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Inventory and assessment of Palaeoarchaean gneiss terrains and detrital zircons in southern West Greenland

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Abstract

Geochemical methods to advance knowledge on the early Earth require supplies of well-preserved >3600 Ma rocks and minerals (e.g. zircon). One of the most important resources for these are small domains within high metamorphic grade Palaeoarchaean gneisses in the Archaean Craton, southern West Greenland. In the Nuuk region these gneisses occur as two terranes (totalling ~3000 km^2) with differences in their pre-3600 Ma histories, and they are tectonically separated from each other by younger rocks. In the north is the *Isukasia terrane* (early metamorphic grade amphibolite facies). It is devoid of pre-3600 Ma in situ partial melt and contains locally well-preserved 3690 and 3810 Ma tonalites, ≥3810 Ma ultramafic rocks plus the tectonically composite Isua supracrustal belt with some locally well-preserved ~3800 and 3710 Ma volcanic and sedimentary materials. To the south is the *Færingehavn terrane* (up to granulite facies in the Palaeoarchaean) that is dominated by 3850–3660 Ma migmatites containing variable amounts of in situ 3660–3600 Ma partial melt. Akilia island in this terrane contains some ~3850 Ma tonalites, mafic and sedimentary rocks – the world's oldest-known sediments. Domains of lower total strain and anatexis on some of other islands near Akilia has left small amounts of well-preserved 3850 Ma tonalite.

There are two less known smaller bodies of Palaeoarchaean rocks north of the Nuuk region. The *Qarliit Tasersuat assemblage* consists of polyphase, migmatitic gneisses with lenses of mafic and siliceous rocks. Two SHRIMP zircon dates reveal 3600–3700 Ma rocks, strongly affected by ~3600 and 2770 Ma metamorphisms. The *Aasivik terrane* also consists of polyphase migmatitic gneisses, and previous SHRIMP U/Pb zircon reconnaissance dating of three samples found components up to ~3600 Ma old with strong reworking in 2720–2550 Ma events.

The West Greenland Archaean Craton is a collage of Palaeo- to Neoarchaean terranes, assembled in several Archaean events. Metasediments *within* post-Palaeoarchaean terranes are devoid of ≥3600 Ma detritus, but are dominated by zircons of the same age as major crust-forming TTG suites of their terrane. Metasediments along terrane boundaries, even those in contact with Palaeoarchaean terranes, contain very few (<5%) ≥3600 Ma detrital zircons. Therefore, these sediments are not a significant resource for ancient zircons. The scarcity of ≥3600 Ma detritus within these sediments supports a model that the Palaeoarchaean bodies (Færingehavn, Isukasia, Qarliit tasersuat and Aasivik) are allochthonous terranes captured within an Archaean accretionary

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system and comprise a tectonic assembly of juvenile crustal blocks with different age. The metasediments in contact with the Isukasia terrane are dominated by ~3070 Ma detrital zircons and were first metamorphosed (along with the adjacent Isukasia terrane) at ~2960 Ma. On the other hand, metasediments in contact with the Færingehavn terrane are dominated by ~2831 Ma detrital zircons, and thus were deposited after sediments in contact with the Isukasia terrane had already been tectonically emplaced and metamorphosed. Although the Palaeoarchaean terranes were incorporated into the Archaean terrane collage at different times, they might have been spawned from a single larger body of ancient crust broken up from ~3500 Ma onwards.

Keywords: Palaeoarchaean; Crustal evolution; Zircons; Greenland

1. Introduction

The world's oldest rocks (3600–4020 Ma – e.g., Black et al., 1971; Bowring and Williams, 1999) and oldest minerals (zircons back to 4000–4400 Ma – e.g., Froude et al., 1983; Wilde et al., 2001; Mojzsis et al., 2001) are windows onto the early Earth. They can provide direct information on geological processes at the great age when they formed. In addition radiogenic isotopic studies on them provide information on the evolution of terrestrial reservoirs (mantle to atmosphere) in the earliest stages of Earth's history. Some of these materials are old enough to show any isotopic anomalies due to short-lived radionuclides (e.g. the ^{146}Sm–^{142}Nd parent daughter system) that were extant in the first few hundred million years of Earth's history – with the potential for providing the clearest information on the evolution of early terrestrial reservoirs. These isotopic studies and more pedestrian geological and non-isotopic geochemical ones require resources of diverse ancient rocks and minerals for study, least corrupted by geological processes in their long life since they formed 4400–3600 million years ago.

Palaeoarchaean rocks now form only about a millionth of Earth's surface, and are all contained in Archaean gneiss complexes. In these complexes they were intruded by younger granitoids and underwent variable but generally strong ductile deformation during multiple amphibolite or granulite facies metamorphism(s) (500–800 °C, 5–10 kbar), which in some complexes culminated in variable degrees of in situ partial melting. The Palaeoarchaean rocks are predominantly tonalitic–granitic "banded grey" gneisses derived from plutonic protoliths, which enclose remnants of ultramafic, gabbroic and supracrustal rocks (e.g., McGregor, 1973; Bridgwater et al., 1976; Collerson and Bridgwater, 1979; Myers, 1988; Black et al., 1986; Schiøtte et al., 1989a; Bowring et al., 1989; Kinny and Nutman, 1996, Nutman et al., 1996, 2000; Bowring and Williams, 1999; Friend et al., 2002a). To varying degrees these rocks have been chemically modified from their igneous and (rarer) sedimentary protoliths during many tectonothermal events (see Nutman, 1986; Rosing et al., 1996 for extreme alteration examples). The state of preservation of these key old materials makes them difficult targets for geochemical and isotopic investigations. Thus all samples from these gneiss complexes are not equal, and detailed fieldwork is necessary to seek out small domains of rocks and minerals least modified by younger geological processes for the greatest success in geochemical studies. A key resource for these materials occurs in southern West Greenland, the subject of this paper.

West Greenland contains the world's most diverse and most studied Palaeoarchaean rock record. The Narryer Gneiss Complex of Western Australia (e.g., Myers, 1988; Nutman et al., 1991; Kinny and Nutman, 1996) is of similar extent to the West Greenland Palaeoarchaean rocks. However, it is very poorly exposed (<0.1% of its likely area), and what rocks are exposed are generally weathered and dominated by closely banded polyphase migmatites, with mostly 3660–3300 Ma components. This makes it a less useful resource for early Earth studies. The Uivak Gneiss Complex in northern Labrador (e.g., Collerson and Bridgwater, 1979) is about as extensive as the Palaeoarchaean rocks in West Greenland. Although it is almost as well exposed, a considerable part of it suffered Neoarchaean granulite facies metamorphism (Collerson and Bridgwater, 1979), plus it is logistically more difficult and expensive to work in. Thus the Labrador Palaeoarchaean rocks have not been studied as intensively as the Greenland ones.

2. Archaean Craton of southern West Greenland

The world's first discoveries of Palaeoarchaean rocks were in the Nuuk district of the Archaean Craton of southern West Greenland (Fig. 1). They were discovered by combined fieldwork (McGregor, 1968, 1973; Bridgwater and McGregor, 1974) and dating by whole rock Pb/Pb, Rb/Sr (Black et al., 1971; Moorbath et al., 1972, 1973) and bulk zircon methods (Baadsgaard, 1973). The majority of these rocks are quartzo-feldspathic orthogneisses and were named the *Amîtsoq gneisses*. The Amîtsoq gneisses are cut by the *Ameralik* (basic) *dykes*, that are now mostly strongly deformed and amphibolitised by Neoarchaean tectonothermal events. The presence of these dykes served as a field tool whereby McGregor (1973) could distinguish the older Amîtsoq gneisses from the younger *Nûk gneisses* that are not cut by the dykes. Despite several subsequent discoveries of Palaeoarchaean rocks around the world, those in the Nuuk region remain a focus for studies of the early Earth. This is because they form a large extent ($\sim$3000 km^2) of well-exposed, unweathered rocks, which has allowed location of small domains least affected by post-Palaeoarchaean geological events. In an overview of these rocks with presentation of many new zircon dates, Nutman et al. (1996) named these Palaeoarchaean rocks collectively the *Itsaq Gneiss Complex*. "Complex" was deliberately chosen to stress the diversity of these rocks and that the "Amîtsoq gneisses" are not cogenetic. Within the best-preserved domains there are fragments of upper mantle peridotites and layered gabbro complexes, granitoids ranging from diorite to granite, mafic and felsic volcanic rocks and sediments (e.g. Bridgwater et al., 1976; Nutman et al., 1996, 2001; Komiya et al., 1999), with ages ranging from $\geq$3800 Ma (Michard-Vitrac et al., 1977; Baadsgaard et al., 1984; Compston et al., 1986; Nutman, 1990; Nutman et al., 1996, 1997a, 2000, 2002; Krogh et al., 2002; Mojzsis and Harrison, 2002a; Crowley, 2003) to 3600 Ma (e.g. Baadsgaard, 1973; Nutman et al., 1996, 2002; Crowley et al., 2002). The diversity and age span of ancient rocks found in West Greenland allows a broad range of early Earth problems to be addressed, and compliments other investigations on the >4000 Ma detrital zircons in Jack Hills and Mt. Narryer metasediments of Western Australia (Maas et al., 1992; Wilde et al., 2001; Mojzsis et al., 2001). North of the Nuuk region there are two lesser known bodies of Palaeoarchaean rocks, the *Aasivik terrane* (Rosing et al., 2001) and what we name here the *Qarliit Tasersuat assemblage* (Fig. 1). There are also units of discounted and unproven Palaeoarchaean rocks (indicated by "x" and "?" symbols, respectively on Fig. 1). All these bodies are covered in this paper, to present a complete inventory of known Palaeoarchaean rocks from southern West Greenland. This inventory concentrates only on the salient characteristics of these rocks and the debates concerning them that are of most significance for geochemical constraints on early Earth history.

The Palaeoarchaean rocks of the Nuuk region were originally thought to be large enclaves within the Meso- to Neoarchaean *Nûk gneisses* (e.g. McGregor, 1973; Taylor et al., 1980). The Palaeoarchaean rocks are certainly cut by intrusions of younger crustally derived granite. However, prior to intrusion of these granites, the Palaeoarchaean rocks were emplaced as mylonite-bounded blocks within a terrane collage (each terrane with their own rock ages and internal early evolution). This collage was assembled in the later part of the Archaean with multiple strong deformation and amphibolite facies metamorphism (Friend et al., 1987, 1988, 1996; Nutman et al., 1989; McGregor et al., 1991; Crowley, 2002). A recent development of the terrane model is that the northern and southern parts of the Itsaq Gneiss Complex of the Nuuk region are recognised as separate terranes with different late Archaean assembly histories (Friend and Nutman, in press-a). In the central-southern part of the Nuuk region is the *Færingehavn terrane*, containing the island of Akilia – the focus of several recent controversies. In the north is the *Isukasia terrane* (previously regarded as the northern end of the Færingehavn terrane), which contains important localities such as the Isua supracrustal belt (Fig. 1).

The terrane assembly model for evolution of the Craton has been extended southwards out of the Nuuk region almost to the border of the Palaeoproterozoic Ketilidian mobile belt (Friend and Nutman, 2001). It is also a likely model for evolution of the northern part of the Craton as well (Friend and Nutman, 1994; Rosing et al., 2001). Sensitive high mass resolution ion micro probe (SHRIMP) U/Pb dating of detrital zircons in two younger Archaean sediments in the Nuuk region found a small proportion of Palaeoarchaean

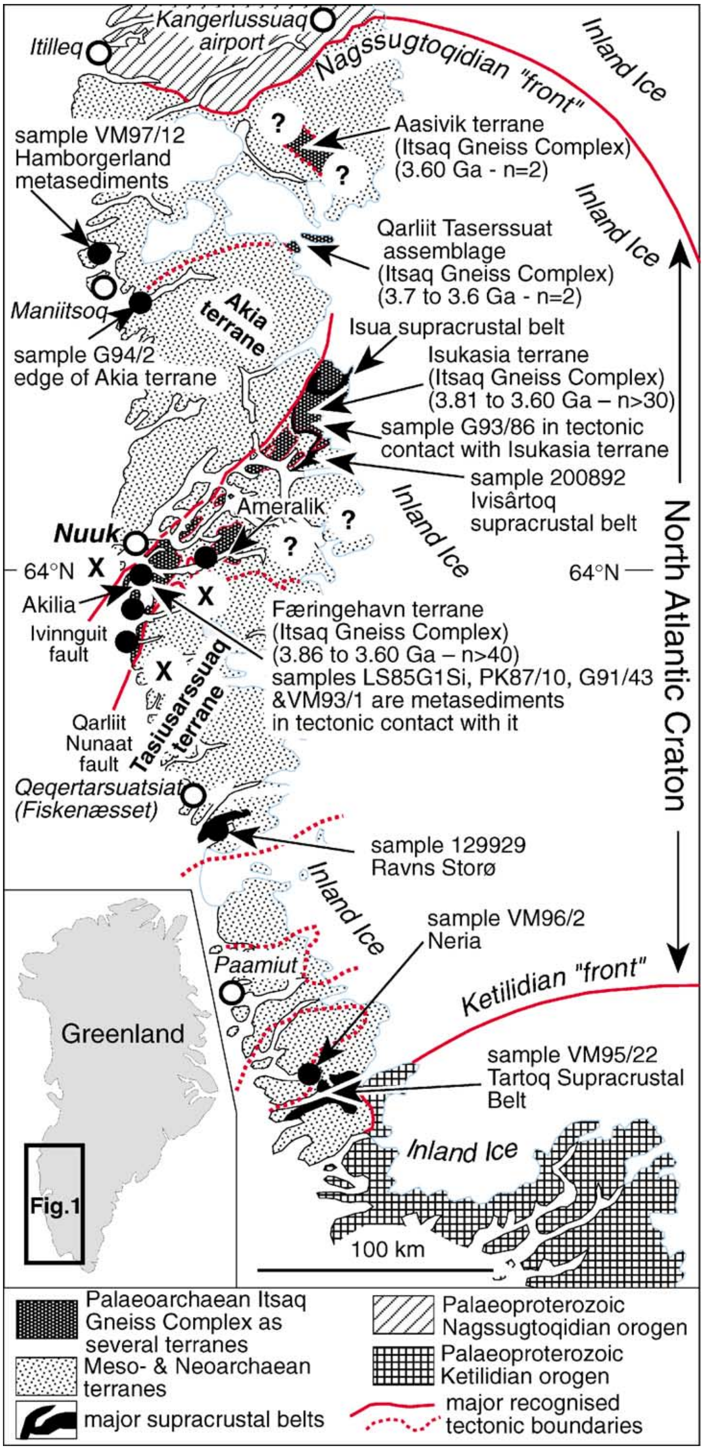

Fig. 1. Geological map of the Archaean craton of West Greenland.

grains (Schiøtte et al., 1988). Therefore, West Greenland metasediments could be a potential resource for very ancient zircons, like the Western Australian sediments from Jack Hills and Narryer. This possibility is pursued in this paper by dating detrital zircons from many metasediments across the whole Craton (Fig. 1). In fact, Palaeoarchaean detrital zircons are found to be rare in all these sediments, thus they do not contain a significant resource of materials for early terrestrial studies. However the detrital zircon age spectra provide further information on the later Archaean tectonic assembly and evolution of the Craton. This is relevant to understanding the geological setting of the Palaeoarchaean rocks in the region.

The dates reported here are by the SHRIMP U/Pb zircon method. General analytical techniques, data calibration and analytical error assessment is described by Stern (1998) and Williams (1998). Nutman et al. (2000) compares dating of some of the Greenland Archaean rocks by both the SHRIMP and IDTIMS methods.

3. Palaeoarchaean rocks

3.1. *Færingehavn terrane*

The Færingehavn terrane is the late Archaean tectonic entity that contains the world's first Palaeoarchaean rock localities. The earliest field and geochemical studies recognised the polyphase, migmatitic nature of these rocks (McGregor, 1973). The predominant quartzo-feldspathic gneisses are derived from granitoids of different composition – tonalite, Fe-rich granite and diorite (McGregor, 1979; O'Nions and Pankhurst, 1974; Nutman et al., 1984a). These gneisses contain a varied suite of ultramafic, mafic and iron-rich siliceous rocks, as lenses up to 1 km long (but generally much smaller). One such body of mafic and siliceous rocks occurs on the southwestern tip of the island Akilia (Fig. 1), and due to the ease of spelling and pronunciation of "Akilia", the name *Akilia association* was adopted for all these bodies (McGregor and Mason, 1977). McGregor and Mason (1977) demonstrated that the Akilia association contained rocks derived from komatiites, basalts, gabbros and banded iron formation. McGregor and Mason (1977) and Griffin et al. (1980) provided integrated field and petrographic evidence that the rocks of the Færingehavn terrane had suffered Palaeoarchaean (ca. 3600 Ma) granulite facies metamorphism, but had been largely retrogressed under amphibolite facies conditions in later events.

Because banded gneisses dominate the Færingehavn terrane, understanding of its Palaeoarchaean evolution is difficult, and is surrounded by controversy. In a few areas of lower deformation, there are rare intrusive relationships between original plutonic components that are the precursors to the banded gneisses. These components were overprinted and infiltrated by variable amounts of Palaeoarchaean partial melt during metamorphism(s) up to granulite facies (Fig. 2a and Nutman et al., 1996, 2000, 2002a; Friend and Nutman, in press-b). At most localities the partial melt was incorporated into the banded gneiss fabric by superimposed strain – and thus is difficult to recognise (Fig. 2b–d). For geochemical studies, the significance of this is that only tiny amounts of the banded grey tonalitic gneisses in the Færingehavn terrane may be regarded as representing single-phase igneous protoliths with closed chemical systems.

Zircon dating of the Færingehavn terrane is summarised in Fig. 3a and b. Nutman et al. (1993, 1996, 1997a, 2000, 2002a) interpreted the multiple ages indicated the presence of rocks of many ages from ca. 3850 to 3600 Ma. Tonalites have a spread of ages from ca. 3850 to 3660 Ma, whereas granites range in age from ca. 3660 to 3600 Ma. Zircons grew again and/or were recrystallised in high-grade metamorphic events starting after 3700 Ma (Figs. 3a and b). Kamber and Moorbath (1998) and Whitehouse et al. (1999) contested this interpretation and proposed that mostly rocks of only 3600–3700 Ma are present. They explained that the abundant >3800 Ma zircons in these rocks, even in the strongly Zr-undersaturated tonalitic compositions, were derived from some distal reservoir with ancient zircons (this reservoir had to be free of significant Nd, Sr and common Pb that could "pollute" their suite of "juvenile" 3700–3600 Ma rocks). In the way we read these papers, it is implicit that they treat their gneiss samples as derived from single phases of igneous rock. However, as explained above, examination in the field shows that this is rarely the case (Nutman et al., 2000, 2002a). Illuminating in this respect is sample 110 999, dated by ion microprobe by both Kinny (1986) and Whitehouse et al. (1999), with very different interpretations (protolith ages of ca. 3820 versus 3660 Ma, respectively). However, 110 999 is clearly a polyphase rock, with neosome (partial melt?) transformed into a subtle layering by later strong deformation (Figs. 2c and d). We regard the polyphase, migmatitic nature of these rocks as the reason for the complex age structure of the zircons from such rocks as 110 999 (see CL (cathodoluminescence) images in Whitehouse et al., 1999).

In some of the rare low strain zones in the banded gneisses of the Færingehavn terrane, relict igneous relationships combined with SHRIMP U/Pb zircon dating were used to show that at several localities ca. 3850 Ma tonalites intrude mafic rocks ± possible banded iron formation – BIF (Nutman, 1990; Nutman et al., 1996, 1997a, 2000, 2002a). At Akilia a tonalite body (Fig. 2e) with surprisingly well-preserved zircons (see CL images in Nutman et al., 2000, 2002a,b) gave a SHRIMP U/Pb zircon age of ca. 3850 Ma (Nutman et al., 1997a, 2000, 2002a,b). If the interpretation of this body as a 3850 Ma discordant intrusion is correct (Nutman et al., 1997a, 2000, 2002a), then the mafic rocks and possible BIF that they cut are the world's oldest-known such rocks. Myers and Crowley (2000) disputed that the tonalite body (Fig. 2e) is locally discordant. Instead, they interpreted that the observed relationship to be caused by heterogeneous strong deformation, with

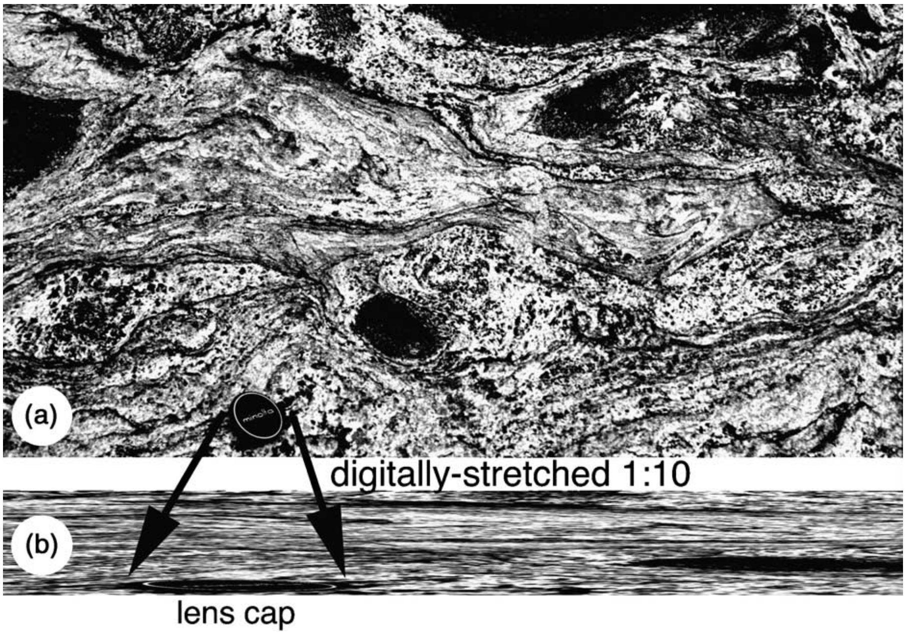

Fig. 2. (a) Early Archaean migmatite structures preserved in orthogneisses from a small part of the Færingehavn terrane that escaped strong Neoarchaean deformation. Note the heterogeneous nature of the gneisses, with older partly assimilated components included within a more leucocratic components – the zircons in such rocks are structurally complex, reflecting the complex geology (Nutman et al., 2000, 2002a). (b) part of image (a) that has been digitally stretched by approximately 1:10. Note that even at this moderate strain all the details of the migmatite are hard to resolve and parts of the rock already appear "homogeneous". (c and d) Field appearance of GGU110999 whose zircons have been dated by ion probe by Kinny (1986) and Whitehouse et al. (1999). Note that it is strongly deformed, with felsic veins or segregations incorporated into the (folded) banding. Thus this sample is polyphase. (e) Locally discordant tonalitic sheet G93/05 (top) cutting (bottom) hornblendite (black) siliceous material (white) and banded amphibolites (banded grey). Note that the edge of the tonalitic sheet clearly truncates the hornblendite – siliceous rock – banded amphibolite lithological boundaries, a relationship we interpret as a relict igneous discordance. The zircons tonalitic sheet G93/05 are structurally simple, with only one generation of (3850 Ma) igneous zircon present (Nutman et al., 2000, 2002a; Mojzsis and Harrison, 2002a). (f) Area of both low Neoarchaean and Palaeoarchaean strain on an island near Akilia. Ca. 5 cm lens cap for scale lies to the left of the mafic body (dark). The tonalite (pale) is homogeneous and non-banded, apart from some neosome development that is concentrated around mafic bodies. Despite this overprint, the tonalite clearly breaks-up the mafic rocks. SHRIMP U/Pb zircon dating of these invasive tonalites has given a date for igneous zircons of ca. 3850 Ma and for metamorphic overgrowths and recrystallisation of ca. 3660 Ma (Nutman, unpublished SHRIMP data).

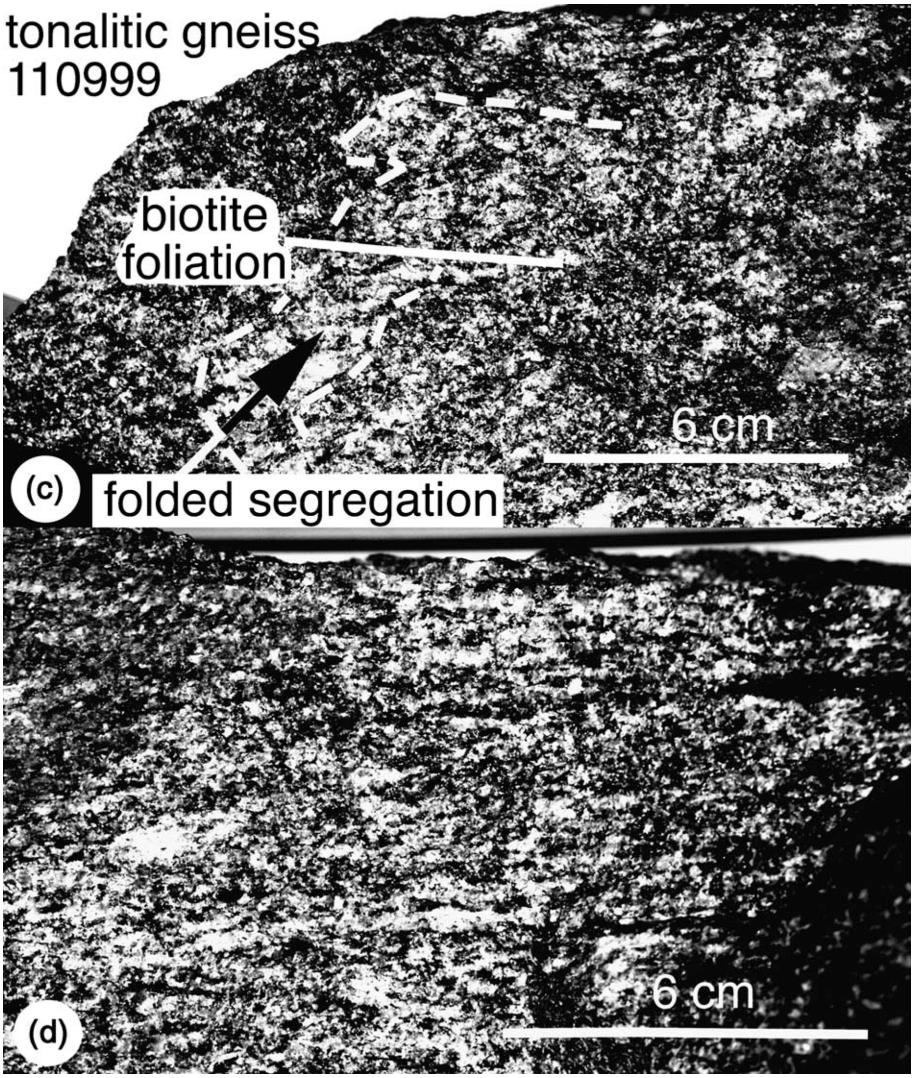

Fig. 2. (*Continued*)

the relative age relationship between the different rock types no longer certain. Strain along the tonalite body is variable, but, at the photographed part, it would appear to us that the tonalite body is less deformed and cuts banded amphibolite, siliceous material and hornblendite. We consider this as a fortuitous preservation of a discordant contact at the margin of an intrusive tonalite sheet. Mojzsis and Harrison (2002a) have now replicated the Nutman et al. (1997a, 2000, 2002a) 3850 Ma age on the sheet (Fig. 2e) and another sheet of similar age is reported by Manning et al. (submitted for publication). Also Krogh et al. (2002) concluded that ca. 3850 Ma rocks are present on Akilia. Thus there is growing support of our work that there are ca. 3850 Ma rocks on Akilia. In a way it is a pity that Akilia has been the sole focus of other workers, because there is evidence on adjacent islands for more ≥3850 Ma rocks (Nutman et al., 1996, 2000, 2002a), making Akilia less of a "special case". A most dramatic demonstration of this is from one of the small islands near Akilia, where in the best area of least strain yet found, ca. 3850 Ma tonalites clearly intrude mafic and ultramafic rocks (Fig. 2f).

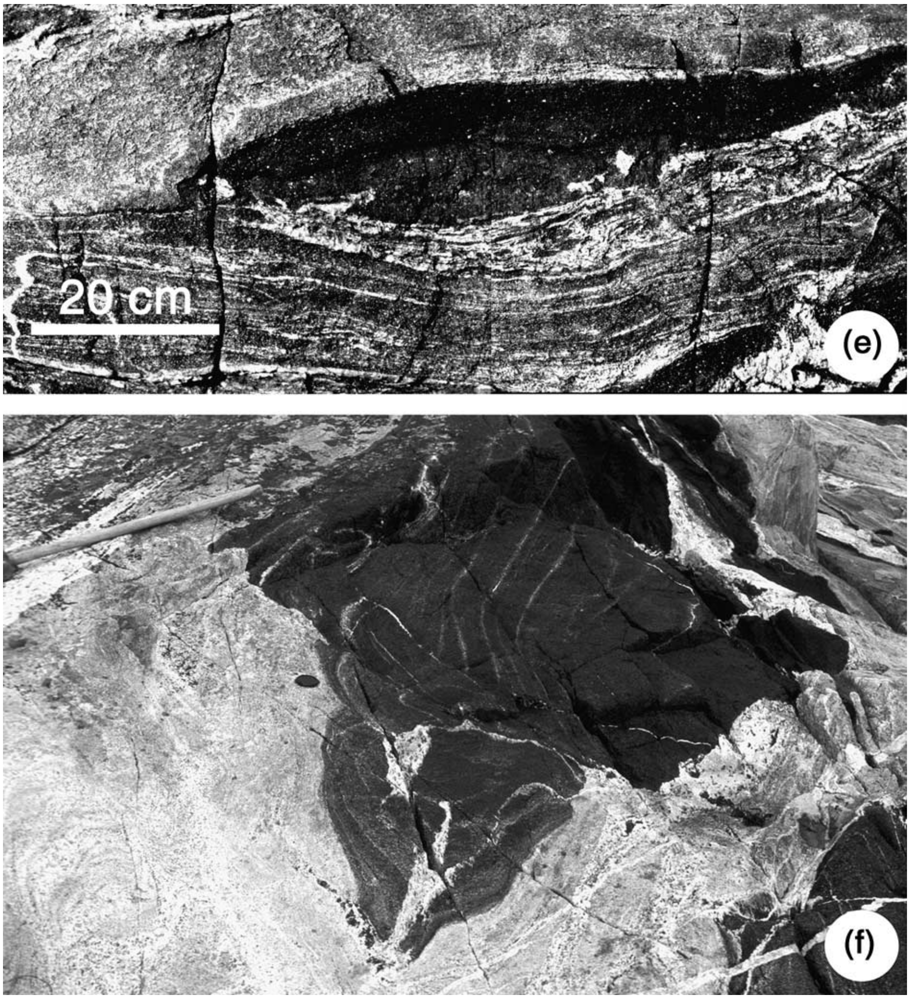

Fig. 2. (*Continued*).

Another strand of the Akilia controversy centres on Akilia's multiply folded unit of siliceous rocks. Nutman (1990) and Nutman et al. (1997a) proposed that this unit contains ≥3850 Ma BIF, whereas Fedo and Whitehouse (2002) interpreted it to be entirely of metasomatic origin and thus irrelevant in study of early Earth environments and the search for life. Fedo and Whitehouse's study concentrated on coarse-grained quartz + amphibole + pyroxene ± garnet parts of the siliceous unit already recognised to be partly of metasomatic origin in previous studies (McGregor and Mason, 1977; Nutman et al., 1997a). Fedo and Whitehouse (2002) did not discuss finer-grained rocks in the unit with magnetite-rich layering that were highlighted by Nutman et al. (1997a) as likely to be BIF. These magnetite-bearing rocks have major and trace element chemistry resembling Archaean BIF (Friend et al., 2002b; Mojzsis and Harrison, 2002b), and are quite different in composition from the materials in the same unit highlighted by Fedo and Whitehouse (2002). Mojzsis et al. (2003) detected mass-independent S isotopic effects in the Akilia BIF. These effects are considered to arise in the early atmosphere, with isotopically anomalous sulfur then rained-out and was trapped in sediments (Farquahar et al., 2000). This S-isotopic evidence further supports a sedimentary origin for the Akilia possible BIF.

Therefore, there is a growing case to be made that rocks as old as 3850 Ma are present in the Færingehavn terrane, and include not only tonalites and mafic rocks, but probably BIF as well. These mafic rocks and BIF, despite their strong structural modification and granulite facies overprint are the oldest-known of their type in the world (older than better preserved ones in the Isua supracrustal belt – see below). As such they provide an imperfect, but unique window on early terrestrial environments at 3850 Ma. Most significantly we would argue they give direct evidence for a retained hydrosphere by 3850 Ma.

Another strand of controversy surrounding the Akilia siliceous unit is whether it contains evidence for life. The unit is much too metamorphosed (up to 800 °C) and deformed to retain any direct evidence via microfossils. Instead, evidence must rely on features such as isotopically light carbon. Preferably this carbon should be trapped in a capsule of a robust mineral formed during diagenesis, so that its isotopic signature can be argued as an original feature. Mojzsis et al. (1996) proposed that phosphate minerals (particularly (hydroxy)apatite) commonly associated in sediments with biogenic material could recrystallise during diagenesis to apatite, forming a capsule that trapped organic carbon throughout progressive metamorphism. Using Akilia sample G91/26 of Nutman and Friend (interpreted by them as a recrystallised BIF) Mojzsis et al. (1996) reported carbon in association with apatite in sufficient abundance to determine accurately its isotopic signature. The carbon was found to be light (δ ^{13}C = −20 to −50) and interpreted to be of ancient (≥3850 Ma) biogenic origin. This would place the emergence of life prior to the *preserved* sedimentary record, and earlier in Earth's history than many would have expected. Like any paradigm-breaking result, this evidence put forward for very early life has been the subject of debate since its announcement (Hayes, 1996 to Lepland et al., 2004).

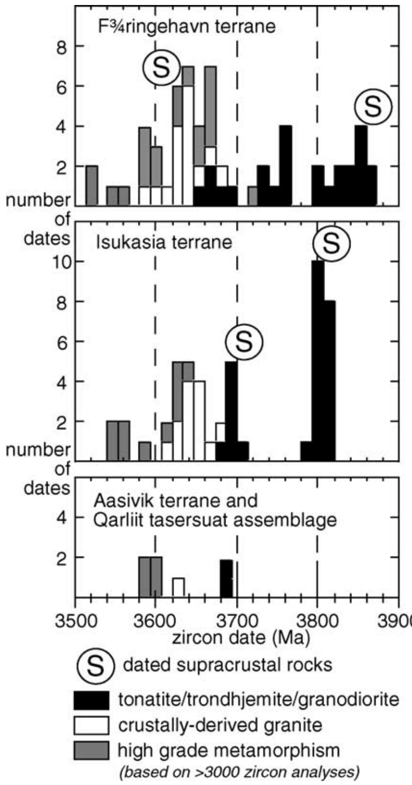

Fig. 3. Summary of U/Pb zircon age determinations of plutonic rocks and events in the Færingehavn terrane, Isukasia terrane, Aasivik terrane and the Qarliit Tasersuat assemblage. In total, over 3000 zircon analyses have been used to derive the ages summarised in this figure. Data are from (Nutman et al., 1993, 1996, 1997a,b, 1999, 2000, 2002a,b; our interpretation of Whitehouse et al., 1999; Crowley et al., 2002; Mojzsis and Harrison, 2002a; Krogh et al., 2002; Crowley, 2003; Friend and Nutman, in press-b).

The Færingehavn terrane contains the augen gneiss suite of iron-rich granitoids and gabbros (McGregor, 1973). This forms about 20% of the terrane south of Ameralik. These rocks featured heavily in early whole rock isotopic studies, because they were recognised to be more homogeneous than the more abundant banded grey tonalitic gneisses, and because in Rb–Sr geochronology their high Rb/Sr ratios gave greater "spread" for the isochron suites constructed from regional collections of gneisses. Zircon dating on these rocks indicates that they have an age of 3630–3640 Ma (Baadsgaard, 1973; Nutman et al., 1996, 2000; Whitehouse et al., 1999). They have trace element chemistry like younger A-type or rapakivi granites (e.g. enriched in Nb, Y, Zr – Nutman et al., 1996). These represent the oldest-know granites of this type in the world, and they probably formed by hybridisation between fractionated mantle melts and melting of older continental crust (Nutman et al., 1984a). This hybrid origin of the augen gneiss suite is reflected in magmatic zircons from gabbros and granites in the suite having different initial $\varepsilon_{\mathrm{Hf}}$ values (see discussion by Nutman et al., 2000 on Hf isotopic data for samples GGU155817 and GGU155820 presented by Vervoort et al., 1996). The Pb isotopic compositions from relicts of igneous K-feldspar in these granites have been long regarded to represent the mantle at 3630–3640 Ma, and hence has been used in modelling early crustal evolution via Pb isotopes (e.g. Kramers and Tolstikin, 1997; Kamber and Moorbath, 1998 – see discussion by McGregor, 2000). This juvenile interpretation of these rocks is inconsistent with their geochemistry and the small amount of Hf isotope data available on their magmatic zircons.

3.2. Isukasia terrane

What we now call the *Isukasia terrane* (Fig. 1) contains the Isua supracrustal belt and adjacent orthogneisses that were the world's second discovery of Palaeoarchaean rocks (Moorbath et al., 1973; Bridgwater and McGregor, 1974). It is a different tectonic slice from the previously discussed Færingehavn terrane (Friend and Nutman, in press-a). From the outset it was realised that the Isukasia terrane contains Palaeoarchaean rocks much less affected by younger Archaean tectonothermal events than the Færingehavn terrane. Bridgwater and McGregor (1974) found abundant cross-cutting relationships preserved between the plutonic components of the orthogneisses, and southwards could document their transition into banded gneisses, which are more like the Færingehavn terrane "Amîtsoq gneisses". This terrane has suffered metamorphism up to only amphibolite facies (Griffin et al., 1980), and is free of in situ partial melt formation (Nutman et al., 1996, 2000). Combined with some low strain zones of appreciable size, this means there is greater scope for finding unequivocal field relationships, to give a broadly accepted geological context to geochronological and geochemical results.

Upon realisation that the ca. 35 km long Isua supracrustal belt with abundant BIF was >3700 Ma (Moorbath et al., 1973), it became the main focus of early terrestrial geology. To a large degree it still remains so (see volume 126 of Precambrian Research). Geological and geochemical investigations of the belt are hampered by locally severe metasomatism combined with generally high strain (Nutman et al., 1984b; Rosing et al., 1996; Myers, 2001). None the less, lower strain zones within the belt contain relict volcanic and sedimentary structures. The first of these including graded layering in felsic metasediments were discovered in the 1970s (McGregor, unpublished data). Nutman et al. (1984b) found more graded layering and definite conglomerates, but proposed that some previously reported conglomerates (Allaart, 1976; Bridgwater et al., 1976) were probably tectonic artefacts. Fedo (2000) revisited this conglomerate debate. The belt contains a large unit of amphibole + chlorite schist known as the "garbenschiefer". Throughout the 1970s and 1980s this was interpreted as a highly altered gabbroic sill (e.g. Bridgwater et al., 1976; Appel, 1977; Nutman et al., 1984b). At the start of the 1990s locally very well-preserved pillow structures were found in the garbenschiefer unit (Maruyama et al., 1992; Komiya et al., 1999). These pillows demonstrated that the garbenschiefer unit consists of generally strongly deformed volcanic rocks. These pillow discoveries were a focus of later studies on the belt (Appel et al., 1998). The better preserved rocks such as these pillows facilitates finding original igneous compositions to understand their petrogenesis and to stereotype their tectonic setting via comparisons with younger mafic suites (Polat et al., 2003).

Early Rb/Sr, Pb/Pb, Sm/Nd whole rock geochronology (e.g. Moorbath et al., 1973; Jacobsen and Dymek, 1987) had insufficient precision to resolve different ages for Isua supracrustal belt rocks. This lack of precision was a handicap in understanding the belt's stratigraphic and tectonic evolution. Unfortunately the first more precise zircon dating was restricted to one locality thus giving a robust age to only one rock in the belt (Michard-Vitrac et al., 1977; Baadsgaard et al., 1984). This rock is a quartzo-feldspathic schist with granitic clasts of debated origin. In the earliest studies this lithology has been interpreted to be a conglomerate (Allaart, 1976), whereas more recently Myers (2001) has interpreted it as a highly altered tonalitic sheet. Our interpretation (e.g. Nutman et al., 1984b, 1997b) is that it is an altered felsic volcano-sedimentary rock, and that the clasts are tectonic boudins. James (1976) interpreted the belt as a rim syncline of volcanosedimentary materials invaginated into the adjacent gneisses. Nutman (1984) interpreted the belt as one or two stratigraphic packages, in tectonic contact with each other and with the surrounding orthogneiss areas. In the early structural studies it was implicit that all rocks in the belt were of similar age. SHRIMP U/Pb zircon dating of rocks from the belt started with Compston et al. (1986) on the same rock previously dated by Michard-Vitrac et al. (1977) and Baadsgaard et al. (1984). Since then, dating of many rocks from different units by SHRIMP has shown that the belt has panels of supracrustal rocks with different ages (ca. 3700 and 3800 Ma) that are in early Archaean tectonic contact with each other (Nutman et al., 1996, 1997b, 2002b). Other workers have also concluded that the belt is tectonically composite (Komiya et al., 1999; Appel et al., 1998; Myers, 2001). For geochemical studies, whether it be on the mafic rocks for mantle evolution or on the metasediments for quantifying bolide flux and searching for signs of early life, the different ages of these rocks must be taken into account. It is also surprising that the first reported possible ^{142}Nd anomaly – resulting from fractionations of Sm very early in Earth's history (Harper and Jacobsen, 1992) is from a unit with a U/Pb zircon age of only ca. 3710 Ma (Nutman et al., 1996, 1997b, 2002b; B. Kamber personal communication to A. Nutman, 2004).

The first Palaeoarchaean age of the Isua BIFs also demonstrated the great antiquity of the hydrosphere (>3700 Ma – Moorbath et al., 1973). Thus the Isua sediments have been searched for evidence for life at 3700–3800 Ma. Some studies have attempted to find microfossils (Pflug and Jaeschke-Boyer, 1979), but most attempts to detect life have been via carbon isotope geochemistry. Schidlowski et al. (1979) reported somewhat negative δ^{13}C values (ca. −20) from bulk samples of Isua graphites, which were interpreted as a more negative δ^{13}C early life signature that had been corrupted by open-system behaviour during subsequent metamorphism. In a graphite-bearing Isua supracrustal belt sample provided by P.W.U. Appel, Mojzsis et al. (1996) reported graphite in association with apatite with δ^{13}C values between −20 and −35. Like the similar result from Akilia island, this was taken as evidence that life existed when the 3700–3800 Ma Isua rocks formed. Van Zuilen et al. (2003) studied unequivocal quartz + magnetite BIF from the Isua belt, and failed to find an association between apatite and graphite. Rosing (1999) extracted fine-grained negative δ^{13}C (−10 to −20) graphite particles from a felsic turbiditic metasediment from the Isua belt. This was interpreted as another line of evidence for life when these rocks formed. Clearly there will be continuing debate over evidence for life in the Isua rocks. However all workers would agree that the Isua belt contains the world's best-preserved surficial 3700–3800 Ma rocks, and thus will remain an important focus for research.

Pb/Pb isotopic compositions of galenas from the Isua supracrustal belt were first reported by Appel et al. (1978). As the then least radiogenic terrestrial Pb compositions these featured in models of early Earth processes. More recent discoveries of galenas from the belt (e.g. Frei and Rosing, 2001) include some with even less radiogenic Pb than the first discoveries. These new discoveries point to early terrestrial crust/mantle reservoirs with μ values as high as 7.7, and thus provide new perspectives in understanding the early Earth.

The orthogneisses surrounding the Isua supracrustal belt have been less studied, but contain a wealth of well-preserved Palaeoarchaean materials. North of the belt there are domains where variably deformed tonalites derived from melting of hydrated mafic crust are intruded by still undeformed crustally derived granite sheets (Nutman, 1984; Nutman and Bridgwater, 1986; Baadsgaard et al., 1986; Crowley et al., 2002). SHRIMP zircon geochronology shows that the tonalites are 3700–3690 Ma old, whereas the granite sheets have ages of 3650–3630 Ma (Nutman et al., 1993, 1996, 2000, 2002b; Crowley et al., 2002).

The superficially similar orthogneisses south of the Isua supracrustal belt are dominated by ca. 3800 Ma tonalites (i.e. 100 million years older than the tonalites to the north – Nutman et al., 1993, 1996, 1999, 2000, 2002b). The ca. 3800 Ma zircon ages have been replicated by Crowley (2003) and Kamber et al. (2003), and no controversy surrounds this age. These tonalites contain rare domains where they have remained little deformed since they were intruded (Nutman et al., 1999). Xe isotopic signatures of well-preserved ca. 3810 Ma igneous zircons from one of these tonalites are best explained by the incorporation of ^{244}Pu into them when they formed, with Earth having accreted with a chondritic ^{244}Pu/^{238}U ratio (Honda et al., 2003). This result also supports other noble gas studies for the great age of Earth's atmosphere.

South of the Isua supracrustal belt there are abundant pods of ultramafic rocks within the orthogneisses, which from SHRIMP U/Pb zircon geochronology can be demonstrated to be Palaeoarchaean in age, probably >3800 Ma (Nutman et al., 1996; Friend et al., 2002a). Rare patches (<1%) with least hydration and metasomatism in these ultramafic rocks consist of almost anhydrous relicts of aluminous spinel bearing dunite and harzburgite. These rocks provide a rare sample of the early Archaean mantle (Friend et al., 2002a), with the

Os isotopic systematics of the spinels confirming the antiquity of these rocks and that they are consistent with the Earth having accreted with a chondritic $^{187}Os/^{188}Os$ ratio (Bennett et al., 2002). Bennett et al. (2002) also pointed out that the "normal" PGE abundance in these rocks indicated that PGE from any post core formation late accretionary veneer that had been well mixed into the mantle by 3800 Ma. Another important suite of rocks south of the Isua supracrustal belt are >3800 Ma layered gabbro-anorthosite, layered ultramafic (including chromitites) complex(es) that occur as scattered fragments within the >3800 Ma tonalites (Chadwick and Crewe, 1986; Nutman et al., 1996; Friend et al., 2002a). For the study of the early mantle, these gabbros are an attractive alternative to the metavolcanic rocks of the Isua supracrustal belt that have been studied in more detail.

3.3. Discounted and unproven Palaeoarchaean rocks in the Nuuk region

The first systematic mapping of the Nuuk region (McGregor, 1973, 1984) designated some rocks to the northwest and southeast of the Færingehavn terrane as smaller enclaves of the Palaeoarchaean rocks in the younger Nûk gneisses (shown by "X" on Fig. 1). Lithologically, these resemble the Palaeoarchaean rocks of the Færingehavn terrane because they are heterogeneous, polyphase, with remnants of amphibolite dykes. However, when examples of these (most infamous at Kangimut Sammisoq in Ameralik) were dated by Rb/Sr, Pb/Pb and Nd/Sm whole rock methods (Jones et al., 1986; Moorbath et al., 1986) they yielded Neoarchaean errochrons with no evidence of long crustal residence from the initial ratios. SHRIMP U/Pb zircon work on some of these contentious gneisses (Kinny, 1987; Schiøtte et al., 1989b) found no Palaeoarchaean zircon in these rocks, even as cores. An understanding for these contentious rocks was found in the new tectonic interpretation that the gneisses of the region consist of disparate, unrelated terranes, which were tectonically assembled towards the end of the Archaean (Friend et al., 1987, 1988). Thus, regional Archaean mylonites separate the contentious "X" localities from the isotopically proven Palaeoarchaean Færingehavn terrane (Nutman et al., 1989). The gneisses at the "X" localities are simply older components in other terranes. In the Tasiusarsuaq terrane these rocks are ca. 2920 Ma old (Kinny, 1987; Schiøtte et al., 1989b), and are inclusions in mostly 2860–2840 Ma gneisses. In the Akia terrane such gneisses are ca. 3200 Ma old and are associated with 3060–2970 Ma gneisses (Baadsgaard and Nutman, unpublished SHRIMP data).

In the east of the Nuuk region there are units totalling 100–200 km^2 marked as "?" on Fig. 1 that in 1970s reconnaissance mapping were designated as Palaeoarchaean (Allaart et al., 1977). A sample of the easternmost of these units has yielded a SHRIMP U/Pb zircon Neoarchaean age of ca. 2820 Ma (L. Schiøtte, unpublished data), whereas the others have yet to be tested by any geochronological method. They are a potential Palaeoarchaean resource worthy of investigation.

3.4. Aasivik terrane

The poorly known Aasivik terrane (Rosing et al., 2001) is the northernmost entity of Palaeoarchaean rocks in the Craton (Fig. 1) and is estimated to underlie ca. 1500 km^2. Like the Isukasia and Færingehavn terranes, it appears to be in late Archaean tectonic contact with adjacent terranes of Neoarchaean gneisses (Rosing et al., 2001). The Aasivik terrane is dominated by garnet-bearing granulite facies polyphase banded orthogneisses and migmatites, partially retrogressed under amphibolite facies conditions. The Aasivik terrane orthogneisses contain lenses and pods of mafic and ultramafic rocks, which might be older inclusions within the gneisses. To date, no definite sedimentary rocks have been identified within the Aasivik terrane. Reconnaissance SHRIMP U/Pb zircon dating in the Aasivik terrane identified protolith components in the orthogneisses up to ca. 3600 Ma, but with abundant new zircon growth at ca. 2700–2550 Ma, equated with the late granulite facies metamorphism and migmatisation overprints (Rosing et al., 2001). This terrane clearly suffered Neoarchaean granulite facies metamorphism. In this respect it is a unique Palaeoarchaean terrane in West Greenland, as the other three terranes suffered only amphibolite facies metamorphism later in the Archaean. The late granulite facies overprint and associated migmatisation reduces the potential of this terrane for studies of the early Earth, because of the greater possibility of elemental and isotopic fractionation in these severe events almost a billion years after the rocks formed.

3.5. Qarliit tasersuat assemblage

The *Qarliit tasersuat assemblage* is an entity of Palaeoarchaean rocks outcropping around the lake Qarliit tasersuat at the edge of the Inland Ice, about 50 km north of the Isua supracrustal belt (Fig. 1). It was discovered in helicopter regional reconnaissance geological mapping (Hall, 1978), and was revisited for further reconnaissance and assessment by McGregor in 1997. It is dominated by amphibolite facies polyphase banded to migmatitic gneisses with a lesser amount of strongly deformed augen granite gneisses also present (Hall, 1978). The scale and complexity of these gneisses makes it hard to separate out components of the original igneous protoliths. These gneisses contain broken up tabular amphibolites, equated with the Mesoarchaean Ameralik dykes cutting the Palaeoarchaean rocks in the Nuuk region. Within the orthogneisses are lenses of magnetite-layered quartz rich rock, interpreted as BIF (Hall, 1978) and lenses of mafic and ultramafic rocks. The BIF contains sufficiently unradiogenic Pb to point to an early Archaean component in the Qarliit tasersuat assemblage (P.N. Taylor, personal communication to C.R.L. Friend, 1984).

To investigate the Palaeoarchaean history of these rocks further, SHRIMP U/Pb zircon dating has been undertaken on two polyphase orthogneisses. VM97/17 is from a small island in the eastern part of Qarliit tasersuat (65°46.75′N, 50°43.24′W). On the island are strongly foliated (probably polyphase) orthogneisses, with scattered enclaves of mafic and hornblendic lithologies, and broken up fragments of homogeneous tabular amphibolite, probably derived from mafic dykes. The specimen was collected adjacent to one of the dyke-like amphibolites. It yielded 150–250 μm long prismatic zircons, some with overgrowths clearly visible in CL imaging. Much of the main component of prismatic zircon displays micron-scale euhedral zoning. In many grains this zoning is still concordant to the exterior of the crystal. In others, the zoning has been truncated by overgrowth of recrystallisation domains. Fourteen analyses were undertaken on 10 grains. Cores and whole prismatic grains yielded mostly concordant dates with $^{207}Pb/^{206}Pb$ dates >3500 Ma (Table 1 , Fig. 4a). The distribution of the $^{207}Pb/^{206}Pb$ dates suggests two or more generations of zircon are present, rather than a single, ca. 3700 Ma, variably disturbed group (Fig. 4a). The three "oldest" determinations have $^{207}Pb/^{206}Pb$ ages around 3670–3700 Ma, and it appears that a 3550–3600 Ma group is also present. Although only a reconnaissance study has been made on this gneiss, it is clearly dominated by >3500 Ma components. Five overgrowths on Palaeoarchaean zircons were also dated. These are typically high U, commonly with low Th/U. They yielded a weighted mean $^{207}Pb/^{206}Pb$ date of 2781 ± 8 Ma (MSWD = 0.45). This is interpreted as dating a thermal event that give rise to either new zircon growth throughout the gneiss or the emplacement of granitic veins of that age.

VM97/18 is from a unit of banded orthogneisses on the east side of Qarliit Tasersuat (65°47.25′N, 50°42.69′W). The sample is of a more homogeneous nature than the average gneisses. The sample yielded abundant 150–250 μm long prismatic zircons, some with overgrowths clearly visible in CL imaging. Much of the main component of prismatic zircon displays micron-scale oscillatory zoning. Twenty-five analyses were undertaken on 20 grains. Cores and whole prismatic grains yielded mostly concordant dates with $^{207}Pb/^{206}Pb$ dates >3500 Ma (Table 1, Fig. 4b). The distribution of $^{207}Pb/^{206}Pb$ dates suggests several age groups of zircons are present. As more zircon analyses were undertaken on this sample than VM97/17, the age distribution of the Palaeoarchaean zircons was examined in more detail. Mixture modelling (Sambridge and Compston, 1994) suggests that three age components are most likely to be present; 3680 ± 9 Ma, 3639 ± 12 Ma and 3599 ± 17 Ma (MSWD = 0.91). Treated as a single group that had suffered ancient loss of radiogenic Pb, the six "oldest" give a $^{207}Pb/^{206}Pb$ date of 3681 ± 7 Ma with a high MSWD of 2.49. No clear morphological features or Th/U values can be distinguished between analyses assigned to different age groups. Clearly the sample is recording a complex Palaeoarchaean history, with components as old as ca. 3700 Ma present. This complex history is in keeping with the banded nature of the unit from which VM97/18 was sampled. VM97/18 zircons also have younger domains. Some have dates intermediate between ca. 3500 and 2800 Ma and are of partly recrystallised grain centres and cores, and are interpreted as disturbed domains within >3500 Ma grains. There are also sites with typically higher U and commonly low Th/U. Five overgrowths plus a single core (analysis 18.2) yielded a weighted mean $^{207}Pb/^{206}Pb$

Table 1
SHRIMP U/Pb zircon analyses

Labels	Grain type	U (ppm)	Th (ppm)	Th/U	Comm. 206/Pb%	238U/206/Pb ratio	207Pb/206Pb ratio	207/206 date (Ma)	% Disc
VM97/17 banded gneiss from the Qarliit Taserssuat assemblage (Itsaq Gneiss Complex)									
1.1	e, osc, p	186	126	0.68	0.11	1.281 ± 0.032	0.3263 ± 0.0018	3600 ± 8	3
2.1	m, h, p, fr	158	144	0.91	0.06	1.331 ± 0.031	0.3216 ± 0.0030	3577 ± 14	1
3.1	m, osc, p, fr	139	30	0.22	0.32	1.228 ± 0.029	0.3480 ± 0.0026	3698 ± 12	4
4.1	e, hd, p	2214	63	0.03	0.01	1.866 ± 0.036	0.1947 ± 0.0013	2782 ± 11	−1
4.2	m, hd, p	1476	33	0.02	0.05	1.868 ± 0.037	0.1939 ± 0.0014	2776 ± 12	0
5.1	e, osc/hd, p	793	141	0.18	1.93	1.442 ± 0.028	0.3203 ± 0.0012	3571 ± 6	−5
6.1	m, rex, p	391	70	0.18	0.08	1.358 ± 0.026	0.3164 ± 0.0014	3552 ± 7	0
6.2	e, hd/osc, p	733	61	0.08	0.08	1.355 ± 0.039	0.3307 ± 0.0025	3620 ± 12	−2
7.1	m, osc, p, fr	402	176	0.44	0.03	1.314 ± 0.037	0.3437 ± 0.0038	3679 ± 17	−1
8.1	e, osc, p, fr	122	29	0.23	0.15	1.334 ± 0.044	0.3478 ± 0.0026	3697 ± 12	−2
9.1	e, hd, ov	994	54	0.05	0.10	1.914 ± 0.046	0.1939 ± 0.0008	2776 ± 7	−2
10.1	r, hd, p	2755	38	0.01	0.03	1.831 ± 0.049	0.1950 ± 0.0015	2785 ± 13	1
10.2	c+r, p	608	25	0.04	0.09	1.776 ± 0.131	0.2191 ± 0.0041	2974 ± 31	−3
10.3	r, hd, p	2508	75	0.03	0.16	1.978 ± 0.036	0.1955 ± 0.0010	2789 ± 9	−5
VM97/18 banded gneiss from the Qarliit Taserssuat assemblage (Itsaq Gneiss Complex)									
1.1	c, osc, p	268	89	0.33	0.02	1.255 ± 0.035	0.3475 ± 0.0029	3696 ± 13	2
2.1	m, rex, p	235	19	0.08	0.18	1.609 ± 0.065	0.2416 ± 0.0025	3131 ± 16	0
3.1	c, osc, p	151	44	0.29	0.18	1.331 ± 0.043	0.3523 ± 0.0061	3717 ± 27	−3
4.1	c, rex, p	183	82	0.45	0.06	1.349 ± 0.035	0.3256 ± 0.0021	3596 ± 10	−1
4.2	r, hd, p	826	53	0.06	6.18	1.994 ± 0.049	0.1897 ± 0.0018	2739 ± 16	−4
5.1	c, h, p	55	78	1.40	0.22	1.287 ± 0.041	0.3356 ± 0.0038	3643 ± 17	2
6.1	m, p	310	47	0.15	0.06	1.287 ± 0.034	0.3393 ± 0.0021	3660 ± 9	1
7.1	m, rex/osc, p	314	25	0.08	0.42	1.390 ± 0.039	0.3358 ± 0.0026	3644 ± 12	−4
8.1	c, rex, p	35	44	1.27	1.94	1.315 ± 0.056	0.3096 ± 0.0071	3519 ± 36	4
8.2	r/rex? p	732	31	0.04	0.03	1.909 ± 0.051	0.1931 ± 0.0009	2768 ± 7	−2
8.3	r/rex? p	615	35	0.06	1.17	1.817 ± 0.044	0.1950 ± 0.0014	2785 ± 12	2
9.1	r, h, p	115	35	0.31	1.95	1.502 ± 0.315	0.3187 ± 0.0036	3563 ± 17	−8
10.1	m, osc, p	370	38	0.10	0.28	1.238 ± 0.030	0.3456 ± 0.0013	3688 ± 6	4
11.1	m, hd, p	883	207	0.23	0.24	1.335 ± 0.031	0.3326 ± 0.0013	3629 ± 6	−1
12.1	c, osc/rex, p	289	62	0.21	0.40	1.294 ± 0.032	0.3416 ± 0.0013	3670 ± 6	1
13.1	m, osc, p	148	11	0.07	6.83	1.315 ± 0.039	0.3337 ± 0.0055	3634 ± 25	0
14.1	m, osc, p	265	68	0.26	0.35	1.282 ± 0.031	0.3451 ± 0.0016	3686 ± 7	1
15.1	c, osc, p	448	99	0.22	0.32	1.357 ± 0.035	0.3377 ± 0.0024	3653 ± 11	−3
16.1	c, osc/rex, p	93	15	0.16	0.05	1.436 ± 0.041	0.2950 ± 0.0032	3444 ± 17	−1
17.1	c, rex, p	180	21	0.12	0.48	1.756 ± 0.046	0.2581 ± 0.0019	3235 ± 12	−10
17.2	r, h, p	532	36	0.07	1.89	1.883 ± 0.046	0.1864 ± 0.0017	2711 ± 15	1
18.1	r, hd, p	393	226	0.58	0.22	1.834 ± 0.044	0.1845 ± 0.0010	2694 ± 9	4
18.2	c, hd, p	2530	178	0.07	0.02	1.763 ± 0.040	0.1931 ± 0.0004	2768 ± 4	5
19.1	r, hd/osc, p	361	46	0.13	0.43	1.874 ± 0.050	0.1923 ± 0.0019	2762 ± 16	0
20.1	r, h, p	1343	75	0.06	2.95	1.819 ± 0.043	0.1953 ± 0.0016	2787 ± 14	1
VM95/22 metasediment, Tartoq Group									
1.1	m, osc/d, p	330	205	0.62	0.03	1.815 ± 0.035	0.2140 ± 0.0005	2936 ± 4	−4
2.1	m, osc/rex, p	198	106	0.54	0.14	1.883 ± 0.045	0.2079 ± 0.0017	2889 ± 13	−5
3.1	e, osc, p	253	317	1.25	0.03	1.796 ± 0.032	0.2090 ± 0.0014	2898 ± 11	−2
4.1	m, osc, p	321	157	0.49	0.10	1.751 ± 0.067	0.2344 ± 0.0027	3082 ± 19	−5
5.1	e, osc, p, fr	174	75	0.43	0.06	1.297 ± 0.025	0.3447 ± 0.0029	3684 ± 13	0
6.1	m, osc, p	24	14	0.60	0.13	1.830 ± 0.077	0.2137 ± 0.0020	2933 ± 15	−4
7.1	m, osc, p	124	84	0.68	<0.01	1.884 ± 0.055	0.2157 ± 0.0020	2949 ± 15	−7

Table 1(*Continued*)

Labels	Grain type	U (ppm)	Th (ppm)	Th/U	Comm. 206/Pb%	238U/206/Pb ratio	207Pb/206Pb ratio	207/206 date (Ma)	% Disc
7.2	rex, hb, p	140	71	0.51	0.07	1.938 ± 0.073	0.2095 ± 0.0013	2901 ± 10	−8
7.3	e, osc, p	54	18	0.33	0.05	1.874 ± 0.079	0.2117 ± 0.0019	2919 ± 15	−5
8.1	m, osc, eq	171	84	0.49	0.07	1.806 ± 0.036	0.2121 ± 0.0022	2921 ± 17	−3
9.1	m, osc, p	168	100	0.59	0.04	1.300 ± 0.027	0.3650 ± 0.0020	3771 ± 8	−2
10.1	m, osc/h, p, fr	187	250	1.34	0.03	1.784 ± 0.042	0.2156 ± 0.0012	2948 ± 9	−3
11.1	e, osc, p	203	153	0.75	1.11	1.967 ± 0.038	0.2003 ± 0.0018	2829 ± 15	−6
11.2	e, osc, p	153	98	0.64	0.06	1.854 ± 0.058	0.2014 ± 0.0009	2837 ± 7	−2
12.1	e, osc, p	139	201	1.45	0.48	1.817 ± 0.037	0.2270 ± 0.0013	3031 ± 9	−7
12.2	e, osc, p	549	193	0.35	0.55	2.454 ± 0.085	0.1767 ± 0.0014	2622 ± 14	−16
13.1	m, osc, p	56	37	0.66	0.55	1.909 ± 0.045	0.1999 ± 0.0018	2825 ± 15	−4
14.1	e, h, eq	253	31	0.12	0.08	1.893 ± 0.033	0.2008 ± 0.0011	2833 ± 9	−3
15.1	m, fr	145	136	0.94	0.17	1.889 ± 0.045	0.2000 ± 0.0010	2826 ± 8	−3
16.1	c, osc, p, fr	102	74	0.72	0.12	1.776 ± 0.045	0.2383 ± 0.0018	3109 ± 12	−7
17.1	m, osc, p	297	143	0.48	0.08	1.838 ± 0.045	0.2130 ± 0.0033	2929 ± 25	−4
18.1	?c, osc, p	352	94	0.27	0.10	1.779 ± 0.056	0.2482 ± 0.0027	3173 ± 17	−9
19.1	e, osc, p	140	51	0.37	0.08	1.806 ± 0.042	0.2045 ± 0.0009	2863 ± 7	−1
19.2	e, osc, p	162	67	0.41	0.06	1.862 ± 0.056	0.2031 ± 0.0008	2851 ± 6	−3
20.1	m, h/rex, fr	248	18	0.07	0.08	1.890 ± 0.034	0.2099 ± 0.0006	2905 ± 5	−6
20.2	e, h/rex, fr	241	84	0.35	0.23	1.850 ± 0.069	0.2065 ± 0.0010	2878 ± 8	−3
21.1	e, h/rex, fr	180	96	0.53	0.26	1.747 ± 0.054	0.2317 ± 0.0048	3064 ± 33	−5
21.2	e, osc/d, p	406	255	0.63	0.87	2.482 ± 0.116	0.2276 ± 0.0021	3035 ± 15	−28
22.1	m, osc, p	162	133	0.82	0.10	1.890 ± 0.044	0.2077 ± 0.0015	2888 ± 12	−5
22.2	e, osc, p	169	109	0.64	0.08	1.821 ± 0.044	0.2082 ± 0.0009	2892 ± 7	−2
23.1	m, osc/hd, p	395	123	0.31	0.05	1.403 ± 0.024	0.3363 ± 0.0013	3646 ± 6	−5
24.1	c, osc/rex, p	158	56	0.36	2.11	2.139 ± 0.125	0.2755 ± 0.0056	3338 ± 32	−26
25.1	e, osc/rex, p	82	26	0.32	0.20	1.756 ± 0.042	0.2127 ± 0.0012	2926 ± 9	−1
26.1	m, osc, p	70	57	0.80	0.58	1.674 ± 0.101	0.2104 ± 0.0027	2909 ± 21	4
27.1	e, osc, p	166	101	0.61	0.06	1.946 ± 0.070	0.1994 ± 0.0014	2821 ± 11	−5
28.1	e, osc, p, fr	195	181	0.93	0.02	1.748 ± 0.053	0.2022 ± 0.0010	2844 ± 8	3
29.1	e, osc/d, p, fr	498	297	0.60	0.10	2.554 ± 0.104	0.1789 ± 0.0010	2643 ± 9	−19
30.1	e, osc, eq	117	69	0.59	0.18	1.803 ± 0.056	0.2090 ± 0.0015	2897 ± 12	−2
31.1	e, osc, p	164	168	1.03	0.15	1.806 ± 0.066	0.2025 ± 0.0013	2846 ± 11	0
32.1	e, osc, p	159	173	1.09	0.10	1.808 ± 0.048	0.2013 ± 0.0012	2837 ± 10	0
33.1	e, osc, p, fr	248	182	0.73	1.37	2.201 ± 0.057	0.2099 ± 0.0022	2905 ± 17	−17
34.1	e, osc, ov, fr	246	155	0.63	0.08	1.797 ± 0.058	0.2086 ± 0.0009	2895 ± 7	−1
35.1	e, osc/d, p	642	162	0.25	0.88	1.504 ± 0.046	0.3104 ± 0.0016	3523 ± 8	−7
36.1	e, osc/d, eq	250	203	0.81	0.16	2.117 ± 0.080	0.1839 ± 0.0009	2688 ± 8	−7
37.1	e, osc/d, p, fr	744	520	0.70	1.11	2.508 ± 0.104	0.1616 ± 0.0069	2473 ± 74	−12
38.1	e, osc, p, fr	86	64	0.74	0.26	1.749 ± 0.109	0.2083 ± 0.0022	2892 ± 17	1
VM96/2 metasediment in Neria, between Neria and Sermiligaarsuk terranes									
1.1	m, h, p	317	136	0.43	0.04	1.774 ± 0.114	0.2141 ± 0.0045	2937 ± 34	−2
2.1	m, osc/h, ov	61	34	0.56	0.01	1.827 ± 0.059	0.2264 ± 0.0024	3027 ± 17	−7
2.2	m, osc/h, ov	118	52	0.44	0.36	1.839 ± 0.055	0.2093 ± 0.0030	2900 ± 23	−3
3.1	m, h, ov/anh	143	43	0.30	0.23	2.146 ± 0.055	0.1962 ± 0.0023	2795 ± 19	−12
4.1	e, osc/h, p	239	126	0.53	0.85	1.895 ± 0.061	0.2051 ± 0.0012	2867 ± 10	−5
5.1	m, h/osc, ov	101	47	0.46	0.11	1.779 ± 0.061	0.2221 ± 0.0038	2996 ± 28	−4
6.1	m, h, p	158	71	0.45	0.04	1.714 ± 0.039	0.2241 ± 0.0025	3011 ± 18	−2
7.1	c, osc, p	64	53	0.82	0.13	1.873 ± 0.047	0.2090 ± 0.0019	2897 ± 15	−5
8.1	e, osc, p, fr	45	33	0.73	0.18	1.786 ± 0.046	0.2204 ± 0.0027	2984 ± 20	−4
8.2	e, rex, p, fr	299	85	0.28	0.09	1.917 ± 0.053	0.1905 ± 0.0015	2746 ± 13	−1
9.1	m, h, p, fr	428	241	0.56	0.11	2.295 ± 0.045	0.2023 ± 0.0009	2845 ± 8	−18
9.2	r, h, p, fr	853	47	0.06	0.06	2.045 ± 0.054	0.1749 ± 0.0005	2605 ± 5	−1

Table 1(*Continued*)

Labels	Grain type	U (ppm)	Th (ppm)	Th/U	Comm. 206/Pb%	238U/206/Pb ratio	207Pb/206Pb ratio	207/206 date (Ma)	% Disc
10.1	m, h, p	274	183	0.67	0.09	1.910 ± 0.037	0.2076 ± 0.0009	2887 ± 7	−6
11.1	e, osc/h, p	336	188	0.56	0.03	1.874 ± 0.053	0.2164 ± 0.0022	2954 ± 16	−7
12.1	m, osc, ov	309	114	0.37	0.01	1.875 ± 0.036	0.2132 ± 0.0013	2930 ± 9	−6
13.1	m, h, p	158	80	0.50	0.18	1.679 ± 0.038	0.2267 ± 0.0028	3029 ± 20	−1
14.1	m, h, p	716	81	0.11	0.07	1.822 ± 0.052	0.2075 ± 0.0014	2886 ± 11	−2
15.1	e, osc, p	824	580	0.70	0.04	2.177 ± 0.071	0.2071 ± 0.0010	2883 ± 8	−16
16.1	h, p, fr	179	152	0.85	0.04	1.769 ± 0.039	0.2212 ± 0.0017	2990 ± 12	−3
17.1	e, h, p, fr	78	67	0.86	0.10	1.760 ± 0.068	0.2260 ± 0.0031	3024 ± 22	−4
18.1	e, osc, p	54	20	0.36	0.17	1.811 ± 0.044	0.2043 ± 0.0031	2861 ± 25	−1
19.1	m, osc, p, fr	352	216	0.62	0.31	1.837 ± 0.037	0.2152 ± 0.0017	2945 ± 13	−5
20.1	m, h/osc, p	144	55	0.38	0.18	1.822 ± 0.098	0.2118 ± 0.0023	2919 ± 18	−3
21.1	c, h, ov	46	27	0.58	0.03	1.679 ± 0.040	0.2279 ± 0.0024	3037 ± 17	−1
22.1	m, hd, ov	514	66	0.13	0.07	1.809 ± 0.069	0.2105 ± 0.0006	2910 ± 5	−3
23.1	r, h, p, fr	51	26	0.51	0.25	1.953 ± 0.055	0.1883 ± 0.0019	2727 ± 17	−2
24.1	m, osc, p, fr	49	47	0.96	0.06	1.808 ± 0.042	0.2020 ± 0.0022	2842 ± 18	0
25.1	m, osc, p	113	88	0.78	0.11	1.687 ± 0.061	0.2207 ± 0.0014	2986 ± 10	1
26.1	m, osc, p, fr	581	301	0.52	0.04	2.277 ± 0.049	0.2063 ± 0.0006	2877 ± 5	−18
27.1	m, h/osc, p	78	64	0.82	0.01	1.665 ± 0.058	0.2273 ± 0.0023	3033 ± 16	0
28.1	e, h, p, fr	165	138	0.84	0.01	1.527 ± 0.201	0.2234 ± 0.0102	3006 ± 75	8
29.1	m, h/osc, p	251	37	0.15	0.02	2.078 ± 0.210	0.1981 ± 0.0027	2811 ± 22	−10
29.2	r? h, p	828	71	0.09	0.06	2.104 ± 0.057	0.1815 ± 0.0010	2667 ± 9	−6
30.1	c, osc, p	38	29	0.76	0.01	1.742 ± 0.186	0.2219 ± 0.0022	2994 ± 16	−2
31.1	c, osc, p	41	19	0.45	0.61	1.766 ± 0.289	0.2239 ± 0.0042	3009 ± 30	−4
32.1	e, hd, p	267	76	0.28	0.13	1.848 ± 0.052	0.2148 ± 0.0017	2942 ± 13	−5
33.1	m, osc/h, p	186	54	0.29	0.14	1.653 ± 0.073	0.2113 ± 0.0016	2916 ± 12	5
34.1	e, hh, p	332	184	0.55	0.27	1.947 ± 0.052	0.2023 ± 0.0017	2845 ± 14	−6
35.1	c, h, p	153	100	0.65	0.04	1.687 ± 0.129	0.2115 ± 0.0091	2917 ± 72	3
36.1	c, h, ov	119	59	0.50	0.02	1.752 ± 0.052	0.2211 ± 0.0016	2989 ± 12	−3
37.1	m, osc, p, fr	34	24	0.71	0.50	1.693 ± 0.077	0.2202 ± 0.0036	2982 ± 27	0
38.1	m, hd, ov	453	126	0.28	0.02	1.417 ± 0.038	0.3296 ± 0.0022	3615 ± 10	−5
38.2	e, hd, ov	582	68	0.12	0.14	1.480 ± 0.042	0.2775 ± 0.0022	3349 ± 12	−1
39.1	m, h, p, fr	390	50	0.13	0.02	1.837 ± 0.623	0.2060 ± 0.0450	2874 ± 408	−3
40.1	e, osc, p	152	45	0.30	0.23	1.822 ± 0.087	0.2011 ± 0.0034	2835 ± 28	0
40.2	m, osc, p	116	73	0.63	0.19	1.846 ± 0.089	0.2171 ± 0.0028	2959 ± 21	−6
41.1	e, osc, p, fr	198	215	1.09	0.07	1.810 ± 0.067	0.2089 ± 0.0019	2897 ± 15	−2
42.1	m, osc/h, p	63	39	0.61	1.16	1.832 ± 0.066	0.2166 ± 0.0059	2956 ± 45	−5
43.1	m, osc, p, fr	95	96	1.01	0.45	1.627 ± 0.057	0.2206 ± 0.0030	2985 ± 22	3
44.1	m, osc, p, fr	89	24	0.27	0.17	1.678 ± 0.063	0.2331 ± 0.0027	3074 ± 19	−2
45.1		198	496	2.50	0.16	1.815 ± 0.061	0.2079 ± 0.0027	2889 ± 21	−2
46.1		261	69	0.27	1.40	1.805 ± 0.069	0.2070 ± 0.0021	2882 ± 16	−1
47.1		458	85	0.19	1.36	1.817 ± 0.061	0.2040 ± 0.0019	2858 ± 16	−1
48.1	e, osc, p, fr	190	89	0.47	0.49	1.906 ± 0.116	0.2030 ± 0.0035	2851 ± 28	−5
49.1	e, h, p	117	41	0.35	0.38	1.935 ± 0.071	0.2112 ± 0.0024	2914 ± 18	−8
50.1	m, h, anh	787	20	0.03	0.06	2.024 ± 0.109	0.1774 ± 0.0007	2629 ± 7	−1
51.1	m, osc/rex, ov	464	58	0.13	0.12	2.077 ± 0.059	0.1795 ± 0.0014	2649 ± 12	−4
52.1	c, osc, oc	274	198	0.72	0.10	1.812 ± 0.056	0.2033 ± 0.0012	2853 ± 10	−1
52.2	r, h, ov	175	76	0.44	0.35	2.054 ± 0.075	0.1896 ± 0.0021	2739 ± 18	−7
53.1	m, h, p, fr	105	77	0.73	0.37	1.816 ± 0.062	0.2062 ± 0.0024	2876 ± 19	−2
54.1	m, osc, ov	426	148	0.35	5.10	2.032 ± 0.070	0.2094 ± 0.0254	2901 ± 211	−11
55.1	m, h, p	115	62	0.54	0.31	1.698 ± 0.050	0.2196 ± 0.0025	2978 ± 19	0
56.1	e, osc, p	133	85	0.64	0.92	1.808 ± 0.064	0.2210 ± 0.0040	2988 ± 30	−5

Table 1(*Continued*)

Labels	Grain type	U (ppm)	Th (ppm)	Th/U	Comm. 206/Pb%	238U/206/Pb ratio	207Pb/206Pb ratio	207/206 date (Ma)	% Disc
57.1		709	468	0.66	0.19	1.905 ± 0.052	0.1799 ± 0.0012	2652 ± 11	3
58.1		325	150	0.46	0.12	1.659 ± 0.065	0.2178 ± 0.0017	2965 ± 13	3
59.1		595	198	0.33	0.05	2.135 ± 0.067	0.1840 ± 0.0011	2689 ± 10	−8
60.1	e, hd, p	954	674	0.71	1.00	2.391 ± 0.080	0.1927 ± 0.0014	2765 ± 12	−18
129929 Metasediment in Ravns Storø supracrustal belt, in Tasiusarssuaq terrane									
1.1	m, osc, p	60	54	0.90	0.18	1.669 ± 0.072	0.2070 ± 0.0059	2882 ± 47	5
2.1	m, osc, p	175	81	0.46	<0.01	1.742 ± 0.046	0.2097 ± 0.0026	2903 ± 20	1
3.1	e, osc, p	102	66	0.64	0.09	1.760 ± 0.083	0.2101 ± 0.0020	2906 ± 16	0
4.1	c/m, osc, p	226	117	0.52	0.05	2.011 ± 0.110	0.2046 ± 0.0049	2864 ± 39	−9
5.1	m, osc, p, fr	190	128	0.67	0.00	1.723 ± 0.072	0.2080 ± 0.0047	2890 ± 37	2
6.1	m, osc, p, fr	94	37	0.39	0.17	1.756 ± 0.099	0.2067 ± 0.0060	2880 ± 48	1
7.1	c/m, osc, p	165	84	0.51	0.03	1.604 ± 0.047	0.2063 ± 0.0033	2877 ± 26	9
8.1	e, osc/rex, p, fr	347	100	0.29	<0.01	1.819 ± 0.079	0.1928 ± 0.0020	2766 ± 17	2
9.1	r, eq, fr	363	88	0.24	<0.01	1.923 ± 0.058	0.1889 ± 0.0036	2733 ± 32	−1
10.1	e, osc, p	92	48	0.52	<0.01	1.668 ± 0.076	0.2115 ± 0.0031	2917 ± 24	4
11.1	e, osc, p, fr	183	117	0.64	<0.01	1.790 ± 0.066	0.2116 ± 0.0020	2918 ± 15	−2
12.1	e, osc, eq	237	45	0.19	<0.01	1.721 ± 0.086	0.1981 ± 0.0017	2810 ± 14	5
13.1	e, osc, eq	282	72	0.26	0.06	1.971 ± 0.082	0.1982 ± 0.0019	2811 ± 16	−6
14.1	e, osc, eq	215	77	0.36	0.08	1.927 ± 0.091	0.2067 ± 0.0029	2880 ± 23	−6
15.1	e, osc/rex, eq	184	23	0.12	0.06	1.905 ± 0.069	0.1897 ± 0.0034	2739 ± 30	−2
16.1	m, osc, eq, fr	242	202	0.83	0.01	1.768 ± 0.089	0.2130 ± 0.0052	2929 ± 40	−1
17.1	e, osc, eq	247	81	0.33	<0.01	1.737 ± 0.075	0.1962 ± 0.0016	2795 ± 13	5
18.1	e, osc, p	155	90	0.58	0.01	1.669 ± 0.051	0.2118 ± 0.0051	2919 ± 40	4
20.1	e, osc, eq, fr	177	70	0.40	0.49	1.563 ± 0.060	0.2103 ± 0.0046	2908 ± 36	10
20.1	e, osc, eq, fr	101	66	0.65	<0.01	1.771 ± 0.070	0.2123 ± 0.0029	2923 ± 22	−2
21.1	m, osc, eq, fr	86	45	0.53	0.12	1.684 ± 0.073	0.2084 ± 0.0038	2893 ± 30	4
22.1	e, osc, eq, fr	67	41	0.62	0.15	1.790 ± 0.068	0.2151 ± 0.0040	2944 ± 30	−2
23.1	rex, fr	311	26	0.08	0.31	1.787 ± 0.068	0.1837 ± 0.0050	2686 ± 46	7
G91/43 metasediment in contact with Færingehavn terrane (Itsaq Gneiss Complex) at Færingehavn									
1.1	c	153	99	0.65	0.49	1.872 ± 0.043	0.1967 ± 0.0028	2799 ± 24	−1
2.1	ov	1487	205	0.14	1.08	4.911 ± 0.100	0.1611 ± 0.0015	2467 ± 16	−52
3.1	c ± r	171	111	0.65	1.13	1.896 ± 0.043	0.1917 ± 0.0030	2757 ± 26	−1
4.1	c	67	39	0.58	0.23	1.888 ± 0.049	0.2015 ± 0.0045	2838 ± 37	−3
4.2	c ± r	76	45	0.60	0.26	1.706 ± 0.063	0.1956 ± 0.0034	2790 ± 29	7
4.3	r	361	4	0.01	0.14	1.899 ± 0.066	0.1815 ± 0.0016	2666 ± 15	2
4.4	r	665	44	0.07	0.13	2.011 ± 0.069	0.1858 ± 0.0009	2706 ± 8	−4
5.1	p	219	177	0.81	1.22	2.051 ± 0.045	0.1900 ± 0.0029	2743 ± 25	−7
6.1	p	112	59	0.53	0.51	1.973 ± 0.047	0.1901 ± 0.0033	2743 ± 29	−4
7.1	c	155	99	0.64	0.47	1.846 ± 0.042	0.1982 ± 0.0028	2811 ± 23	−1
9.1	c	150	93	0.62	0.19	1.809 ± 0.064	0.2008 ± 0.0025	2832 ± 20	0
9.2	r	2099	540	0.26	0.35	6.115 ± 0.208	0.1643 ± 0.0009	2500 ± 9	−61
10.1	p	96	53	0.55	0.03	1.974 ± 0.071	0.1851 ± 0.0020	2699 ± 18	−2
11.1	ov	1560	509	0.33	0.06	4.619 ± 0.157	0.1715 ± 0.0007	2572 ± 7	−51
12.1	c	154	98	0.64	0.26	1.784 ± 0.063	0.1948 ± 0.0021	2783 ± 18	3
13.1	c	100	58	0.59	0.06	1.772 ± 0.063	0.2066 ± 0.0019	2879 ± 15	0
14.1	c	160	79	0.49	0.19	1.759 ± 0.062	0.1987 ± 0.0018	2816 ± 15	3
15.1	c	55	31	0.57	0.47	1.779 ± 0.067	0.2003 ± 0.0055	2828 ± 46	2
16.1	ov	949	18	0.02	1.89	2.903 ± 0.099	0.1770 ± 0.0015	2625 ± 14	−27

Table 1(*Continued*)

Labels	Grain type	U (ppm)	Th (ppm)	Th/U	Comm. 206/Pb%	238U/206/Pb ratio	207Pb/206Pb ratio	207/206 date (Ma)	% Disc
PK87/10 metasediment in contact with Færingehavn terrane (Itsaq Gneiss Complex) on Sagdlerssua sample presented in Friend et al. (1996)									
1.1	c	41	21	0.51	0.39	1.704 ± 0.021	0.1992 ± 0.0024	2820 ± 20	6
2.1	c	76	46	0.60	0.13	1.808 ± 0.017	0.2000 ± 0.0015	2826 ± 12	0
2.2	r	456	10	0.02	0.09	1.898 ± 0.011	0.1879 ± 0.0005	2724 ± 5	0
3.1	c	125	90	0.72	0.07	1.817 ± 0.014	0.2022 ± 0.0011	2844 ± 9	−1
3.2	c	548	10	0.02	0.02	1.891 ± 0.010	0.1910 ± 0.0005	2751 ± 4	−1
4.1	c	300	60	0.20	0.08	1.944 ± 0.012	0.1889 ± 0.0007	2733 ± 6	−2
5.1	c	102	68	0.67	0.09	1.848 ± 0.016	0.2037 ± 0.0014	2856 ± 11	−2
5.2	r	389	4	0.01	0.07	1.884 ± 0.011	0.1844 ± 0.0006	2693 ± 5	2
6.1	c	48	23	0.48	0.31	1.823 ± 0.022	0.1991 ± 0.0022	2819 ± 18	0
7.1	c	126	108	0.86	0.15	1.857 ± 0.015	0.2011 ± 0.0012	2835 ± 10	−2
7.2	c	251	280	1.12	0.05	1.826 ± 0.012	0.2003 ± 0.0008	2828 ± 6	0
8.1	c	85	55	0.64	0.27	1.794 ± 0.017	0.1986 ± 0.0018	2815 ± 15	1
9.1	r	857	48	0.06	0.04	1.956 ± 0.010	0.1871 ± 0.0005	2717 ± 4	−2
10.1	r	1185	13	0.01	0.01	1.931 ± 0.010	0.1872 ± 0.0004	2717 ± 3	−1
11.1	c	150	37	0.25	0.08	1.898 ± 0.014	0.1930 ± 0.0012	2768 ± 10	−1
11.2	r	953	20	0.02	0.01	1.914 ± 0.010	0.1864 ± 0.0004	2711 ± 4	0
12.1	c	165	109	0.66	0.17	1.853 ± 0.013	0.2014 ± 0.0011	2838 ± 9	−2
12.2	r	406	43	0.11	0.04	1.899 ± 0.011	0.1867 ± 0.0006	2713 ± 6	1
13.1	c	50	25	0.50	0.13	1.826 ± 0.019	0.2032 ± 0.0021	2852 ± 17	−1
14.1	c	133	55	0.41	0.20	1.830 ± 0.013	0.1976 ± 0.0012	2806 ± 10	0
14.2	r	780	27	0.04	0.02	1.892 ± 0.010	0.1875 ± 0.0005	2720 ± 4	1
15.1	c	141	106	0.75	0.15	1.843 ± 0.014	0.2013 ± 0.0012	2837 ± 10	−2
18.1	c	129	72	0.56	0.10	1.822 ± 0.014	0.2011 ± 0.0012	2835 ± 10	−1
16.1	c	105	78	0.74	0.16	1.815 ± 0.015	0.2023 ± 0.0015	2845 ± 12	−1
16.2	c	780	16	0.02	0.02	1.895 ± 0.010	0.1893 ± 0.0005	2736 ± 4	0
VM93/1 metasediment in contact with Færingehavn terrane (Itsaq Gneiss Complex) central Ameralik									
2.2	c	203	165	0.82	0.40	1.953 ± 0.046	0.1991 ± 0.0015	2818 ± 12	−5
2.3	c	32	<1	<0.01	1.37	1.941 ± 0.088	0.2013 ± 0.0058	2837 ± 48	−6
3.2	c	78	35	0.45	0.36	1.800 ± 0.102	0.2276 ± 0.0043	3035 ± 31	−6
4.2	c	69	37	0.54	1.07	1.681 ± 0.045	0.2111 ± 0.0047	2914 ± 37	3
4.3	c	194	84	0.43	0.54	1.774 ± 0.070	0.2147 ± 0.0015	2941 ± 12	−2
5.3	r ± c	19	1	0.07	3.93	2.078 ± 0.128	0.1957 ± 0.0095	2791 ± 82	−9
5.4	c	141	93	0.66	0.68	1.876 ± 0.047	0.2010 ± 0.0023	2834 ± 18	−3
9.2	c	429	44	0.10	0.69	1.982 ± 0.114	0.1985 ± 0.0031	2814 ± 25	−7
11.1	r	244	1	<0.01	0.25	2.005 ± 0.076	0.1877 ± 0.0013	2722 ± 12	−4
11.2	c	21	12	0.55	1.75	1.699 ± 0.153	0.2245 ± 0.0065	3013 ± 47	−1
12.1	c	126	104	0.83	0.51	1.940 ± 0.086	0.2027 ± 0.0018	2848 ± 15	−6
12.2	r	35	<1	0.01	1.26	1.941 ± 0.077	0.1913 ± 0.0062	2753 ± 55	−3
13.1	r	32	<1	<0.01	2.21	1.914 ± 0.054	0.1871 ± 0.0061	2717 ± 55	0
14.1	c	54	64	1.19	0.79	1.799 ± 0.054	0.2043 ± 0.0048	2861 ± 39	0
15.1	r	21	<1	<0.01	2.64	1.905 ± 0.084	0.1871 ± 0.0058	2717 ± 52	0
15.2	c	125	110	0.88	0.43	1.814 ± 0.057	0.2035 ± 0.0028	2854 ± 22	−1
15.3	r	18	<1	<0.01	3.05	1.913 ± 0.096	0.1926 ± 0.0069	2764 ± 60	−2
16.1	c	129	43	0.33	0.33	1.796 ± 0.050	0.2093 ± 0.0021	2900 ± 17	−2
17.1	c	60	32	0.53	0.61	1.723 ± 0.047	0.2219 ± 0.0040	2995 ± 29	−1
17.2	r	24	<1	<0.01	3.27	2.021 ± 0.084	0.1834 ± 0.0075	2684 ± 69	−3
LS85G1Si metasediment in contact with Færingehavn terrane (Itsaq Gneiss Complex) outer Ameralik									
1.1	e, osc, p	143	76	0.53	0.15	1.561 ± 0.073	0.2570 ± 0.0019	3229 ± 12	−1
2.1	e, osc, p	256	126	0.49	1.57	2.261 ± 0.256	0.1927 ± 0.0021	2765 ± 18	−15

Table 1(*Continued*)

Labels	Grain type	U (ppm)	Th (ppm)	Th/U	Comm. 206/Pb%	238U/206/Pb ratio	207Pb/206Pb ratio	207/206 date (Ma)	% Disc
3.1	e, osc, p, fr	70	39	0.56	0.31	1.779 ± 0.072	0.2076 ± 0.0025	2887 ± 20	0
4.1	e, osc, p	125	95	0.76	0.23	1.780 ± 0.068	0.2006 ± 0.0015	2831 ± 12	2
5.1	c, osc, p	229	181	0.79	0.34	1.749 ± 0.085	0.2046 ± 0.0011	2863 ± 9	2
6.1	e, osc, p	60	24	0.40	0.25	1.659 ± 0.066	0.2069 ± 0.0026	2881 ± 21	6
7.1	e, osc, stubby	159	87	0.55	0.22	1.731 ± 0.078	0.2047 ± 0.0011	2864 ± 9	3
8.1	c, osc, ov	68	37	0.55	0.43	1.698 ± 0.085	0.2154 ± 0.0017	2946 ± 13	1
8.2	r, h, ov	489	10	0.02	0.06	1.996 ± 0.045	0.1776 ± 0.0018	2630 ± 17	0
8.3	r, h, ov	369	8	0.02	0.40	2.140 ± 0.047	0.1797 ± 0.0020	2650 ± 19	−7
9.1	psc, fr	54	32	0.59	4.70	1.708 ± 0.100	0.2056 ± 0.0073	2871 ± 59	4
10.1	c, osc, fr	216	181	0.84	0.28	1.873 ± 0.108	0.1987 ± 0.0015	2816 ± 12	−2
10.2	r, h, p, fr	511	11	0.02	0.05	1.987 ± 0.050	0.1787 ± 0.0017	2641 ± 15	0
11.1	e, osc, p	474	125	0.26	0.05	1.490 ± 0.060	0.2645 ± 0.0010	3274 ± 6	1
12.1	e, osc, p	336	136	0.40	0.06	1.761 ± 0.075	0.1966 ± 0.0009	2799 ± 7	4
13.1	e, osc, p	131	65	0.50	0.20	1.247 ± 0.068	0.3692 ± 0.0086	3788 ± 36	0
14.1	e, osc, p	132	51	0.39	0.04	1.621 ± 0.074	0.2015 ± 0.0013	2838 ± 11	9
15.1	c, osc, p	61	47	0.78	0.28	1.771 ± 0.094	0.2080 ± 0.0024	2890 ± 19	0
15.2	r, h, p	555	12	0.02	0.01	1.933 ± 0.042	0.1775 ± 0.0008	2630 ± 8	2
16.1	c, osc, p, fr	109	41	0.38	0.24	1.807 ± 0.079	0.1998 ± 0.0016	2825 ± 13	1
16.2	r, h, p, fr	583	13	0.02	0.07	2.031 ± 0.053	0.1792 ± 0.0013	2645 ± 12	−2
17.1	c, osc, p, fr	190	156	0.82	0.18	1.587 ± 0.245	0.2200 ± 0.0134	2981 ± 101	6
17.2	r, h, p, fr	546	4	0.01	0.10	2.134 ± 0.054	0.1675 ± 0.0013	2533 ± 13	−2
18.1	m, osc, p, fr	343	220	0.64	<0.01	1.815 ± 0.034	0.2000 ± 0.0009	2826 ± 8	0
19.1	e, osc, p, fr	166	181	1.10	0.21	1.745 ± 0.040	0.2027 ± 0.0019	2848 ± 15	3
20.1	e, osc/h, p	80	62	0.78	0.33	1.905 ± 0.073	0.1930 ± 0.0028	2768 ± 24	−2
20.2	m, osc, p	52	48	0.92	0.11	1.851 ± 0.054	0.1986 ± 0.0042	2815 ± 35	−1
21.1	e, osc, p, fr	125	87	0.70	<0.01	1.511 ± 0.039	0.2593 ± 0.0032	3243 ± 20	1
22.1	e, osc/h, p	58	47	0.82	0.03	1.559 ± 0.053	0.2562 ± 0.0037	3223 ± 23	−1
23.1	e, osc, p, fr	106	80	0.76	0.55	1.802 ± 0.063	0.2024 ± 0.0041	2846 ± 34	0
24.1	m/c, osc, p, fr	114	54	0.48	0.13	1.808 ± 0.047	0.2059 ± 0.0018	2873 ± 14	−1
25.1	m, osc, ov, fr	72	50	0.70	<0.01	1.755 ± 0.044	0.2037 ± 0.0028	2856 ± 23	2
26.1	e, osc, p	135	80	0.59	0.38	1.686 ± 0.043	0.2237 ± 0.0025	3008 ± 18	0
27.1	m, osc, p	178	188	1.06	0.25	1.832 ± 0.043	0.1919 ± 0.0024	2758 ± 20	2
28.1	c, osc, ov	92	39	0.42	0.08	1.553 ± 0.040	0.2570 ± 0.0026	3229 ± 16	−1
27.2	m, osc, p	337	155	0.46	0.99	2.087 ± 0.053	0.1995 ± 0.0039	2822 ± 32	−11
29.1	m/c, osc, p, fr	241	205	0.85	0.13	1.706 ± 0.037	0.2174 ± 0.0016	2961 ± 12	0
200892 Felsic volcanic in Ivisaartoq supracrustal belt within the Kapisilik terrane (Friend and Nutman, in press)									
1.1	e, osc, fr	16	193	12	0.7040	1.638 ± 0.073	0.2304 ± 0.0047	3055 ± 33	1
2.1	e, osc, fr	13	184	14	0.9360	1.748 ± 0.156	0.2320 ± 0.0069	3066 ± 48	−5
3.1	m/c, osc, eq	11	112	10	1.2340	1.703 ± 0.092	0.2290 ± 0.0075	3045 ± 54	−2
4.1	e, osc, p, fr	9	94	10	0.5610	1.632 ± 0.093	0.2368 ± 0.0071	3099 ± 49	0
5.1	m/c, osc, eq	13	164	12	0.9170	1.651 ± 0.067	0.2356 ± 0.0051	3090 ± 35	−1
6.1	m, osc, fr	13	138	10	1.3990	1.735 ± 0.088	0.2302 ± 0.0069	3053 ± 49	−4
7.1	e, osc, eq, fr	17	126	7	0.3280	1.656 ± 0.071	0.2313 ± 0.0045	3061 ± 31	0
8.1	m, osc, eq	16	140	9	0.4740	1.650 ± 0.092	0.2304 ± 0.0048	3055 ± 34	0
9.1	m, osc, eq, fr	15	154	11	1.3430	1.687 ± 0.085	0.2297 ± 0.0075	3050 ± 54	−2
10.1	m/c, osc, eq	15	136	9	0.7280	1.706 ± 0.098	0.2328 ± 0.0043	3072 ± 30	−3
11.1	e, osc, p/eq	11	62	6	0.5260	1.681 ± 0.076	0.2361 ± 0.0075	3094 ± 52	−3
12.1	e, osc, p, fr	21	174	8	0.4710	1.672 ± 0.086	0.2348 ± 0.0046	3085 ± 32	−2
13.1	e, osc, p, fr	15	177	12	0.8690	1.601 ± 0.080	0.2257 ± 0.0075	3022 ± 54	4
14.1	e, osc, p, fr	41	145	4	0.2980	1.675 ± 0.058	0.2322 ± 0.0033	3068 ± 23	−2
15.1	e, osc, p, fr	30	496	16	0.3810	1.661 ± 0.071	0.2354 ± 0.0043	3089 ± 29	−2

Table 1(*Continued*)

Labels	Grain type	U (ppm)	Th (ppm)	Th/U	Comm. 206/Pb%	238U/206/Pb ratio	207Pb/206Pb ratio	207/206 date (Ma)	% Disc
16.1	e, osc, p, fr	14	151	11	1.0660	1.682 ± 0.104	0.2310 ± 0.0051	3059 ± 36	−2
17.1	e, osc, p, fr	16	171	11	0.7800	1.651 ± 0.082	0.2346 ± 0.0047	3083 ± 33	−1
18.1	e, osc, p, fr	16	144	9	0.4500	1.625 ± 0.072	0.2437 ± 0.0041	3144 ± 27	−2
19.1	m, osc, ov	21	563	27	0.3510	1.645 ± 0.058	0.2320 ± 0.0034	3066 ± 23	0
20.1	e, osc, p	13	237	18	0.4200	1.592 ± 0.067	0.2262 ± 0.0056	3026 ± 40	4
21.1	e, osc, p	8	70	8	0.7310	1.718 ± 0.102	0.2369 ± 0.0068	3100 ± 47	−5
22.1	e, osc, p	8	54	7	0.5770	1.677 ± 0.085	0.2416 ± 0.0144	3131 ± 98	−4
23.1	e, osc, p, fr	19	255	13	0.4670	1.683 ± 0.077	0.2335 ± 0.0066	3076 ± 46	−2
G93/086 metasediment in contact with Isukasia terrane (Itsaq Gneiss Complex)									
1.1	c, osc	147	177	1.20	0.23	1.365 ± 0.066	0.3232 ± 0.0051	3585 ± 25	−1
2.1	c, osc	153	139	0.90	0.08	1.373 ± 0.077	0.3639 ± 0.0049	3766 ± 21	−6
3.1	r, hd	762	117	0.15	0.07	1.725 ± 0.053	0.2142 ± 0.0014	2938 ± 11	0
4.1	osc, p	1018	375	0.37	0.05	1.570 ± 0.072	0.2574 ± 0.0031	3231 ± 19	−2
5.1	osc, p	638	143	0.22	0.07	1.528 ± 0.075	0.2719 ± 0.0023	3317 ± 13	−2
6.1	c, osc	814	86	0.11	0.10	1.685 ± 0.064	0.2304 ± 0.0107	3055 ± 77	−2
7.1	m, osc, p, fr	641	158	0.25	0.08	1.525 ± 0.090	0.2695 ± 0.0042	3303 ± 25	−2
8.1	m, hd, ov	678	65	0.10	0.04	1.777 ± 0.081	0.2199 ± 0.0026	2980 ± 19	−3
9.1	c, osc, ov	289	79	0.27	0.08	1.540 ± 0.080	0.2722 ± 0.0032	3319 ± 19	−3
10.1	m, osc, p, fr	791	188	0.24	0.10	1.727 ± 0.067	0.2340 ± 0.0033	3080 ± 22	−4
11.1	e, osc, p	1399	164	0.12	0.04	1.862 ± 0.065	0.2024 ± 0.0016	2846 ± 13	−3
11.2	e, osc, p	443	264	0.60	0.06	1.780 ± 0.132	0.2309 ± 0.0060	3058 ± 42	−6
12.1	m, osc, p	627	131	0.21	0.06	1.603 ± 0.137	0.2493 ± 0.0050	3180 ± 32	−2
13.1	m, osc, p	262	61	0.23	0.34	1.629 ± 0.069	0.2945 ± 0.0056	3442 ± 30	−10
14.1	e, osc, p	703	31	0.04	0.02	1.735 ± 0.053	0.2293 ± 0.0006	3047 ± 4	−4
15.1	e, osc, p	1109	175	0.16	0.05	1.732 ± 0.086	0.2309 ± 0.0034	3058 ± 24	−4
16.1	m, osc, ov	255	83	0.33	0.17	1.422 ± 0.083	0.3319 ± 0.0052	3626 ± 24	−5
17.1	osc/hd, ov	340	203	0.60	0.08	1.526 ± 0.059	0.2928 ± 0.0015	3433 ± 8	−5
18.1	osc/hd, p	1486	188	0.13	0.05	1.750 ± 0.065	0.2111 ± 0.0017	2914 ± 13	0
19.1	hd, p	1533	217	0.14	0.06	1.878 ± 0.265	0.2105 ± 0.0077	2910 ± 60	−5
19.2	hd, p	1336	204	0.15	0.02	1.729 ± 0.078	0.2085 ± 0.0022	2894 ± 17	2
20.1	m, osc, p	207	115	0.55	0.22	1.596 ± 0.068	0.2949 ± 0.0043	3444 ± 23	−9
21.1	m, osc, p	277	42	0.15	0.10	1.579 ± 0.070	0.2875 ± 0.0069	3404 ± 38	−7
22.1	m, osc, p	679	192	0.28	0.13	1.412 ± 0.116	0.2464 ± 0.0031	3162 ± 20	9
23.1	m, hb, ov	121	37	0.31	0.47	1.662 ± 0.092	0.2385 ± 0.0069	3110 ± 47	−2
24.1	m, osc, ov	200	78	0.39	0.14	1.502 ± 0.123	0.2882 ± 0.0074	3408 ± 40	−3
25.1	m, osc, p, fr	275	96	0.35	0.09	1.319 ± 0.278	0.2499 ± 0.0034	3184 ± 21	14
26.1	e, osc, ov	781	145	0.19	0.03	1.665 ± 0.076	0.2329 ± 0.0024	3072 ± 17	−1
27.1	c, osc, ov	467	63	0.13	0.13	1.838 ± 0.132	0.2175 ± 0.0024	2963 ± 18	−5
28.1	m, osc, ov	405	62	0.15	0.09	1.671 ± 0.267	0.2282 ± 0.0124	3039 ± 90	0
29.1	r, hd, ov	808	104	0.13	0.03	1.602 ± 0.113	0.2136 ± 0.0027	2933 ± 21	7
30.1	r, hd, ov	805	101	0.13	0.02	1.723 ± 0.065	0.2138 ± 0.0011	2934 ± 8	1
G94/02 metasediment at probable northern edge of Akia terrane from Garde et al. (2001)									
1.1	ov	2271	15	0.01	0.02	2.046 ± 0.040	0.1640 ± 0.0008	2497 ± 9	3
2.1	ov	1141	10	0.01	0.41	2.065 ± 0.041	0.1687 ± 0.0006	2544 ± 6	0
3.1	ov	3348	52	0.02	0.01	2.072 ± 0.040	0.1679 ± 0.0005	2537 ± 5	0
4.1	ov	2511	62	0.02	0.01	2.107 ± 0.043	0.1654 ± 0.0003	2511 ± 3	0
5.1	ov	1550	16	0.01	0.02	2.068 ± 0.057	0.1674 ± 0.0016	2532 ± 17	0
6.1	c	65	36	0.55	0.53	1.450 ± 0.041	0.2534 ± 0.0040	3206 ± 25	6
7.1	c	899	511	0.57	0.03	1.764 ± 0.038	0.2131 ± 0.0012	2930 ± 9	−1
8.1	ov	4479	25	0.01	<0.01	2.055 ± 0.041	0.1703 ± 0.0007	2560 ± 7	0

Table 1(*Continued*)

Labels	Grain type	U (ppm)	Th (ppm)	Th/U	Comm. 206/Pb%	238U/206/Pb ratio	207Pb/206Pb ratio	207/206 date (Ma)	% Disc
9.1	ov	662	10	0.01	0.08	2.120 ± 0.044	0.1631 ± 0.0041	2489 ± 43	0
9.2	ov	7138	61	0.01	0.01	2.113 ± 0.040	0.1660 ± 0.0011	2518 ± 11	−1
10.1	ov	1258	17	0.01	0.04	2.151 ± 0.044	0.1690 ± 0.0005	2548 ± 5	−3
11.1	ov	1036	24	0.02	0.09	2.137 ± 0.043	0.1662 ± 0.0006	2519 ± 6	−2
12.1	c	171	68	0.40	0.59	1.617 ± 0.051	0.2474 ± 0.0024	3168 ± 15	−2
13.1	c	179	131	0.73	0.15	1.600 ± 0.037	0.2492 ± 0.0036	3180 ± 23	−2
14.1	ov	785	13	0.02	0.06	2.150 ± 0.034	0.1680 ± 0.0028	2538 ± 29	−3
15.1	c	102	41	0.40	0.67	1.632 ± 0.054	0.2425 ± 0.0052	3137 ± 35	−2
16.1	c	246	154	0.63	0.45	2.041 ± 0.052	0.2324 ± 0.0015	3069 ± 10	−16
17.1	ov	734	42	0.06	0.08	1.895 ± 0.080	0.1838 ± 0.0028	2687 ± 25	2
18.1	c	114	66	0.58	0.65	1.648 ± 0.044	0.2479 ± 0.0025	3172 ± 16	−4
19.1	c	141	32	0.23	0.22	1.695 ± 0.044	0.2431 ± 0.0032	3140 ± 21	−5
VM97/12 metasediment in Tuno terrane (?composite), north of Akia terrane, Hamborgerland									
1.1	e, osc, p	306	159	0.52	0.07	1.868 ± 0.057	0.1944 ± 0.0015	2780 ± 13	−1
2.1	e, osc, p	144	75	0.52	0.02	1.672 ± 0.064	0.2218 ± 0.0022	2994 ± 16	1
3.1	m, osc, p	231	183	0.79	0.14	1.643 ± 0.045	0.2176 ± 0.0010	2963 ± 7	3
4.1	e, osc, ov	229	199	0.87	0.06	1.763 ± 0.061	0.2056 ± 0.0009	2871 ± 7	1
5.1	c, osc, ov	119	48	0.40	0.02	1.658 ± 0.052	0.2203 ± 0.0020	2983 ± 15	2
6.1	e, osc, p	197	106	0.54	0.05	1.729 ± 0.064	0.2101 ± 0.0011	2907 ± 9	1
7.1	m, osc, p	205	114	0.56	0.04	1.694 ± 0.063	0.2210 ± 0.0009	2988 ± 6	0
8.1	e, osc, p	132	88	0.66	0.20	1.704 ± 0.061	0.2167 ± 0.0024	2956 ± 18	1
9.1	e, osc, p	242	127	0.53	0.07	1.663 ± 0.047	0.2193 ± 0.0033	2976 ± 24	2
10.1	e, osc, p, fr	575	399	0.69	0.14	1.818 ± 0.054	0.1985 ± 0.0008	2814 ± 7	0
11.1	e, osc, p	412	679	1.65	0.06	1.771 ± 0.055	0.2022 ± 0.0017	2844 ± 14	2
12.1	e, osc, p	143	200	1.39	0.03	1.733 ± 0.070	0.2085 ± 0.0019	2894 ± 15	2
13.1	m, osc, ov	194	73	0.38	0.05	1.696 ± 0.073	0.2239 ± 0.0104	3009 ± 77	−1
14.1	e, osc, p, fr	181	83	0.46	0.09	1.747 ± 0.053	0.2038 ± 0.0014	2857 ± 11	2
15.1	m, osc, p	215	87	0.40	0.13	1.822 ± 0.057	0.2219 ± 0.0035	2994 ± 25	−6
16.1	e, osc, p	257	134	0.52	0.05	1.795 ± 0.037	0.2056 ± 0.0017	2871 ± 14	−1
17.1	m, osc/h, ov	127	95	0.75	0.05	1.687 ± 0.066	0.2221 ± 0.0013	2996 ± 9	0
18.1	m, osc/h, ov	313	296	0.95	0.01	1.637 ± 0.048	0.2256 ± 0.0020	3021 ± 14	2
19.1	e, hb, p, fr	81	49	0.61	<0.01	1.702 ± 0.056	0.2312 ± 0.0073	3060 ± 52	−3
20.1	e, osc, p	411	251	0.61	0.01	1.804 ± 0.044	0.2067 ± 0.0015	2880 ± 12	−1
21.1	m, osc, p, fr	141	57	0.40	0.01	1.833 ± 0.051	0.2260 ± 0.0036	3024 ± 26	−7
22.1	m, osc, p	126	62	0.49	0.02	1.701 ± 0.043	0.2262 ± 0.0012	3025 ± 8	−1
23.1	m, osc, p	306	172	0.56	0.04	1.758 ± 0.065	0.2217 ± 0.0014	2993 ± 10	−3
24.1	m, osc, p	238	190	0.80	0.11	1.868 ± 0.066	0.2048 ± 0.0015	2865 ± 12	−3
25.1	m, osc, fr	176	71	0.40	0.10	1.801 ± 0.045	0.2218 ± 0.0018	2994 ± 13	−5
26.1	e, osc, p	111	46	0.41	0.06	1.928 ± 0.049	0.1889 ± 0.0012	2732 ± 11	−1
27.1	e, osc, p	199	91	0.46	<0.01	1.790 ± 0.040	0.2029 ± 0.0032	2849 ± 26	0
28.1	m, osc, p	164	69	0.42	0.05	1.901 ± 0.061	0.2042 ± 0.0034	2860 ± 27	−5
29.1	e, osc, fr	627	30	0.05	0.12	2.051 ± 0.046	0.1947 ± 0.0011	2782 ± 9	−2
30.1	m, osc, ov	422	369	0.87	0.04	1.767 ± 0.044	0.2171 ± 0.0014	2959 ± 10	−2
31.1	e, osc, p	395	335	0.85	0.06	1.826 ± 0.038	0.2080 ± 0.0006	2890 ± 5	−3
32.1	e, osc, p	391	170	0.43	0.09	1.821 ± 0.039	0.2189 ± 0.0018	2973 ± 13	−5
33.1	e, osc, p	112	47	0.42	0.25	1.827 ± 0.047	0.2183 ± 0.0017	2968 ± 13	−5
34.1	e, osc, p	286	114	0.40	0.07	1.749 ± 0.060	0.2225 ± 0.0030	2999 ± 22	−3
35.1	m, osc, p	386	169	0.44	0.02	1.802 ± 0.039	0.2158 ± 0.0014	2950 ± 11	−3
36.1	e, osc, p	194	96	0.50	0.07	1.771 ± 0.042	0.2214 ± 0.0022	2991 ± 16	−3
37.1	m, osc, p	496	544	1.10	0.04	1.832 ± 0.037	0.2057 ± 0.0011	2872 ± 9	−2
38.1	e, osc, p	796	1257	1.58	0.13	1.853 ± 0.073	0.1985 ± 0.0010	2813 ± 9	−1

Table 1(*Continued*)

Labels	Grain type	U (ppm)	Th (ppm)	Th/U	Comm. 206/Pb%	238U/206/Pb ratio	207Pb/206Pb ratio	207/206 date (Ma)	% Disc
39.1	m, osc, p	152	58	0.38	0.22	1.830 ± 0.042	0.2153 ± 0.0027	2945 ± 20	−5
40.1	e, osc, p	225	36	0.16	0.07	1.897 ± 0.049	0.1981 ± 0.0017	2810 ± 14	−3
41.1	e, osc, p	473	205	0.43	0.07	1.807 ± 0.038	0.2242 ± 0.0015	3011 ± 11	−6
42.1	e, osc, p, fr	171	72	0.42	0.04	1.782 ± 0.045	0.2201 ± 0.0020	2981 ± 15	−4
43.1	m, hd, p	629	135	0.22	0.06	1.945 ± 0.041	0.1877 ± 0.0010	2722 ± 9	−2
44.1	e, osc, p	415	87	0.21	0.05	1.885 ± 0.044	0.2023 ± 0.0016	2845 ± 13	−4
45.1	e, osc, p	260	217	0.83	0.02	1.790 ± 0.041	0.2184 ± 0.0009	2969 ± 6	−4
46.1	e, osc, p	286	104	0.36	0.05	1.818 ± 0.077	0.2152 ± 0.0028	2945 ± 21	−4
47.1	e, osc, p	276	307	1.11	0.04	1.849 ± 0.051	0.1995 ± 0.0010	2822 ± 8	−1
48.1	e, osc, p	571	237	0.42	0.07	1.869 ± 0.050	0.1976 ± 0.0012	2807 ± 10	−2
49.1	m, osc, ov	306	126	0.41	0.07	1.678 ± 0.039	0.2196 ± 0.0027	2977 ± 20	1
50.1	m, osc, ov	323	128	0.40	0.21	1.754 ± 0.038	0.2210 ± 0.0019	2988 ± 14	−3

p: prismatic, ov: equant and ovoid, e: end, m: middle, r: overgrowth, c: core, rex: recrystallised, osc: oscillatory finescale zoning, s: sector zoning, h: homogeneous, f: fragment, anh: anhedral, turb: turbid *corrected with*3600 Ma *model Pb of Cumming and Richards, 1975 of likely igneous age*.

date of 2767 ± 5 Ma (MSWD = 1.44). Two other overgrowths (including analysis 18.2 over core 18.1) gave a weighted mean ^{207}Pb/^{206}Pb date of 2698 ± 16 Ma (MSWD = 0.85). The 2767 ± 5 Ma and 2698 ± 16 Ma dates are interpreted as recording thermal events affecting Palaeoarchaean orthogneisses.

The SHRIMP U/Pb zircon results reported here, together with the early Pb whole rock isotopic work of P.N. Taylor, are clear demonstrations that the Qarliit tasersuat assemblage contains Palaeoarchaean rocks. However, they have been strongly overprinted by later Archaean tectonothermal events under amphibolite facies, to a degree similar to that shown by the Færingehavn terrane in the Nuuk region. Study of the Qarliit tasersuat assemblage is clearly at a reconnaissance stage, and with the results so far indicating it is worthy of further investigation.

4. Ages of detrital zircons in Meso-Neoarchaean metasediments

The initial purpose of dating zircons from metasediments throughout the Archaean craton of West Greenland was to look for abundant very old (>3800 Ma) detrital zircons to form another resource for study of the earliest Earth. In fact these metasediments have very little Palaeoarchaean detritus, but their detrital zircon age distributions provide a useful insight into Meso-Neoarchaean tectonic evolution of the craton.

The first SHRIMP U/Pb zircon provenance study of Neoarchaean metasediments (then assigned to the Malene supracrustals) from the Nuuk region was by Schiøtte et al. (1988). Their study of two samples from islands near the mouth of Ameralik (fjord) near Nuuk demonstrated that the zircons were sourced from rocks of several ages from the Palaeoarchaean (very rare grains) to ca. 2840 Ma. These grains showed evidence of recrystallisation and overgrowth development in thermal events from ca. 2700 to 2550 Ma. This early study is considerably expanded upon here by looking at the provenance of eleven Neoarchaean metasedimentary rocks from across the craton (Fig. 1). The purpose of this is two-fold. First, to search for any metasedimentary units with a large proportion of Palaeoarchaean zircons, which could be a useful geochemical resource for early Earth studies. Secondly, by looking at the age of provenance within and across several tectonostratigraphic terranes, to gain further understanding of Neoarchaean terrane assembly. For the latter, different structural setting for the chosen samples is noted. Some are from supracrustal units along the tectonic boundaries *between* terranes, whereas others are from supracrustal units *within* single terranes. Most of the analyses yielded close to concordant dates (Table 1). However, some sites were markedly discordant and/or had appreciable common Pb content (assessed on the basis of measured ^{204}Pb). Thus, in the account of detrital zircon age provenance below (and in Fig. 5), the data has been filtered to remove duplicate analyses of the

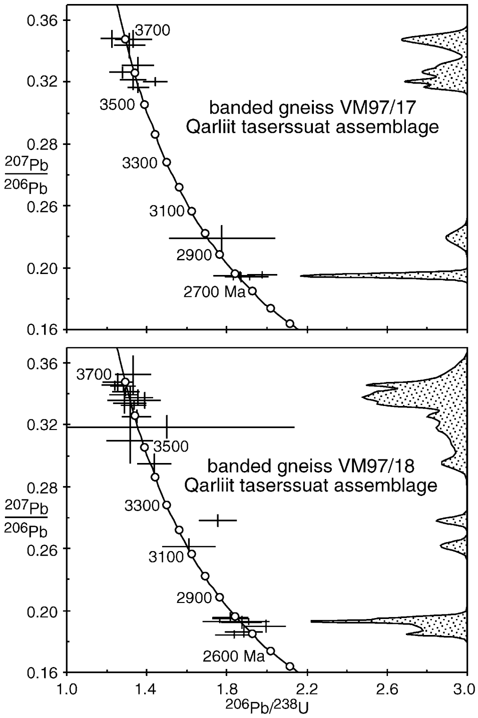

Fig. 4. $^{238}U/^{206}Pb$ vs. $^{207}Pb/^{206}Pb$ "Tera-Wasserburg" concordia plot for zircon analyses of orthogneisses VM97/17 and VM97/18 from the Qarliit taserssuat assemblage. Errors are depicted at the 1σ level and have been corrected for (minor) common Pb based on measured $^{204}Pb/^{206}Pb$. Insets are cumulative probability plots of $^{207}Pb/^{206}Pb$ dates in each sample.

same grains, sites >10% normal discordant and >5% reverse discordant, and with high common Pb (comm. $^{206}Pb > 2\%$). Also only undoubted detrital zircons are plotted in Fig. 5. All doubted detrital zircons plus zircon clearly crystallised in situ during metamorphism are not plotted. The full, unfiltered data set is given in Table 1. Table 2 summarises $^{207}Pb/^{206}Pb$ dating of zircons from these sediments. For polymodal populations, age components were extracted using mixture modelling (Sambridge and Compston, 1994). For unimodal populations, a weighted mean date is presented of the (volcanic) detritus. The column "youngest detrital component" gives the age of the group of zircons considered to be the youngest volcanic and/or sedimentary component. For samples with polymodal age spectra, the next five columns list in decreasing significance the age components extracted by mixture modelling. The final column signifies whether Palaeoarchaean grains were detected or not.

4.1. Tartoq Group – at a terrane boundary at the south of the Neria block?

The Tartoq Group occurs in the southern end of the Archaean craton (Fig. 1). It is a multiply folded belt over 40 km long and up to 5 km wide consisting of upper greenschist to lower amphibolite facies mafic rocks (some with relict pillow structures) and quartzofeldspathic schists derived from volcanic and/or sedimentary rocks (Higgins and Bondesen, 1966; Higgins, 1968; Berthelsen and Henriksen, 1975; Evans and King, 1993). Gneisses north of the Tartoq Group in the Neria and Paamiut blocks have yielded SHRIMP U/Pb zircon dates between 2940 and 2840 Ma (Nutman and Kalsbeek, 1994; Friend and Nutman, 2001). The gneisses south of the Tartoq Group are assigned as the *Sermiligaarsuk block* (Friend and Nutman, 2001) and have yielded SHRIMP U/Pb zircon ages up to 3000 Ma (Nutman and McGregor, unpublished SHRIMP U/Pb zircon data). Gneisses of this age are not encountered to the north until the Nuuk region. Thus it is possible that the Tartoq Group coincides with a terrane boundary.

A discordant gneiss sheet of tonalitic composition intruding amphibolites in the northern part of the belt yielded a SHRIMP U/Pb zircon date of 2944 ± 7 Ma (Nutman and Kalsbeek, 1994). This provides a minimum age for part of the Tartoq Group. SHRIMP U/Pb zircon dating of quartz-sericite schist sample VM95/21 from the southern part of the belt (at 61°27.2′N, 48°47.4′W) is reported here. The sample is from a 20–50 m thick unit (from the middle of the outcrop in Fig. 25 of Berthelsen and Henriksen, 1975). Despite deformation, the schists preserve possible relict volcano-sedimentary structures. The sample yielded abundant prismatic subhedral to rounded zircons. Detrital/volcano-sedimentary components in these zircons display a polymodal age distribution (Fig. 5). The youngest group, with oscillatory zoning parallel to the exterior of prismatic grains, is 2842 ± 6 Ma (2σ) by mixture modelling (Fig. 5). Other grains give dates back to almost 3800 Ma (Tables 1 and 2) with 2900–2940 Ma grains most important – these correspond with dates obtained on orthogneisses to the north of the Tartoq Group (Nutman and Kalsbeek, 1994; Friend and Nutman, 2001). For grains with ages between 3050 and 3200 Ma a source is not known in nearby orthogneisses. Palaeoarchaean zircons occurring in the sample appear to be relatively rare, forming <5% of the population. Presently a local source for these old zircons in the adjacent orthogneisses has not yet been detected. The youngest oscillatory-zoned zircons in VM97/21 with a date of 2842 ± 6 Ma are interpreted as giving the time of deposition of the volcano-sedimentary unit. If this interpretation is correct, then the Tartoq Group must contain unrelated volcano-sedimentary sequences of different ages, because amphibolites within the northern part of the belt are cut by a 2944 ± 7 Ma tonalitic sheet (Nutman and Kalsbeek, 1994), and thus are ca. 100 million years older.

4.2. Metasediments between the Neria and Sermiligaarsuk blocks

On the northeast side of Neria (fjord), a strongly deformed unit of amphibolites and mica schists separates the Neria block (affected by granulite facies metamorphism) from the Sermiligaarsuk block (only ever affected by amphibolite facies metamorphism). Thus this amphibolite and mica schist unit is interpreted to coincide with a tectonic boundary. The dated mica schist (from 61°40.1′N, 48°37.2′W) yielded abundant prismatic to rounded generally small (<100 μm) zircons. In CL images many zircons show vestiges of oscillatory zoning which may be parallel to grain margins or strongly truncated. Dull, homogeneous zircon forms overgrowths on some grains and forms all of some equant grains. Detrital zircons give a polymodal age pattern with mixture modelling components ranging from 2857 ± 12 (2σ) Ma to Palaeoarchaean (only one detected), with components at 2857 ± 12 and 2987 ± 12 Ma most abundant (Tables 1 and 2, Fig. 5). A 2857 ± 12 Ma is interpreted as the maximum deposition age of the metasediment. The dull homogeneous in CL images overgrowths and equant grains have higher U, lower Th/U and give younger dates of ca. 2750 and 2650 Ma. These are interpreted to have grown during metamorphisms affecting the metasediment.

4.3. Ravns Storø belt within the Tasiusarsuaq terrane

The Ravns Storø belt within the southern part of the Tasiusarsuaq terrane was mapped and described by Andersen and Friend (1973). The belt is ca. 30 km long and up to 5 km wide, and consists of multiply deformed amphibolite facies schists and gneisses, affected by amphibolite facies metamorphism. Most of the rocks are amphibolites, which in lower strain domains preserve relict pillow and pillow breccia structures (Andersen and Friend, 1973). Less common in the belt are units of biotite ± garnet quartzo-feldspathic gneisses, which have been interpreted as being derived from sediments or felsic volcanic rocks (Andersen and Friend, 1973). SHRIMP U/Pb dating of a discordant tonalite gneiss sheet (sample VM90/4) cutting amphibolites at the southern end of the belt gave a date of 2878 ± 10 Ma (Friend and Nutman, 2001), which gives a minimum depositional age for at least parts of the belt. Zircons from biotite ± garnet gneiss GGU129929 (62°41′N, 50°12.5′W) were dated to use the Ravns Storø belt to examine provenance within the Tasiusarsuaq terrane. Zircons are mostly subhedral prisms with widely preserved oscillatory zoning parallel to the grain edges. This is disturbed by some recrystallisation domains, but later metamorphic overgrowths are rare or absent. Analyses on best preserved oscillatory-zoned zircon with low U abundance yielded mostly concordant dates with a weighted mean $^{207}Pb/^{206}Pb$ date of 2908 ± 13 Ma (MSWD = 0.52). Other sites, with generally higher U, lower Th/U, and located on the edges of grains yielded a spread of $^{207}Pb/^{206}Pb$ dates from ca. 2820 to 2690 Ma. These are interpreted as domains of disturbance and recrystallisation of a unimodal 2908 ± 13 Ma zircon population. The unimodal age distribution of least modified zircon in this sample combined with the unrounded nature of the grains is interpreted as giving the age of deposition of a volcano-sedimentary protolith of gneiss GGU129929, and to indicate it was derived solely from locally sourced volcanic debris. An age of deposition of 2908 Ma is within the range of TTG protolith ages in the Tasiusarsuaq terrane of ca. 2840–2920 Ma.

4.4. Metasediments in contact with the Færingehavn terrane

Units of quartz + plagioclase + biotite ± garnet ± cordierite ± sillimanite paragneisses with associated banded amphibolites (formerly assigned to the Malene supracrustals – McGregor, 1973) occur in con-

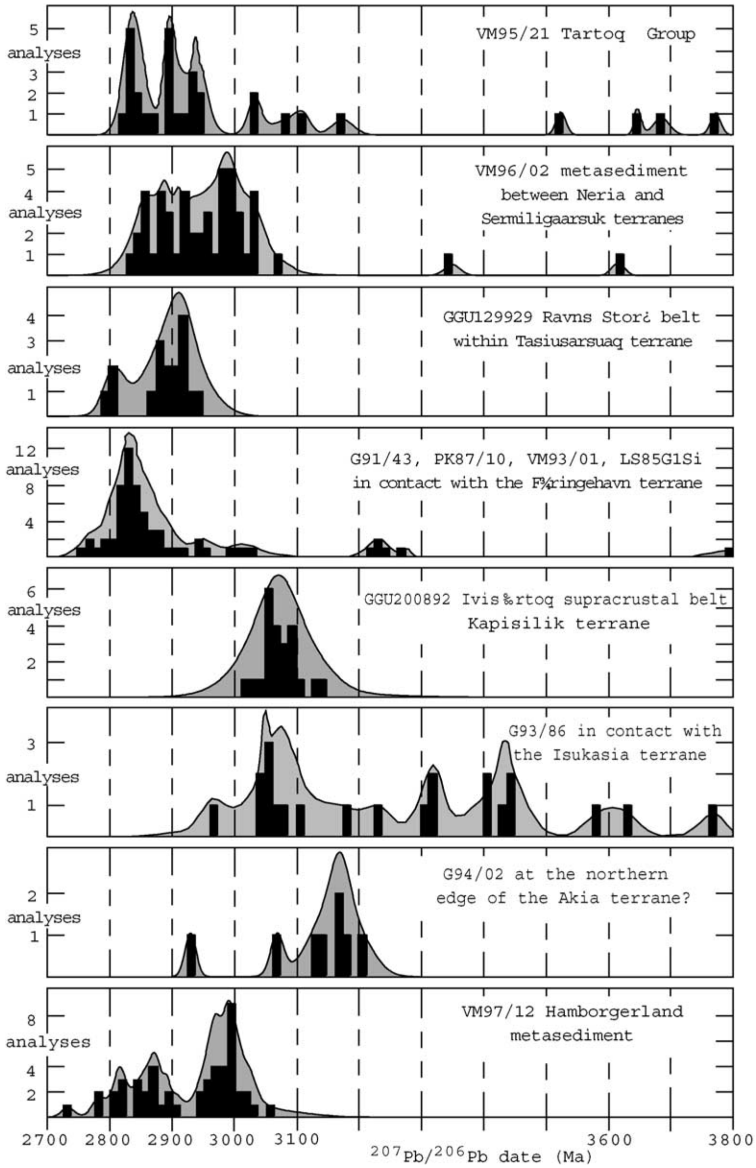

Fig. 5. Cumulative probability (background) and histogram plots for detrital zircon dates for metasediments across the Archaean Craton. The complete data from Table 1 has been filtered to produce this figure, using criteria described in Section 4 of the text.

tact with the Palaeoarchaean rocks of the Færingehavn terrane. All contacts between the paragneiss and amphibolite units and the Færingehavn terrane are now tectonic (Nutman et al., 1989). The chemistry of the paragneisses (e.g. elevated Nb and Zr abundances) suggests their protoliths were dominated by sediments derived from weathering of igneous material with A-type chemistry (Dymek and Smith, 1990). Zircon dating results from four paragneisses up to 50 km distant from each other (Fig. 1; samples G93/43 – 63°41.92′N, 51°32.42′W; PK87/10 – exact location not known – from east side of the island of Sagdlerssua, VM93/1 – 64°13.08′N, 50°41.07′W and LS85G1Si – 64°3.33′N, 51°36.33′W – Table 1) are discussed together here because they give broadly similar results. Sample LS85G1Si is a re-analysis of one of the paragneisses originally dated by Schiøtte et al. (1988) – prior to the benefits of CL imaging for guiding analyses. New dating on this sample has been guided by CL imaging, and was referenced to Temora 1, one of the U/Pb calibration standards currently used by the ANU SHRIMP laboratory. This has resulted in less "smear" in the age spectra, plus generally concordant age determinations. Results for metasediment PK87/10 are from Friend et al. (1996).

The zircons are variable in character. Some are clearly rounded, with dramatic truncation of their internal oscillatory zoning. However, many zircons are only slightly rounded prisms with oscillatory zoning parallel to the crystal faces. Some grains have developed metamorphic overgrowths and/or recrystallisation domains. In all four samples, the better preserved, non-rounded, oscillatory-zoned zircon coincides in age, with the mixture modelling component age (dominant) of 2831 ± 6 Ma (Tables 1 and 2, Fig. 5). From their age consistency and grain morphology these are interpreted as giving the age of a volcanic component within these metasediments. We interpret this as the age of the A-type igneous component in these metasediments proposed by Dymek and Smith (1990). The pre-2831 Ma zircons form a polymodal age distribution. Those up to ca. 3230 Ma could be sourced from the nearby Tasiusarsuaq and Akia terranes. Despite these metasediments being in contact with the Palaeoarchaean Færingehavn terrane, only rare Palaeoarchaean detrital grains were detected in them (see Schiøtte et al., 1988 and grain 13 in our analysis of LS85G1Si with a $^{207}Pb/^{206}Pb$ date of 3788 Ma – Table 1).

4.5. *Ivisaartoq supracrustal belt – within the Kapisilik terrane of the Nuuk region*

Fieldwork and SHRIMP U/Pb zircon dating in the northeast of the Nuuk region led Friend and Nutman (in press-a) to propose that the region contains additional tectonostratigraphic terranes to the four originally recognised in the 1980s (Friend et al., 1988). The *Kapisilik terrane* contains supracrustal rocks dominated by mafic and ultramafic rocks derived from pillow lavas within more voluminous orthogneisses with ages between ca. 3070 and 2970 Ma. These supracrustal rocks are best developed in the *Ivisaartoq*

Table 2

Sample and unit	Youngest detrital comp. (Ma)	Important component no. 1 (Ma)	Important component no. 2 (Ma)	Important component no. 3 (Ma)	Important component no. 4 (Ma)	Important component no. 5 (Ma)	Palaeoarchaean detritus
VM95/22 Tartoq Group	2842 ± 6	2842 ± 6	2896 ± 8	2938 ± 8	3122 ± 20	3034 ± 18	yes
VM96/2 Neria	2857 ± 12	2987 ± 12	2857 ± 12	2886 ± 14	3034 ± 18	2938 ± 16	yes
GGU129929 Ravns Storø	2908 ± 13						No
LS85G1Si, PK87/10, G91/43, VM93/1 in contact with Færingehavn terrane	2831 ± 6	2831 ± 6	2868 ± 10	2950 ± 14	3230 ± 16	3011 ± 28	Yes
GGU200892 Ivisaartoq belt	3075 ± 15						No
G93/86 in contact with Isukasia terrane	3048 ± 8	3432 ± 14	3073 ± 28	3048 ± 8	3316 ± 20	3217 ± 32	Yes
G94/2 at probable edge of Akia terrane	ca. 3180						No
VM97/12 Hamborgerland	2815 ± 8 or 2866 ± 8	2991 ± 8	2965 ± 6	2866 ± 8	2815 ± 8	2893 ± 10	No

supracrustal belt (Hall and Friend, 1979; Chadwick, 1986), an arcuate belt 30 km long and up to 5 km wide. The thin continuation of the Ivisartoq belt to the west is intruded on its south side by a foliated granodiorite (sample G91/92) with a SHRIMP U/Pb zircon date of 3070 ± 9 Ma (Friend and Nutman, in press). This gives the minimum age of the Ivisaartoq supracrustal rocks. Presently then we consider that the Kapisilik terrane consists of 3080–2960 Ma rocks. The northern edge of the Kapisilik terrane is in tectonic contact with the southern edge of the Palaeoarchaean Isukasia terrane. Friend and Nutman (in press-a) reported that a zircon population of oscillatory-zoned prisms and crystal fragments from a felsic volcano-sedimentary schist (sample GGU200892) within the Ivisartoq supracrustal belt yielded a unimodal age distribution (reproduced in Fig. 5). All analyses gave a weighted mean $^{207}Pb/^{206}Pb$ date of 3075 ± 15 Ma. Clearly these zircons are derived from a simple (volcanic) source within the Kapisilik terrane.

4.6. Metasediment in tectonic contact with the Isukasia terrane

The southern edge of the Isukasia terrane consists of strongly deformed Palaeoarchaean gneisses, which incorporate either granitic veins or neosome developed at ca. 2960 Ma, perhaps when the Isukasia and Kapisilik terranes first docked (Friend and Nutman, in press-a). Within these strongly deformed gneisses are agmatite trains of layered gabbro and ultramafic rocks of definite Palaeoarchaean age (Chadwick and Crewe, 1986; Nutman et al., 1996), plus more continuous units of amphibolites and schists, which during regional mapping in the 1980s were interpreted to be also Palaeoarchaean in age. The unit that has been visited (at 64°55.83′N, 50°0.42′W) is in an isoclinal synformal keel, with mafic/ultramafic rocks in the centre, underlain by garnet + biotite siliceous schists and by strongly deformed to mylonitic Palaeoarchaean orthogneisses of the Isukasia terrane. The siliceous schist unit is 5–10 m thick and is interpreted to be of sedimentary origin. Siliceous schist G93/86 gave a low yield of small (mostly 50–100 μm across) prismatic to oval zircons. Many of the grains appear completely structureless and dull in CL images, with less than half of them (mostly of more prismatic habit) displaying small structural cores of oscillatory-zoned zircon. The oscillatory-zoned cores yielded a polymodal age distribution from ca. 3020 Ma up to almost 3800 Ma (Fig. 5). Eight cores have dates of <3100 Ma, whereas only three >3550 Ma cores were detected. Within this population 3048 ± 8 Ma and 3073 ± 28 Ma (2σ) components were extracted by mixture modelling (Table 2). A 3073 ± 28 Ma is indistinguishable from the 3075 ± 15 Ma zircons from the volcano-sedimentary sample GGU200892 dated from the Ivisaartoq supracrustal belt of the Kapisilik terrane (see above). Thus the sample G93/86 has detrital zircons, which could have been derived from both the Kapisilik and Isukasia terranes. Our preferred interpretation of the synformal keel supracrustal unit at the G93/86 site is a klippe of structurally overlying Kapisilik terrane. Lower Th/U, higher U overgrowths and whole grains that appear dull and featureless in the CL images yield a weighted mean $^{207}Pb/^{206}Pb$ date of 2968 ± 10 Ma. This is interpreted to record amphibolite facies metamorphism and deformation, perhaps associated with the original docking the Kapisilik and Isukasia terranes.

4.7. Metasediments at the northern edge of the Akia terrane

The northern boundary of the Akia terrane is poorly understood. However, south of the mouth of the fjord Søndre Isortoq (ca. 65°25′N) the gneisses definitely belong to the northern part of the Akia terrane, with protolith dates of 2970–3250 Ma, ca. 2970 Ma granulite facies metamorphism and little evidence from zircon geochronology of younger thermal events (Garde et al., 2000). North of the mouth of Søndre Isortoq other metamorphic and protolith ages occur, which led Friend and Nutman (1994) to speculate that a separate tectonic entity, the *Tuno terrane*, lies north of the Akia terrane (Fig. 1). Garde et al. (2000) reported SHRIMP U/Pb zircon dating on amphibolite facies metasediment sample G94/02 from the north of the mouth of Søndre Isortoq (at 65°24.80′N, 52°31.4.0′W). The sample is from a narrow, approximately NNE-striking belt of metasediments carrying garnet ± sillimanite ± staurolite ± cordierite. Garde et al. (2000) suggested this metasediment belt might mark the northern edge of the Akia terrane. Zircon results for G94/02 from Garde et al. (2000) are summarised here. Most analyses fall into two groups. The first is of rounded grains and cores interpreted to be detrital in origin and with ages between 2900 and 3200 Ma. Due to the small amount of analyses in this sample, mixture modelling was not undertaken. Second, there is a group of homogeneous grains plus overgrowths with a weighted mean $^{207}Pb/^{206}Pb$ date of 2546 ± 6 Ma, interpreted as giving the age of a metamorphic event. The young metamorphic age supports the suggestion this metasedimentary unit marks the northern edge of Akia terrane rather than being within it.

4.8. Hamborgerland metasediments north of the Akia terrane

In the coastal region of the northern part of the Archaean craton, there are major units of granulite facies anatectic garnet ± sillimanite paragneisses and mafic rock. These are most abundant on the large island of Hamborgerland, but are known from only reconnaissance mapping (Allaart et al., 1978). A sample of metasediment VM97/12 from NW Hamborgerland at 65°35.97′N, 52°50.56′W was dated. The sampled unit is a garnet-rich sillimanite-bearing layer about 1 m thick. The zircons are up to 200 μm long and variably rounded. A large minority of grains show only minimal rounding, whereas others are strongly corroded and/or rounded, plus some are sheathed in overgrowths and recrystallisation domains. This is in keeping with the migmatitic, granulite facies condition of this sample. As much as possible, recrystallisation domains and overgrowths were avoided during analysis. The age distribution of the zircons is polymodal, with most grains having ages between 2850 and 3000 Ma (Tables 1 and 2, Fig. 5). No Palaeoarchaean detrital grains were detected amongst the sixty analysed. A few analyses left after filtering create mixture modelling component ages of 2732 and 2782 Ma – these are known ages of magmatism and high grade metamorphism in the vicinity (Friend and Nutman, 1994). A minor component also occurs at 2815 Ma, but it is uncertain whether this is due to tectonothermal disturbance or if it is the youngest detrital component. However, we interpret a larger component at 2866 ± 8 Ma (2σ) as a detrital component. Thus a definite maximum age of deposition of the sediment is 2866 ± 8 Ma, but it is possible that it was deposited after 2815 Ma.

5. Conclusions and discussion

5.1. Diversity of the Greenland Palaeoarchaean record

There is much yet to be extracted from the Greenland Palaeoarchaean record. However on the basis of present results several key points emerge:

(1) All rocks are not equal. Tectonothermal overprints on the Palaeoarchaean rocks have reduced most of the rocks into unusable, strongly deformed, chemically modified tectonites of composite origin (e.g. Nutman et al., 1984b, 1996, 2000, 2002a; Myers, 2001). The earliest of such events are recorded at ca. 3800 Ma (Nutman et al., 1999, 2002b; Friend et al., 2002a) and continued throughout the Archaean. The time span and abundance of these events makes unravelling of isotopic fractionations caused in open-system behaviour particularly difficult. This means that careful fieldwork is necessary to select materials which will give both the most definitive age determinations and which can be presented as the geologically least disturbed materials for cutting edge geochemical investigation. If these crucial first steps are skipped over, then skilled geochemical studies will not produce definitive results on the protoliths and the history of the terrestrial reservoirs from where they were derived.

(2) It is unlikely that rocks better preserved than in the northern part of the Isukasia terrane (in and around the Isua supracrustal belt) will be found. These rocks are still unique globally in the degree of low superimposed late Archaean strain, and the lack of in situ migmatisation. These rocks will remain a key resource for understanding early terrestrial evolution via field and geochemical studies.

(3) However, the best-preserved part of the Isukasia terrane does not contain a complete inventory of West Greenland's Palaeoarchaean geological record. The Færingehavn terrane to the south contains some unique resources. Notable is the 3630–3640 Ma augen gneiss suite, which is the world's oldest-known suite with A-type granite compositions. These rocks have been recognised for more than thirty years (McGregor, 1973). There is now growing acceptance that the Færingehavn

terrane does contain ca. 3850 Ma tonalites, but there is less consensus in the geological community whether these tonalites actually intrude adjacent mafic and siliceous rocks. However, we contend that relationships such as shown in Fig. 2e and f combined with SHRIMP U/Pb dating make a strong case for some of the mafic and siliceous rocks being >3850 Ma (Nutman et al., 2002a). If this interpretation is correct, then possible BIF on Akilia are a highly significant resource, because at ≥3850 Ma they would be the world's oldest-known sedimentary rocks and would indicate a retained hydrosphere at the start of earth's known volcanic and sedimentary record.

(4) The division of the Palaeoarchaean rocks of the Nuuk region into two Neoarchaean tectonic entities (Friend and Nutman, in press-a) clarifies previously noted differences in these rocks north-south across the region (Griffin et al., 1980; Nutman et al., 1993, 1996, 2000). For example, the Isukasia terrane contains ca. 3800 and 3690 Ma TTG rocks and has only ever experienced amphibolite facies metamorphism, whereas in the Færingehavn terrane ca. 3850, 3760 and 3690–3660 Ma TTG are important and significant parts suffered Palaeoarchaean granulite facies metamorphism (Fig. 3a and b).

(5) Initial whole rock Nd isotopic compositions on Palaeoarchaean rocks representing juvenile crust in West Greenland show supra-CHUR initial whole rock $^{143}Nd/^{144}Nd$ isotopic compositions (Hamilton et al., 1983; Bennett et al., 1993; Moorbath et al., 1997). Similar results on ultramafic and mafic rocks from Labrador were obtained by Collerson et al. (1991). There is now general consensus this reflects very early fractionation events in the Earth's mantle, rather than being due solely to open-system behaviour during the subsequent long crustal residence of these rocks. The magnitude of this early fractionation and any changes in convection regimes in the early mantle (Collerson et al., 1991; Bennett et al., 1993) continues to be assessed by isotopic studies on new occurrences of "best" Palaeoarchaean rocks from West Greenland. The search for ^{142}Nd anomalies caused by early terrestrial fractionation of ^{146}Sm with a short half life (Harper and Jacobsen, 1992; to Caro et al., 2003) adds another tool to assess these early terrestrial fractionation events using West Greenland rocks.

(6) It should be noted that >90% of the publications on the West Greenland Palaeoarchaean have concentrated on the Isua supracrustal belt and a few other notorious small localities such as Akilia (island). There are large amounts of the Greenland Palaeoarchaean that is still very poorly known, apart from reconnaissance field studies and a few SHRIMP U/Pb zircon determinations to check their antiquity. This gap in knowledge is being redressed by a major geochronological project in a collaboration between ANU and Hiroshima University. Furthermore, with the exception of two publications (Rosing et al., 2001 on the Aasivik terrane and Hall, 1978 on the Qarliit tasersuat assemblage), *all* studies have been undertaken on the Palaeoarchaean rocks actually within the Nuuk district. Therefore, the region still has potential to discover resources useful for geochemical studies of the early Earth.

(7) The Palaeoarchaean record is distributed in West Greenland as four late Archaean tectonometamorphic entities – the Færingehavn terrane, Isukasia terrane, Qarliit tasersuat assemblage and the Aasivik terrane (Fig. 1). Sufficient U/Pb zircon dating has been undertaken on all four of these entities to recognise differences in their late Archaean metamorphic histories. Thus the Isukasia terrane shows tectonothermal events at ca. 2960 and 2680 Ma, in common with the 3075–2960 Ma Kapisilik terrane to the south, whereas the Færingehavn terrane shows first Neoarchaean metamorphisms at ca. 2720 and 2680 Ma, the Qarliit tasersuat assemblage shows a metamorphism at ca. 2760 Ma, and the Aasivik terrane shows metamorphism at ca. 2720 Ma. This suggests that these Palaeoarchaean bodies were assembled into the Neoarchaean terrane collage at separate times (Friend and Nutman, in press).

(8) Despite these Neoarchaean differences, all four entities show some similarities in their later Palaeoarchaean history. Thus ca. 3690 Ma TTG has been recognised in three of the entities (Fig. 3), whereas all four record a history of granite formation and/or migmatisation with high grade metamorphism between ca. 3660 and 3600 Ma. We contend that although they were assembled separately into a Neoarchaean terrane

collage, they might have originally been spawned from a single block of Palaeoarchaean crust (accreted between 3850 and 3660 Ma with common metamorphism and granite production from 3660 to 3550 Ma – Nutman et al., 1993, 2004) that was rifted apart and dispersed in the Mesoarchaean. Because of this, we extend the term *Itsaq Gneiss Complex* to embrace all Palaeoarchaean bodies of rock in West Greenland, although they now reside in several Neoarchaean tectonic entities.

(9) In the metasediments examined across the extent of the Archaean Craton, Palaeoarchaean detrital grains are extremely rare. As most major meta-sedimentary units have been examined, it is unlikely that any contain a significant resource for ancient zircon grains for geochemical study. The best resource of this type still remains sediments from the Jack Hills in Western Australia (Wilde et al., 2001; Mojzsis et al., 2001).

5.2. *Implications for West Greenland Archaean Neoarchaean crustal evolution*

(1) The rarity of Palaeoarchaean detrital grains in the metasediments, even those in contact with Palaeoarchaean rocks, reinforces the interpretation of the Palaeoarchaean terranes as entirely allochthonous bodies, tectonically emplaced in the Neoarchaean and unrelated to neighbouring gneisses (Friend et al., 1988; Nutman et al., 1989).

(2) There is a difference in provenance style between metasediments sampled from *within* tectonostratigraphic terranes and those *along* tectonic terrane boundaries. Sediments *within* terranes (GGU129929 from the Ravns Storø belt in the Tasiusarsuaq terrane, GGU200892 from the Ivisartoq supracrustal belt in the Kapisilik terrane) have a unimodal age distribution for their detrital zircons, which commonly show little rounding. This suggests they are derived from nearby volcanic sources and formed during the growth of their terrane. Sediments definitely *along* terrane boundaries (VM96/02 between the Neria and Sermiligaarsuk blocks, four sediments in contact with the Færingehavn terrane and G93/86 in contact with the Isukasia terrane) show polymodal age distributions for their detrital zircons. These sediments commonly contain some detrital zircons with ages unknown in the two terranes they lie between. Added to this could be VM95/21 from the Tartoq Group and VM97/12 from the Hamborgerland metasediments. In both cases the lack of modern detailed geological investigations means that their tectonic setting is unclear. However, from their polymodal detrital zircon age patterns, it is possible that these samples are also from units coinciding with terrane boundaries.

(3) The age distribution of metasediments in contact with the Palaeoarchaean Færingehavn and Isukasia terranes is different. The youngest detrital zircons in sample G93/86 in contact with the Isukasia terrane are 3060–3080 Ma, whereas in four samples in contact with the Færingehavn terrane the youngest detrital zircons are always 2830–2840 Ma. The metasediment G93/86 and the nearby edge of the Isukasia terrane (Friend and Nutman, in press-a) was already metamorphosed by 2960 Ma, before the sediments in contact with the Færingehavn terrane had even been deposited. This reinforces the different later tectonic evolution of the Færingehavn and Isukasia terranes.

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