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Formation of Earth's early Archaean continental crust

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Abstract

Subduction of oceanic crust at an unusually low-angle has been proposed as a model for the growth of continental crust older than about 2.5 Ga. At modern zones of low-angle-, or flat-subduction, magmatic additions to new crust come from partial melting of both the subducting oceanic crust (slab) and the thin wedge of mantle above the slab. Evidence for both a slab and wedge source is preserved in most late Archaean (3.0–2.5 Ga) terrains, but we find little evidence that a mantle wedge contributed to crustal growth prior to ∼3.1 Ga. This lack of evidence in part reflects a dearth of exposed crust aged between 3.0 and 3.3 Ga, but also suggests that subduction enriched mantle source regions did not develop before ∼3.3 Ga and possibly not before 3.1 Ga. In contrast to most modern terrains and some late-Archaean terrains, early Archaean (>∼3.3 Ga) continental crust evolved through direct melting of thick mafic crust. We invoke a process of subduction that does not include the development of a mantle wedge, and call this process *Archaean flat-subduction* to distinguish it from modern low-angle subduction.

Keywords: Archaean; Crustal evolution; Magmatism; Low-angle subduction; Enriched mantle

1. Introduction

Opinion on how Earth's earliest continental crust evolved is strongly divided. A popular uniformitarian view is that it formed in the Archaean (>2.5 Ga) as it has today (e.g. Lowe, 1994; Kusky and Polat, 1999), primarily through accretion of volcanic arc material. However, features that characterise modern convergent margins (e.g. high-pressure metamorphic belts, ophiolites, etc.) are absent from much of the Archaean record (e.g. Condie, 1997) and this has led to an alternative view that modern-style plate margin processes played little role in Archaean crustal evolution (Hamilton, 1998). It has also been suggested that crust-forming processes evolved throughout the Archaean (e.g. Davies, 1995, 1998; De Wit, 1998). While brittle and partially hydrated oceanic plates might have both existed and interacted in the early Archaean (here taken as ∼3.3 Ga and older), modern-style subduction and arc accretion processes may only have emerged towards the late Archaean (De Wit, 1998). Earlier crust may have evolved through melting of very thick, or perhaps stacked, mafic crust (Davies, 1995, 1998; De Wit, 1998).

It is widely believed that the Archaean mantle was hotter than modern mantle and underwent more extensive melting, which would have resulted in oceanic crust that spread at a higher rate and was thicker, warmer and more buoyant than modern oceanic crust—and would have strongly resisted subduction (e.g. Abbott and Hoffmann, 1984; Bickle, 1986; Hoffman and Ranalli, 1988; Abbott et al., 1994). An

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emerging view is that if subduction occurred in the Archaean then faster moving, more buoyant oceanic crust subducted at a much lower angle than it typically has done in modern subduction zones (e.g. Abbott and Hoffmann, 1984; Martin, 1986; Abbott et al., 1994).

Oceanic crust at today's convergent margins is typically subducted at a *steep* angle of ≥30° (e.g. Gutscher et al., 2000a), with extreme inclinations of 70° recorded at some subduction zones (e.g. Vanuatu—Peate et al., 1997). Subduction at angles <30° is less common and has been variously referred to as *low-angle*-, *shallow*- or *flat*-subduction (e.g. Jarrard, 1986; Abbott et al., 1994; Gutscher et al., 2000a, b). Conditions favouring truly *flat*-subduction, where the slab actually approaches a horizontal plane, occur at only ∼10% of today's convergent margins (Abbott et al., 1994; Gutscher et al., 2000a) but that figure may have decreased from 99% at ∼3.0 Ga (Abbott et al., 1994).

At modern convergent plate margins, crustal growth has typically occurred via accretion of arc-related material (e.g. Rudnick, 1995) largely derived from the wedge of peridotitic mantle between the steeply subducting oceanic slab and the overriding crustal plate. This mantle wedge has partially melted as large ion lithophile element (LILE)-rich volatiles are released from the slab (Fig. 1). The result is basaltic and andesitic 'calc-alkaline' arc magmas enriched in LILE.

In the less common case of modern low-angle-, and flat-subduction, the slabmelts to produce silica-, Na- and LILE-rich magmas (Kay, 1978) called adakites (Fig. 1; Drummond and Defant, 1990). This

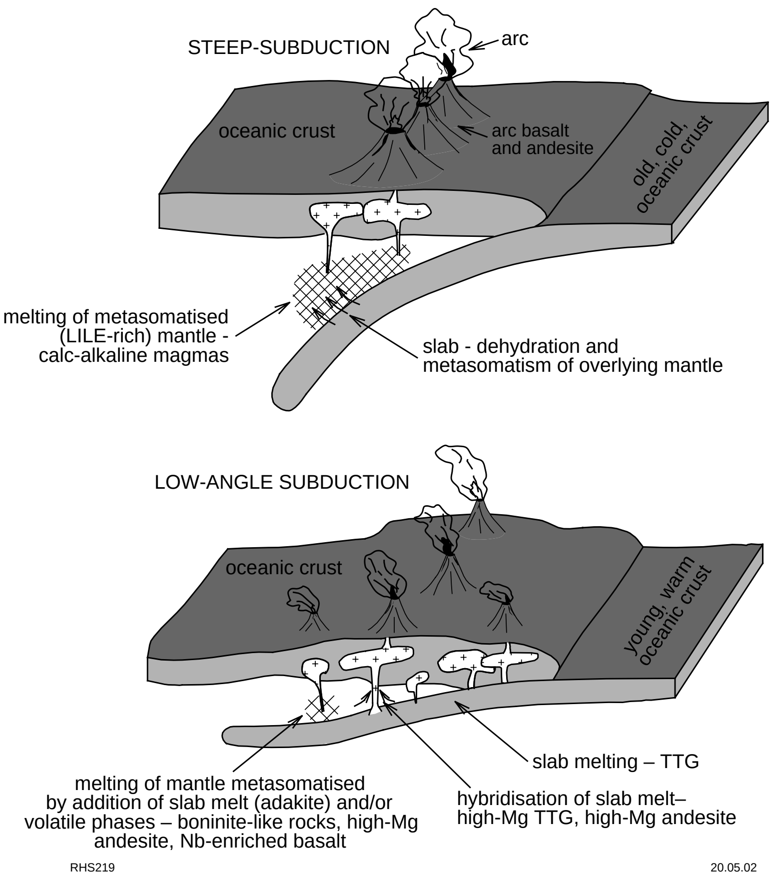

Fig. 1. Modern-style steep- and low-angle subduction, showing contrasting magma sources—note that a mantle wedge still occurs during flat- and low-angle subduction and slab-derived melts typically interact significantly with that wedge.

high-pressure melting occurs in the presence of garnet and amphibole, but commonly not plagioclase, and leaves adakite notably depleted in heavy rare earth elements (e.g. Yb) and enriched in Sr compared to calc-alkaline rocks (Drummond and Defant, 1990). Gutscher et al. (2000b) suggest that the main cause of very low subduction angles is the subduction of unusually thick (plateau) oceanic crust, but in some cases it is more likely the result of subduction of anomalously young and hot oceanic crust (e.g. Drummond and Defant, 1990; Defant and Drummond, 1993). Defant and Kepezhinskas (2001) note that some adakites have been produced during fast subduction (i.e. with high convergence rates), which, according to Jarrard (1986) and Gutscher et al. (2000b), should typically also be at a lower angle. Other mechanisms that may lead to adakite production, but that don't necessarily involve low-angle-, or flat-subduction, include arc–arc collision, the initiation of subduction, or slab-tearing (Defant and Kepezhinskas, 2001). Importantly, however, Gutscher et al. (2000a) noted that ~80% of modern zones of flat-subduction are linked to adakite magmatism.

Thus, a low-angle-, or flat-subduction model predicts that adakite-like rocks should be very common in the Archaean, and this is certainly the case. Granitic rocks with adakite-like compositions form the tonalite-trondhjemite-granodiorite (TTG) series that volumetrically dominates many areas of Archaean crust (Martin, 1987; Drummond and Defant, 1990). The complimentary dearth of calc-alkaline arc ('wedge-derived') basalt and andesite, particularly in early Archaean terrains, is consistent with highly buoyant Archaean oceanic crust, and suggests that steep-subduction was unlikely to have been an important process throughout much of the early Archaean.

In modern-day slab-derived magmas, high $Mg^{\#}$ ($Mg/(Mg + Fe_{tot})$), Cr and Ni provide strong evidence for interaction with a mantle wedge, and this is most obvious in high-Mg andesite, Mg-rich adakite, and adakite−Nb-enriched basalt associations (Kay, 1978; Defant and Drummond, 1993; Mahlburg Kay et al., 1993; Yogodzinski et al., 1995; Kelemen, 1995; Kepezhinskas et al., 1997; Rapp et al., 1999). The mantle wedge, therefore, is still an important contributor to arc-related crustal growth during modern low-angle- and flat-subduction. In this paper we show that, unlike the case at modern subduction (steep, low-angle or flat) settings, there is very little evidence for any interaction between slab-derived magmas and mantle material prior to c. 3.1 Ga. One reason for this might be that styles of subduction involving a mantle wedge, including low-angle- and flat-subduction, may not be appropriate for early Archaean crustal growth. We propose an alternative process whereby one slice of oceanic crust is subducted (or thrust) beneath another, in a way that totally excludes a mantle wedge. We refer to this process as *Archaean flat-subduction*, both to distinguish it from modern flat-subduction and to emphasize that subduction of hydrated oceanic crust remains the fundamental process.

2. Mantle-wedge interaction and the Archaean rock record

2.1. Mg-rich adakite

Adakites and TTGs share characteristics such as high Al_2O_3, Sr, Na_2O/K_2O and La/Yb and low Yb, which suggest that both formed through high-pressure melting of basaltic crust (e.g. Drummond and Defant, 1990; Martin, 1999). Modern-day adakites are restricted to convergent margins, so there has been a strong tendency to link TTGs to analogous subduction environments in the Archaean (e.g. Drummond and Defant, 1990; Martin, 1987, 1999; Drummond et al., 1996).

Very few modern-day adakites are purely high-pressure melts of basaltic slabs. Their ascent through a peridotitic mantle wedge results in interaction with the wedge, and this is clearly demonstrated by systematic shifts to higher $Mg^{\#}$ and lower SiO_2 (Kay, 1978; Rapp et al., 1999) (Fig. 2). Some late Archaean TTG suites show this trend, but very few, if any, pre-3.0 Ga suites show this evidence for mantle interaction, even in the most mafic end-members (Smithies, 2000) (Fig. 2).

Smithies (2000) and Smithies and Champion (2000) interpreted the $Mg^{\#}$ versus SiO_2 range for TTG to suggest that the mantle wedge contributed to petrogenesis only in the late Archaean (post c. 3.0 Ga), and even then, was not always involved. Martin and Moyen (2002) also examined the petrogenesis of TTG and considered a greater range of elements (including Ni, Cr and Sr). These authors concluded that there was actually a systematic increase in the contribution of

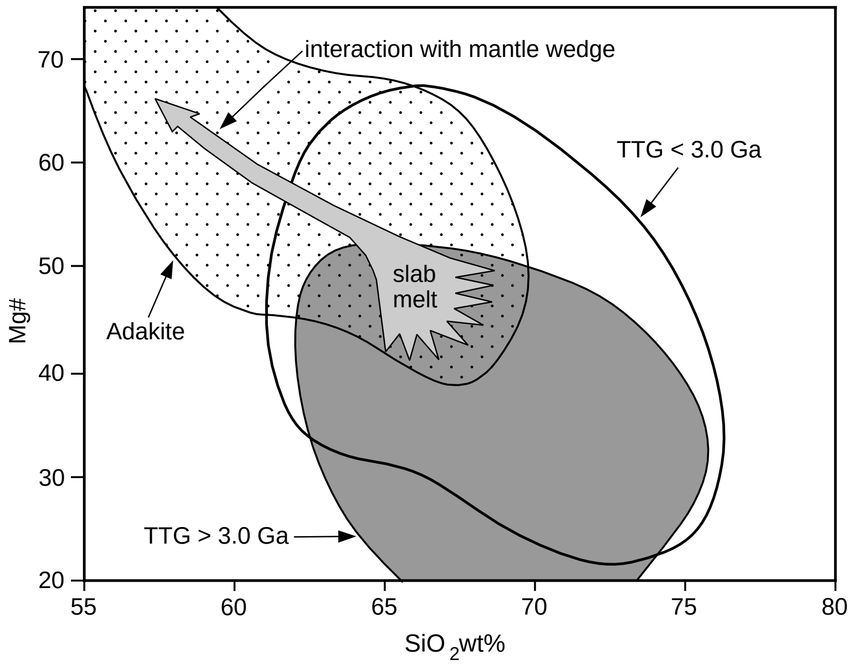

Fig. 2. Variation of Mg# with SiO_2 for modern adakites and for Archaean TTG (modified after Smithies, 2000). Broad arrow shows the typical affect of contamination of adakite by mantle. TTGs older than c. 3.0 Ga do not show this trend but form a broad field at lower Mg# and higher SiO_2. Compositional variation within that field is controlled primarily by varying degrees of melting of a hydrated mafic source and by plagioclase and hornblende fractionation.

mantle peridotite to TTG petrogenesis from the early to late Archaean. Their data, however, can also be interpreted to show that the concentrations of Cr and Ni do not increase with decreasing age until c. 3.0 Ga, and show a marked increase after 3.0 Ga (although it must also be noted that there are relatively few data for Ni and Cr compared to the other elements that Martin and Moyen (2002) considered). Similarly, with the addition of data for the 3.45 Ga primitive TTGs from the Shaw Granitoid Complex of the Pilbara Craton (Sr up to 900 ppm—Bickle et al., 1983, 1993), the concentration of Sr can also be interpreted to show no increase with decreasing age from c. 3.5 Ga until c. 2.75 Ga, with a marked increase after that (Fig. 3). Consequently, an alternative interpretation of the data presented by Martin and Moyen (2002) is that they reflect little, if any, TTG-mantle interaction before c.3.0 Ga, but that such interaction rapidly became important in the late Archaean.

Defant and Kepezhinskas (2001) have suggested that one reason why Archaean TTG suites show little or no evidence for interaction with a mantle wedge might be that the very high volumes of melt driven off the thick slab swamped the thin mantle wedge. This might explain why low-Mg# (Cr, Ni) TTGs dominate early Archaean crust but not why high-Mg# TTGs are virtually absent. Periods of low slab-melt/mantle peridotite ratio, such as at the early or late stages of slab-melting, or at marginal zones between asthenospheric mantle and the adakite-swamped mantle wedge, should have provided an opportunity for extensive interaction of slab-melts with the mantle wedge. Such interaction may have also been expected at the cessation of subduction as asthenospheric mantle moved back to reclaim the wedge, and enriched mantle created in this way would have been a suitable source for subsequent sanukitoid magmatism (see below). There is very little evidence that this occurred in the early Archaean.

An important conclusion of Martin and Moyen (2002) was that the pressure at which TTGs were produced was much lower before c. 3.5 Ga than in

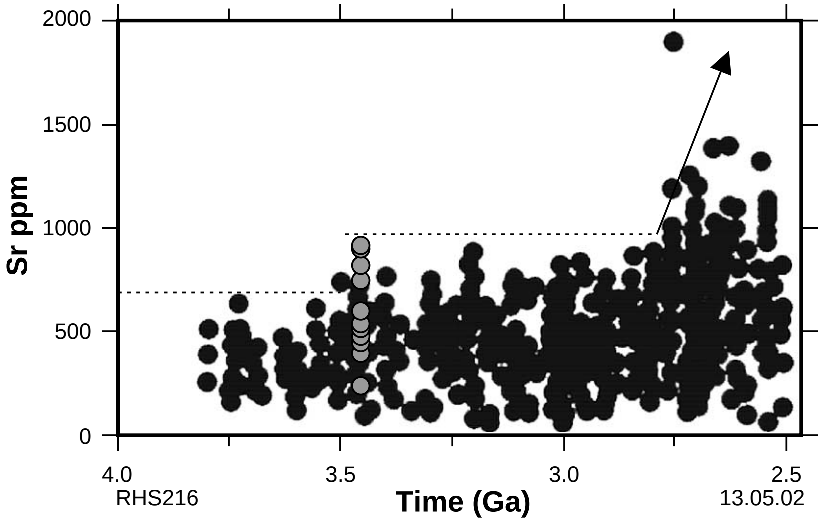

Fig. 3. Plot showing variation in Sr concentration in TTGs (grey dots) throughout the Archaean. This diagram has been modified after Martin and Moyen (2002) by the addition of data for c. 3.45 Ga TTGs from the Pilbara Craton (Bickle et al., 1983, 1993).

the late Archaean. This is indicated by significantly lower Sr concentrations (<600 ppm) in rocks older than 3.5 Ga, reflecting a plagioclase-rich residual assemblage in their genesis (Fig. 3). Consequently, regardless of the subduction angle, pre-3.5 Ga subducted slabs may have melted at a level shallow enough to have effectively precluded any interaction between slab-melts and mantle (Martin and Moyen, 2002). This melting is envisaged to have occurred at shallower depths than the peridotite solidus in the corresponding mantle wedge, although that solidus may also have occurred at a shallower level in a hotter early Archaean mantle than it does in today's mantle. Importantly, when data for the 3.45 Ga TTGs from the Pilbara Craton (Bickle et al., 1983, 1993) are added to the Sr versus time plot of Martin and Moyen (2002), it appears that between 3.5 Ga and c. 2.75 Ga, maximum Sr concentrations were not only much higher than those for pre-3.5 Ga TTGs, but they remain rather constant at ∼900–1000 ppm (Fig. 3). These concentrations are higher that average modern-day adakite values of ∼700 ppm (e.g. Drummond et al., 1996; Martin, 1999). Thus, the source for post- 3.5 Ga TTGs probably contained less residual plagioclase, and melting typically occurred at a deeper level than was the case for the >3.5 Ga TTG, but there is still no clear evidence that TTG magmas interacted within mantle peridotite until ∼3.0 Ga.

2.2. High-Mg andesites

High-Mg andesite is rare in the Archaean, but the intrusive equivalent, high-Mg diorite or sanukitoid, forms a minor but widespread component of most late Archaean terrains. High Mg#s (commonly > 60) and high Cr and Ni contents require a mantle source but high LILE concentrations are believed to reflect either a subduction-modified source or mantle contamination of a slab-melt (Shirey and Hanson, 1984; Evans and Hanson, 1997; Rapp et al., 1999; Smithies and Champion, 2000).

Average Archaean continental crust also has a LILE-rich composition similar to that of high-Mg andesite (Kelemen, 1995), and this has been used to argue that interaction between slab-melts and a mantle wedge played a key role in forming Archaean continental crust (Shimoda et al., 1998; Tatsumi, 2001). However, where high-Mg diorite is recognised, it typically forms <5% of outcrop, and it is too rich in Th, U and K to represent magmas from which the more common Archaean felsic rocks (TTG) evolved. Furthermore, virtually all documented high-Mg diorites are younger than 3.0 Ga (Smithies and Champion, 2000) (the 3.2 Ga Kaap Valley Pluton in South Africa (Robb et al., 1984) shares some features of high-Mg diorite (Smithies, 2000) and may be a sole pre-3.1 Ga example of such rocks). Also, it is more likely that the high-Mg andesitic composition of Archaean crust merely reflects the average of abundant Archaean rock-types—TTG, basalt and komatiite (e.g. Taylor and McLennan, 1985). High-Mg diorite magmatism does indicate that the mantle wedge may have contributed to the growth of Archaean continental crust, but the contribution was very small ($\ll$5%) and this evidence is restricted to the late Archaean.

2.3. Boninites

Boninites have low Ti, and Gd/Yb and high Al_2O_3/TiO_2 reflecting a refractory source, but high LILE concentrations and high La/Gd suggest subsequent source enrichment (e.g. Sun and Nesbitt, 1978; Crawford et al., 1989). Only five cases of Archaean rocks with these signatures have been documented—from the c. 3.7 Ga Isua greenstones of southern Greenland (Polat et al., 2002), c. 3.5 Ga greenstones of the Barberton region of South Africa (Parman et al., 2001), the c. 2.95 Ga Mallina Basin in the central Pilbara Craton (Smithies, 2002) and the c. 2.7 Ga greenstones of the Abitibi (Kerrich et al., 1998; Wyman, 1999) and Opatica (Boily and Dion, 2002) regions of Canada. A further example of such rocks might be high La/Gd, Yb/Gd basalts recently identified within the c. 3.12 Ga Whundo Group in the western part of the Pilbara Craton (work in progress—Geological Survey of Western Australia and Geoscience Australia)

The challenge with Archaean rocks of boninite-like composition is to identify compelling evidence that the observed LILE and LREE enrichments truly reflect an enriched mantle source rather than simply crustal contamination during ascent. Because both enrichment processes include crustal material, distinguishing between the two may be hard or impossible based on the geochemistry of the boninite-like rocks alone. The Abitibi suites, however, appear to have formed in an oceanic environment and so their LILE-enrichments are unlikely to be the result of crustal assimilation (Kerrich et al., 1998; Wyman, 1999), and the composition of exposed crust in the Pilbrara Craton cannot explain LILE- and LREE-enrichments in the Mallina rocks (Smithies, 2002). The petrogenesis of boninite-like rocks of the 3.12 Ga Whundo Group has not yet been evaluated, although they form the lowest part of a mixed sequence of basalts with both ocean floor and calc-alkaline compositional characteristics (work in progress—Geological Survey of Western Australia and Geoscience Australia).

Although a subduction enriched source has been suggested for the early Archaeaen boninite-like rocks of the Isua and Barberton greenstone belts (Parman et al., 2001; Polat et al., 2002), the possibility that these rocks have simply assimilated felsic crust can not be discounted. Parman et al. (2001) suggest that crustal contamination of the boninite-like komatiites of the Barberton region lacks support because the nature of the crust upon which the magmas were erupted is not known and because the magmas themselves contain no crustal xenoliths. However, crustal xenoliths are not frequently described from komatiites, even from those shown to be contaminated (e.g. Arndt and Jenner, 1986). Also, the early Archaean evolution of the Barberton region and of the eastern part of the Pilbara Craton (Western Australia) is remarkably similar (e.g. Zegers et al., 1998), and the recognition that c. 3.5 Ga mafic and ultramafic rocks of the latter were erupted through, and contaminated by, felsic crust (e.g. Green et al., 2000), clearly indicates that a similar fate for the Barberton boninite-like komatiites should be properly evaluated.

Polat et al. (2002) showed that the boninitic signature of the Isua boninite-like rocks can be reproduced simply by assimilation of slightly older, regionally available TTG crust. However, the exposed geology provides no evidence for a direct relationship between the boninite-like rocks and the TTG crust, and so Polat et al. (2002) invoke a modern-style subduction setting with a mantle source region enriched through interaction with TTG-like slab-melt. It is worth noting, however, that c. 3.8 Ga TTGs show the least evidence for mantle interaction (Martin and Moyen, 2002), a feature contrary to that expected if the mantle source component of the boninite-like rocks was indeed enriched by TTG-like slab-melts. It is also important to note that the Isua greenstones outcrop over an area of $<140\,km^2$ and form one of the most deformed Archaean sequences. Cas et al. (2001) show that rock relationships in the area are extremely difficult, if not impossible, to confidently assess. Consequently, any relationship between local TTG crust and the boninite-like rocks might be impossible to either prove or disprove.

2.4. Nb-enriched basalt

Nb-enriched basalt combines high LILE concentrations with Nb and Ta concentrations (and Nb/La ratios) higher than those in typical arc- and continental basalts. Many modern examples of flat-subduction show a strong association between adakites and Nb-enriched basalts (Defant and Drummond, 1993; Drummond et al., 1996), and these basalts are attributed to melting of a mantle wedge that has been contaminated by slab-melt (Kepezhinskas et al., 1997; Sajona et al., 2000). Wyman et al. (2000) note that the common recognition of Nb-enriched basalt within Archaean terrains would provide good evidence for ancient mantle-wedge processes, but there are only two documented occurrences (Wyman and Hollings, 1998; Wyman et al., 2000); both are from the Superior Province and are 3.0 Ga or younger.

3. low-angle-, and flat-subduction in the Archaean?

Low-angle-, and flat-subduction has been proposed as a model for Archaean crustal growth, based primarily on the suggestion that Archaean oceanic crust typically met the physical requirements (i.e. was thick, warm and buoyant), and on the observation that Archaean TTGs are compositionally similar to felsic partial melts of recently subducted basaltic crust (e.g. Drummond and Defant, 1990; Martin, 1999). This model would explain the abundance of TTGs (as slab-melts) and the complementary dearth of rocks similar to modern-day, calc-alkaline arc products. However, even in modern zones of low-angle-, and flat-subduction there is still significant evidence that the mantle wedge has contributed to arc formation by way of a range of Mg- and LILE-rich magmas reflecting interaction between slab-melts and the wedge (e.g. Kay, 1978; Drummond and Defant, 1990; Kelemen, 1995; Kepezhinskas et al., 1997).

For the late Archaean, the best evidence for the development of a subduction-modified mantle comes from the Abitibi subprovince (Superior Province) and the central part of the Pilbara Craton. Both contain boninite-like rocks and high-Mg andesites (sanukitoids) while the Abitibi subprovince also contains Nb-enriched basalt and TTGs that show evidence of mantle interaction (e.g. high $Mg^{\#}$, Cr, Ni: Feng and Kerrich, 1992).

Convincing evidence of a subduction-modified mantle is considerably harder to find in the igneous rock record of terranes that are older than ∼3.1 Ga, although the boninite-like rocks of the Barberton and Isua greenstone belts must be viewed as possible evidence. At least for the period between 3.1 and 3.3 Ga, this may in part reflect a relative dearth of preserved outcrop. However, for the early Archaean, and possibly also the period between 3.1 and 3.3 Ga, we suggest that either the mantle was not frequently enriched through recycling of crust at subduction zones, or that such ‘enriched’ sources made no contribution to crustal growth at subduction zones. Either case contrasts strongly with processes at most modern convergent margins, including low-angle-, and flat-subduction zones, where subduction modified mantle wedge has provided a primary source for new or recycled crust. Modern-style subduction processes, including flat-subduction, are not thoroughly appropriate models for the evolution of early Archaean crust, and are probably not even universally applicable for late Archaean crustal evolution.

4. Discussion—towards a better model

Thermal considerations indicate that early Archaean mafic crust was typically too thick, warm and buoyant to subduct steeply, if it could have subducted at all (e.g. Abbott and Hoffmann, 1984; Hoffman and Ranalli, 1988; Abbott et al., 1994). The estimated thickness of Archaean oceanic crust ranges between ∼15 and 45 km (Bickle, 1986; Abbott et al., 1994; Ohta et al., 1996), significantly thicker than typical modern oceanic crust (∼7 km—e.g. Hoffman and Ranalli, 1988), and thicker than most modern oceanic plateau crust (15–20 km—Gutscher et al., 2000b). Even on the modern Earth, oceanic plateau crust is very difficult to subduct and may instead be obducted or even ‘injected’ (i.e. tectonically underplated) into the lithosphere (Tarney, 1992). If oceanic crust was subducted in the early Archaean, it would have almost certainly have done so at a very low-angle. The actual mechanism of subduction may also have differed from those that are believed to drive modern flat-subduction. According to Gutscher et al. (2000a), in the rare case of modern flat-subduction the typically steep angle of subduction shallows to a near horizontal trajectory as normal thickness oceanic slab is replaced by thicker oceanic (including plateau) crust. In the Archaean, however, it is unlikely that any form of subduction was typically facilitated by a preceding phase of steep subduction. We suggest that buoyant Archaean oceanic crust was not dragged into a subduction zone, but was pushed by very high spreading rates (e.g. Hoffman and Ranalli, 1988), with one edge thrust beneath, or even into, the other. This model excludes a mantle wedge altogether and is similar in some respects to earlier models proposed by Wilks (1988), De Wit (1998) and Davies (1992). The process represents underthrusting or tectonic underplating, and we refer to it here as 'Archaean flat-subduction'. Under these circumstances, compositional differentiation of earliest mafic crust can only have proceeded through direct internal melting of the thickened crust.

Two constraints on the evolution of early crust are that the voluminous TTG suites require a considerably more voluminous mafic source, and that melting of that source must have occurred at depths great enough to stabilise garnet (garnet amphibolite or eclogite) (∼40 km or more: Wyllie et al., 1997), even in the early Archaean. Zegers and van Keken (2001) estimated that melting between 3.48 and 3.42 Ga alone contributed a volume of TTG to the Pilbara Craton equivalent to a layer up to10 km thick. Up to ∼20% partial melting of hydrated basalt produces TTG-like melts (Rapp et al., 1999), so 10 km of TTG crust equates to a ∼50 km thick basaltic source region that must have resided at depths of 40 km or more. Either the crust (including a lower 50 km of basaltic material) was at least 90 km thick, or a very large amount of mafic crust was cycled through the melting zone over a ∼60 million-year period. Such estimates would obviously be even higher if the amount of partial melting of basaltic crust is less than the assumed 20%. Furthermore, the common occurrence in Pilbara granites of inherited zircons with ages older than c. 3.48 Ga (Van Kranendonk et al., 2002), and Nd-isotopic data (Fig. 4), indicate that TTG production was certainly not confined to the period between 3.48 and 3.42 Ga, and so these estimates should be considered a minimum. Indeed, seismic refraction data shows the Pilbara Craton to be between 30 and 35 km with an average density consistent with a felsic composition (Drummond, 1988). Geochronological and Nd-isotopic data suggest that the vast majority of crust in the eastern part of the Pilbara Craton was produced before c. 3.4 Ga, and has simply been recycled since then (Fig. 4 and Champion and Smithies, in prep). To generate a maximum combined 35 km thickness of TTG crust would have required a combined thickness of up to ∼170 km of mafic crust (assuming 20% melting of basaltic crust) to have been below the garnet-present melting zone before c. 3.4 Ga. Clearly, the actual amount of mafic crust below the garnet-present melting zone at any single stage was almost certainly considerably less than 170 km. However, the volume of mafic crust indicated in these rough calculations is clearly inconsistent with any form of magmatic underplating. Rather, it strongly suggests that the mafic source for the TTG was continuously processed through a garnet-present melting zone, in a way very much like oceanic crust is cycled through modern subduction zones.

Consequently, we suggest that early continental crust evolved through direct partial melting of thick (perhaps overthickened; e.g. De Wit, 1998), and continuously replenished, hydrated oceanic crust, by a process of Archaean flat-subduction, with little or no contribution from subduction-enriched mantle. One possible schematic representation of the development of thick earlier Archaean mafic crust is presented in Fig. 5. The process of Archaean flat-subduction may have been locally, or perhaps more generally, facilitated by ductile lower crustal weak zones in the thick oceanic crust or in the evolving, thickened proto-continental crust. It has been suggested that modern continental crust that is thick and hot (e.g. Basin and Range Province, Tibetan Plateau) may develop a middle- or lower-crustal viscous layer capable of significant flow on geological time scales (e.g. Ranalli and Murphy, 1987; Burchfiel et al., 1989; Block and Royden, 1990; Clark and Royden, 2000). According to Hoffman and Ranalli (1988), Archaean oceanic crust with a thickness of ≥25 km would have a similar rheological structure. We speculate that Archaean flat-subduction may, in some cases, have been initiated at lower-crustal weak zones as high ocean-plate spreading rates (Bickle, 1986; Hoffman and Ranalli, 1988) pushed thick buoyant oceanic crust together.

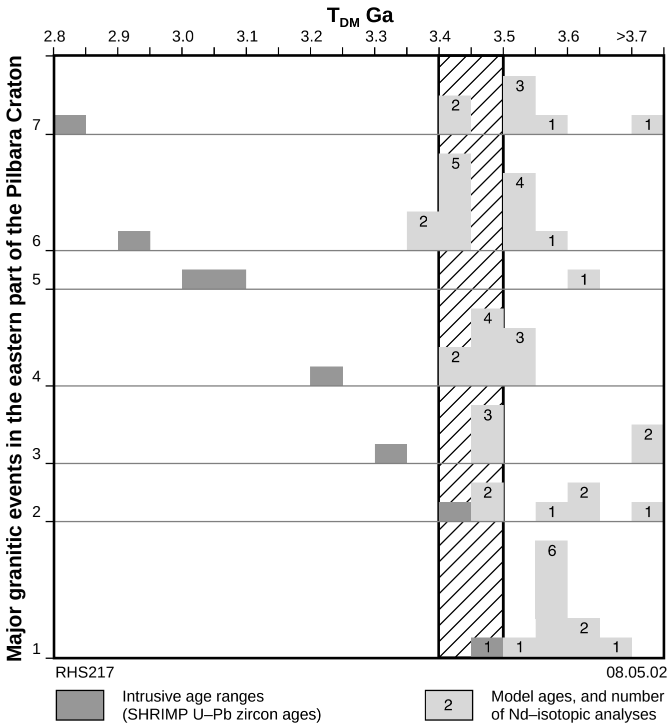

Fig. 4. Variation in the depleted mantle model ages (T_{DM}) determined on granites from seven distinct felsic intrusive periods in the eastern part of the Pilbara Craton (Nd-isotopic data—Bickle et al., 1989, 1993; Geological Survey of Western Australian and Geoscience Australia, unpublished data; and Champion and Smithies, in prep). According to Zegers and van Keken (2001), melting between 3.48 and 3.42 Ga (hatched area), contributed a volume of TTG equivalent to a layer up to10 km thick, however, it is clear from this T_{DM} plot that considerable felsic crust was also generated prior to that period.

Partial melting in the lower portion of the thick crust leads to TTG magmatism, which subsequently forms the earliest felsic continental crust (Fig. 5a). The dense, melt (TTG)-depleted material that remains (eclogite) is returned to the mantle through a process of delamination, which may proceed via a process of negative diapirism (e.g. Davies, 1995, 1998; Zegers and van Keken, 2001) or may also be initiated along lower-crustal weak zones, possibly as a result of further Archaean flat-subduction (Fig. 5b). If Archaean oceanic plates were faster moving that their modern counterparts (e.g. Abbott and Hoffmann, 1984), tectonic thickening of mafic crust may also have been rapid, with the potential to produce felsic crust at a high rate. Faster moving Archaean plates also suggest that horizontal 'ridge push' would have been a dominant force and any 'gap' created by delamination may have been readily filled both from beneath, by mantle upwelling, and from the sides by further Archaean flat-subduction (Fig. 5b). These competing processes provide both a fertile source for further TTG magmatism (i.e. cycling hydrous oceanic crust into the melting zone) and heat (upwelled mantle) to melt that source. Voluminous mafic volcanism forms a large proportion of the basal successions of early Archaean greenstones, and was a feature that occurred in several distinct stages throughout the 3.48–3.42 Ga evolution of the Pilbara Craton (Hickman, 1983; Van Kranendonk et al., 2002). These mafic rocks are intercalated with felsic volcanic rocks on a craton-wide scale, but neither mafic nor felsic rocks shows any clear evidence for mantle-wedge type processes (e.g. Green et al., 2000). Zegers and van Keken (2001) have suggested that these rocks might also be a result of delamination induced mantle upwelling events.

Our model is an attempt to explain only the relative temporal distribution of diagnostic igneous rock types throughout the Archaean, with considerably lesser at-

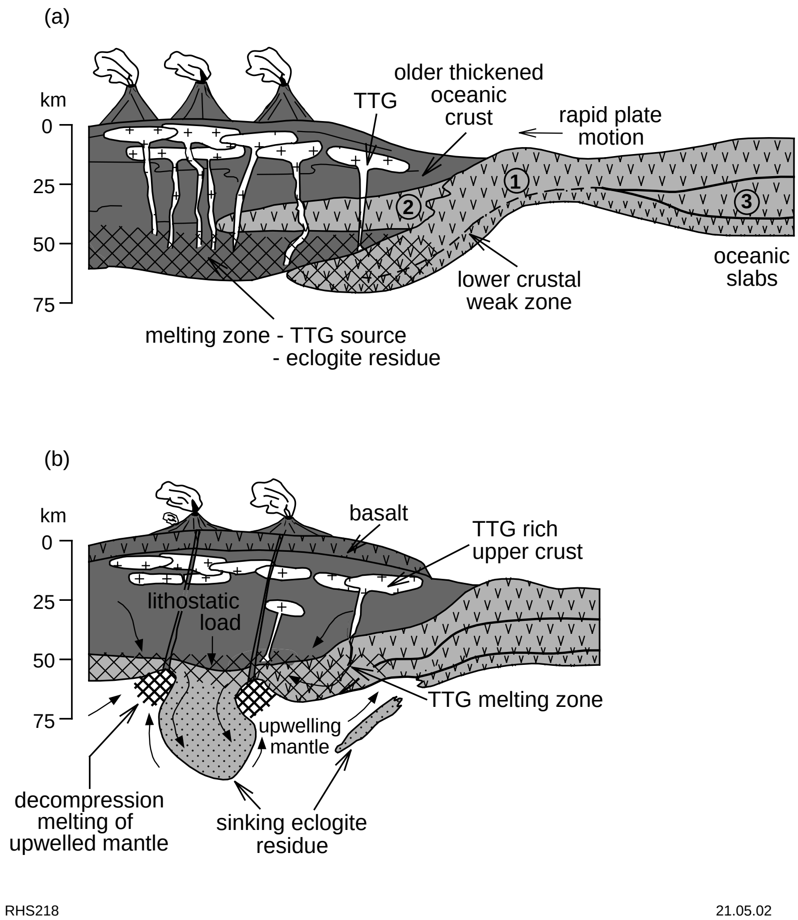

Fig. 5. Cartoon showing possible mode of early Archaean crustal evolution by Archaean flat-subduction. TTG crust is produced as the lower part of thickened mafic crust melts and is converted to eclogite. (a) Shows oceanic crust (slabs) subducted beneath (1) slightly older thickened mafic (oceanic) crust. It also shows Archaean flat-subduction initiated at lower-crustal weak zones (slabs 2 and 3) speculated to develop in thick and warm crust (see text). Eclogite is delaminated (b) and replaced by "TTG-fertile" oceanic crust from the sides and from above (Lithostatic loading). Upwelled mantle provides both heat for subsequent TTG production, and a source of basalt for the supercrustal greenstone successions.

tention paid to other lines of evidence (e.g. structure) pertaining to early crustal growth. The scenario presented here does not exclude the possibility that modern-style subduction (steep or flat, but accompanied by a wedge) occurred before c. 3.1 Ga, or indeed, before 3.3 Ga; boninite-like rocks such as those of the Barberton and Isua greenstone belts (Parman et al., 2001; Polat et al., 2002) may suggest that it might have. However, the instances of such events must have been sufficiently low that preservation of diagnostic evidence has not yet been clearly identified. The scenario is also simplistic, and it seems likely that the warmer Archaean Earth worked in at least two complimentary ways to ensure that interaction between slab-melts and mantle peridotite was not a dominant process before c. 3.1 Ga—by causing slab-melting at lower pressures (Martin and Moyen, 2002) and by producing thicker, buoyant and faster moving oceanic crust. By the end of the Archaean (∼2.5 Ga), however, oceanic crust was less buoyant and slab-melting occurred at higher pressures such that processes resembling modern-style low-angle subduction operated frequently enough to ensure that evidence was preserved in most terrains.

Many crustal growth models have suggested a significant increase in the growth rate of continental crust around 3.0 Ga (e.g. McLennan and Taylor, 1982). We suggest that the more or less coincident emergence of evidence for a subduction-modified mantle contribution to crustal growth marks a fundamental change in the actual process of crustal growth, with the beginning, in earnest, of modern-style mantle-wedge formation and contribution to juvenile crust.

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