[American Journal of Science, Vol. 310, November, 2010, P. 775–790, DOI 10.2475/09.2010.01]

American Journal of Science

NOVEMBER 2010

THE ARCHEAN-PROTEROZOIC BOUNDARY: 500 MY OF TECTONIC TRANSITION IN EARTH HISTORY

KENT C. CONDIE* and CRAIG O'NEILL**

ABSTRACT. Changes in the solid Earth at the end of the Archean fall into two categories. First are those related to cooling of the mantle and include a decrease in both komatiite abundance and MgO content, a decrease in Ni/Fe ratio in banded iron formation, and increases in incompatible element ratios (such as Nb/Yb, La/Yb, Zr/Y, La/Sm and Gd/Yb) in non-arc type basalts. A second group of changes is related to the extraction of continental crust from the mantle and stabilization of major cratons at 2.7 to 2.5 Ga. These include an increase in Nb/Th ratio and ε_{Nd}(T) of non-arc basalts; significant increases in large-ion lithophile and high-field strength elements and a decrease in Sr in continental crust, which reflect a shift in magma types from TTG (tonalite-trondhjemite-granodiorite) to calc-alkaline; a prominent increase in the maximum values of $\delta^{18}O$ of zircons from granitoids after the end of the Archean; a major peak in gold reserves is found at or near 2.7 Ga; and a peak in Re/Os depletion ages from mantle xenoliths at 2.7 Ga consistent with widespread thickening of the continental lithosphere at this time.

All of these changes may be related to the widespread propagation of plate tectonics at the end of the Archean. Subduction produces continental crust in numerous arcs, which rapidly collide to form supercratons. Oceanic slabs sinking into the deep mantle could increase the production rate of mantle plumes, as well as increase the heat flux from the core, which warms the newly arrived slabs. The cooling of the deep mantle would begin after 2.5 Ga and continue until about 2.4 Ga when a 200-My slowdown in plate tectonics begins. This may be the reason for the rapid drop in temperature of the mantle recorded by basalts and komatiites. When plate tectonics comes back on track at about 2.2 Ga, Archean supercratons break up and are dispersed.

INTRODUCTION

It is well established that the late Archean was a time of widespread preservation of continental crust, and most investigators equate this with rapid growth of one or more cratons (McCulloch and Bennett, 1994; Stein and Hofmann, 1994; Davies, 1995; Condie, 1998). However, both the tectonic regime during which this crust formed and the cause of rapid crustal growth remain controversial (Condie and Benn, 2006). From the late Archean onwards, subduction zones were probably the chief sites of continental crust production. However, today, continental crust is rapidly recycled into the mantle at subduction zones and recent estimates of recycling rates suggest that continental production rate and recycling rate are about the same magnitude (each about 2.5 km^3/yr; Scholl and von Huene, 2007). To preserve a large volume of continental crust in the late Archean, it is critical that cratons formed rapidly, protecting most of this new crust from recycling into the mantle.

* Department of Earth and Environmental Science, New Mexico Tech, Socorro, New Mexico 87801 USA; kcondie@nmt.edu

** GEMOC, Department of Earth and Planetary Sciences, Macquarie University, Sydney, NSW 2109, Australia

Although many changes have been proposed at the Archean-Proterozoic boundary, only a limited number are well documented in terms of magnitude and age (Taylor and McLennan, 1985; Condie, 2005a). Those changes related to the atmosphere-ocean-biosphere systems will be considered in a separate paper. The changes in the solid Earth fall into two groups. The first may be related to rapid cooling of the mantle after 2.5 Ga. The second group are those changes that are related to a peak in the growth and stabilization of cratons at about 2.7 Ga. Due to the sparsity of isotopic ages in the so called "crustal age gap" at 2.4 to 2.2 Ga (Condie and others, 2009), it is difficult to track the rate of changes between the late Archean and the early Paleoproterozoic. The significance of a crustal age gap between 2.4 and 2.2 Ga has been discussed by Condie and others (2009) in terms of a possible slowdown of planetary magmatism and plate tectonics.

It is the purpose of this study to briefly review those changes at the end of the Archean that are well documented in the solid Earth and to discuss their significance in terms of changing planetary dynamics at this time.

Changes Related to a Decrease in Mantle Temperature

Decrease in komatiite abundance.—Komatiites (ultramafic lavas) are thought to track the temperature of the mantle, and more specifically that part of the mantle from which mantle plumes arise (Arndt and others, 2008, 2009). The frequency of komatiites in greenstone belts as a function of age is given in figure 1. The term greenstone is used herein to refer to basalt-dominated successions of submarine supracrustal rocks. This graph, first published in De Wit and Ashwal (1997), has been updated based on new data from greenstone stratigraphic sections.

In this figure and some of the subsequent figures there are major gaps in data, especially at 2.4 to 2.2 and 1.5 to 0.8 Ga. The gap at 2.4 to 2.2 Ga may reflect a global

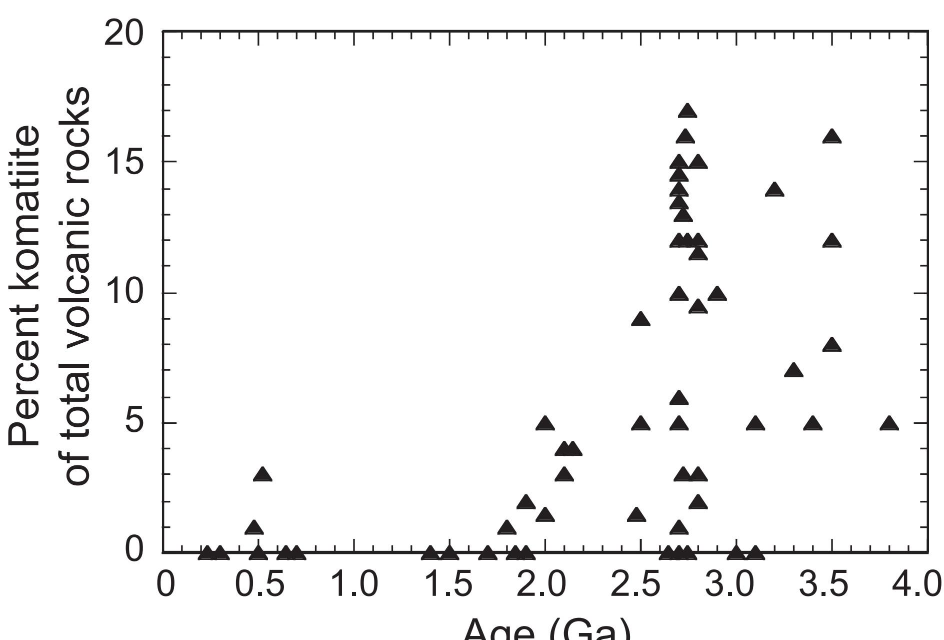

Fig. 1. Percent of komatiites in greenstones versus age. Updated after De Wit and Ashwal (1997).

slowdown in magmatism and plate tectonics as discussed by Condie and others (2009). Hence, the scarcity of igneous rocks during this time window. Because some types of geochemical data are available for rocks between 1.5 and 0.8 Ga, the absence of data in this time interval appears to reflect the absence of published data. This does not bias the conclusions of this paper, however, in that the data gap is long after the end of the Archean.

The results in figure 1 clearly show a major drop in the percent of komatiites after 2.5 Ga. An important feature of the graph is that it shows an abrupt rather than a gradual change in komatiite frequency at the end of the Archean. The decrease does not follow the calculated gradual cooling curve of the mantle as predicted by the decay of radiogenic heat sources.

Decrease in the MgO content of komatiites.—The average MgO content of komatiites is one way to track the average eruptive temperature of the magma and of the mantle sources, which are likely mantle plumes (Arndt and others, 2008). Each point on figure 2 is the average MgO content for komatiites from a given greenstone belt. The results show a clear decrease in the average MgO content of komatiites at the end of the Archean. A graph of maximum MgO content shows a similar decrease. As pointed out by Arndt and others (2008), it is unlikely that komatiites with >30 percent MgO represent pure liquids, and they probably contain cumulus olivine. However, a decrease in MgO is apparent after 2500 Ma, even ignoring the sites with >30 percent MgO. As with komatiite frequency, the decrease in MgO at the end of the Archean also reflects a decrease in mantle temperatures at this time. However, the lack of komatiite data between 2400 and 2200 Ma makes it difficult to ascertain how rapidly this change occurred after 2500 Ma.

Although most data indicate that Archean komatiites crystallized from dry magmas, some investigators propose wet magmas, which do not require as high eruption

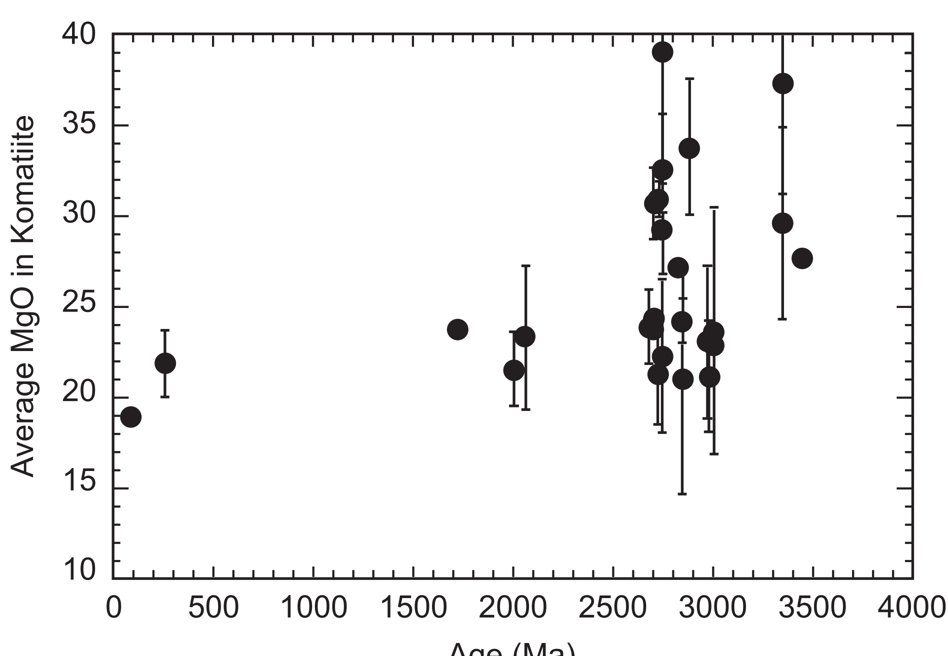

Fig. 2. Average MgO content versus age for komatiites. Data from sources given in Arndt and others (2008) and compiled by the senior author.

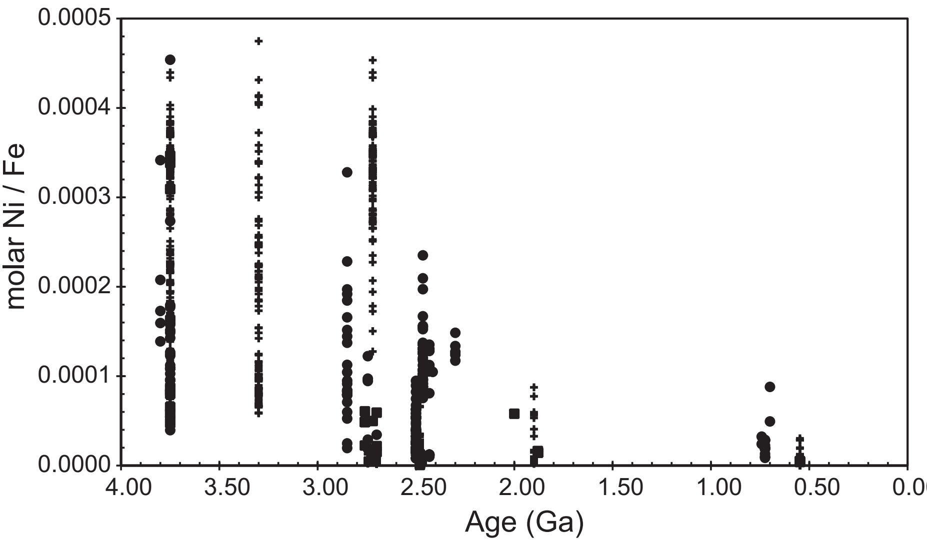

Fig. 3. Ni/Fe mole ratios in banded iron formation versus age. After Konhauser and others (2009).

temperatures as their dry counterparts (Parman and others, 2001; Grove and Parman, 2004; Arndt and others, 2008). In either case, however, the change in MgO at the end of the Archean would reflect a drop in mantle temperature, although not as large if the magmas were wet.

Decrease in Ni/Fe ratio in banded iron formation.—As recently documented by Konhauser and others (2009), the Ni/Fe ratio in banded iron formation decreases rather abruptly at the end of the Archean (fig. 3). The authors attribute this decrease to a reduced flux of Ni into the oceans as a consequence of cooling of the mantle and a corresponding decrease in the rate of eruption of komatiites. As with changes in komatiite frequency and MgO content, the drop in Ni/Fe ratio of banded iron formation attests to a probable drop in temperature of mantle sources near the end of the Archean.

Incompatible elements tracking the degree of melting.—There is a noticeable increase in incompatible element ratios such as Nb/Yb, La/Yb, Zr/Y, La/Sm and Gd/Yb in non-arc oceanic greenstone basalts at the end of the Archean (Condie, 2003, 2005b). This is illustrated with the Nb/Yb ratio in figure 4. Non-arc greenstones, which are identified using rock association and geochemistry, include ocean ridge, oceanic plateau and oceanic island basalts, some of which may have been erupted, at least locally, onto or near continental crust. Since we have not included data that may be caused by crustal contamination, the changes probably reflect a decrease in the degree of melting (Condie, 2005b; Pearce, 2008) in mantle sources, which implies a decrease in mantle temperature at the end of the Archean. Because of the scarcity of greenstones between about 2.4 and 2.2 Ga, the change in degree of melting cannot be closely monitored, but it occurs over a timeframe of no more than 300 My at the beginning of the Paleoproterozoic.

Changes Related to the Growth of Cratons

Incompatible elements and Nd isotopes.—The Nb/Th ratio in non-arc oceanic basalts has been used as a proxy for the growth of continental crust, assuming that continental crust has been extracted from the mantle leaving a depleted trace element signature in

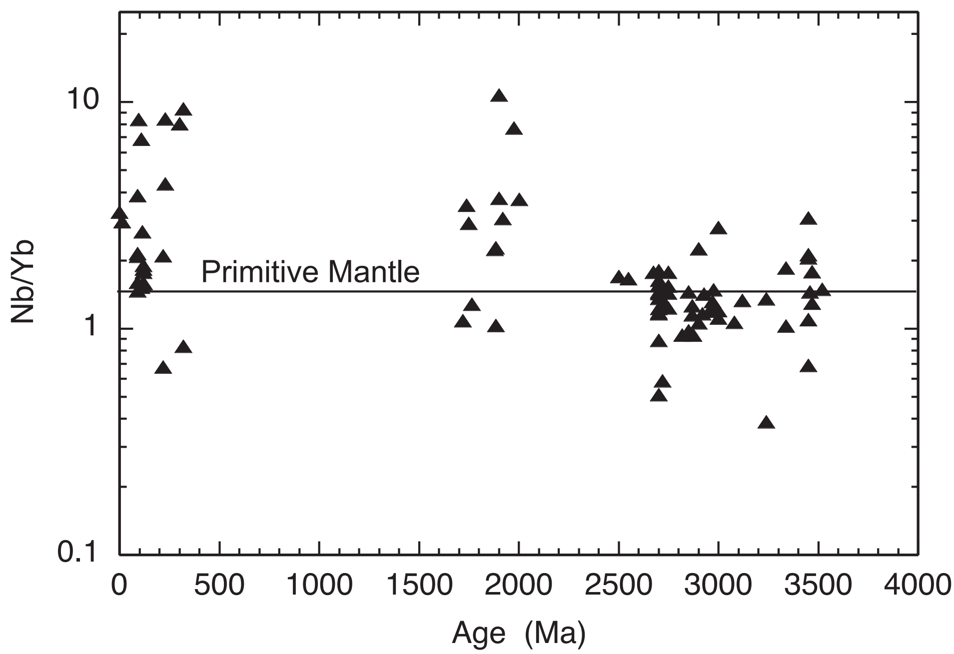

Fig. 4. Nb/Yb versus age for non-arc oceanic basalts. Updated after Condie (2003 and 2005b).

the residual mantle (Sylvester and others, 1997; Collerson and Kamber, 1999; Condie, 2003). Also recording extraction of continental crust are other element ratios involving Nb such as Nb/La and Nb/Ta. The probability that at least some non-arc basalts come from mantle plume sources confirms the existence of a depleted component in plume sources, at least since 2.7 Ga (Kempton and others, 2000; Condie, 2005b). Prior to 2.5 Ga, the Nb/Th ratio scatters around the primitive mantle value, suggesting that both slightly enriched and slightly depleted reservoirs existed in the Archean mantle (Condie, 2003) (fig. 5). However, only after 2.5 Ga does the depleted mantle reservoir begin to grow in volume as reflected by the increasing Nb/Th ratios. This, in turn, may reflect the widespread onset of subduction in the late Archean as discussed below. Epsilon Nd(T) values in non-arc type basalts and associated TTGs (tonalite-trondhjemite-granodiorite) also show an increase in maximum value after the end of the Archean (fig. 6). Bennett (2004) explains this in terms of rapid growth of continental crust in the late Archean causing a change in Sm/Nd ratio of mantle sources.

Incompatible elements in TTGs.—As shown by Condie (2008), significant increases in large-ion lithophile and high-field strength elements and a decrease in Sr in continental crust at the end of the Archean reflect a shift in magma series from TTG control to calc-alkaline control. The distinction between calc-alkaline and TTG suites is based on major element trends such as the K-Na-Ca graph of Martin (1993), various silica variation diagrams, Sr content, Eu anomaly distribution, and low contents of heavy REE and Y (Condie, 2008). The shift from TTG to calc-alkaline control is evident in element ratios such as K_2O/Na_2O, $(La/Yb)_n$, Sr/Y, Th/U, Eu/Eu* and Nb/Ta (four of these ratios shown in fig. 7). Decreases in $(La/Yb)_n$ and Sr/Y and possibly an increase in K_2O/Na_2O in Paleoproterozoic upper crust reflect an increasing calc-alkaline plutonic component. The drop in La/Yb reflects a decrease in garnet content of the mafic source, and the large drop in Sr/Y in the calc-alkaline component (due to

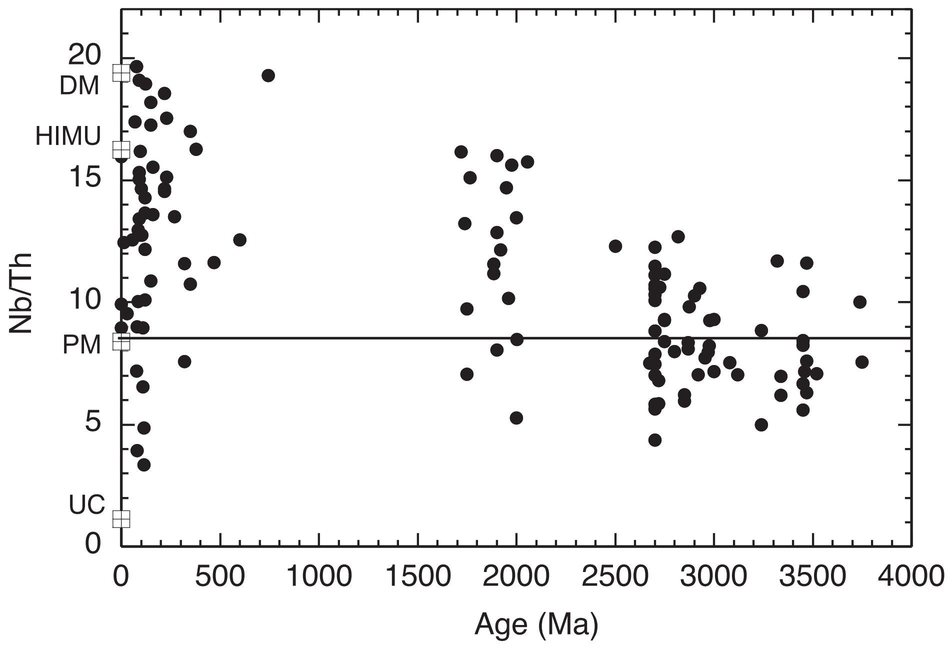

Fig. 5. Nb/Th versus age for non-arc oceanic basalts. Updated after Condie (2003). DM, depleted mantle; HIMU, high mu source; PM, primitive mantle; UC, upper continental crust.

restite plagioclase and/or fractional crystallization) controls the decreases in upper crustal Sr/Y with time (Condie, 2008). Although changes in Eu/Eu* all lie within one standard deviation of mean values, there is a suggestion of a drop in this ratio at the end of the Archean, consistent with plagioclase fractionation. A decrease in Eu/Eu* also occurs in banded iron formations at the end of the Archean (Kato and others, 2006), and probably reflects a combination of a decrease in the temperature of hydrothermal springs on the seafloor and a decrease in contamination of banded iron formation with continent-derived sediments.

These chemical changes appear to reflect chiefly a decrease in TTG magma production and a corresponding increase in calc-alkaline magma production at convergent plate margins after the end of the Archean. Furthermore, late Archean subduction zones may have differed from younger subduction zones in that they produced a thick mafic lower crust that served as a TTG magma source, and they did not give rise to significant volumes of calc-alkaline magma. Another viewpoint advocates that Archean TTGs are produced in oceanic crust by partial melting of descending slabs (Martin, 1993; Martin and Moyen, 2002).

One way to thicken mafic crust in the late Archean is by plate jams in subduction zones caused by the greater buoyancy of oceanic plates produced during a late Archean global thermal event. Modern subduction zones became widespread in the early Paleoproterozoic as magma production shifted from thickened crust or descending slabs to mantle wedges, where the calc-alkaline suite is produced. Large-ion lithophile element enrichment in mantle wedges reflects some combination of devolatilization of descending slabs and their associated subducted sediments.

Increase in $\delta^{18}O$ in granitoid zircons.—Valley and others (2005) and Valley (2008) have shown a prominent increase in the maximum values of $\delta^{18}O$ of zircons from granitoids after the end of the Archean (fig. 8). Oxygen isotopes are especially sensitive

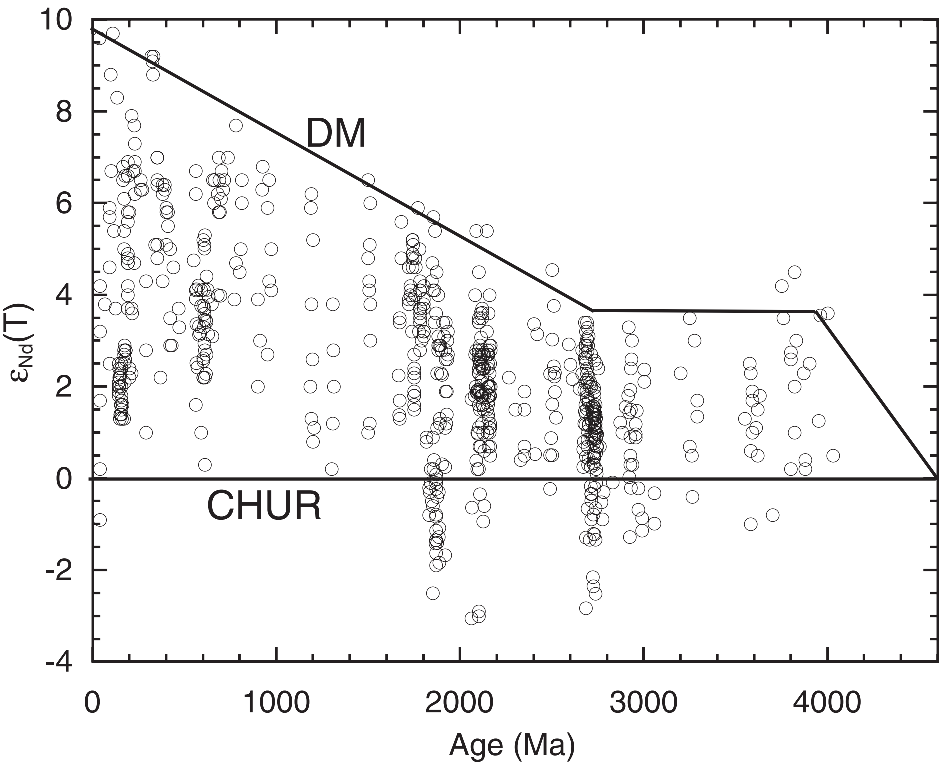

Fig. 6. Epsilon Nd(T) of basalts from non-arc oceanic greenstones and from associated TTGs as a function of age. Modified after Bennett (2004). DM, depleted mantle; CHUR, chondritic uniform reservoir.

to the incorporation of source material for granitoids that have interacted with the hydrosphere, such as shales. The increase in $\delta^{18}O$ after the end of the Archean may reflect the rapid growth and stabilization of cratons in the late Archean, which provided a framework for weathering and recycling of hydrated surface rocks into magma sources of granitoids after 2.5 Ga. Prior to the end of the Archean, felsic crust may have been rapidly recycled into the mantle, thus allowing only minimal interaction with the hydrosphere.

Increase in gold reserves.—Primary gold occurs dominantly in three tectonic settings: orogenic gold, volcanic-hosted massive sulfides (VHMS), and iron oxide copper-gold (IOCG) deposits. A major peak in orogenic and VHMS gold reserves occurs at or near 2.7 Ga (Groves and others, 2005) (fig. 9, orogenic reserves only). As discussed by Goldfarb and others (2001), orogenic gold deposits are produced by hydrothermal fluids, in response to high thermal fluxes at convergent plate margins. VHMS gold deposits are formed chiefly in submarine settings in back arc basins. In contrast to subduction-related gold, IOCG gold deposits are associated with continental rifts and alkaline granitoids. Interestingly, the largest known IOCG gold deposit is the Carajas deposit in Brazil formed at 2.6 Ga, just after the 2.7-Ga juvenile crust peak. Since orogenic and VHMS gold deposits are related directly to subduction and production of juvenile continental crust (Goldfarb and others, 2001; Groves and others, 2005), it is not surprising to find major gold reserves at 2.7 Ga.

Thickening of the Archean lithosphere.—The ^{187}Re-^{187}Os isotopic system has proved valuable in dating episodes of magma extraction from the mantle (Walker and others,

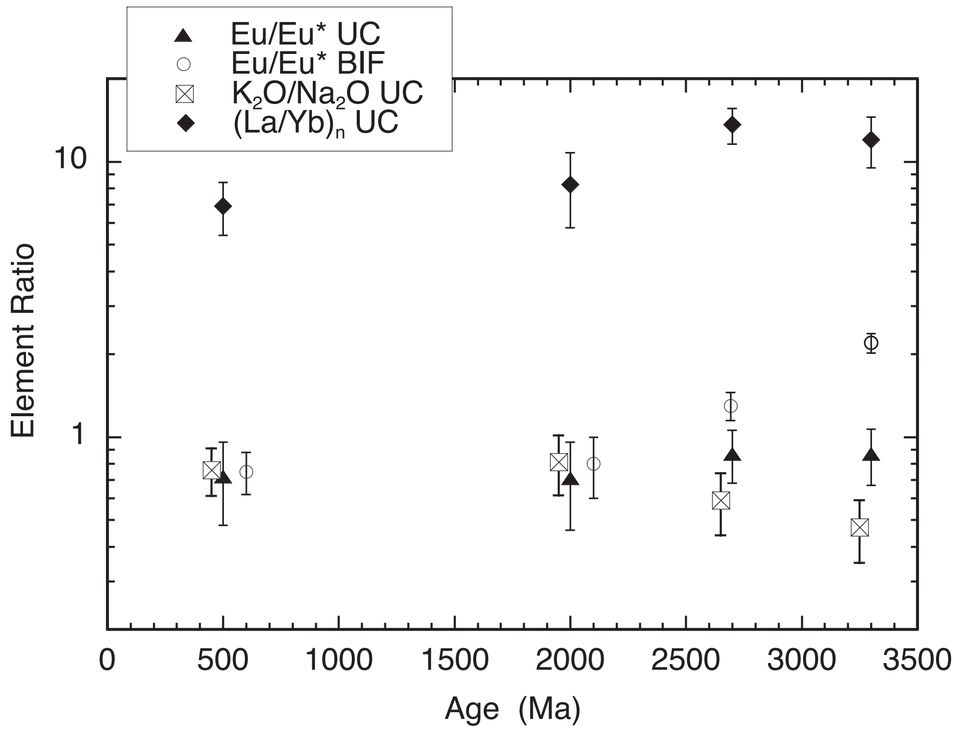

Fig. 7. Various element ratios versus age. Each value is an average with 1s errors. UC, upper continental crust; BIF, banded iron formation. $Eu/Eu^* = Eu_n/(Sm_n \times Gd_n)^{1/2}$, where n refers to chondrite normalized values. References: Condie (1993; 2008 and references therein).

1989; Pearson and others, 2002; Carlson and others, 2005). The key feature of the Re-Os system is that Os is compatible in the residue during mantle melting, whereas Re is moderately incompatible. Hence, the residue left after melt extraction has lower Re, but greater Os concentration than either fertile mantle or mantle melts. The restitic lherzolite or harzburgite left in the mantle is useful in constraining the age of melt extraction, known as a Re depletion age (T_{RD}).

Re-Os depletion ages from mantle xenoliths coming from beneath Archean cratons suggest that much of the thick Archean mantle lithosphere formed in the late Archean (Irvine and others, 2001; Carlson and others, 2005). The distribution of Archean T_{RD} ages from mantle xenoliths shows a prominent peak at 2.7 Ga consistent with widespread thickening of the continental lithosphere at this time (fig. 10). Although most of the data come from the Kaapvaal craton in southern Africa (which formed between 3.5 and 2.7 Ga; Pearson and others, 1995), many xenoliths from other Archean cratons also have T_{RD} ages between 2.7 and 2.65 Ga. This suggests that the thickened lithospheric roots of cratons that formed before 2.7 Ga were produced chiefly by a mantle event around 2.7 Ga, such as underplating with mantle plumes.

Consistent with the thick Archean mantle lithosphere forming as a depleted plume underplate is the major element composition of mantle xenoliths coming from beneath cratons older than 2.7 Ga. Most are high in FeO and MgO, similar to the compositions of basalts derived from young depleted mantle plumes (Herzberg, 2004). However, the jury is still out on the question of whether these sources are remnants of mantle plume heads accreted to the bottom of the lithosphere or if they

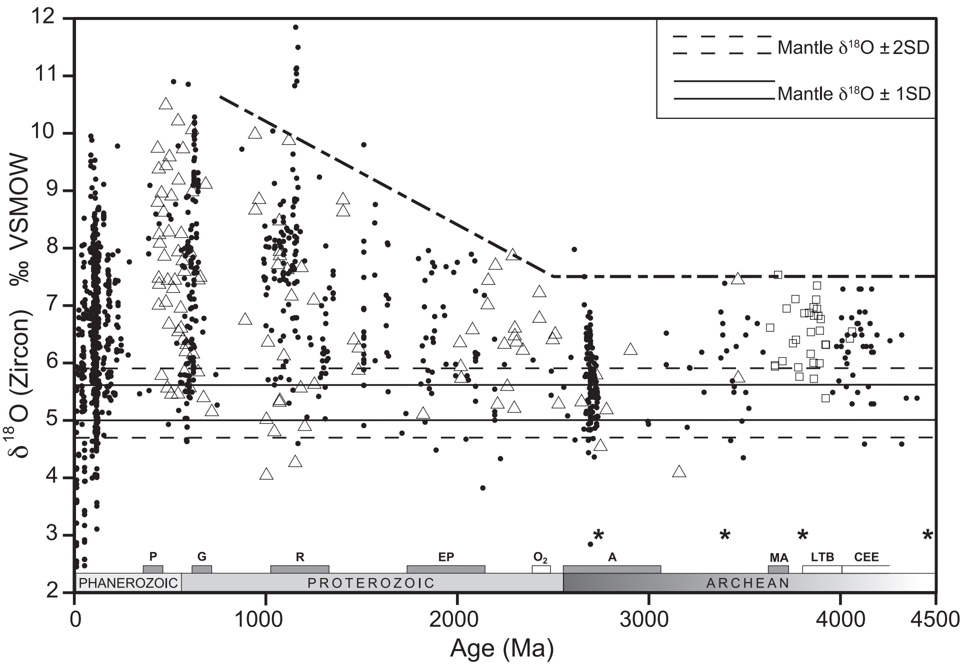

Fig. 8. $\delta^{18}O$ in zircons from granitoids versus age. Symbols: dots, mostly multigrain zircons from bedrock samples; triangles/squares, single detrital zircons. See Valley and others (2008) for more information. Figure courtesy of John Valley.

formed by the tectonic stacking of subducted oceanic slabs, or some combination thereof (Wittig and others, 2008; Wyman and Kerrich, 2009; Arndt and others, 2009). Whatever the mechanism by which thick Archean lithosphere formed, most of it appears to have been produced over a short time interval of less than 100 My in the late Archean.

Another feature consistent with a thick root beneath Archean cratons is the absence of a seismic high-velocity layer in the lower crust of most Archean cratons. As suggested by Durrheim and Mooney (1991), an underplate of thick mantle lithosphere may have prevented mafic additions to the base of overlying Archean crust. Because post-Archean crust is not protected by thick mantle lithosphere, mafic magmas may have continued to underplate this crust by either subduction or plume-related processes.

DISCUSSION

All of the above changes near the Archean/Proterozoic boundary seem to track one or both of two events that happened over a 50 to 300 My time window beginning just after 2.5 Ga. The first is the rapid cooling of the mantle. Second is the production and preservation of stable continental crust between 2.7 and 2.5 Ga, as a supercontinent or as one or more supercratons as suggested by Bleeker (2003). A summary of the changes at the end of the Archean and their probable relationship to these two events is given in table 1.

Much has been written about the nature and cause of a probable global event in the late Archean (Stein and Hofmann, 1994; Condie, 1998; Bedard, 2006). One model that has been widely discussed is a catastrophic global mantle plume event, which is triggered by collapse and sinking of oceanic slabs through the 660 km discontinuity

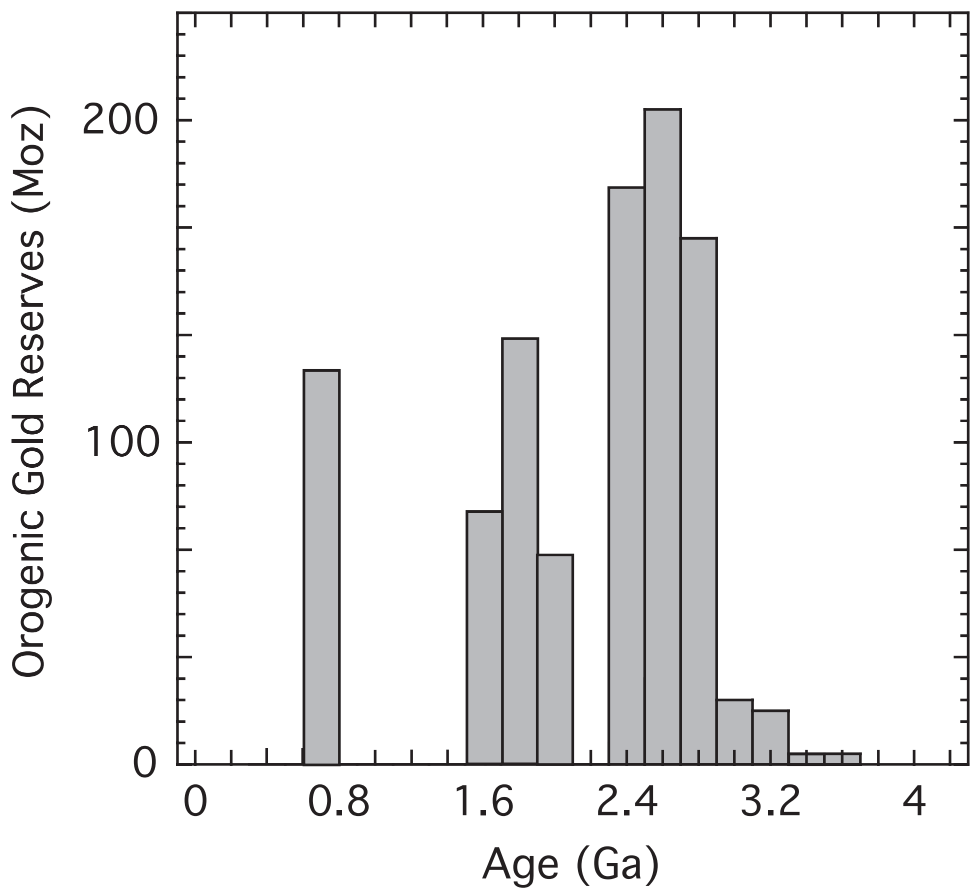

Fig. 9. Distribution of orogenic gold reserves by age. Modified after Groves and others, 2005.

(Tackley, 1997; Condie, 1998). This is based on the assumption that the 660-km phase transition has a large negative Clapeyron slope, and thus is a robust boundary in the hotter Archean mantle. However, recent experimental determinations of the Mg-perovskite reaction show that it is only slightly negative and may not have slowed the rate of slab sinking into the lower mantle, even in a hotter Earth (Katsura and others, 2003; Fei and others, 2004). Hence, the catastrophic sinking slab model is no longer as attractive as an explanation for the 2.7 Ga event.

Based on geodynamic modeling, O'Neill and others (2007) have suggested that plate tectonics did not begin all at once, but rather went through an intermediate state (the episodic regime) during which it turned on and off several times before becoming permanently established. As suggested by Condie and others (2009), the last turn on may have been at about 2200 Ma, following a 200-My slow down between 2.4 and 2.2 Ga. The geologic record suggests that plate tectonics may have begun, at least locally, by about 3 Ga (Condie and Kroner, 2008), but that the first really widespread plate tectonics developed in the late Archean at 2.7 to 2.5 Ga. Particularly relevant is a large drop in the ratio of non-arc to arc oceanic basalts from about 0.7 in the early Archean to less than 0.5 at 2.7 Ga (fig. 11). Could widespread propagation of subduction at 2.7 Ga be responsible for the rapid growth of cratons in the late Archean followed by cooling of the mantle?

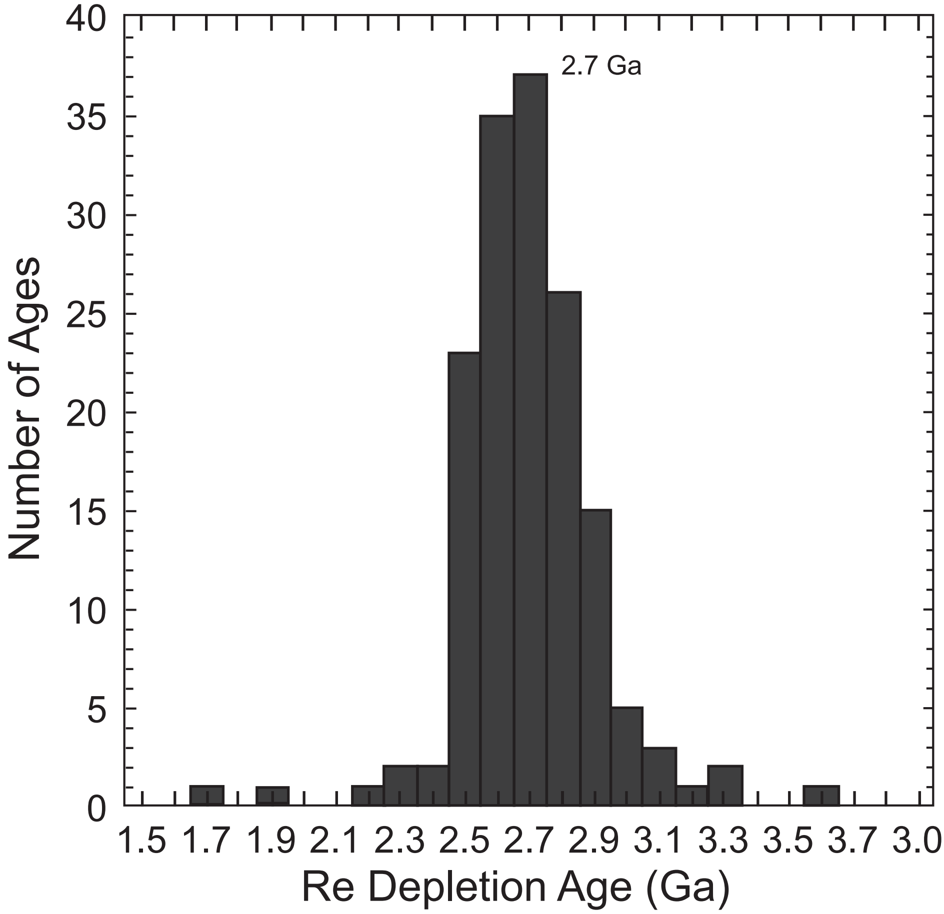

Fig. 10. Histogram showing frequency of Re/Os depletion ages of xenoliths of Archean mantle lithosphere. Data courtesy of Graham Pearson and Rick Carlson (Pearson and others, 2002; Carlson and others, 2005).

Table 1

Summary of tracers of events in the late Archean and early Paleoproterozoic

Table 1

Summary of Tracers of Events in the Late Archean and Early Paleoproterozoic

Event	**Tracer**
Stabilization of Cratons	1. Changes in incompatible element ratios in TTGs and continental crust 2. Peak in orogenic gold reserves at 2.7 Ga 3. Production of thick lithosphere at 2.7 Ga 4. Increase in $\delta^{18}O$ in granitoid zircons 5. Increase in Nb/Th and $\varepsilon_{Nd}(T)$ in non-arc oceanic basalts
Cooling of the Deep Mantle	1. Decrease in frequency of komatiites 2. Decrease in MgO content of komatiites 3. Decrease in Ni/Fe in banded iron formation 4. Increase in Nb/Yb and similar element ratios in non-arc oceanic basalts

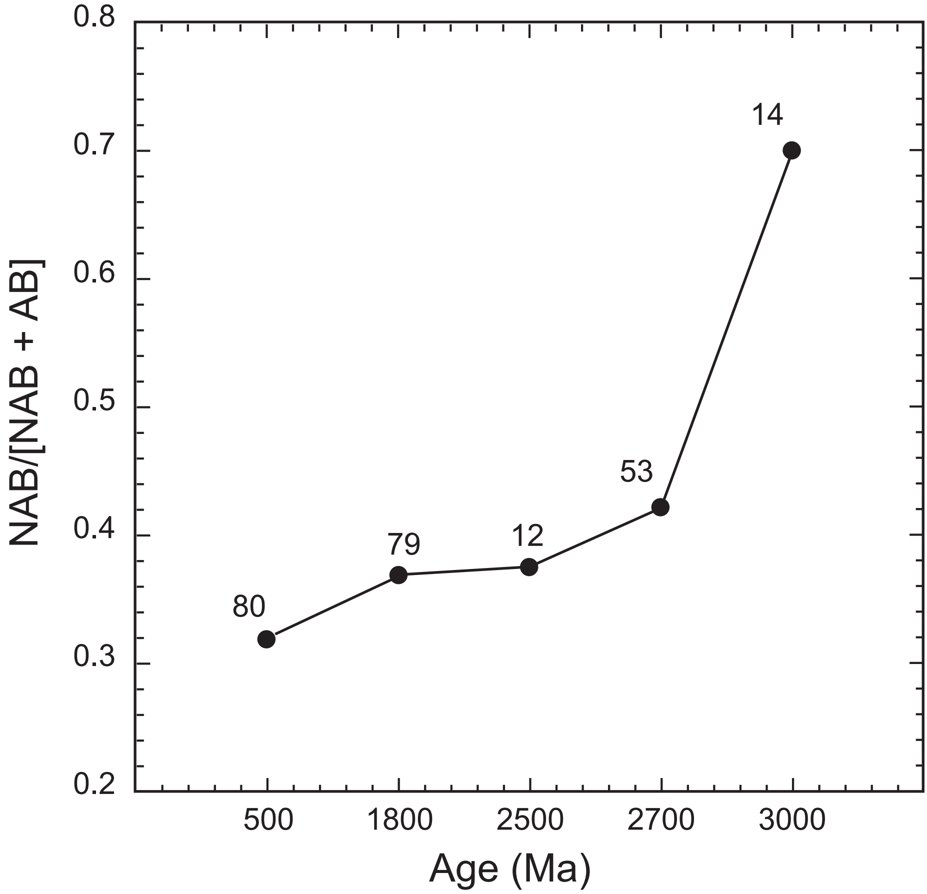

Fig. 11. Ratio of oceanic non-arc to arc type basalts versus age. Each labeled point is the number of greenstone locations. Data updated from Condie (2003 and 2005b). NAB, non-arc basalt; AB, arc basalt.

Any model for the late Archean must also include an explanation for a possible crustal age gap at 2.4 to 2.2 Ga (Condie and others, 2009). One possible scenario is shown in figure 12. The simplest relationship is that the onset of widespread subduction at 2.7 Ga produces continental crust in numerous arcs, which rapidly collide to form large cratons and perhaps a supercontinent. The cratons survive for two reasons: 1) they are underplated with thick lithosphere either from mantle plumes or stacking of buoyantly subducted slabs, and 2) the growth rate of continental crust exceeds the recycling rate back into the mantle, possibly reflecting buoyant subduction or colliding oceanic plateaus that produce "log jams" in subduction zones.

Oceanic slabs sinking to the D″ layer could destablize this layer by thickening it resulting in an enhanced production rate of mantle plumes (Davies, 1999) (fig. 12). This could increase the temperature of the upper mantle in the late Archean, in agreement with a recent model of Komiya (2007). Enhanced plume production should produce more oceanic plateaus and these plateaus could form the nuclei of cratons (Bedard, 2006) or they could collide with arcs. In either case, they would contribute to rapid production of the early cratons. A widespread LIP (large igneous province) event at 2.45 Ga (Heaman, 1997) may have been triggered by the last of the

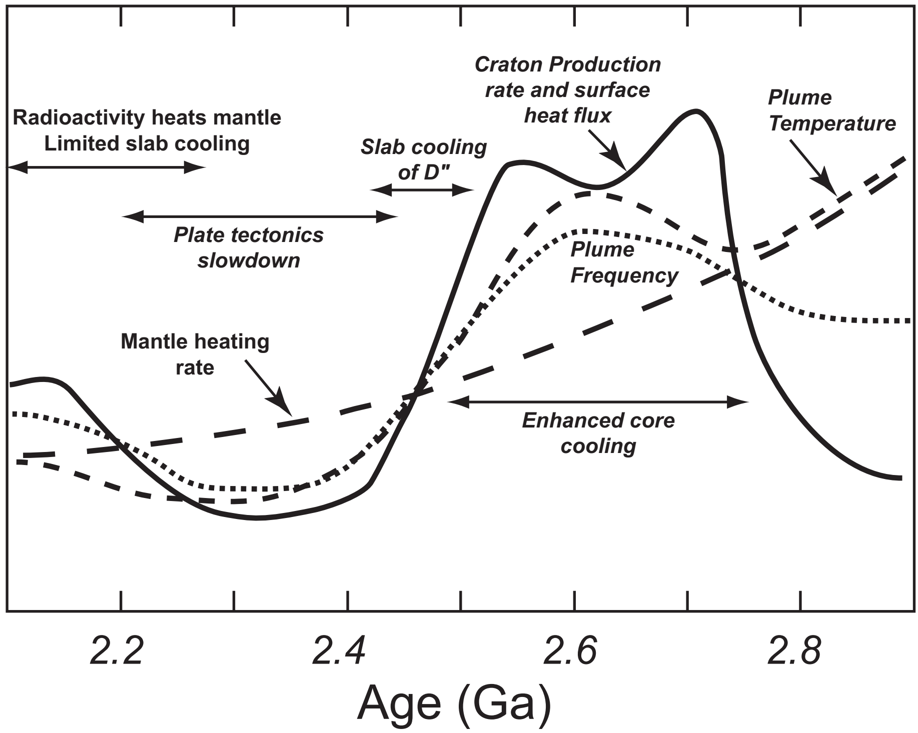

Fig. 12. Thermal model for changes in the mantle near the end of the Archean and beginning of the Paleoproterozoic.

slabs associated with the onset of widespread subduction at 2.7 to 2.5 Ga, which sank to the D″ layer. The key requirement is that all of this must happen in a 200 My time window between 2.7 and 2.5 Ga (fig. 12).

The arrival of cool, depleted slabs at the base of the mantle may also increase the heat flux from the core, which warms the newly arrived slabs (Korenaga, 2006). The cooling of the deep mantle would begin after 2.5 Ga and continue until about 2.4 Ga when the plate tectonics slowdown is initiated (fig. 12). This offers a possible explanation for the rapid drop in temperature of the mantle inferred from the changes in komatiite production rate and MgO levels and the increases in Nb/Yb in non-arc oceanic basalts. Also, supporting a cool lower mantle during this time window is a sparsity of LIPS between 2.4 and 2.2 Ga (Ernst and Buchan, 2001). When plate tectonics comes back on track at about 2.2 Ga, the first supercratons (or supercontinent) formed at 2.7 Ga, break-up, and are then dispersed (Bleeker, 2003). Also at this time, the temperature of plume sources is re-established along the normal mantle heating rate curve of the deep mantle (fig. 12). We are currently testing a geodynamic model for this scenario, which will be published as a separate paper.

This model is not free from problems. One of the uncertainties is how fast depleted oceanic slabs sink into the deep mantle to cool plume sources. The enhanced subduction activity at 2.7 Ga would produce many sinking slabs, and if these slabs sink too fast, cooling of deep plume sources could start before 2.5 Ga. A related question is how to keep the plume sources cool for 200 My (2.4-2.2 Ga), when the rate of slab sinking would be greatly reduced as plate tectonics enters a global slowdown mode.

Our preliminary modeling suggests that it may take this long for the slabs to equilibrate with the temperature in the deep mantle.

CONCLUSIONS

Important, well-documented geochemical changes in the crust-mantle system are found at the end of the Archean. Because of the lack of data between 2.4 and 2.2 Ga, it is not possible to estimate the rate at which the changes occurred. They may have taken place immediately after 2.5 Ga in a time period less than 100 My, or they may have extended over 100 to 300 My between 2.5 and 2.2 Ga. In either case, they may be related to one or both of two events: cooling of the deep mantle and formation of the first widespread cratons. The first major cratons (or supercratons) probably formed in the late Archean around 2.7 Ga, whereas cooling of the deep mantle began after 2.5 Ga. Both of these events could be related to a widespread onset of plate tectonics in the late Archean followed by a slowdown of plate tectonics between 2.4 and 2.2 Ga. Rapid formation of cratons at 2.7 Ga is probably tied to high production rates and relatively rapid collision of arcs, together with enhanced production rates and collision of oceanic plateaus and arcs. The cause of cooling of the deep mantle after 2.5 Ga, however, is problematical, but may be related to the arrival of large volumes of cool, depleted slabs in the lower mantle soon after the 2.7 to 2.5 Ga episode of widespread subduction.

References

Arndt, N. T., Barnes, S. J., and Lesher, C. M., 2008, Komatiiite: Cambridge, United Kingdom, Cambridge University Press, 488 p.

Arndt, N. T., Coltice, N., Helmstaedt, H., and Gregoire, M., 2009, Origin of Archean subcontinental lithosphere mantle: some petrological constraints: Lithos, v. 109, p. 61–71, doi: 10.1016/j.lithos.2008.10.019.

Bédard, J. H., 2006, A catalytic delamination-driven model for coupled genesis of Archean crust and subcontinental lithospheric mantle: Geochimica et Cosmochimica Acta, v. 70, p. 1188–1214, doi: 10.1016/j.gca.2005.11.008.

Bennett, V. C., 2004, Compositional evolution of the mantle: Treatise on Geochemistry, v. 4, chapter 13, p. 493–519, doi: 10.1016/B0-08-043751-6/02013-2.

Bleeker, W., 2003, The late Archean record: a puzzle in ca. 35 pieces: Lithos, v. 71, p. 99–134, doi: 10.1016/j.lithos.2003.07.003.

Carlson, R. W., Pearson, D. G., and James, D. E., 2005, Physical, chemical and chronological characteristics of continental mantle: Reviews of Geophysics, v. 43, RG1001, 10.1029/2004RG000156.

Collerson, K. D., and Kamber, B. S., 1999, Evolution of the continents and the atmosphere inferred from Th-U-Nb systematics of the depleted mantle: Science, v. 283, p. 1519–1522, doi: 10.1126/science.283.5407.1519.

Condie, K. C., 1993, Chemical composition and evolution of the upper continental crust: contrasting results from surface samples and shales: Chemical Geology, v. 104, p. 1–37, doi: 10.1016/0009-2541(93)90140-E.

——— 1998, Episodic continental growth and supercontinents: a mantle avalanche connection?: Earth and Planetary Science Letters, v. 163, p. 97–108, doi: 10.1016/S0012-821X(98)00178-2.

——— 2003, Incompatible element ratios in oceanic basalts and komatiites: tracking deep mantle sources and continental growth rates with time: Geochemistry Geophysics Geosystems, v. 4 (1), 1005, doi: 10.1029/2002GC000333.

——— 2005a, Earth as an Evolving Planetary System: Amsterdam, Elsevier, 447 p.

——— 2005b, High field strength element ratios in Archean basalts: a window to evolving sources of mantle plumes: Lithos, v. 79, p. 491–504, doi: 10.1016/j.lithos.2004.09.014.

——— 2008, Did the character of subduction change at the end of the Archean? Constraints from convergent-margin granitoids: Geology, v. 36 (8), p. 611–614, doi: 10.1130/G24793A.1.

Condie, K. C., and Benn, K., 2006, Archean geodynamics: similar to or different from modern geodynamics?, *in* Benn, K., Mareschal, J. C., and Condie, K. C., editors, Archean geodynamics and environments: American Geophysical Union, Monograph 164, p. 47–59.

Condie, K. C., and Kroner, A., 2008, When did plate tectonics begin? Evidence from the geologic record: Geological Society of America Special Paper 440, p. 281–295, doi: 10.1130/2008.2440(14).

Condie, K. C., O'Neill, C., and Aster, R., 2009, Evidence and Implications for a widespread magmatic shutdown for 250 My on Earth: Earth and Planetary Science Letters, v. 282, p. 294–298, doi: 10.1016/j.epsl.2009.03.033.

Davies, G. F., 1995, Punctuated tectonic evolution of the earth: Earth and Planetary Science Letters, v. 136, 363–379, doi: 10.1016/0012-821X(95)00167-B.

——— 1999, Dynamic Earth Plates, Plumes and Mantle Convection: Cambridge, United Kingdom, Cambridge University Press, 458 p.

De Wit, M. J., and Ashwal, L. D., 1997, Greenstone Belts: Oxford, Oxford University Press, 809 p.
Durrheim, R. J., and Mooney, W. D., 1991, Archean and Proterozoic crustal evolution: evidence from crustal seismology: Geology, v. 19, p. 606–609, doi: 10.1130/0091-7613(1991)019<0606:AAPCEE>2.3.CO;2.
Ernst, R. E., and Buchan, K. L., 2001, Large mafic magmatic events through time and links to mantle-plume heads: Geological Society of America Special Paper 352, p. 483–566, doi: 10.1130/0-8137-2352-3.483.
Fei, Y., Van Orman, J., Li, J., van Westrenen, W., Sanloup, C., Minarik, W., Hirose, K., Komabayashi, T., Walter, W., and Funakoshi, K., 2004, Experimentally determined postspinel transformation boundary in Mg_2SiO_4 using MgO as an internal pressure standard and its geophysical implications: Journal of Geophysical Research, v. 109, doi: 10.1029/2003JB002562.
Goldfarb, R. J., Groves, D. I., and Gardoll, S., 2001, Orogenic gold and geologic time: A global synthesis: Ore Geology Reviews, v. 18, p. 1–75, doi: 10.1016/S0169-1368(01)00016-6.
Grove, T. L., and Parman, S. W., 2004, Thermal evolution of the Earth as recorded by komatiites: Earth and Planetary Science Letters, v. 219, p. 173–187, doi: 10.1016/S0012-821X(04)00002-0.
Groves, D. I., Vielreicher, R. M., Goldfarb, R. J., and Condie, K. C., 2005. Controls on the heterogeneous distribution of mineral deposits through time: Geological Society, London, Special Publication, v. 248, p. 71–101, doi: 10.1144/GSL.SP.2005.248.01.04.
Heaman, L. M., 1997, Global mafic magmatism at 2.45 Ga: remnants of an ancient large igneous province? Geology, v. 25, p. 299–302, doi: 10.1130/0091-7613(1997)025<0299:GMMAGR>2.3.CO;2.
Herzberg, C., 2004, Geodynamic information in peridotite petrology: Journal of Petrology, v. 45, p. 2507–2530, doi: 10.1093/petrology/egh039.
Irvine, G. J., Pearson, D. G. and Carlson, R. W., 2001, Lithospheric mantle evolution of the Kaapvaal craton: A Re-Os isotope sudy of peridotite xenoliths from Lesotho kimberlites: Geophysical Research Letters, v. 28, 2505–2508, doi: 10.1029/2000GL012411.
Kato, Y., Yamaguchi, K. E., and Ohmoto, H., 2006, Rare earth elements in Precambrian banded iron formation: secular changes of Ce and Eu anomalies and evolution of atmospheric oxygen: Geological Society of America, Memoir, v. 198, p. 269–289, doi: 10.1130/2006.1198(16).
Katsura, T., Yamada, H., Shinmei, T., Kubo, A., Ono, S., Kanzaki, M., Yoneda, A., Walter, M. J., Ito, E., Urakawa, S., Funakoshi, K. and Utsumi, W., 2003, Post-spinel transition in Mg_2SiO_4 determined by high *P-T* in situ X-ray diffractometry: Physics of the Earth and Planetary Interiors, v. 136, p. 11–24, doi: 10.1016/S0031-9201(03)00019-0.
Kempton, P. D., Fitton, J. G., Saunders, A. D., Nowell, G. M., Taylor, R. N., Hardarson, B. S., and Pearson, G., 2000, The Iceland plume in space and time: a Sr-Nd-Pb-Hf study of the North Atlantic rifed margin: Earth and Planetary Science Letters, v. 177, 255–271, doi: 10.1016/S0012-821X(00)00047-9.
Komiya, T., 2007, Material circulation through time: Chemical differentiation within the mantle and secular variation of temperature and composition of the mantle, *in* Yuen, D. A., Maruyama, S., Karato, S., and Windley, B. F., editors, Superplumes: Beyond Plate Tectonics: Berlin, Springer-Verlag, p. 187–234.
Konhauser, K. O., Pecoits, E., Lalonde, S. V., Papineau, D., Nisbet, E. G., Barley, M. E., Arndt, N. T., Zahnle, K., and Kamber, B. S., 2009, Oceanic nickel depletion and a methanogen famine before the great oxidation event: Nature, v. 458, p. 750–754, doi: 10.1038/nature07858.
Korenaga, J., 2006, Archean geodynamics and the thermal evolution of Earth: American Geophysical Union Monograph 164, p. 7–32, doi: 10.1029/164GM03.
Martin, H., 1993, The mechanisms of petrogenesis of the Archean continental crust-comparison with modern processes: Lithos, v. 30, 373–388, doi: 10.1016/0024-4937(93)90046-F.
Martin, H., and Moyen, J.-F., 2002, Secular changes in tonalite-trondhjemite-granodiorite composition as markers of the progressive cooling of Earth: Geology, v. 30, p. 319–322, doi: 10.1130/0091-7613(2002)030<0319:SCITTG>2.0.CO;2.
McCulloch, M. T., and Bennett, V. C., 1994, Progressive growth of the Earth's continental crust and depleted mantle: geochemical constraints: Geochimica et Cosmochimca Acta, v. 58, p. 4717–4738, doi: 10.1016/0016-7037(94)90203-8.
O'Neill, C., Lenardic, A., Moresi, L., Torsvik, T. H., and Lee, C.-T. A., 2007, Episodic Precambrian subduction: Earth and Planetary Science Letters, v. 262, p. 552–562, doi: 10.1016/j.epsl.2007.04.056.
Parman, S. W., Grove, T. L., and Dann, J. C., 2001, The production of Barberton komatiites in an Archean subduction zone: Geophysical Research Letters, v. 28, p. 2513–2516, doi: 10.1029/2000GL012713.
Pearce, J. A., 2008, Geochemical fingerprinting of oceanic basalts with applications to ophiolite classification and the search for Archean oceanic crust: Lithos, v. 100, p. 14–48, doi: 10.1016/j.lithos.2007.06.016.
Pearson, D. G., Carlson, R. W., Shirey, S. B., Boyd, F. R., and Nixon, P. H., 1995, Stabilization of Archean lithospheric mantle: A Re-Os isotope study of peridotite xenoliths from the Kaapvaal craton: Earth and Planetary Science Letters, v. 134, 341–357, doi: 10.1016/0012-821X(95)00125-V.
Pearson, N. J., Alard, O., Griffin, W. L., Jackson, S. E., and O'Reilly, S. Y., 2002, In-situ measurement of Re-Os isotopes in mantle sulfides by laser ablation multicollector-inductively coupled mass spectrometry: Analytical methods and preliminary results: Geochimica et Cosmochimica Acta, v. 66, p. 1037–1050, doi: 10.1016/S0016-7037(01)00823-7.
Scholl, D. W., and von Huene, R., 2007, Crustal recycling at modern subduction zones applied to the past-Issues of growth and preservation of continental basement crust, mantle geochemistry, and supercontinent reconstruction: Geological Society of America, Memoir, v. 200, p. 9–32, doi: 10.1130/2007.1200(02).
Stein, M., and Hofmann, A. W., 1994, Mantle plumes and episodic crustal growth: Nature, v. 372, p. 63–68, doi: 10.1038/372063a0.
Sylvester, P. J., Campbell, I. H., and Bowyer, D. A., 1997, Niobium/uranium evidence for early formation of the continental crust: Science, v. 275, p. 521–523, doi: 10.1126/science.275.5299.521.
Tackley, P. J., 1997, Effects of phase transitions on three-dimensional mantle convection, *in* Crossley, D. J., editor, Earth's Deep Interior: Amsterdam, Gordon and Breach, p. 273–336.

Taylor, S. R., and McLennan, S. M., 1985, The Continental Crust: its Composition and Evolution: Oxford, Blackwell Scientific, 312 p.

Valley, J. W., 2008, The origin of habitats: Geology, v. 36, p. 911–912, doi: 10.1130/focus112008.1.

Valley, J. W., Lackey, J. S., Cavosie, A. J., Clechenko, C. C., Spicuzza, M. J., Basei, M. A. S., Bindeman, I. N., Ferreira, V. P., Sial, A. N., King, E. M., Peck, W. H., Sinha, A. K., and Wei, C. S., 2005, 4.4 billion years of crustal maturation: oxygen isotope ratios of magmatic zircon: Contributions to Mineralogy and Petrology, v. 150, p. 561–580, doi: 10.1007/s00410-005-0025-8.

Walker, R. J., Carlson, R. W., Shirey, S. B., and Boyd, F. R., 1989, Os, Sr, Nd and Pb isotope systematics of southern African peridotite xenoliths: Implications for chemical evolution of subcontinental mantle: Geochimica et Cosmochimica Acta, v. 53, p. 1583–1595, doi: 10.1016/0016-7037(89)90240-8.

Wittig, N., Pearson, D. G., Webb, M., Ottley, C. J., Irivine, G. J., Kopylova, M., Jensen, S. M., and Nowell, G. M., 2008, Origin of cratonic lithospheric mantle roots: A geochemical study of peridotites from the North Atlantic Craton, West Greenland: Earth and Planetary Science Letters, v. 274, p. 24–33, doi: 10.1016/j.epsl.2008.06.034.

Wyman, D., and Kerrich, R., 2009, Plume and arc magmatism in the Abitibi subprovince: Implications for the origin of Archean continental lithospheric mantle: Precambrian Research, v. 168, p. 4–22, doi: 10.1016/j.precamres.2008.07.008.