridgwater, D., McGregor, V. R. & Myers, J. S. 1974: A horizontal tectonic regime in the Archaean of Greenland and its implication for crustal thickening. *Precambrian Res.* **1,** 179–197.

Bridgwater, D., Keto, L., McGregor, V. R. & Myers, J. S. 1976: Archaean gneiss complex of Greenland. *In* Escher, A. & Watt, W. S. (edit.) *Geology of Greenland,* 18–75. Copenhagen: Geol. Surv. Greenland.

Bridgwater, D., Collerson, K. D. & Myers, J. S. 1978: The development of the Archaean gneiss complex of the North Atlantic region. *In* Tarling, D. H. (edit.) *Evolution of the Earth's crust.* London: Academic, 19–69.

Bridgwater, D., Allaart, J. H., Baadsgaard, H., Collerson, K. D., Ermanovics, I., Gorman, B. E., Griffin, W. L., Hanson, G., McGregor, V. R., Moorbath, S., Nutman, A. P., Taylor, P. N., Tveten, E. & Watson, J. 1979: International field work on Archaean gneisses in the Godthåbsfjord-Isua area, southern West Greenland. *Rapp. Grønlands geol. Unders.* **95,** 66–71.

Bridgwater, D., Allaart, J. H., Schopf, J. W., Klein, C., Walter, M. R., Barghoorn, E. S., Strother, P., Knoll, A. H. & Gorman, B. E. 1981: Microfossil-like objects from the Archaean of Greenland: a cautionary note. *Nature, Lond.* **289,** 51–53.

Bridgwater, D., Nutman, A. P., Schiøtte, L. & Ryan, B. (in press): The Archaean of the Saglek fiord – Hebron fiord area of northern Labrador: progress report. *Current Res. Geol. Surv. Can.*

Brown, M., Friend, C. R. L., McGregor, V. R. & Perkins, W. T. 1981: The late Archaean Qôrqut granite complex of southern West Greenland. *J. geophys. Res.* **86,** 10617–10632.

Cawthorn, R. G. & Davies, G. 1982: Possible komatiitic affinity of the Bushveld Complex, South Africa. *In* Arndt, N. T. & Nisbet, E. G. (edit.) *Komatiites.* London: Allen & Unwin, 526 pp.

Chadwick, B. 1981: Field relations, petrography and geochemistry of Archaean amphibolite dykes and Malene supracrustal amphibolites, northwest Buksefjorden, southern West Greenland. *Precambrian Res.* **14,** 221–259.

Chadwick, B. 1985: Contrasting styles of tectonism and magmatism in the late Archaean crustal evolution of the northeastern part of the Ivisârtoq region, inner Godthåbsfjord, southern West Greenland. *Precambrian Res.* **27,** 215–238.

Chadwick, B. & Coe, K. 1983: Descriptive text to 1:100 000 sheet Buksefjorden 63 V.1 Nord, 70 pp. Copenhagen: Grønlands geol. Unders.

Chadwick, B. & Crewe, M. A. 1982: Field work in the Ujaragssuit nunât area, Ivisârtoq region, inner Godthåbsfjord, southern West Greenland. *Rapp. Grønlands geol. Unders.* **110,** 57–63.

Chadwick, B. & Nutman, A. P. 1979: Archaean structural evolution in the northwest of the Buksefjorden region, southern West Greenland. *Precambrian Res.* **9,** 199–226.

Chadwick, B., Crewe, M. A. & Park, J. F. W. 1983: Field work in the north of the Ivisârtoq region, Godthåbsfjord, southern West Greenland. *Rapp. Grønlands geol. Unders.* **115,** 49–56.

Cobbold, P. R. & Quinquis, H. 1980: Development of sheath folds in shear regimes. *J. struct. Geol.* **2,** 119–126.

Colvine, A. C., Andrews, A. J., Cherry, M. E., Durocher, M. E., Fyon, A. J., Lavigne, M. J. Jr., Macdonald, A. J., Soussan, M., Poulsen, K. H., Springer, J. S. & Troop, D. G. 1984: An integrated model for the origin of Archaean lode gold deposits. Ontario Geological Survey Open File Report 5524, 98 pp.

Compton, P. 1978: Rare earth evidence for the origin of the Nûk gneisses, Buksefjorden region, southern West Greenland. *Contr. Miner. Petrol.* **66,** 283–293.

Dimroth, E. 1982: The oldest rocks on Earth: stratigraphy and sedimentology of the 3.8 billion years old Isua supracrustal sequence. *In: Sedimentary geology of the highly metamorphosed Precambrian complexes,* 16–27. Moscow: Nauka.

Drever, J. I. 1974: Geochemical model for the origin of Precambrian iron formation. *Bull. geol. Soc. Am.* **85,** 1099–1106.

Dymek, R. F., Boak, J. L. & Kerr, M. T. 1983: Green micas in the Archaean Isua and Malene rocks, southern West Greenland, and the occurrence of barian-chromian muscovite. *Rapp. Grønlands geol. Unders.* **112,** 71–82.

Emslie, R. F. 1978: Anorthosite massifs, Rapakivi granites and late Proterozoic rifting in N. America. *Precambrian Res.* **7,** 61–98.

Escher, A., Sørensen, K. & Zeck, H. P. 1976: Nagssugtoqidian mobile belt in West Greenland. *In* Escher, A. & Watt, W. S. (edit.) *Geology of Greenland,* 76–95. Copenhagen: Geol. Surv. Greenland.

Friend, C. R. L. & Hall, R. P. 1977: Field work in the Ivisârtoq area, inner Godthåbsfjord, southern West Greenland. *Rapp. Grønlands geol. Unders.* **85,** 54–60.

Friend, C. R. L., Nutman, A. P. & McGregor, V. R. (unpublished): Late-Archaean tectonic intercalation of amphibolite and granulite facies terrains in the Faeringehavn – Tre Brødre area, Buksefjorden, southern West Greenland.

Froude, D. O., Ireland, T. R., Kinny, P. D., Williams, I. S., Compston, W., Williams, I. R. & Myers, J. S. 1983: Ion microprobe identification of 4,100–4,200 My old terrestrial zircons. *Nature, Lond.* **304,** 616–619.

Fyfe, W. S. 1970: Some thoughts on granitic magmas. *In* Newall, G. & Rast, N. (edit.) *Mechanisms of igneous intrusion,* 201–216. Liverpool: Gallery Press.

Garde, A. A., Hall, R. P., Hughes, D. J., Jensen, S. B., Nutman, A. P. & Stecher, O. 1983: Mapping of the Isukasia sheet, southern West Greenland. *Rapp. Grønlands geol. Unders.* **115,** 20–29.

Garde, A. A., Larsen, O. & Nutman, A. P. 1986: Dating of late Archaean crustal mobilisation south of Qugssuk, Godthåbsfjord, southern West Greenland. *Rapp. Grønlands geol. Unders.* **128,** 23–36.

Gill, R. C. O. & Bridgwater, D. 1976: The Ameralik dykes of West Greenland, the earliest known basaltic rocks intruding stable continental crust. *Earth planet Sci. Lett.* **29,** 276–282.

Gill, R. C. O. & Bridgwater, D. 1979: Early Archaean basic magmatism in West Greenland: The geochemistry of the Ameralik dykes. *J. Petrol.* **20,** 695–726.

Gill, R. C. O., Bridgwater, D. & Allaart, J. H. 1981: The geochemistry of the earliest known basic volcanic rocks, at Isua , West Greenland: a preliminary investigation. *Spec. Publ. geol. Soc. Australia* **7,** 313–325.

Graham, C. M. & England, P. C. 1976: Thermal regimes and regional metamorphism in the vicinity of overthrust faults: an example of shear heating and inverted metamorphic zonation from southern California. *Earth. planet. Sci. Lett.* **31,** 142–152.

Griffin, W. L., McGregor, V. R., Nutman, A. P., Taylor, P. N. & Bridgwater, D. 1980: Early Archaean granulite facies metamorphism south of Ameralik, West Greenland. *Earth planet. Sci. Lett.* **50,** 59–74.

Hall, R. P. 1978: An occurrence of ironstone enclaves east of Sukkertoppen, southern West Greenland. *Rapp. Grønlands geol. Unders.* **90,** 57–60.

Hall, R. P. & Friend, C. R. L., 1979: Structural evolution of the Archaean rocks in Ivisârtoq and the neighbouring inner Godthåbsfjord region, southern West Greenland. *Geology* **7,** 311–315.

Hall, R. P., Hughes, D. J. & Friend, C. R. L. 1985: Geochemical evolution and unusual pyroxene chemistry of the MD tholeiite dyke swarm from the Archaean craton of southern West Greenland. *J. Petrol.* **26,** 253–282.

Hamilton, P. J., O'Nions, R. K., Evensen, N. M., Bridgwater, D. & Allaart, J. H. 1978: Sm-Nd isotope investigations of the Isua supracrustals and implications for mantle evolution. *Nature, Lond.* **272,** 41–43.

Hamilton, P. J., O'Nions, R. K., Evensen, N. H., Bridgwater, D. & Nutman, A. P. 1983: Sm-Nd studies of Archaean metasediments and metavolcanics from West Greenland and their implications for the Earth's early history. *Earth planet. Sci. Lett.* **62,** 263–272.

Hutton, D. H. W. 1979: Tectonic slides: a review and reappraisal. *Earth Sci. Rev.* **15,** 151–172.

James, H. L. 1966: Chemistry of the iron-rich sedimentary rocks. *Prof. Pap. U.S. geol. Surv.* **440–W,** 61 pp.

James, N. P. 1979: Shallowing-upward sequences in carbonates. *In* Walker, R. G. (edit.) *Facies models. Geoscience Canada Reprint Series* **1,** 145–158.

James, P. R. 1975a: Deformation of the Isua supracrustal belt, southern West Greenland. *Rapp. Grønlands geol. Unders.* **75,** 56–58.

James, P. R. 1975b: Field mapping of Bjørneøen and the adjacent coast of Nordlandet, Godthåbsfjord, southern West Greenland. *Rapp. Grønlands geol. Unders.* **75,** 58–62.

James, P. R. 1976: Deformation of the Isua block, West Greenland: a remnant of the earliest stable continental crust. *Can. J. Earth Sci.* **13,** 816–823.

Kalsbeek, F. & Taylor, P. N. 1983: Anatectic origin of mid-Proterozoic granite dyke in the Isukasia area, West Greenland, Pb-Pb and Rb-Sr isotope evidence. *Rapp. Grønlands geol. Unders.* **115,** 38–42.

Kalsbeek, F. & Taylor, P. N. 1985: Age and origin of early Proterozoic dolerite dykes in South-West Greenland. *Contr. Miner. Petrol.* **89,** 307–316.

Kalsbeek, F., Bridgwater, D. & Zeck, H. P. 1978: A 1950±60 Ma Rb-Sr whole-rock isochron age from two Kangâmiut dykes and the timing of the Nagssugtoqidian (Hudsonian) orogeny in West Greenland. *Can. J. Earth Sci.* **15,** 1122–1128.

Keto, L. 1970. *Isua, a major iron ore discovery in Greenland.* København: Kryolitselskabet Øresund A/S.

Krogh, E. J. 1977: Origin and metamorphism of iron formations and associated rocks, Lofoten-Vesterålen, N. Norway. I. The Vestpolltind Fe-Mn deposit. *Lithos* **10,** 243–255.

Lambert, R. St. J. & Holland, J. G. 1976: Amîtsoq gneiss geochemistry: preliminary observations. *In* Windley, B. F. (edit.) *The early history of the Earth,* 191–201. London: Wiley.

Lambert, R. St. J. & Simons, J. G. 1969: New K/Ar age determinations from southern West Greenland. *Rapp. Grønlands geol. Unders.* **19,** 68–71.

McGregor, V. R. 1973: The early Precambrian gneisses of the Godthåb district, West Greenland. *Phil. Trans. R. Soc.* **A273,** 343–358.

McGregor, V. R. 1979: Archean gray gneisses and the origin of the continental crust: Evidence from the Godthåb region, West Greenland. *In* Barker, F. (edit.) *Trondhjemites, dacites and related rocks,* 169–204. Amsterdam: Elsevier.

McGregor, V. R. & Mason, B. 1977: Petrogenesis and geochemistry of metabasaltic and metasedimentary enclaves in the Amîtsoq gneisses, West Greenland. *Am. Miner.* **62,** 887–904.

McGregor, V. R., Bridgwater, D. & Nutman, A. P. 1983: The Qârusuk dykes: post-Nûk, pre-Qôrqut granitoid magmatism in the Godthåb region, southern West Greenland. *Rapp. Grønlands geol. Unders.* **112,** 101–112.

Mehnert, K. R. 1968: *Migmatites and the origin of granitic rocks.* 393 pp. Amsterdam: Elsevier.

Michard-Vitrac, A., Lancelot, J., Allègre, C. J. & Moorbath, S. 1977: U-Pb ages on single zircons from the early Precambrian rocks of West Greenland and the Minnesota River Valley. *Earth planet. Sci. Lett.* **35,** 449–453.

Moorbath, S., O'Nions, R. K., Pankhurst, R. J., Gale, N. H. & McGregor, V. R. 1972: Further rubidium-strontium age determinations on the very early Precambrian rocks of the Godthaab district, West Greenland. *Nature, phys. Sci.* **240,** 78–82.

Moorbath, S., O'Nions, R. K. & Pankhurst, R. J. 1973: Early Archaean age for the Isua iron formation, West Greenland. *Nature, phys. Sci.* **245,** 138–139.

Moorbath, S., O'Nions, R. K. & Pankhurst, R. J. 1975: The evolution of Early Precambrian crustal rocks at Isua, West Greenland – geochemical and isotopic evidence. *Earth planet. Sci. Lett.* **27,** 229–239.

Moorbath, S., Allaart, J. H., Bridgwater, D. & McGregor, V. R. 1977: Rb-Sr ages of very early Archaean supracrustal rocks and Amîtsoq gneisses at Isua. *Nature, Lond.* **270,** 43–45.

Moorbath, S., Taylor, P. N. & Goodwin, R. 1981: Origin of granitic magma by crustal remobilisation: Rb-Sr and Pb-Pb geochronology and isotope geochemistry of the late Archaean Qôrqut Granite Complex of southern West Greenland. *Geochim. cosmochim. Acta* **45,** 1051–1060.

Nagy, B., Zumberge, J. E. & Nagy, L. A. 1975: Abiotic, graphitic microstructures in micaceous metaquartzite about 3760 million years old from southwestern Greenland: implications for early Precambrian microfossils. *Proc. natn. Acad. Sci. U.S.A.* **72,** 1206–1209.

Nagy, B., Nagy, L. A., Zumberge, J. E., Sklarew, D. S. & Anderson, P. 1977: Indications of a biological and biochemical evolutionary trend during the Archaean and early Proterozoic. *Precambrian Res.* **5,** 109–120.

Nutman, A. P. 1980: A field and laboratory study of the Early Archaean rocks of northwest Buksefjorden region, southern West Greenland. 464 pp. Unpublished Ph.D. thesis, University Exeter.

Nutman, A. P., 1982a: Further work on the early Archaean rocks of the Isukasia area, southern West Greenland. *Rapp. Grønlands geol. Unders.* **110,** 49–54.

Nutman, A. P., 1982b: Investigations on amphibolite facies orthogneisses, amphibolites and leucogabbros on Akugdlerssuaq, inner Godthåbsfjord. *Rapp. Grønlands geol. Unders.* **110,** 72–77.

Nutman, A. P., 1984: Early Archaean crustal evolution of the Isukasia area, southern West Greenland. *In* Kröner, A. & Greiling, E. (edit.) *Precambrian tectonics illustrated,* 79–93. Stuttgart: E. Schweizerbart'sche Verlagsbuchhandlung.

Nutman, A. P. & Bridgwater, D. 1983: Deposition of Malene supracrustal rocks on an Amîtsoq basement in outer Ameralik, southern West Greenland. *Rapp. Grønlands geol. Unders.* **112,** 43–51.

Nutman, A. P. & Bridgwater, D. (in press): Early Archaean Amîtsoq tonalites and granites from the Isukasia area, southern West Greenland: development of the earliest-known sial. *Contr. Miner. Petrol.*

Nutman, A. P., Bridgwater, D., Dimroth, E., Gill, R. C. O. & Rosing, M. 1983: Early (3700 Ma) Archaean rocks of the Isua supracrustal belt and adjacent gneisses. *Rapp. Grønlands geol. Unders.* **112,** 5–22.

Nutman, A. P., Allaart, J. H., Bridgwater, D., Dimroth, E. & Rosing, M. 1984a: Stratigraphic and geochemical evidence for the depositional environment of the Early Archaean Isua supracrustal belt, southern West Greenland. *Precambrian Res.* **25,** 365–396.

Nutman, A. P., Bridgwater, D., & Fryer, B. J. 1984b: The iron-rich suite from the Amîtsoq gneisses of southern West Greenland: early Archaean plutonic rocks of mixed crustal and mantle origin. *Contr. Miner. Petrol.* **87,** 24–34.

Oehler, D. Z. & Smith, J. W. 1977: Isotopic composition of reduced and oxidized carbon in early Archaean rocks from Isua, Greenland. *Precambrian Res.* **5,** 221–228.

O'Nions, R. K. & Pankhurst, R. J., 1974: Rare-earth element distribution in Archaean gneisses and anorthosites, Godthåb area, West Greenland. *Earth planet. Sci. Lett.* **22,** 328–338.

Oskvarek, J. D. & Perry, E. C., Jr. 1976: Temperature limits on the early Archaean ocean from oxygen isotope variations in the Isua supracrustal sequence, West Greenland. *Nature, Lond.* **259,** 192–194.

Perry, E. C., Jr. & Ahmad, S. N. 1977: Carbon isotope composition of graphite and carbonate minerals from 3.8-AE metamorphosed sediments, Isukasia, Greenland. *Earth planet. Sci. Lett.* **36,** 280–284.

Pflug, H. D. 1978: Yeast-like microfossils detected in oldest sediments of the Earth. *Naturwissenschaften* **65,** 611–615.

Pflug, H. D. & Jaeschke-Boyer, H. 1979: Combined structural and chemical analysis of 3,800-Myr-old microfossils. *Nature, Lond.* **280,** 483–486.

Radulescu, D. P. 1966: Rhyolites and secondary ultrapotassic rocks in the subsequent Neogene volcanism from the East Carpathians. *Bull. Volcanol.* **29,** 427–434.

Rivalenti, G. 1975: Chemistry and differentiation of mafic dykes in an area near Fiskenaesset, West Greenland. *Can. J. Earth Sci.* **12,** 721–730.

Rosing, M. 1983: A metamorphic and isotopic study of the Isua supracrustals, West Greenland. Unpublished cand. scient. thesis, Copenhagen University, 139 pp.

Schidlowski, M., Appel, P. W. U., Eichmann, R. & Junge, C. E. 1979: Carbon isotope geochemistry of the 3.7 × 10^9-yr-old Isua sediments, West Greenland: implications for the Archaean carbon and oxygen cycles. *Geochim. cosmochim. Acta* **43,** 189–199.

Sighinolfi, G. P. 1974: Geochemistry of early Precambrian carbonate rocks from the Brazilian Shield: implications for Archaean carbonate sedimentation. *Contr. Miner. Petrol.* **46,** 189–200.

Smith, G. M. & Dymek, R. F. 1983: A description and interpretation of the Proterozoic Kobbefjord Fault Zone, Godthåb district, West Greenland. *Rapp. Grønlands geol. Unders.* **112,** 113–127.

Wagner, P. A. 1982: Geochronology of the Ameralik dykes at Isua, West Greenland. Unpublished M.Sc. thesis, University Alberta, 83 pp.

Walker, R. G. 1984: Turbidites and associated coarse clastic deposits. *In* Walker, R. G. (edit.) *Facies models* (2nd ed.) *Geoscience Canada Reprint Series* **1,** 219–251.

Wells, P. R. A. 1980: Thermal models for the magmatic accretion and subsequent metamorphism of continental crust. *Earth planet. Sci. Lett.* **46,** 253–263.

Windley, B. F. 1969: Evolution of the early Precambrian basement complex of southern West Greenland. *Spec. Pap. geol. Ass. Can.* **5,** 155–161.

Winkler, H. G. F. 1974: *Petrogenesis of metamorphic rocks.* 319 pp. New York: Springer-Verlag.

Wyllie, P. J. 1977: Crustal anatexis: an experimental review. *Tectonophysics* **43,** 41–71.

ISUA SUPRACRUSTAL BELT

compiled by A. P. Nutman

1:40 000

Contour interval 50 m

Topography based on 1:20 000 unpublished maps prepared photogrammetrically by Aerokort A/S and Grønlands Geologiske Undersøgelse from aerial photographs taken in 1968, and triangulation by the Geodetic Institute, Denmark.
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QUATERNARY

- Ice and perennial snow
- Moraine on ice
- Undifferentiated moraine
- Fluviatile and glacial-fluviatile deposits

PROTEROZOIC

- G GRANITE DYKE
- BASIC DYKE

LATE ARCHAEAN (3400-2500 Ma)

- PEGMATITE DYKES
- DEFORMED GRANITOID SHEETS
- TASERSSUAQ TONALITE COMPLEX
- MULTIPHASE BANDED GNEISSES. May contain an early Archaean component locally
- UNDIFFERENTIATED AMPHIBOLITES
- TARSSARTÔQ DYKES. Ultramafic in composition
- TARSSARTÔQ DYKES. Basic in composition

EARLY ARCHAEAN AMÎTSOQ GNEISSES (3400-3800 Ma)

- GREY (TONALITIC) AND WHITE (GRANITIC) GNEISSES
- " " " " " " dominantly white gneiss
- " " " " " " with hornblendite bodies
- " " " " " " weakly deformed
- " " " " " " strongly deformed
- PEGMATITE SHEETS
- GREY GNEISS SHEETS IN SUPRACRUSTAL UNITS
- INALUK DYKE LOCALITIES

EARLY ARCHAEAN SUPRACRUSTAL ROCKS (*c.* 3800 Ma)

ISUA SUPRACRUSTAL BELT

- Sequence A Felsic formation – predominantly layered felsic metasediments
- " Calc-silicate formation – carbonates, quartzites, calc-silicate rocks and amphibolites
- " Upper banded iron formation – predominantly magnetite iron formation and metachert
- " Variegated schist formation – amphibolites, felsic rocks, metapelites and metacherts
- " Lower banded iron formation – predominantly magnetite iron formation
- " Amphibolite formation – predominantly banded amphibolites
- " Undifferentiated Variegated schist formation to Amphibolite formation
- Sequence B Mica schist formation – predominantly Fe-Mg rich mica schists
- " Felsic formation – predominantly layered felsic metasediments
- Garbenschiefer amphibolite – intrusive, Mg and Al rich basic rock
- Ultramafic rocks – metadunites and metaperidotites, probably intrusive
- Way-up from graded bedding

AKILIA ASSOCIATION

- Amphibolite
- Banded iron formation and metachert
- Ultramafic rocks
- Leucoamphibolite
- Biotite amphibolite and felsic rocks

- Sharp boundary
- Inferred boundary
- Transitional boundary
- Foliation and compositional layering
- Foliation in dyke
- Mineral lineation and shape fabrics
- " " " " " in dyke
- Fold axis
- Dyke or pegmatite inclination
- Crenulation cleavage
- Subvertical fault
- Inclined fault
- Sense of displacement of fault
- gr Grønlandite
- tk Soapstone
- Fe Iron at Isukasia

Kartografi: Annette Thorning Rasmussen

Trykt ved Geodætisk Institut 1985.