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## Abstract

Eoarchaean juvenile crust formed as 'proto-arcs'. The northern side of the Isua supracrustal belt is an archetypal proto-arc, with ≥3720 Ma boninites, c. 3720 Ma basalts and gabbros, 3720–3710 Ma andesites, diorites and mafic tonalites, 3710–3700 Ma intermediate-felsic volcanic and sedimentary rocks and 3700–3690 Ma chemical sedimentary rocks. On its northern side there is an extensive body of 3700–3690 Ma tonalite. During its evolution, the c. 3700 Ma Isua volcanic–sedimentary assemblage was partitioned into tectonic slices, with intercalation of mantle dunites with pillow basalts, prior to intrusion of c. 3710 Ma quartz diorites. Partitioning also occurred at 3690–3660 Ma, when the 30–20 million years life of the c. 3700 Ma Isua proto-arc was terminated by juxtaposition with the c. 3800 Ma terrane that occurs along the south of the Isua supracrustal belt. The trace element chemistry for all the ≥3720–3700 Ma mafic to intermediate volcanic rocks indicates fluid-fluxing mantle melting. The c. 3690 Ma tonalites have signatures showing melting of garnet-bearing mafic (eclogite) sources. The Isua c. 3700 Ma assemblage developed at an intra-oceanic convergent plate boundary, and it has a life-cycle broadly analogous to (but not identical to) an oceanic island arc eventually accreted against older crust.



## Keywords

boundary, plate, ma, isua, southern, 3700, west, c, greenland, evolution, convergent, arc, proto, eoarchaean, emergence, GeoQuest


## Disciplines

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# The emergence of the Eoarchaean proto-arc: evolution of a *c*. 3700 Ma convergent plate boundary at Isua, southern West Greenland

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**Abstract:** Eoarchaean juvenile crust formed as 'proto-arcs'. The northern side of the Isua supracrustal belt is an archetypal proto-arc, with ≥3720 Ma boninites, *c*. 3720 Ma basalts and gabbros, 3720-3710 Ma andesites, diorites and mafic tonalites, 3710-3700 Ma intermediate-felsic volcanic and sedimentary rocks and 3700-3690 Ma chemical sedimentary rocks. On its northern side there is an extensive body of 3700-3690 Ma tonalite. During its evolution the *c*. 3700 Ma Isua volcanic-sedimentary assemblage was partitioned into tectonic slices, with the intercalation of mantle dunites with pillow basalts, prior to intrusion of *c*. 3710 Ma quartz diorites. Partitioning also occurred at 3690-3660 Ma, when the 30-20 million years life of the *c*. 3700 Ma Isua proto-arc was terminated by juxtaposition with the *c*. 3800 Ma terrane that occurs along the south of the Isua supracrustal belt. The trace element chemistry for all the ≥3720-3700 Ma mafic to intermediate volcanic rocks indicates fluid-fluxing mantle melting. The *c.* 3690 Ma tonalites have signatures showing melting of garnet-bearing mafic (eclogite) sources. The Isua *c*. 3700 Ma assemblage developed at an intra-oceanic convergent plate boundary, and that it has a life cycle broadly analogous (but not identical to) an oceanic island arc eventually accreted against older crust.

**Keywords:** Eoarchaean proto-arc, Juvenile crust, Island arcs, Convergent plate boundaries, Weathering

Strontium, Nd and Hf radiogenic isotope data from the rare surviving oldest rocks (4000-3600 Ma; approximately a millionth of the present crust) indicate that they represent largely juvenile Eoarchaean crust (Moorbath, 1975; Hamilton *et al.*, 1978; Baadsgaard *et al.*, 1986a; Collerson *et al.*, 1991; Bennett *et al.*, 1993, 2007; Caro *et al.*, 2005; Hiess *et al.*, 2009). One school of thought proposes that surviving Eoarchaean crust, exemplified by that found in the Itsaq Gneiss Complex of Greenland, formed at convergent plate boundaries due to fluid-fluxing of mantle and/or melting of subducted mafic crust, i.e. in an environment akin to modern island arcs (e.g., Nutman *et al.*, 1999, 2007; Polat *et al.*, 2002; Polat & Hoffman, 2003; Jenner *et al.*, 2009; Nagel *et al.*, 2012). Alternatively, based solely on the Hf isotopic signatures of zircons from ancient orthogneisses, a non-plate tectonic mechanism of Eoarchaean crust formation was proposed by Næraa *et al.* (2012). Other tectonic settings for crust formation have been provided by Shirey & Richardson (2011) and Dhuime *et al.* (2012).

However, other lines of evidence besides radiogenic isotopic signatures are important in elucidating the origin and evolution of crust. This is because these signatures only indicate whether newly-formed crust is juvenile or involves melting of older continent-like crust, and reveals nothing concerning the *melting mechanism* and *tectonic setting* of crust formation. Thus structural geology data (e.g., Nutman, 1984, Nutman *et al.*, 2002; Komiya *et al.*, 1999), whole rock geochemistry signatures (e.g., Nutman et al., 1999; Polat & Hofmann, 2003; Hoffmann *et al.*, 2011a) and accurate and precise absolute dates (particularly by U-Pb zircon methods) should be integrated with the isotopic data (e.g., Bennett *et al.*, 2007; Hoffmann *et al.,* 2011b). Consequently, a multidisciplinary data set offers the best constraints on the geodynamic setting of crust formation and provides insight into the meaning of changing palaeogeographic settings (such as marine transgressions and regressions). In this paper we follow this holistic approach, and apply it to the *c*. 3700 Ma portion of the Isua supracrustal belt (West Greenland; Fig. 1), which is the world's best preserved example of Eoarchaean crust formation. For the Isua supracrustal belt we conclude that scenarios analogous with, but not identical to, island-arc settings best fit the multidisciplinary data. We introduce here the term *proto-arc* for these ancient rock assemblages. We then explore how in detail the architecture of proto-arcs might have differed from that of Phanerozoic arcs.

## The oldest geological record and the Isua supracrustal belt

### *The Eoarchaean geological record*

The rare occurrences of 4000-3600 Ma (Eoarchaean) rocks reflect the small volume of them that have survived more than 3.5 billion years of plate tectonics, weathering and erosion (Nutman, 2006 and references therein). Such old rocks occur as variably preserved remnants in several gneiss complexes around the world (the Itsaq Gneiss Complex of Greenland, Uivak Gneisses of Labrador, Narryer Gneiss Complex of Western Australia, Acasta Gneisses of the Northwestern Territories and gneisses in the Anshan area of northeastern China; for summaries of these see Schiøtte *et al.*, 1989; Nutman *et al.*, 1991, 1996; Kinny & Nutman, 1996; Bowring & Williams, 1999; Iizuka *et al.*, 2007; Liu *et al.*, 2007; O'Neil *et al.*, 2007 and Horie *et al.*, 2010). The ~3000 $km^2$ Itsaq Gneiss Complex of the Nuuk district of southern West Greenland is the best exposed with some areas of relatively low strain and low metamorphic grade (Nutman *et al*. 1996 and references therein). These attributes mean that data from theItsaq Gneiss Complex are important in of studies towards a better understanding of the Eoarchaean Earth.

### *Itsaq Gneiss Complex*

Much of the Itsaq Gneiss Complex was affected by 3660-3600 Ma high-grade metamorphism and ductile deformation (Griffin *et al.*, 1980, Nutman *et al.*, 1996; Friend & Nutman 2005a), such that by *c*. 3600 Ma its lithologies had been widely converted into strongly deformed, multi-component, amphibolite-granulite facies gneisses (McGregor, 1973; Nutman *et al.*, 2000). An additional complication in interpreting the Itsaq Gneiss Complex is that it occurs as tectonic slivers bounded by folded Meso- and Neoarchaean meta-mylonites, within a collage of younger Archaean terranes that were assembled into their present configuration by the end of the Archaean (Friend *et al.*, 1987, 1988; Nutman *et al.*, 1989, Nutman & Friend, 2007; Crowley, 2002; Friend & Nutman, 2005b).

The later Archaean tectonothermal overprints mean that at most localities it is hard to extract any detailed information concerning the early history of the Itsaq Gneiss Complex. Fortunately, around the *Isua supracrustal belt* at the Itsaq Gneiss Complex's northern extremity (Fig. 1), there are domains that are least affected by later Archaean

tectonothermal overprinting which provide a window onto the early Earth's crust. This allows (a) study of the earliest tectonic evolution (e.g. Nutman, 1984; Nutman *et al.*, 1984, 2002, Komiya *et al.*, 1999; Nutman & Friend, 2009), (b) provides the most robust geochemical and isotopic interpretations because individual Eoarchaean crustal components can be sampled separately (e.g., Baadsgaard et al., 1986a; Nutman et al., 1996, 1999; Friend *et al.*, 2002; Crowley *et al.*, 2002; Polat *et al.*, 2002; Polat & Hofmann, 2003; Bennett *et al.*, 1993, 2003, 2007; Hiess *et al.*, 2009), (c) gives insight into volcanic and sedimentary environments, (d) perhaps containing evidence of the earliest life habitats (Schidlowski *et al.*, 1979; Rosing, 1999; Dauphas *et al.*, 2004; Nutman *et al.*, 2010; Craddock & Dauphas, 2011) and (e) allows understanding of Earth's early climate (Rosing *et al.*, 2010; Nutman *et al.*, 2012a).

*The Isua supracrustal belt*

Within the Itsaq Gneiss Complex the most important unit of supracrustal rocks is the Isua supracrustal belt. The belt is c. 35 km long by up to 2.5 km wide (Fig. 1). To the east it is covered by the edge of the Inland Ice. To the north and south, the belt is bounded by metaplutonic complexes, with that on the north dominated by 3700-3690 Ma tonalite protoliths with *c*. 3650 Ma granite sheets, and that to the south dominated by 3820-3795 Ma tonalite protoliths (Nutman *et al*., 1996, 1997, 2000; Crowley *et al*., 2002; Crowley, 2003; Nutman & Friend, 2009). In southern part of the belt, intrusive tonalite sheets are more common than in the north, and intrude already deformed metabasaltic rocks. To the west the belt is in tectonic contact with Mesoarchaean gneisses and supracrustal rocks (Nutman & Friend, 2009).

Despite that the Isua supracrustal belt largely escaped Neoarchaean deformation, most of the belt was already strongly deformed in the Eoarchaean. This means that outcrop-scale compositional layering is mostly of transposed tectonic origin (Nutman *et al.*, 1984, 1996, 2002, 2007; Myers, 2001). Thus, it is only in a few places that primary volcanic and sedimentary structures have survived (e.g., Nutman *et al.*, 1984; 1997, 2007; Komiya *et al.*, 1999; Solvang, 1999; Furnes *et al.*, 2007). Plutonic rocks are much less common within the belt, but include some gabbros and ultramafic rocks that were derived from both layered gabbro intrusion cumulates and from the mantle (e.g., Dymek *et al.*, 1988; Friend & Nutman, 2011).

Initial age constraints of the Isua supracrustal belt rocks were Rb-Sr whole rock Eoarchaean errorchrons (linear scatters of data which give an approximate ages with

errors of >100 million years) for orthogneiss components invading and proximal to the belt (Moorbath *et al.*, 1972, 1977), a whole rock Pb-Pb 3710 ± 70 Ma errorchron on a banded iron formation unit (Moorbath *et al.*, 1973), and a Sm-Nd 3770 ± 42 Ma errorchron for a mixed suite of felsic and mafic rocks (Hamilton *et al.*, 1978). Within this early geochronological framework, it was considered that the Isua supracrustal rocks were all related, albeit partitioned by early tectonic breaks (Nutman *et al.*, 1984). Subsequently, with uncertainties of ≤ ± 10 Ma, SHRIMP U-Pb zircon dating demonstrated that the belt contained supracrustal rocks varying in age by 100 million years. Hence the belt's southern part is dominated by *c.* 3800 Ma rocks, whereas along its northern and central portions there are *c.* 3700 Ma rocks (Fig. 1; Nutman *et al.*, 1996, 1997). Nutman *et al.* (1997) proposed that these unrelated sequences were separated by Eoarchaean mylonites, and this basic principle has stood independent tests of integrated detailed field work with zircon U-Pb geochronology by Crowley (2003) and Kamber *et al.* (2005) and has been expanded upon in more recent work (Nutman & Friend, 2009; Nutman *et al.*, 2002, 2009; Friend & Nutman, 2010). The younger, *c*. 3700 Ma supracrustal and plutonic rocks along the northern side of the Isua supracrustal belt overall have the most areas of low rare low strain of all the Itsaq Gneiss Complex is therefore the most amenable one to provide information on the origin and evolution of Eoarchaean crust.

## Detailed evolution of the *c*. 3700 Ma portion of the Isua supracrustal belt

### *Lithologies and U-Pb zircon geochronology*

The *c*. 3700 Ma portion of the Isua supracrustal belt comprises tectonically imbricated slices of mostly strongly deformed, amphibolitised pillow lavas and lesser amounts of gabbro and cumulate ultramafic rocks, felsic schists with andesite-dacite protoliths, metasedimentary mica-rich schists derived from volcanic sources, chemical sedimentary rocks and depleted mantle dunite (e.g., Allaart, 1976; Nutman *et al.*, 1984, 2009, Polat *et al.*, 2002; Polat & Hofmann, 2003; Bohlar *et al.*, 2004, 2005; Kamber *et al*., 2005; Friend *et al.*, 2007; Friend & Nutman, 2011).

The amphibolites have two main variants, a high Mg/Fe one that in early studies was known as the *garbenschiefer unit* (e.g., Allaart *et al.*, 1976, Gill et al., 1981; Nutman *et al.*, 1984) and the other is a lower Mg/Fe type. These form mutually exclusive mappable units (Nutman *et al.*, 1984, 2002; Nutman & Friend, 2009). The high Mg/Fe variety (garbenschiefer) has boninitic affinities, whereas the lower Mg/Fe

varieties are geochemically akin to arc basalts (showing features such as enrichment of the light rare earth elements and Nb and Ti depletions; Polat *et al., 2002;* Polat & Hoffman, 2003; see below). Both varieties are dominated by volcanic protoliths, with sporadic but overall rare preservation of pillow structure (e.g., Komiya *et al.*, 1999; Furnes *et al.*, 2007, 2009; Friend & Nutman, 2010). In a rare low strain area in the west of the belt (Fig. 2), layered gabbro with amphibolitised relict igneous texture occurs within boninitic pillow lavas and contains high Th/U igneous zircons with an age of 3717 ± 11 Ma (Nutman & Friend, 2009). Boninitic amphibolites with relict pillow structure are also cut by a 3712 ± 6 Ma hypabyssal mafic tonalite sheet (Nutman & Friend, 2009). These precise U-Pb zircon minimum ages indicate that the boninitic amphibolites are ≥3720-3710 Ma old, in accord with less precise whole rock Sm-Nd and Lu-Hf age determinations on them of 3756 ± 200 Ma to 3825 ± 160 Ma (Hamilton *et al.*, 1978; Hoffmann *et al.*, 2010). The 3717 ± 11 Ma leucogabbro sheet that cuts boninitic metabasalts has arc-basalt chemical affinities (sample G07/26 – below), and might give an absolute age determination on the regional lower Mg/Fe basalts. In the northern end of the belt, 3717 ± 6 Ma diorites (Friend & Nutman, 2010) intrude lower Mg/Fe amphibolites.

Ultramafic rocks occur as lenses within the *c*. 3700 Ma amphibolite sequences. Most of these are highly altered and modified, and now occur as variably carbonated amphibole, serpentine and locally talc-bearing schists. Some of these represent dismembered, altered cumulate portions of gabbro intrusions (Dymek *et al.*, 1988), but one strand of ultramafic schists in the western end of the belt (Figs. 2 and 3) contains two large pods (*c.* 50 x 100 m and 100 x 500 m) of variably-altered but locally fresh dunite, which have depleted mantle chemical signatures, such as depletion of the light rare earth elements, extremely low Al, Ca and Ti contents and being dominated by c. $Fo_{90}$ olivine (Friend & Nutman, 2011).

Layered felsic mica schists are most extensive in the eastern end of the belt (Fig. 1). They locally preserve agglomerate structures and graded layering (Nutman *et al.*, 1984, 1997) and are thus most likely of volcanic or volcano-sedimentary origin. These rocks contain 3710-3700 Ma igneous zircons, showing they are marginally younger than intercalated mafic volcanic rocks (Nutman *et al*., 2009). Mafic mica-garnet rich schists are found interspersed with the felsic schists, and are derived from weathered mafic igneous sources (Nutman *et al.*, 1984, 2012a; Bohlar *et al*., 2005). The mafic schists have given sparse yields of small prismatic zircons with

3710-3700 Ma ages (Kamber *et al.*, 2005) – confirming that they are coeval with the associated felsic mica schists.

Major units of non-detrital sedimentary rocks occur with the *c*. 3700 Ma volcanic rocks (Moorbath *et al.*, 1973; Allaart, 1976). These yield sparse, small (typically *c*. 50 μm long), euhedral, oscillatory-zoned zircons, with ages of 3695-3690 Ma (Nutman *et al.*, 2002, 2009). These are interpreted as small amounts of volcanogenic material incorporated into the chemical sedimentary rocks, possibly as ash, and thereby give a proxy for their true age (Nutman *et al.*, 2002, 2009). These chemical sedimentary rocks vary from chert, to layered quartz + magnetite banded iron formation, to calc-silicate rocks and dolomitic marbles. The cherts and banded iron formations have always been accepted as having sedimentary protoliths (onwards from Moorbath *et al.*, 1973 and Allaart, 1976). However the protoliths for the calc-silicate rocks and marbles have been disputed. Some authors suppose that they are entirely metasomatic in origin (e.g. Rosing *et al.*, 1996), contesting earlier interpretations that at least some have sedimentary/surficial protoliths (Allaart, 1976; Nutman *et al.*, 1984). The sedimentary interpretation is supported by modern detailed geochemical and stable isotope studies because at least some of them display a seawater-like REE+Y (rare earth element + yttrium) trace element signature (Friend *et al.*, 2007), and because they display Fe isotopic fractionation indicative of biomediation (Craddock & Dauphas, 2011). Thus although it is widely acknowledged that the belt has been subjected to metasomatism with carbonate veining, there is increasing evidence that at least some of the carbonates are of sedimentary/surficial origin. It also now appears that biomediation probably played a part in their development (Craddock & Dauphas, 2011; Nutman *et al.*, 2010).

≥3700 Ma supracrustal rocks of the Isua supracrustal belt, locally intruded by 3720-3710 Ma quartz diorites and mafic tonalites, are in early tectonic contact with a northern domain of 3700-3690 Ma tonalitic gneisses intruded by *c*. 3650 Ma granite sheets (Fig. 1; Nutman *et al.*, 1996, 2000, 2002; Crowley *et al.*, 2002). This 3700-3690 Ma tonalite domain, or a volcanic carapace eroded off it, could be representative of the distal source for the rare, small oscillatory-zoned zircons in the chemical sedimentary rocks of the *c.* 3700 Ma part of the belt.

*Submarine and subaerial environments*

In low strain zones scattered throughout the belt, ≥3720 Ma both boninitic and lower Mg/Fe amphibolites locally preserve pillow structures. In a low strain zone in the east of the belt (Fig. 4, locality A), more evolved mafic-intermediate volcanic rocks also preserve pillow structures (Solvang, 1999), and these have yielded a single oscillatory-zoned prismatic zircon with an age of 3709 ± 9 Ma (Nutman *et al.*, 2010). This indicates that submarine conditions continued until at least *c*. 3710 Ma.

Based on the facing direction of the relict pillow structures in low strain zone A on Fig. 4, the *c*. 3710 Ma mafic-intermediate pillowed rocks are stratigraphically the lowest, and face up to a unit of chemical sedimentary rocks. Sparse 3695-3690 Ma volcanogenic zircons within the chemical sedimentary rocks are taken as giving their time of deposition (Fig. 4; Nutman *et al.*, 2002, 2009). The structural and stratigraphic top of the chemical sedimentary rocks is terminated by a mylonite, above which there is an allochtonous body of ≥3720 Ma boninitic amphibolites. The boundary between the underlying pillowed mafic-intermediate volcanic rocks and the chemical sedimentary rocks is not a tectonic break (albeit it is deformed). Our preferred interpretation, supported by geochemical evidence given below, is that the underlying pillowed submarine volcanic rocks were uplifted, subaerially exposed, weathered and eroded, prior to a transgression and deposition of the chemical sedimentary rocks on top of them at 3695-3690 Ma.

The accrued geological, geochronological and geochemical data are evidence for contrasting volcanic and sedimentary environments, from (deep) marine for the early basalts to at least locally subaerial between 3710 and 3695 Ma and then submarine again for the 3695-3690 Ma chemical sedimentary rocks.

*Whole rock geochemistry of igneous rocks*

Polat *et al.*, (2002) and Polat & Hofmann (2003) showed that with careful sampling, Isua mafic igneous rocks with close to original igneous geochemical composition can be selected. This is borne out in the mafic igneous (M), felsic igneous (F) and weathering (W) ternary plot of Ohata & Arai (2007; Fig. 5; see figure caption for more information on this plot), where 'fresh' well-preserved metabasaltic rocks of both categories plus leucocratic metagabbros of the lower Mg/Fe suite coincide with the igneous trend in this diagram (Table 1, Fig. 5). The centres of well-preserved pillows in basalts of lower Mg/Fe island arc affinity show $\delta^{18}O$ values of +9.9 to +6.3 (Furnes *et al.,* 2007), typical of those found in younger submarine settings, with the

variation considered to reflect hydrothermal effects immediately following eruption. Likewise, well-preserved (i.e. with relict igneous texture) more evolved igneous rocks fall exactly on the igneous trend M-F igneous trend on the M-F-W plot; 3720-3710 Ma diorites and mafic tonalites (samples 229404, G07/25 and G07/28) and the c. 3690 Ma tonalites (such as sample G07/41; Table 1, Fig.5). Indication of the freshness of these rocks is given by 'mantle-like' whole rock $\delta^{18}O$ values between +7 to +8 for samples 229404 and 225893 (Baadsgaard *et al.*, 1986b).

Alteration of the igneous chemistry of the c. 3700 Ma igneous rocks could have been by metasomatism or Eoarchaean weathering. Superimposed metasomatism can be due to focused-flow fluid-dominant regimes during later tectonothermal events (e.g. Nutman *et al.*, 1984; Rosing *et al.*, 1996). Field evidence such as coarse-grained quartz or carbonate veining and growth of coarse mica can discriminate such cases, and such rocks have been avoided in sampling (e.g., Polat & Hofmann, 2003; Jenner *et al.*, 2009; Hoffmann *et al.,* 2010). Strong ductile deformation with distributed fluid flow may also provide an opportunity for marked modification in the chemistry of the rocks. This can be assessed by comparing the chemistry of Isua metavolcanic rocks from low and high ductile strain domains. In the rare low strain domains pillow structures are preserved, but generally Eoarchaean (i.e., prior to the intrusion of the ~3500 Ma Ameralik dykes) ductile high strain converted the 'boninitic' rocks into magnesian chlorite + orthoamphibole + plagioclase schists, whereas the lower Mg/Fe basalts were converted into strongly layered hornblende + plagioclase amphibolites. On an MFW plot, the compositions of the strongly deformed rocks from both categories are congruent with their less deformed counterparts (Fig. 5 and its legend). This demonstrates that the superimposed ductile deformation did not modify the major element chemistry of the rocks to the extent it could be detectable on such plots. Immobility is also apparent for elements such as the REE and HFSE that are used to discriminate different petrogenesis of igneous suites. This is shown by comparing the trace element pattern for low strain pillowed amphibolites such as G05/32, -33 with highly strained, flaggy amphibolite G93/18 from the same unit (Fig. 6a).

A second scenario to modify whole rock chemistry is weathering, particularly in a subaerial setting. At localities A and B (Fig. 4) the major element chemistry of the *c*. 3710 Ma volcanic rocks unconformably overlain by chemical sedimentary rocks of deviates from igneous trends in the MFW plot, with a strong dispersion from the M-F trend towards W. Such deviations from undisturbed geochemical trends are focused

306 stratigraphically underneath a younger, overlying chemical sedimentary unit.
307 Furthermore, 3710-3700 Ma volcano-sedimentary rocks that are low strain and lacking
308 in evidence of metasomatism via veining show variable but mostly considerable
309 deviation towards W off the M-F igneous trend. This suggests that prior to or during
310 deposition, material forming them had been subjected to variable degrees of subaerial
311 weathering (Nutman *et al.*, 2012a; Fig. 5).
312 Our selection of mafic whole rock geochemical data (Table 2) replicates the
313 results of the more extensive studies of Polat *et al*. (2002) and Polat & Hofmann
314 (2003), with enrichment of the LREE over the HREE and negative Nb, Ta and Ti
315 anomalies (Fig. 6A). Using combinations of elements considered as immobile (at least
316 in rock-dominated tectonic systems), location of the data in the volcanic arc field in
317 plots such as Nb/Yb versus Th/Yb (Fig. 6B) indicate the importance of fluid-fluxing of
318 upper mantle sources in forming the boninites and the lower Mg/Fe basalts that are of
319 arc affinity (see Eggins, 1993 for explanation of the fluid fluxing mechanism
320 geochemical signatures). Rare gabbro bodies (such as G07/26 with a U-Pb zircon age
321 of 3717 ± 11 Ma that occurs with boninitic amphibolites) also have arc-basalt affinity
322 (Fig. 6). The same applies for somewhat more evolved volcanic rocks such as the
323 slightly younger (3709 ± 9 Ma) varieties around locality A in Fig. 4 – which have a
324 more andesitic affinity (Table 1, Fig. 6).
325 On the $SiO_2$ versus MgO plot the older (3720-3710 Ma) plutonic rocks are
326 mostly low $SiO_2$ and high MgO (Fig. 7A). These can be broadly classified as
327 quartz-diorites, apart from G07/25 that is a high-Mg tonalite. On discriminant
328 diagrams such as Y versus Sr/Y and $Yb_{(N)}$ versus $La/Yb_{(N)}$ the lower silica 3720-3710
329 Ma plutonic rocks fall in the field of typical arc rocks (Fig. 7B,C). Most likely their
330 origin was dominated by melts produced by fluid fluxing of mantle (see Martin *et al.*,
331 2005). It should be noted also that these rocks have trace element patterns and Nb/Yb
332 and Th/Yb very similar to c. 3710 Ma andesites (such as samples at locality A, Fig. 4;
333 see the data for plutonic rocks G07/28 and G05/25 placed onto Fig. 6). This indicates
334 that they are simply plutonic and volcanic manifestations of the same igneous event.
335 The 3700-3690 Ma tonalites that form the large allochthonous mass north of
336 the belt are geochemically distinct, with higher $SiO_2$ and much lower MgO (Fig. 7A).
337 In Y versus Sr/Y and $Yb_{(N)}$ versus $La/Yb_{(N)}$ discriminant plots (Figs. 7B,C), some of
338 the samples fall in the field of typical adakites whereas others lie in the region where
339 adakites and typical arc compositions overlap. Based on this alone, the younger suite

could have formed by melting of mafic rocks under high pressure (eclogite or garnet amphibolite; c.f., Martin, 1986; Martin *et al.*, 2005) or by (extensive) fractional crystallisation of more mafic arc magmas (MacPherson et al., 2006). However there is a major volume problem with the latter model, because the tonalite-trondhjemite-granodiorite suites form approximately 90% of the Eoarchaean crust. It is now documented that early tectonic intercalation has juxtaposed upper mantle and upper crustal rocks (Nutman et al., 1996; Friend et al, 2010). Why then is there not ample evidence of l*ower crustal* cumulates complementing the tonalite-trondhjemite-granodiorite suites in the Eoarchaean geological record? This volume problem is resolved if the tonalite-trondhjemite-granodiorite suites formed by partial melting of mafic rocks buried by underthrusting into the stability field of eclogite. In which case, the complementary very dense resitites reside in the mantle.

*≥3650 Ma tectonism*

Mapping combined with U-Pb zircon geochronology shows that within the *c*. 3700 Ma part of Isua, rocks were tectonically intercalated prior to intrusion of the *c*. 3510 Ma Ameralik (metabasic) dykes (Nutman *et al.*, 1984, 2007, 2009; Komiya *et al.*, 1999; Nutman & Friend, 2009). There are several episodes of tectonic intercalation. Thus at the eastern margin of northwestern portion of the belt (Fig. 2), *c*. 3720 Ma low Mg/Fe amphibolites of island arc basalt affinity with rare relicts of pillow structure (e.g., Furnes *et al.*, 2007) are tectonically intercalated with upper mantle dunites (Friend & Nutman, 2011). This tectonic intercalation is older than the intrusion of the 3717 ± 6 Ma diorites to the east (Friend & Nutman, 2010). Within the northwestern part of the belt, *c*. 3690 Ma banded iron formation is tectonically intercalated with felsic schists bearing 3724-3712 Ma volcanic or sedimentary zircons (G05/28+29 in Fig. 2). In turn, these intercalated rocks occur as a tectonic sliver within ≥3717 Ma boninitic amphibolites. In the eastern part of the belt, a tectonic (folded thrust) contact between *c*. 3690 Ma metachert with banded iron formation and tectonically overlying boninitic amphibolite has been intruded by a gabbro sheet (Fig. 4). Along strike to the southwest, where this sheet lies entirely within pillowed boninitic amphibolites, it has yielded sparse high Th/U 3643 ± 6 Ma zircons, interpreted as giving the magmatic age (sample G97/87 in Nutman *et al.*, 2002). This indicates that the thrusting of the boninitic rocks over *c*. 3690 Ma chemical sedimentary rocks occurred before 3643 Ma.

There are also younger Eoarchaean shear zones and tectonic breaks in the Isua supracrustal belt. The most important of these is the tectonic break separating the *c*. 3800 Ma rocks in the south of Isua from the *c*. 3700 Ma rocks to the north (Fig. 1). This break has been interpreted as a tectonically reactivated suture, marking the juxtaposition of *c*. 3800 and *c*. 3700 Ma terranes. This occurred between *c*. 3690 Ma (youngest rocks unique to the northern terrane) and 3660 Ma, the age of the first rocks common to both terranes (Crowley, 2003; Nutman & Friend, 2009). North of, and sub-parallel to this important boundary, there are other mylonites. These mylonites partition the *c*. 3700 Ma assemblage into sub-terranes, and thereby also truncate earlier tectonic breaks described above (Nutman & Friend, 2009). These are commonly marked by synkinematic pegmatites with flaser structure that yield ages of 3650-3640 Ma (Crowley *et al.*, 2002; Nutman *et al.*, 2002). These breaks are regarded as relating to extensional partitioning of crust that had been thickened a short time before during the juxtaposition of the 3800 Ma and 3700 Ma terranes (Nutman *et al.*, 2012b).

## Discussion

### *Juvenile crust*

All samples from the *c.* 3700 Ma supracrustal assemblage and the tonalite domain to the north have whole rock initial $\varepsilon_{Nd}$ values of $\geq +1$ and near-chondritic zircon and whole rock initial $\varepsilon_{Hf}$ values of *c.* $0 \pm 1$ (Hamilton *et al.*, 1978; Jacobsen & Dymek, 1987; Baadsgaard *et al.*, 1986a; Moorbath, 2005; Polat *et al.*, 2002; Hiess *et al*., 2009; Rizo *et al.*, 2011; Hoffmann *et al.*, 2011b, Rizo *et al.*, 2011, Næraa *et al.*, 2012). The single exception is the ≥3720 Ma boninitic metabasalts, with initial $\varepsilon_{Hf}$ values of up to + 12 (Hoffmann *et al.*, 2010). Furthermore, all zircons within sedimentary rocks of the *c*. 3700 Ma terrane are 3740-3690 million years old, in stark contrast to the *c*. 3800 Ma terrane to the south where detrital grains up to 3940 million years old have been recorded in sedimentary rocks (Nutman *et al.*, 2009). This shows that the *c*. 3700 Ma assemblage is essentially juvenile – with no detectable input from older (3940-3800 Ma) sialic crust exposed immediately south of the suture running through the Isua supracrustal belt (Fig. 1).

Geochemical evidence such as depletion of Nb and Ta (Fig. 6) indicates that mafic to intermediate igneous rocks in the *c*. 3700 Ma terrane involved fluid-fluxing of upper mantle sources (Polat *et al.*, 2002; Polat & Hofmann, 2003). Igneous rocks represented

by the extensive 3700-3690 Ma tonalite outcrops north of the Isua supracrustal belt have geochemical signatures (e.g. high $La/Yb_{(N)}$ and low $Yb_{(N)}$) that according to widely accepted interpretations (Martin, 1986; Martin *et al.*, 2005; Hoffmann *et al.*, 2011a) are indicative of melting eclogites and/or high pressure granulites. Combined with the U-Pb zircon geochronology (Nutman & Friend, 2009) this shows a temporal progression of melting mechanisms from early mantle fluid-fluxing (≤3720-3710 Ma) to melting high pressure mafic rock sources (3700-3690 Ma).

*Tectonically active environment*

Development of the igneous assemblages took place in a tectonically active environment, such that by 3717 Ma, metabasalts had been tectonically intercalated with slices of upper mantle dunite (Fig. 2; Friend & Nutman, 2010, 2011), and by 3643 Ma andesites and basalts capped by 3690 Ma chemical sedimentary rock had ≥3720 Ma boninites thrust over them (Localities A and B in Fig. 4). Further evidence for the tectonically active environment comes from the variable degree of *c*. 3700 Ma chemical weathering displayed by both 3710-3700 Ma volcano-sedimentary rocks (Nutman *et al.*, 2012a) and by ≥3710 igneous rocks (Fig. 5). As demonstrated by Nesbitt *et al*. (1997), *highly variable* degrees of weathering within a single terrane are indicative of tectonic instability.

*Apparent sea level*

All of the early (≥3710 Ma) mafic to intermediate volcanic rocks, preserved in rare low strain zones have relict pillow structures (Nutman *et al.*, 1984; Komiya *et al.*, 1999; Solvang, 1999; Furnes *et al.*, 2007). Thus early volcanism was all submarine. 3710-3700 Ma felsic and mafic volcano-sedimentary rocks display variable and locally intense weathering, indication of emergence of the source regions (Nutman *et al.*, 2012a). Likewise, it is demonstrated here that *c*. 3710 Ma pillowed volcanic rocks were strongly weathered, prior to deposition on top of them of 3695-3690 Ma chemical sedimentary rocks. This shows emergence, followed a transgression between 3710-3695 Ma. Therefore, throughout the ≥3720 Ma-3690 Ma evolution of the supracrustal assemblage there were changes in relative sea level. These apparent sea level changes can be because of progressive accumulation of the igneous suites so that they eventually emerged above sea level. However, syn-igneous tectonic thickening of the crust was at least partly responsible for the emergence, as demonstrated by ≥3717

Ma tectonic intercalation of upper mantle dunites and harzburgites within crustal sequences (Fig. 2; Friend & Nutman, 2010, 2011). However after the emergence, there was clearly a transgression followed by the deposition of chemical sedimentary rocks, indicating a relative rise in sea level between 3710-3695 Ma.

*Comparison with Phanerozoic intra-oceanic arc and plateau settings*

Because the *c*. 3700 Ma Isua assemblage consists of essentially juvenile crust at the time it formed, two possible settings for its formation are at an intra-oceanic convergent plate boundary and within an intra-oceanic mafic plateau. The integrated data set allows clear discrimination between these possibilities. The ≥3720-3690 Ma igneous activity took place in a tectonically active setting, with tectonic intercalation of crustal and mantle rocks *during* the development of the package, not just in superimposed, younger, unrelated, tectonic events. This crust mantle intercalation would not be observed during the development of an intra-oceanic plateau, distal from convergent plate boundaries, but it does occur at intra-oceanic convergent plate boundaries (Shervais, 2001 and references therein). Geochemically, there is a temporal shift from mafic, to intermediate and then felsic igneous rocks (Nutman & Friend, 2009). The earlier mafic-intermediate igneous suites (≥3720-3710 Ma) have geochemical signatures indicating melting of depleted mantle by fluid fluxing (Polat *et al.*, 2002; Polat & Hofmann, 2003; Hoffmann *et al.,* 2011b and data in this paper). The younger 3700-3690 Ma tonalites on the north side of the belt are classic Archaean calc-alkaline rocks, with geochemical signatures indicating melting of high pressure (garnetiferous) mafic rocks. Calc-alkaline magmatism appears with the maturing of intra-oceanic island arcs (e.g. Shervais, 2001). On the other hand, adiabatic decompression dry melting in mantle diapirs is the essential processes in the formation of most igneous rocks in intra-oceanic within-plate settings (Farnetani & Richards, 1994 and references therein).

Based then on geochemical criteria, the *c*. 3700 Ma Isua assemblage has characteristics expected at convergent plate boundary settings and shows many of the features recognised in the birth, death and resurrection life cycle of suprasubduction zone ophiolites (arc assemblages; Dilek & Polat, 2008; Friend & Nutman, 2010). As summarised by Stern & Bloomer (1992) and Shervais (2001), key features of this cycle as seen in *c*. 3700 Ma Isua are: Tectonically active, with early mafic rocks produced from the upper mantle by fluid fluxing and melt overprinting from the ‘subducted’ slab

474 to produce submarine boninites and arc basalts, then subaerial emergence with the
475 transition of magmas towards more felsic calc-alkaline varieties and the production of
476 felsic volcano-sedimentary and sedimentary aprons from the emergent volcanic
477 centres.
478
479 *Eoarchaean 'proto-arcs' based on Isua crust formation at c. 3700 Ma*
480 It is likely that the internal architecture of Eoarchaean intra-oceanic island arcs
481 was markedly different from those in the Phanerozoic, albeit geodynamically they
482 occupied the same niche; *i.e.* the removal of 'excess' oceanic first-generation crust
483 from Earth's surface by the formation of second-generation intermediate to felsic
484 juvenile crust. For these Eoarchaean arc-like features we here introduce the term
485 '*proto-arcs*'– to stress that they might have had an architecture very different from
486 modern arc systems. One marked difference could have been the lack of a discrete,
487 long-lived single subduction zone to transport oceanic crust well below the
488 asthenosphere. Instead, mature proto-arcs might have been marked by imbricate
489 packages of intercalated oceanic crust, uppermost mantle and earliest mafic proto-arc
490 crust, that extended only into the uppermost mantle (de Wit, 1986; Nutman *et al*.,
491 2007; Nagel *et al.*, 2012).
492 The different thermal state of the Eoarchaean Earth is why *c*. 3700 Ma
493 convergent plate boundaries were unlikely to be the same as modern intra-oceanic
494 subduction channel + arc assemblages. One important factor to consider is the relative
495 buoyancy of 'oceanic' mafic crust versus the asthenosphere. On the modern Earth,
496 subduction continues because older, cold oceanic crust becomes denser than the
497 asthenosphere, and upon rupture will sink through it. Thus it is widely recognised that
498 during the life of intra-oceanic subduction systems, the conserved plate is mostly under
499 tension, commonly with the roll-back of the subduction zone (Collins, 2002; Cawood,
500 2005). With time, in modern systems that are not sediment starved, this will lead to the
501 overstepping of the fore-arc volcanic apron by younger arc igneous rocks (e.g.,
502 Cawood *et al*., 2011 and references therein). If in the Eoarchaean the oceanic crust
503 typically remained buoyant relative to the asthenosphere (because oceanic crust was
504 hotter? younger?) then the commonest style of Phanerozoic intra-oceanic subduction
505 could not operate. However, in order to compensate for oceanic crust production along
506 spreading centres, Eoarchaean convergent plate boundaries must have existed. An
507 initial compressional intra-oceanic rupture zone would have intercalated mantle with

still-buoyant oceanic crust. At the earliest stage of proto-arc development (this stage is *not* illustrated in Fig. 8), this would have promoted the formation of magmas by fluid-fluxing of the upper mantle to produce the Eoarchaean equivalents of the boninites and arc tholeiites, as seen in the earliest stages of Phanerozoic intra-oceanic arc development (Sterner & Bloomer, 1992; Shervais, 2001). Thereafter, the buoyancy of the underthrust oceanic crust would have promoted a compressional tectonic regime, because of the inability of that crust to undergo Phanerozoic-style subduction (*i.e.*, sink into the mantle on its own accord). Crustal shortening in this compressional regime would have been accommodated by tectonic intercalation (Fig. 8A; e.g. deWitt, 1985; Nutman *et al.*, 2007). Pressure increase at the base of this growing imbricated pile combined with temperature increase by conduction would have caused melting first in the stability field of garnet amphibolite (and/or high pressure granulite) and then rutile-bearing eclogite (Hoffmann *et al.*, 2011a). This temporal progression in changing geochemistry is in keeping with garnet-signatures only being present in the youngest rocks of the *c*. 3700 Ma Isua assemblage – the 3700-3690 Ma high $SiO_2$ – low MgO tonalites, which display trace element signatures such as high $La/Yb_{(N)}$ and low $Yb_{(N)}$ (Fig. 7C).

Another feature of this model is that the earliest proto-arc components – largely mafic volcanic rocks, gabbros with intercalations of upper mantle rocks – were preserved in the crust by being locked in a greater volume of more silicic rocks formed by the melting of underthrust eclogitised mafic crust. Thus in the prot-arc model, oceanic (MORB-like) crust will have a much lower crustal preservation potential than the early mafic arc rocks. This is because there is a much greater probability of it being partially melted at depth to produce garnetiferous restites that remain in the mantle, than for it being tectonically emplaced and preserved into the crust. No doubt this genre of model will require further testing, but at present it satisfies not only geochemical and isotopic constraints, but geochronological and structural geology ones as well.

An important difference between Archaean proto-arcs and modern arc systems would have been the bias between different melting mechanisms (Martin, 1986). Although the melting of eclogitised slab does occur locally on Earth today, recycling of oceanic crust into the deep mantle with fluid-loss is a much more important process. For this reason Phanerozoic island arc magmatism produces largely basaltic-andesitic suites, with lesser amounts of more silicic rocks. In the Phanerozoic, the production of

542 more silicic compositions is predominantly in continental arcs, where the silicic
543 compositions show isotopic evidence of recycling of the older continental rocks (e.g.,
544 Hildreth & Moorbath, 1988). However, in the Eoarchaean, the greater importance of
545 melting of mafic eclogite led to silicic rocks such as tonalites being the dominant
546 product of proto-arc magmatism, with only lesser amounts of slightly older basalts and
547 andesites preserved as enclaves and tectonic slices within them. Hence in contrast to
548 Phanerozoic silicic arc magmas, those formed in the Eoarchaean are predominantly
549 juvenile additions to the crust, and therefore contributing to net growth of the crust
550 more effectively.
551
552

*Eoarchaean proto-arcs and mantle isotopic signatures*

It is increasingly recognized that that there is a change in the pattern of $^{143}Nd$ and $^{176}Hf$ isotopic evolution in the mantle, as reflected in the isotopic signatures of Archaean juvenile granitoids and mantle-derived rocks, with an inflection in both long-lived systems occurring between 3600 Ma and 3200 Ma (c.f., Bennett *et al.,* 1993, Hiess *et al.,* 2009, Næraa *et al.*, 2012). Pre-3500 Ma, mantle derived samples are characterised by near chondritic $\varepsilon_{Hf}$ and positive $\varepsilon_{Nd}$ values. Post 3200 Ma, both systems are characterized by a quasi-linear evolution with a well defined correlation of positive $\varepsilon_{Hf}$ and $\varepsilon_{Nd}$ values (Vervoort *et al.*, 2011).

The near chondritic Eoarchaean $\varepsilon_{Hf}$ compositions likely reflect the limited extent of Hadean and early Archaean crust and thus its minimal impact on mantle isotopic evolution, but may also be related changes in arc processes. As previously discussed, the chemistry of the basalts must at least in part reflect depleted mantle overprinted by subduction components. During the proto-arc era and in contrast to younger time periods, any subducted sediments contributing to Isua arc development were likely of similar Hf and Nd isotopic composition to the mantle. Although Sm and Nd and to a lesser extent Lu and Hf can be fractionated by subduction fluids and melts, there was insufficient time to generate significant isotopic variations in the sedimentary component prior to subduction. Furthermore, the efficacy of crustal extraction and mixing of depleted mantle domains might have been hindered by the lack of well-developed, subduction systems resulting in slower transfer and homogenisation of isotopic signatures into the upper mantle.

A significant difference between Eoarchaean proto-arcs versus younger arc systems, is the widespread partial melting of recycled mafic crust, giving rise to distinctive Archean tonalite-trondhjemite-granodiorite suites and resulting in the formation of extensive garnet ± clinopyroxene (with very high Lu/Hf and high Sm/Nd) restites in the upper mantle (Nutman *et al.*, 1999). These restites would develop highly radiogenic Hf and Nd isotopic compositions. The behavior of these residues in the Archean mantle, and their possible influence on mantle isotope characteristics have been little considered. Here we speculate that in the Archaean, refractory residues from TTG genesis may have been resistant to mixing in the upper mantle. If this where the case, the storage of these garnetiferous residues would result in delaying the depletion of Lu/Hf and Sm/Nd and as such buffering upper mantle Nd and Hf isotopic

signatures. Although melting of such restites directly would not produce boninitic composition melts, we further note that *c*. 3900 Ma restites (complementary to the oldest Itsaq Gneiss Complex tonalites) could generate a mantle source with the distinctive radiogenic $\varepsilon_{Hf}$ and $\varepsilon_{Nd}$ characteristics noted for the ≥3720 Ma Isua boninites by Hoffmann *et al.* (2010).

## Conclusions

(1) The 3720-3690 Ma isotopically juvenile crustal assemblage at Isua shows a progression from mafic to felsic combined with geochemical signatures indicating generation initially by fluid fluxing ± melt transport through the depleted upper mantle succeeded by partial melting of garnetiferous metamorphosed mafic rocks.
(2) Within the 3720-3690 Ma period there was tectonic intercalation of crustal and mantle rocks, indicating a tectonically-active environment *during* crust formation.
(3) Tectonic activity juxtaposed rocks ranging in age from ≥3720 Ma to 3690 Ma, that had formed in different environments ranging from deep to shallow water and subaerial.
(4) The geochemical characteristics of the 3720-3690 Ma syn-tectonic Isua assemblage show melting mechanisms known in post-Archaean arc settings, and are best viewed as 'proto-arcs' having formed along Eoarchaean convergent plate boundaries.
(5) The Eoarchaean proto-arcs had the same geodynamic function as arcs today, i.e. the 'disposal' of oceanic crust production coupled with the formation of buoyant intermediate-felsic crust. In our evolving working model, we suggest that their tectonic architecture was likely different from modern arcs, with broad crustal imbrication zones without mass transfer below the asthenosphere, i.e., lacking a discrete subduction zone recycling oceanic crust below the upper mantle.

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910

911 **Table captions**

912 Table 1. Summary of c. 3700 Ma crustal evolution in the Isua area, southern West Greenland.

913

914 Table 2. Whole rock geochemistry of *c*. 3700 Ma Isua rocks. See footnotes for analytical
915 techniques and sources. Laser –ICP-MS trace element analyses were undertaken using fused
916 discs.

917

918 **Figure captions**

919 Figure 1. Geological map of the Isua area, emphasising the *c*. 3700 Ma rocks and the tectonic
920 contact with the c. 3800 Ma terrane to the south. Areas shown in Figures 2 and 4 are indicated.

921

922 Figure 2. Geological map of part of the northwestern Isua supracrustal belt. U-Pb zircon age
923 constraints (Nutman & Friend, 2009 and Friend & Nutman, 2010 and references therein) are
924 given with 95% confidence uncertainties.

925

926 Figure 3. (A) Dunite from the large mantle-derived ultramafic pod in the northeastern part of
927 Fig. 2 (pen for scale, bottom right). Fine-grained olivine matrix contains larger orientated
928 olivine grains (running left-right in the photograph). This fabric formed at high temperature in
929 the mantle, not in the crust during amphibolite facies metamorphism (Friend & Nutman, 2011).
930 (B) Back scattered electron image of dunite. Grey = forsteritic olivine; black = secondary
931 serpentine and white = secondary magnetite. The small amount of secondary magnetite reflects
932 the very high Mg/Fe ratio of the rock. Note the polygonal texture.

933

934 Figure 4. Geological map of part of the eastern Isua supracrustal belt. U-Pb zircon age
935 constraints (Nutman & Friend, 2009 and Friend & Nutman, 2010 and references therein) are
936 given with 95% confidence uncertainties.

937

938 Figure 5. Isua *c.* 3700 Ma rocks in the mafic igneous (M) – felsic igneous (F) – weathering (W)
939 plot of Ohata & Arai (2008). See Ohata & Arai (2008) for how these parameters are

940 constructed. This plot is applicable for all igneous compositions from basalts to granites. This
941 plot includes data from a figure in Nutman *et al.* (2012a).
942
943 Figure 6. Geochemistry plots of Isua *c.* 3700 Ma mafic and intermediate igneous rocks. (A)
944 Primitive mantle normalised trace element diagrams (normalised with values of Sun &
945 McDonough, 1989), (B) Nb/Yb versus Th/Yb, with fields from Pearce (2003).
946
947 Figure 7. Geochemistry of *c.* 3700 Ma felsic igneous rocks: (A) $SiO_2$ versus MgO, (B) Y versus
948 Sr/Y and (C) $Yb_{(N)}$ versus $La/Yb_{(N)}/Yb_{(N)}$. TTG = tonalite-trondhjemite-granodiorite. The
949 fields in the trace element diagrams are taken from Martin (1986). Rare earth elements are
950 normalised to chondrite, using the values of Sun & McDonough (1989).
951
952 Figure 8. Cartoon cross sections comparing *mature* Eoarchaean proto-arc and Phanerozoic
953 island arc structure. The proto-arc is shown at the dominant tonalite-dacite producing phase,
954 *after* arc initiation when fluid fluxing of a mantle wedge occurred.
955

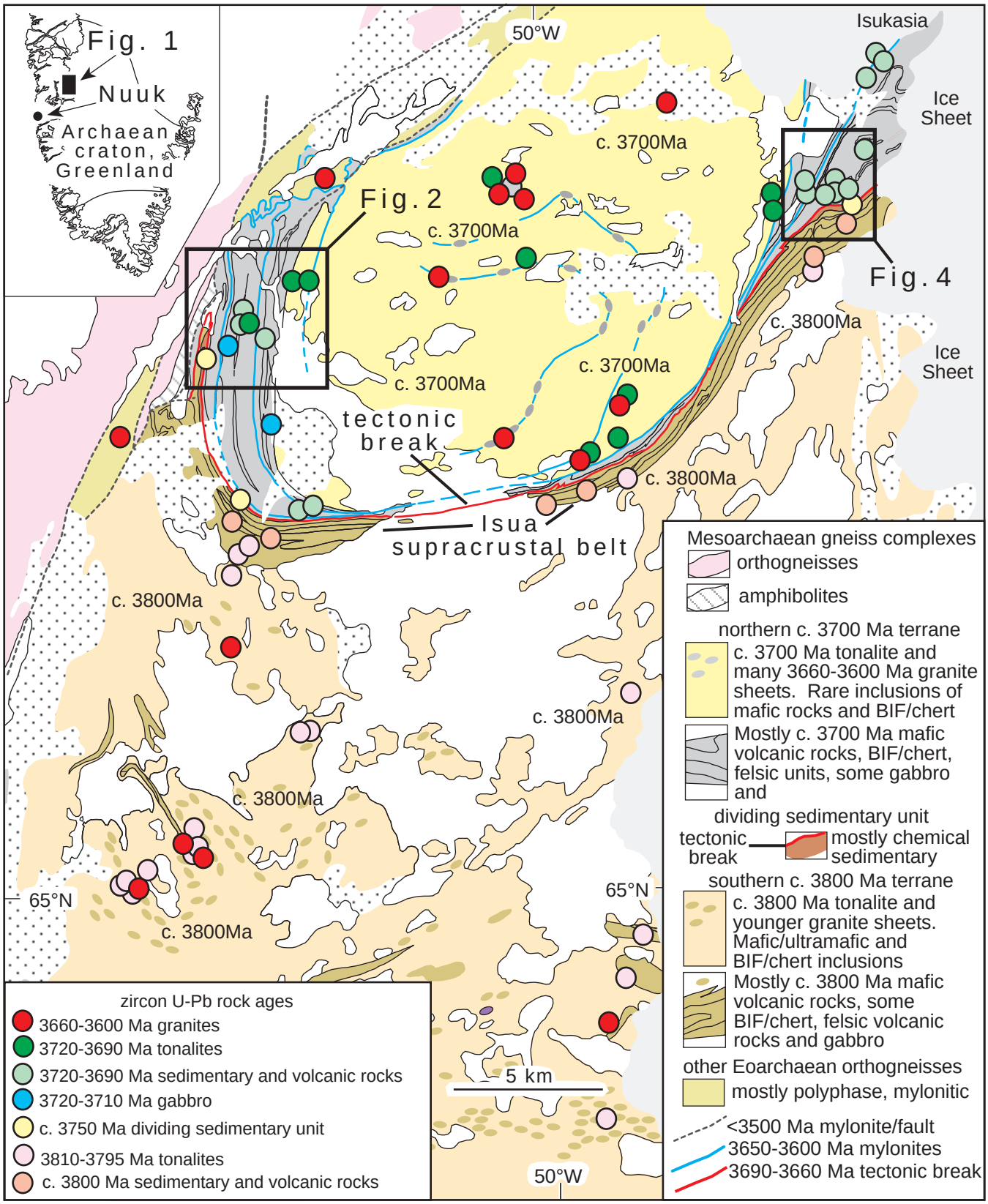
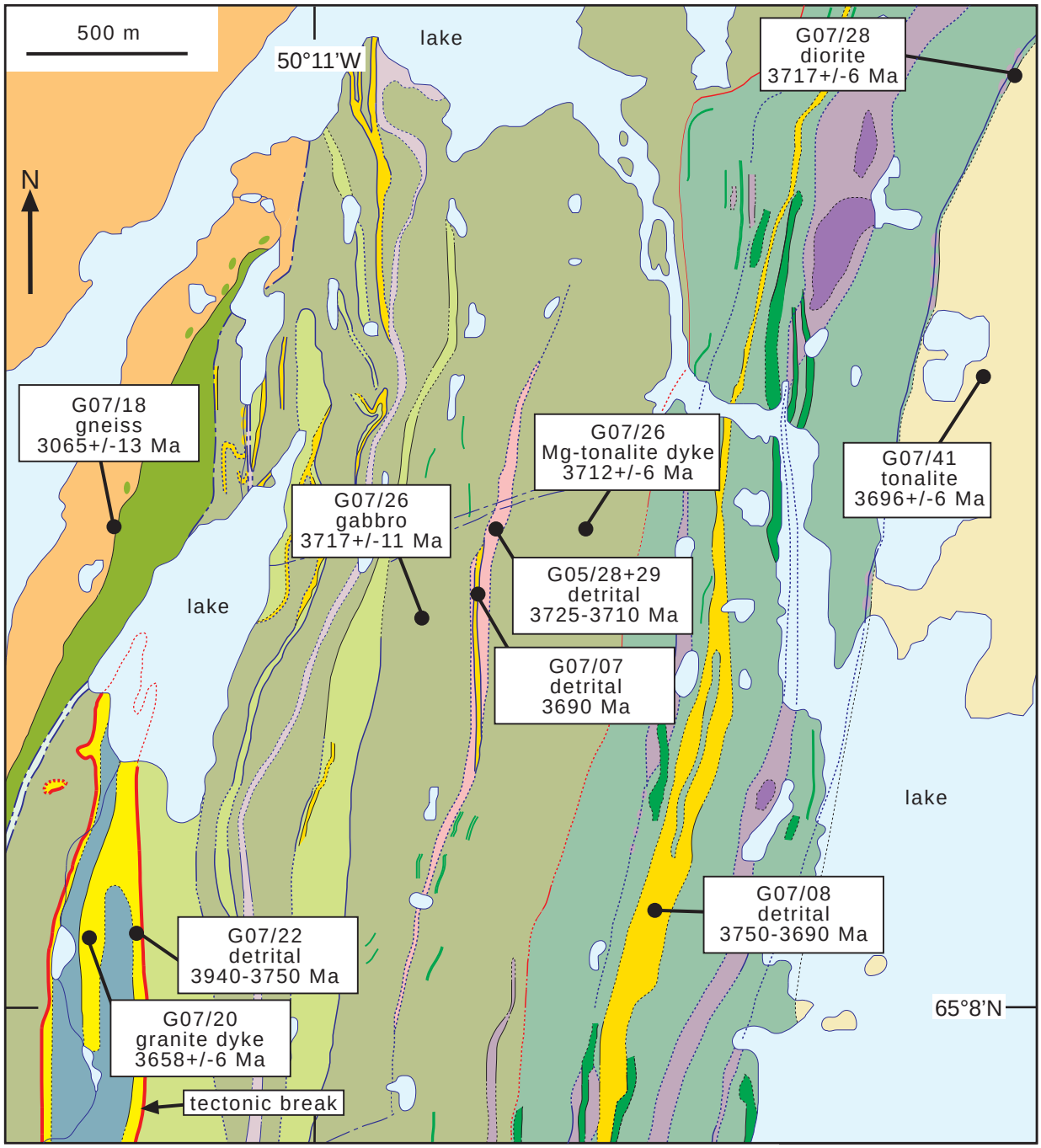


Fig. 1
Nuuk
Archaean craton, Greenland
Fig. 2
Fig. 4
50°W
Isukasia
Ice Sheet
c. 3700Ma
c. 3800Ma
tectonic break
Isua supracrustal belt
65°N
5 km
zircon U-Pb rock ages
3660-3600 Ma granites
3720-3690 Ma tonalites
3720-3690 Ma sedimentary and volcanic rocks
3720-3710 Ma gabbro
c. 3750 Ma dividing sedimentary unit
3810-3795 Ma tonalites
c. 3800 Ma sedimentary and volcanic rocks
Mesoarchaean gneiss complexes
orthogneisses
amphibolites
northern c. 3700 Ma terrane
c. 3700 Ma tonalite and many 3660-3600 Ma granite sheets. Rare inclusions of mafic rocks and BIF/chert
Mostly c. 3700 Ma mafic volcanic rocks, BIF/chert, felsic units, some gabbro and
dividing sedimentary unit
tectonic break
mostly chemical sedimentary
southern c. 3800 Ma terrane
c. 3800 Ma tonalite and younger granite sheets. Mafic/ultramafic and BIF/chert inclusions
Mostly c. 3800 Ma mafic volcanic rocks, some BIF/chert, felsic volcanic rocks and gabbro
other Eoarchaean orthogneisses
mostly polyphase, mylonitic
<3500 Ma mylonite/fault
3650-3600 Ma mylonites
3690-3660 Ma tectonic break




post-Mesoarchaean faults and mylonites

Mesoarchaean orthogneisses

Mesoarchaean amphibolites

Ameralik dykes 3510 Ma

mylonites associated with 3650-3600 Ma interleaving

3690-3660 Ma tectonic break between northern and southern terranes

Dividing sedimentary unit deposited between 3750-3690 Ma

cherts, siliceous carbonates and BIF, with minor detritus including 3940-3750 Ma zircons

Southern terrane ca. 3800 Ma

>3800 Ma amphibolites of mostly island-arc tholeiite affinity

Northern terrane 3720-3690 Ma

imbricate faults internal to the northern terrane

BIF and cherts with mostly 3700-3690 Ma zircons

3720-3710 Ma tonalites

3710-3700 Ma felsic volcanic and sedimentary rocks

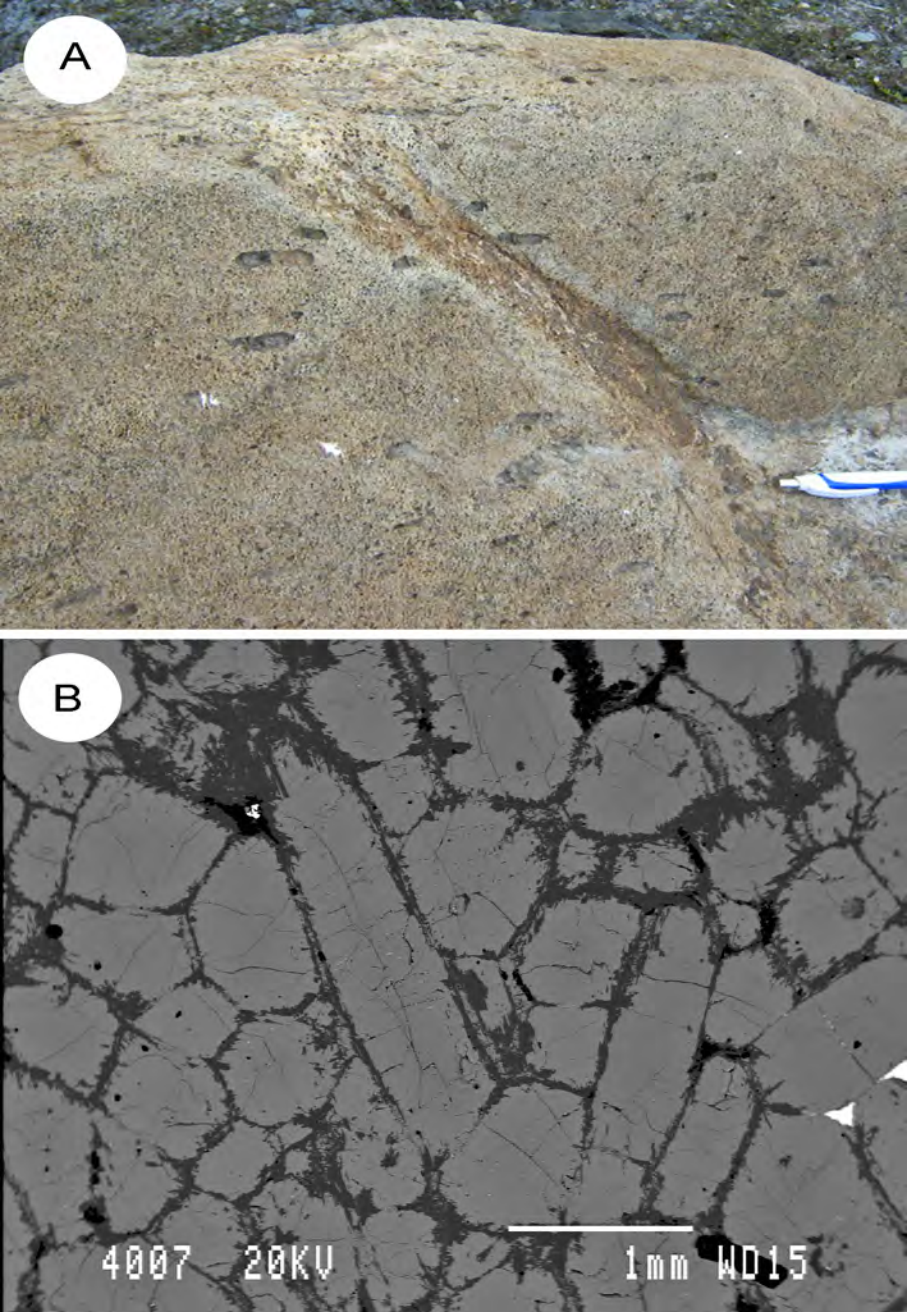
ca. 3710-3700 Ma mafic and intermediate volcanic rocks

ca. 3715 Ma amphibolites of mostly island-arc tholeiite and picrite affinity

>3715 Ma amphibolites of boninitic chemical affinity

>3715 Ma schistose ultramafic rocks

>3715 Ma relict mantle dunitic rocks


A
B
4007 20KV
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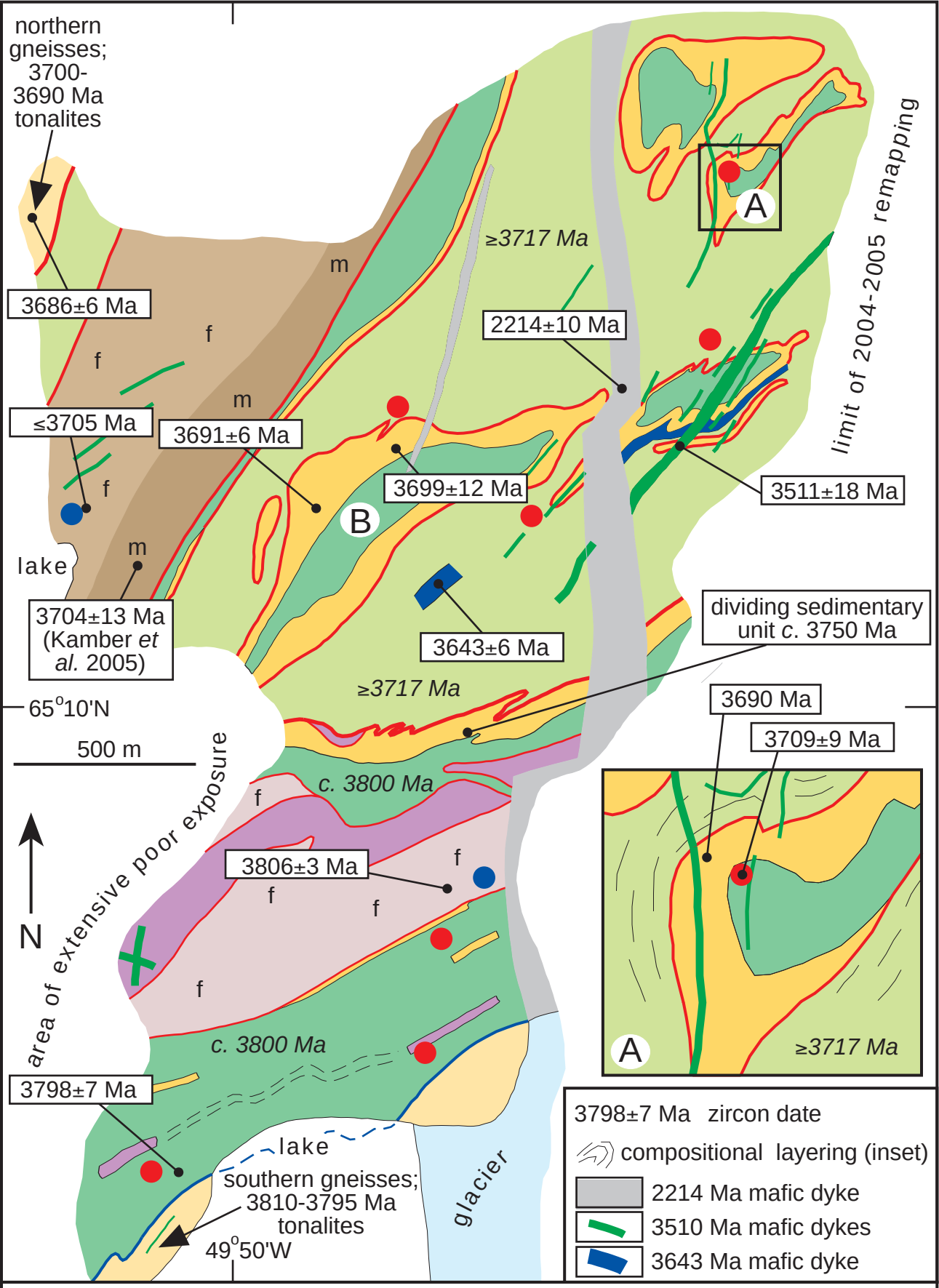
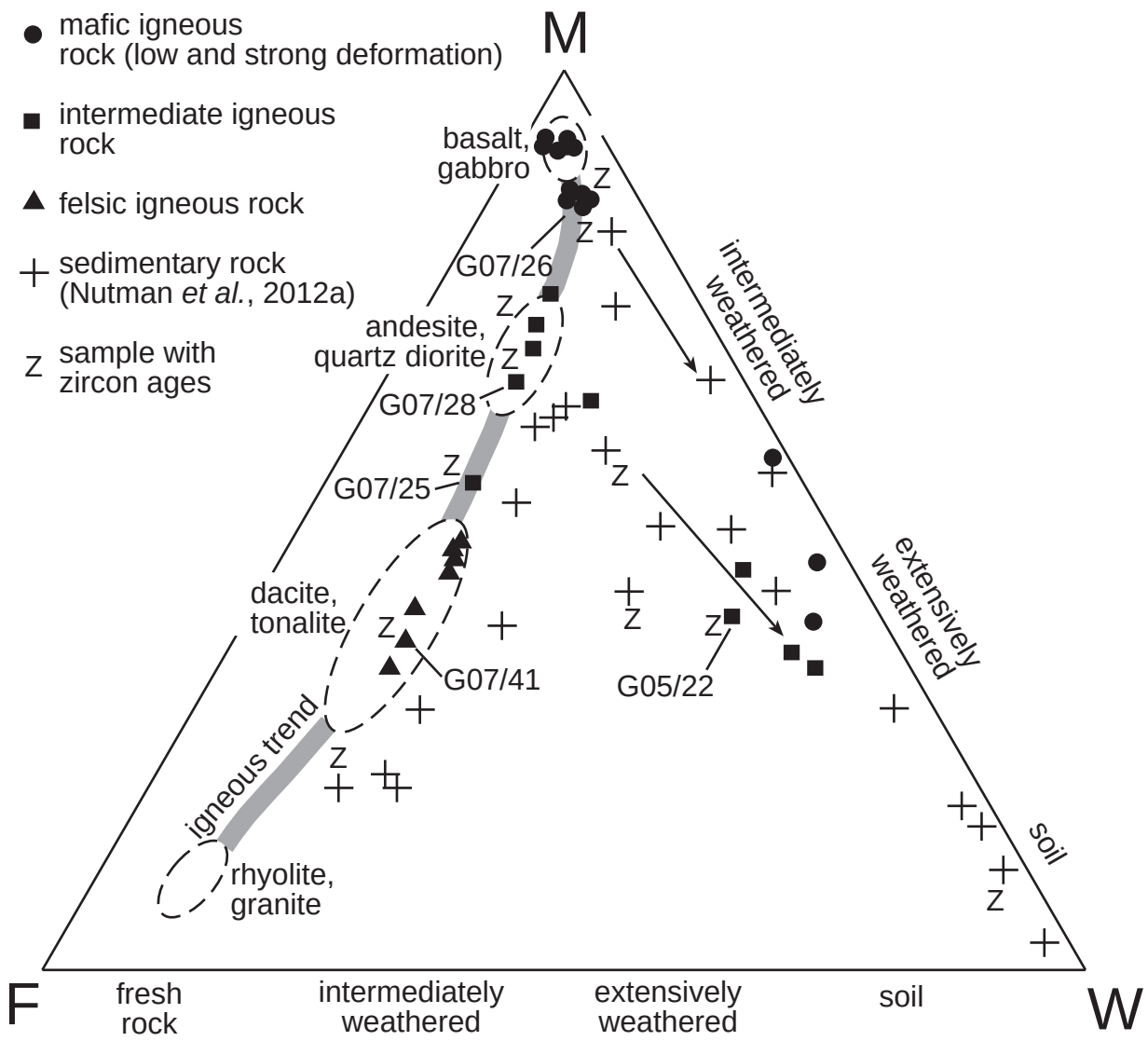
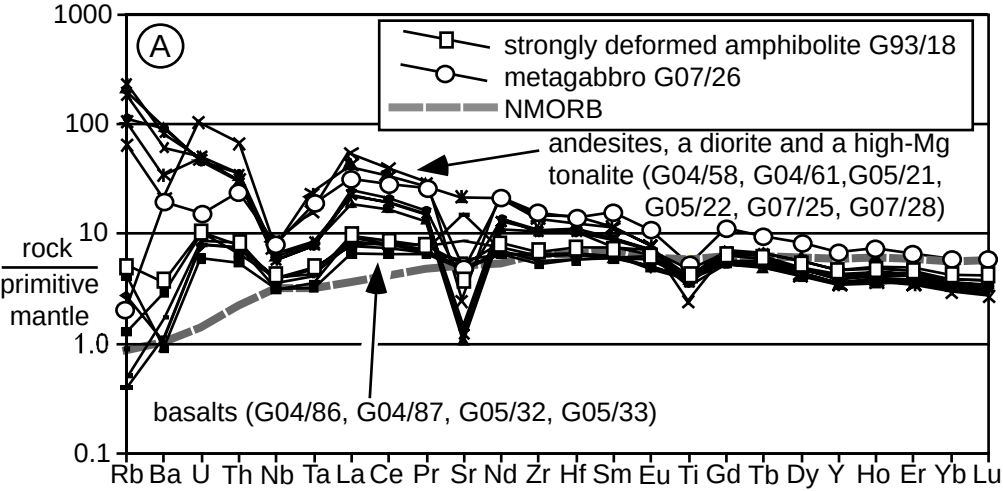
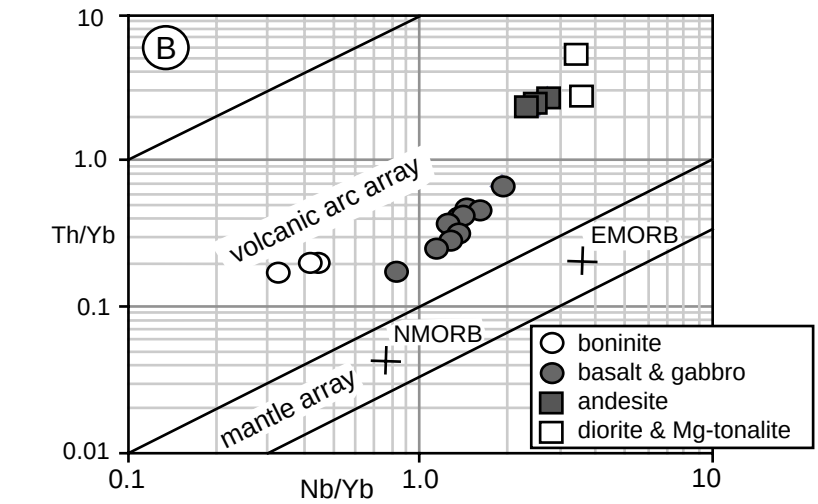
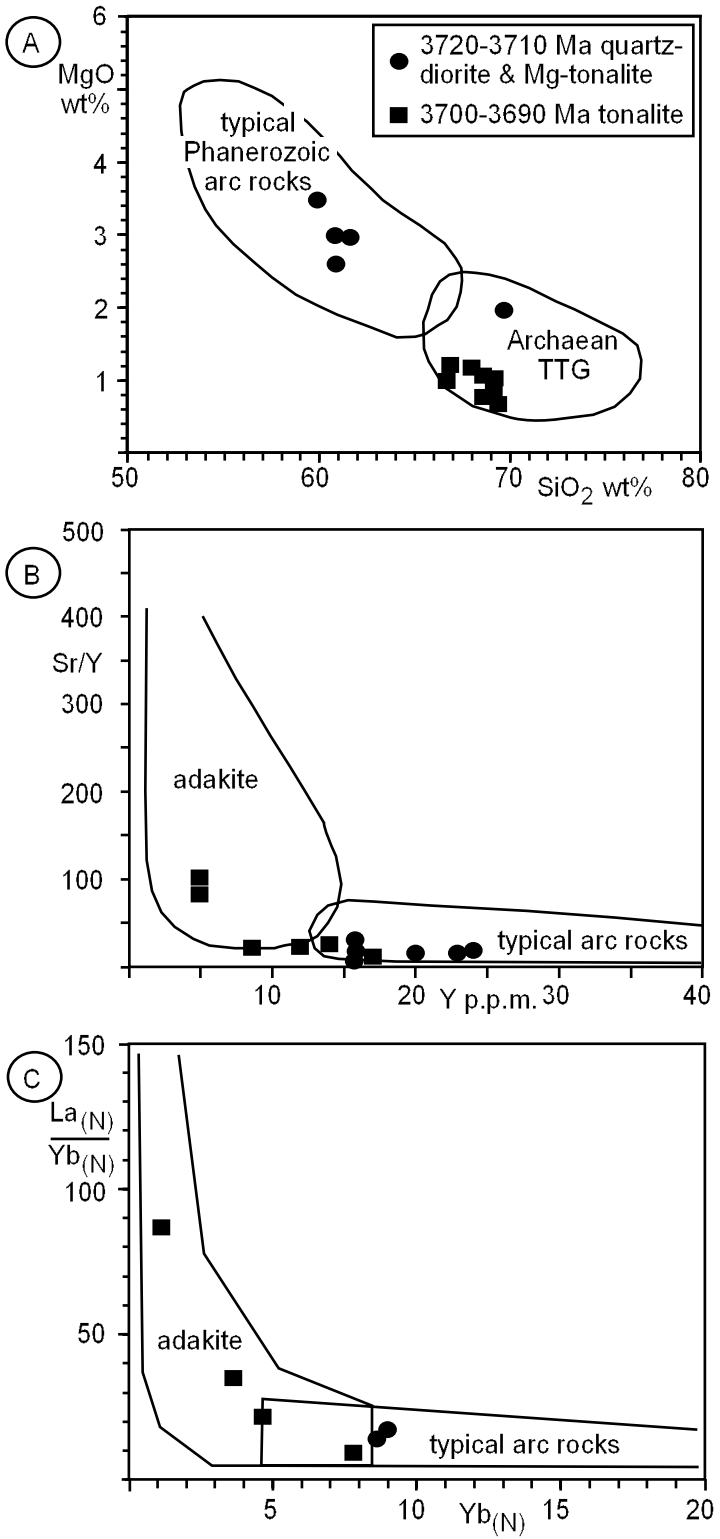
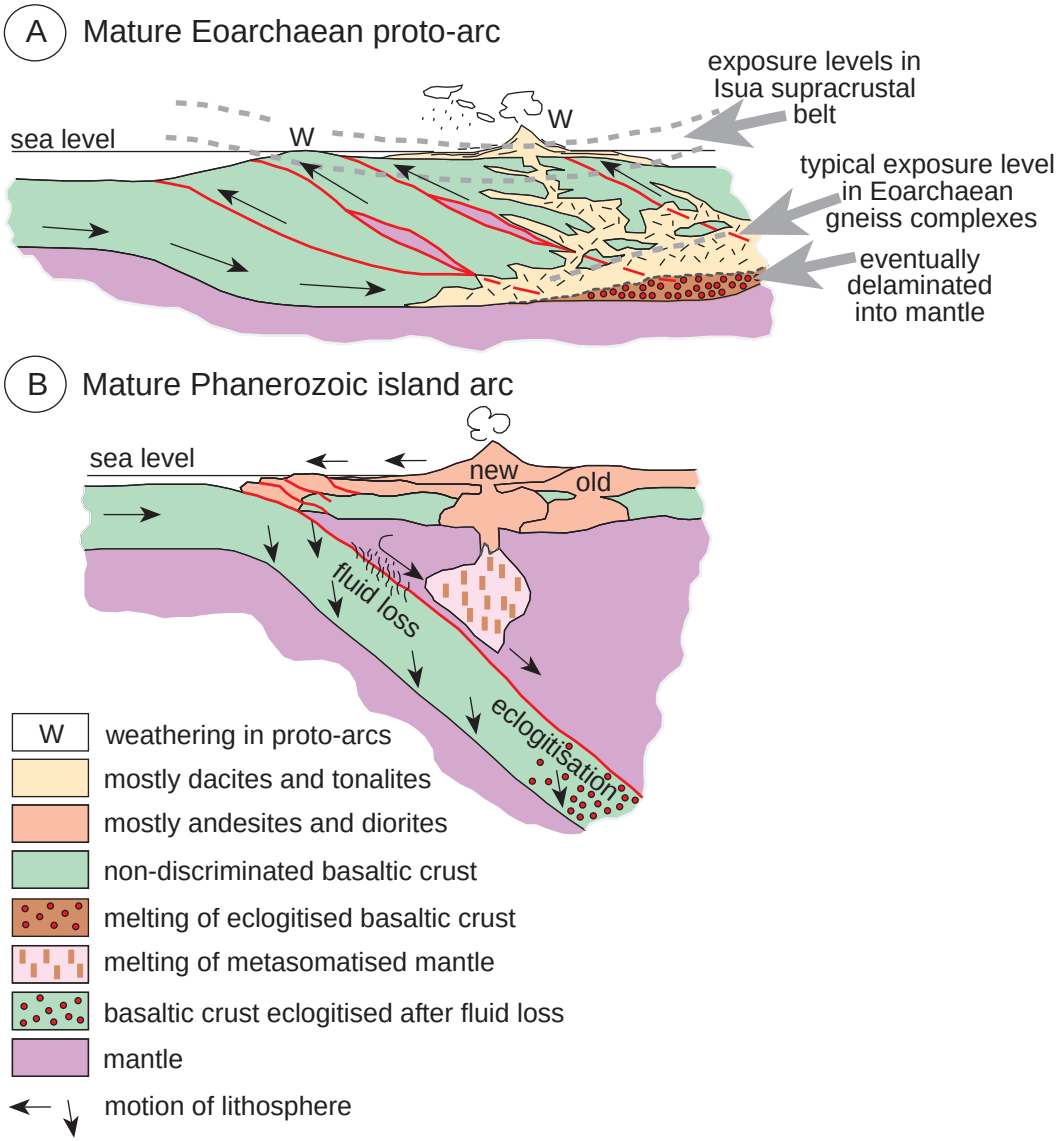

northern gneisses; 3700-3690 Ma tonalites
3686±6 Ma
≤3705 Ma
3691±6 Ma
3704±13 Ma (Kamber et al. 2005)
lake
≥3717 Ma
2214±10 Ma
3699±12 Ma
3643±6 Ma
3511±18 Ma
limit of 2004-2005 remapping
dividing sedimentary unit c. 3750 Ma
3690 Ma
3709±9 Ma
≥3717 Ma
65°10'N
500 m
N
area of extensive poor exposure
c. 3800 Ma
3806±3 Ma
3798±7 Ma
lake
southern gneisses; 3810-3795 Ma tonalites
49°50'W
glacier
3798±7 Ma zircon date
compositional layering (inset)
2214 Ma mafic dyke
3510 Ma mafic dykes
3643 Ma mafic dyke
boninitic amphibolites
IAB amphibolites and andesites
rare pillow structure relicts show that volcanic rocks are important protoliths
felsic/mafic schists - rare volcanic structure relicts show volcanic origin
chemical sedimentary rocks
ultramafic rocks
orthogneisses
Eoarchaean boundaries:
tectonic
intrusive
stratigraphic


mafic igneous rock (low and strong deformation)
intermediate igneous rock
felsic igneous rock
sedimentary rock (Nutman et al., 2012a)
Z sample with zircon ages
M
basalt, gabbro
andesite, quartz diorite
dacite, tonalite
rhyolite, granite
igneous trend
G07/26
G07/28
G07/25
G07/41
G05/22
intermediately weathered
extensively weathered
soil
F
fresh rock
intermediately weathered
extensively weathered
soil
W


A
strongly deformed amphibolite G93/18
metagabbro G07/26
NMORB
andesites, a diorite and a high-Mg tonalite (G04/58, G04/61,G05/21, G05/22, G07/25, G07/28)
basalts (G04/86, G04/87, G05/32, G05/33)
rock / primitive mantle
1000
100
10
1.0
0.1
Rb Ba U Th Nb Ta La Ce Pr Sr Nd Zr Hf Sm Eu Ti Gd Tb Dy Y Ho Er Yb Lu


B
volcanic arc array
EMORB
NMORB
mantle array
boninite
basalt & gabbro
andesite
diorite & Mg-tonalite
Th/Yb
Nb/Yb
10
1.0
0.1
0.01
0.1
1.0
10


A
MgO wt%
6
4
3
2
1
50
60
70
80
SiO2 wt%
3720-3710 Ma quartz-diorite & Mg-tonalite
3700-3690 Ma tonalite
typical Phanerozoic arc rocks
Archaean TTG
B
Sr/Y
500
400
300
200
100
10
20
30
40
Y p.p.m.
adakite
typical arc rocks
C
La(N)/Yb(N)
150
100
50
5
10
15
20
Yb(N)
adakite
typical arc rocks


A
Mature Eoarchaean proto-arc
exposure levels in
Isua supracrustal
belt
sea level
W
W
typical exposure level
in Eoarchaean
gneiss complexes
eventually
delaminated
into mantle
B
Mature Phanerozoic island arc
sea level
new
old
fluid loss
eclogitisation
W
weathering in proto-arcs
mostly dacites and tonalites
mostly andesites and diorites
non-discriminated basaltic crust
melting of eclogitised basaltic crust
melting of metasomatised mantle
basaltic crust eclogitised after fluid loss
mantle
motion of lithosphere