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The role of serpentine in preferential craton formation in the late Archean by lithosphere underthrusting

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Abstract

Cratons form the cores of continents and were formed within a narrow window of time (2.5–3.2 Gy ago), the majority having remained stable ever since. Petrologic evidence suggests that the thick mantle roots underlying cratons were built by underthrusting of oceanic and arc lithosphere, but paradoxically this requires that the building blocks of cratons are weak even though cratons must have been strong subsequent to formation. Here, we propose that one form of thickening could be facilitated by thrusting of oceanic lithospheres along weak shear zones, generated in the serpentinized upper part of the oceanic lithosphere (crust + mantle) due to hydrothermal interaction with seawater. Conductive heating of the shear zones eventually causes serpentine breakdown at ~ 600 °C, shutting down the shear zone and culminating in craton formation. However, if shear zones are too thin, serpentine breakdown and healing of the shear zone occurs too soon and underthrusting does not occur. If shear zones are too thick, serpentine breakdown takes too long so healing and lithospheric thickening is not favored. Shear zone thicknesses of ~18 km are found to be favorable for craton formation. Because the maximal depth of seawater-induced serpentinization into the lithosphere is limited by the depth of the isotherm for serpentine breakdown, shear zone thicknesses should have increased with time as the Earth's heat flux and depth to the serpentine breakdown isotherm decreased and increased, respectively, with time. We thus suggest that the greater representation of cratons in the late Archean might not necessarily be explained by preferential recycling in the early Archean but may simply reflect preferential craton formation in the late Archean. That is, our model predicts that the early Archean was too hot, the Phanerozoic too cold, and the late Archean just right for making cratons.

Keywords: craton; cratonic mantle; underthrusting; serpentine; Archean; continental crust

1. Introduction: the uniqueness of cratons

Cratons are unique. They are the cores of continents that formed primarily within a narrow window of time in the Archean (2.5–3.2 Gy; Fig. 1A) and have for the most part remained stable ever since. They are underlain by cold mantle roots, which are distinctive in that their thicknesses and chemical compositions cluster tightly around ~200 km and refractory compositions equivalent to 30–40% melt depletion, respectively (Jordan, 1978; Boyd, 1989; Menzies, 1990; Pearson et al., 1995; O'Reilly et al., 2001; Arndt et al., 2002; Griffin et al., 2003; Herzberg, 2004; King, 2005; Lee, 2006; Bernstein et al., 2007; Simon et al., 2007). Cratonic mantle is thus thicker and more melt depleted, as a whole, than any post-Archean lithospheres.

These thick, melt-depleted roots are thought to be responsible for the longevity of cratons. The most important property is that melt depletion decreases the Ca, Al and Fe content of cratonic mantle, providing chemical buoyancy that compensates for the craton's negative thermal buoyancy (Boyd and McAllister, 1976; Jordan, 1978, 1979; Boyd, 1989; Kelly et al., 2003; Lee, 2003). However, buoyancy alone is not enough to preserve cratons for eons. Cratonic mantle must also be inherently strong so that they do not gravitationally collapse or

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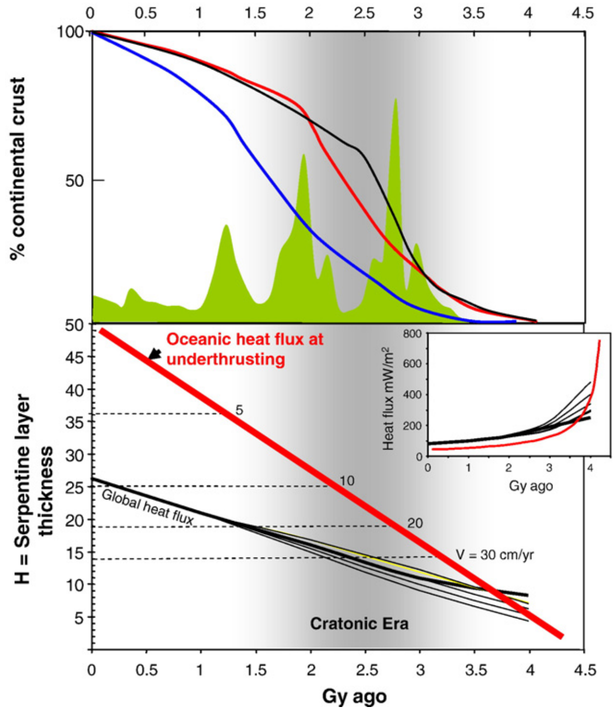

Fig. 1. A. Various cumulative crustal growth curves and U–Pb crystallization ages of zircons having juvenile oxygen isotopic signatures; all data are adapted from Hawkesworth and Kemp (2006). B. This figure shows how the thickness H of the serpentine layer, which we equate to the shear layer, in oceanic lithosphere might change throughout Earth's history. It is essentially a plot of the 600 °C isotherm. Solid black lines represent predictions using thermal histories expressed in terms of global average heat flux (see inset), which is a minimum bound because subducting lithosphere is cold and hence has a lower heat flux. The red solid line represents a linear extrapolation between H inferred from the thermal state of average subducting oceanic lithosphere (100 My old) and $H=0$ in the earliest Archean; corresponding heat flux is shown in inset. Because it is the thermal state of oceanic lithosphere just prior to subduction that is of concern here, the red line is the more appropriate curve. Note, that all curves should converge to very low values of H in the earliest Archean. Horizontal dashed lines represent optimum H for a given plate velocity V (taken from Fig. 4B). Intersection of these lines with thermal history curves represent optimum time for craton formation.

become thermally eroded (Lenardic and Moresi, 1999; King, 2005). Extensive melt depletion should have also resulted in dehydration, and because of the sensitivity of rheology to water content, dehydration would have greatly increased the intrinsic viscosity of cratonic mantle, making them resistant to disruption (Pollack, 1986; Hirth et al., 2000).

The tight clustering of cratonic root thicknesses, compositions, and formation times suggest that craton-forming mechanisms were fundamentally different from continent-forming processes in the Phanerozoic (Fig. 1A). The origin of cratons has thus been the subject of much speculation. In one class of hypotheses, cratonic mantle forms in place by high pressure (5–8 GPa) melting in large plume heads or major mantle overturns (Davies, 1995; Pearson et al., 1995; Herzberg, 1999; Arndt et al., 2002; Griffin et al., 2003). These plume scenarios are motivated by the fact that in order to attain 30–40% melting at 200 km depth (~8 GPa), very deep and hot melting is required. In the second class, the extensively melt-depleted nature of cratonic mantle is suggested to have formed by hydrous flux melting in arc environments, so cratonic mantle formed by thickening of sub-arc continental lithospheric mantle (Parman et al., 2004; Horodyskyj et al., 2007). Finally, the third class invokes low pressure melting in plumes or mid-ocean ridges (2–6 GPa), followed by thickening and underthrusting of these relatively low pressure melt residues to the greater depths characterizing cratons today (Helmstaedt and Schulze, 1989; de Wit et al., 1992; Bernstein et al., 1998; Walter, 1999; Canil, 2004; Herzberg, 2004; Lee, 2006; Bernstein et al., 2007).

All of the above scenarios could have operated preferentially in the Archean because a hotter Earth should promote more extensive melt extraction and hence generation of chemically buoyant mantle, which are the ingredients of cratonic mantle. If so, it follows that as the Earth cooled, the production of chemically buoyant mantle would have decreased with time and hence craton generation would also decrease with time. However, none of these models explains the absence of cratons ~3.3 Gy and earlier. In a hotter Earth, even more extensive melting (and hence more cratons) is predicted in the earliest Archean, so their absence >3.3 Gy ago is conspicuous and requires that all such cratons were recycled back into the mantle. This paper considers an alternative to recycling, that is, craton formation itself is confined to a narrow window of time. We show below that in addition to the secular change in lithospheric mantle compositions, accretion mechanisms may have been temperature-dependent, and if craton formation is controlled to some extent by accretion, the formation and assembly of cratons were favored between 2 and 3 Gy.

2. Evidence for an accretionary origin

Although all three classes of craton-forming scenarios could have operated, there is growing circumstantial evidence that accretionary processes may have operated. For example, field geologic studies show that at least some parts of cratons appear to be amalgamations of linear belts of accreted terranes (Hoffman, 1989; Condie, 1990; de Wit et al., 1992; Condie and Selverstone, 1999; Chamberlain et al., 2003). Large-scale accretion might be expected to lead to dipping seismic reflectors extending well into the cratonic mantle. No such reflectors have yet been seen in the well-studied Kaapvaal craton in South Africa (James et al., 2001; Shirey et al., 2002), but seismic reflection studies in the Slave craton (in Canada) have revealed a shallowly dipping reflector, which extends into the mantle and extrapolates upwards to major crustal sutures or thrust faults (Bostock, 1998; Bostock, 1999).

It is also noteworthy that eclogite xenoliths in kimberlites erupted through cratons have non-mantle-like $^{18}O/^{16}O$ signatures, which are reasonably explained by having a low-temperature alteration overprint sometime in their histories (Jacob et al., 1994; Barth et al., 2001). Diamonds, having eclogitic inclusions (Shirey et al., 2002), are also characterized by distinctly low $^{13}C/^{12}C$ signatures compared to mantle carbon and thus may suggest a biogenic origin for the carbon (Deines et al., 2001). There is debate, of course, over whether kineticallylimited high-temperature processes could explain some of the unusual oxygen and carbon isotopic compositions (Cartigny et al., 1998). However, none of these processes seems able to explain the non-mass-dependent fractionation of sulfur isotopes in sulfide inclusions in eclogitic diamonds (Farquhar et al., 2002). Such fractionations are thought to be possible only by photochemical reactions in the atmosphere. Thus, it is widely thought that the protoliths of these eclogite xenoliths were hydrothermally altered oceanic crusts, which were underthrust during craton-forming events.

There is also a growing view that although cratonic peridotites now reside at pressures as high as 7 GPa (~210 km), their average melt extraction pressures are likely to be lower than 5 GPa (150 km). Garnet in cratonic peridotites is often interpreted as having formed by exsolution of high temperature orthopyroxenes or as a product of melt-refertilization (Cox et al., 1987; Saltzer et al., 2001; Simon et al., 2003; Malkovets et al., 2007) and therefore the presence of garnet may not reflect high average pressures of melting. If such garnet had been present as a primary residual phase during melting, cratonic peridotites would be expected to have higher Al, Sc, Y, and heavy rare earth element contents due to the preference of garnet for these elements; instead, these elements are highly depleted, hence melting must have proceeded into the spinel stability field or to the extent of garnet exhaustion (Bernstein et al., 1998; Canil, 2004; Lee, 2006; Simon et al., 2007). Another commonly cited evidence for high pressure origins of melting is the apparent low FeO contents of many cratonic peridotites (Herzberg, 1993; Pearson et al., 1995; Herzberg, 1999; Griffin et al., 2003; Herzberg, 2004). This is because the solidus temperature increases with pressure and FeO becomes more incompatible with increasing temperature so that high pressure melting generates high FeO content melts and low FeO content mantle residues. The problem, however, is that many of the samples with low FeO content are also anomalously enriched in SiO_2 due to orthopyroxene contents in excess of that expected by anhydrous melting. Although the explanation for Si- and orthopyroxene-enrichment is still debated (Kesson and Ringwood, 1989; Rudnick et al., 1994; Kelemen et al., 1998; Herzberg, 1999; Griffin et al., 2003; Lee et al., 2003; Lee, 2006; Simon et al., 2007), what seems likely is that orthopyroxene-enrichment correlates with low FeO contents due to the lower FeO of orthopyroxene compared to olivine. If Si-enriched samples are excluded or corrected for orthopyroxene-enrichment, the FeO contents of cratonic peridotites are found to be higher and consistent with average pressures of melting lower than 3–5 GPa (Fig. 2). Finally, the clustering of cratonic peridotite Mg#s around 92 (Mg#=molar Mg/(Mg+Fe)×100) compared to 89 for fertile asthenospheric mantle and 90 for Phanerozoic continental lithospheric mantle (Bernstein et al., 2007), requires 30–40% melt extraction from a pyrolitic type mantle. If the entire cratonic mantle had formed in one big plume event, the 30–40% melting implied at 7–8 GPa would necessitate extremely deep melting, perhaps down into the transition zone, because the slope of the mantle or melting adiabats becomes subparallel to that of the solidus at increasing pressures. There appears to be no evidence for such deep melting. It is more likely that the 30–40% melting occurs near the low pressure end of a polybaric melting column (Bernstein et al., 2007).

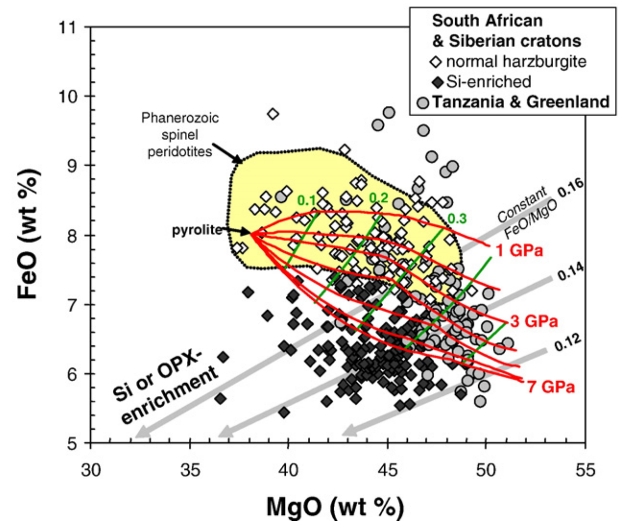

Fig. 2. FeO versus MgO (wt.%) in low-temperature cratonic peridotites (high-temperature sheared peridotites not included) modified from other publications (Lee et al., 2003; Lee, 2006). Yellow shaded region represents Phanerozoic peridotite xenoliths. Symbols refer to cratonic peridotites as shown in legend. In particular, those from South Africa that are enriched in Si and orthopyroxene (opx) are denoted as black diamonds (Si-enriched samples represent those that deviate from the partial melting curve in Mg# and modal olivine space (Boyd, 1989)). Red curves represent isobaric melting curves of a pyrolite starting composition for pressures ranging from 1 to 7 GPa (Walter, 1999, 2003). Straight green lines crosscutting the partial melting curves represent melting degrees at 0.1 intervals. Gray diagonal lines pointing towards the lower left hand corner represent predicted trajectories of Si-enrichment assuming no change in the ratio of FeO/MgO in the peridotite (labels on these trajectories represent constant FeO/MgO ratios). Note that the low FeO contents of Si-enriched peridotites may not necessarily indicate high pressure melting, but may instead be product of orthopyroxene addition.

Collectively, such data suggest that cratonic peridotites might represent just the upper parts of polybaric melting columns beneath ridges or plumes and that these highly depleted mantle slivers were later forced downwards to their current depths in cratons (Helmstaedt and Schulze, 1989). Clearly, each line of evidence outlined above is highly debatable, but the collective evidence is enough to warrant further consideration of the possibility of craton formation by the underthrusting of oceanic lithospheric mantle segments.

3. Difficulties in the accretionary hypothesis

Is the accretion hypothesis for making cratons physically plausible? In order to ensure longevity, cratons must have been strong since formation, but accretion and underthrusting is only possible if the building blocks of cratons are weak enough to deform and be transported along thrust faults. Cooper et al. (2006) showed that only by lowering the coefficient of friction significantly below that of dry rock could underthrusting of oceanic lithospheres be possible to make cratons. The implication was that hydrated lithosphere or shear zones were necessary for thickening to occur, but they pointed out that the existence of such weak zones later competes against the preservation of cratons once formed, essentially challenging the "heart and soul" of a craton. However, Cooper et al. suggested that if these shear zones were to dehydrate immediately after underthrusting terminated, the coefficient of friction would be driven back up to anhydrous values (effectively healing the fault), ensuring the craton a long life. Exactly how this occurs forms the remaining focus of this paper.

4. Serpentinite shear zones

We propose that the weak shear zones, which enable underthrusting of oceanic lithosphere, derive from the upper part of the oceanic lithosphere, which had experienced hydrothermal alteration by seawater prior to underthrusting (Fig. 3A). Such alteration would give rise to chloritized oceanic crust and serpentinized lithospheric mantle. In modern oceanic lithosphere, these alteration zones can extend well beyond 15 km depth into the oceanic lithosphere (Ranero et al., 2003). For old and cold oceanic lithosphere, serpentinization may even extend to 30–40 km depth along pre-existing fractures generated during seafloor spreading or new fractures generated during plate bending just before subduction (Hacker et al., 2003; Rüpke et al., 2004; Li and Lee, 2006; Lee and Chen, 2007). Since the Earth is now believed to have had oceans throughout most of its history (Watson and Harrison, 2005), serpentinized (or chloritized) zones may have also formed throughout most of Earth's history. We assume that even minor serpentinization (e.g., on grain boundaries or fractures) is enough to sufficiently weaken the upper part of the oceanic lithosphere to form a shear zone (Escartin et al., 2001).

Once underthrusting begins, these cold serpentinized shear zones will heat up by thermal diffusion from the hotter surroundings (such as the over-riding plate). Serpentine breaks down at ~600–700 °C (Fig. 3B) and transforms into olivine, shutting down the shear zone so that underthrusting terminates. If the shear zone is too thin, the hotter upper plate will soon conductively heat the shear zone and shut the shear zone down before sufficient underthrusting can take place; as a consequence, lithospheric thickening is not permitted. If the shear zone is too thick, heating will take too long and the shear zone does not heal before the lithosphere is underthrust deep into the mantle. In such a case, the downgoing oceanic lithosphere continues to subduct away. There should thus be an optimum shear zone thickness for craton formation by underthrusting. If we assume cratons today look more or less like when they were formed, the depth extent of thrusting is manifested by the present day thickness of cratons (~200 km). To make a craton, we assume that the minimum amount of slip during underthrusting is ~100–200 km, corresponding to perfectly vertical

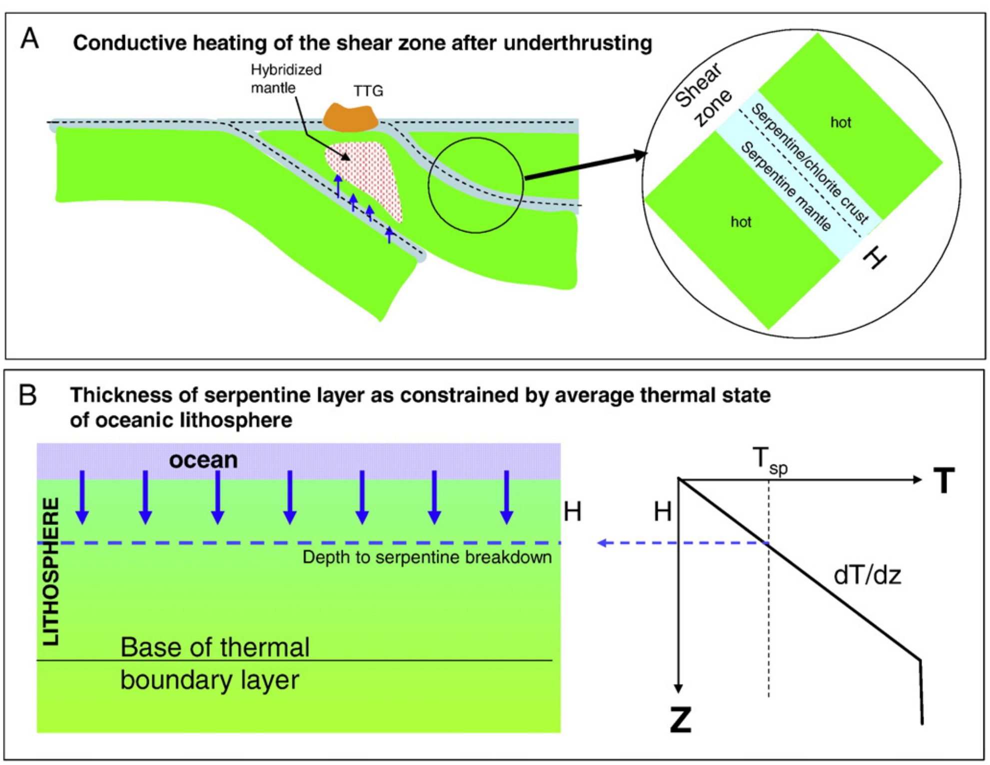

Fig. 3. A. Cartoon showing how the accretion of oceanic lithospheric mantle results in thick cratonic mantle. Motion is facilitated along weak shear zones of thickness *H*. More speculatively, TTG represents tonalite–trondjhemite–granodiorite magmas that could have formed by hydrous flux melting of basaltic crust as they get heated from the overlying lithosphere. B. The origin of the shear zones are suggested here to represent the upper part of the lithospheric mantle that has been hydrothermally altered by seawater. Depth to serpentinized zone *H* is shown by horizontal blue dashed line on the left figure. Right figure schematically shows a temperature profile for oceanic lithosphere with temperature gradient d*T*/d*z*. T_{sp} represents the temperature of serpentine breakdown. *H* corresponds to the depth at which temperature equals T_{sp}. Gradual shading in color of the lithosphere is shown to qualitatively describe the more fertile nature of the asthenosphere and deep part of the lithosphere.

underthrusting (which of course is probably unreasonably steep). As for the maximum amount of slip, according to Hoffman (1989), Archean accreted terranes are typically between 300 and 500 km wide, which means that the maximum slip distance (corresponding to flat subduction, e.g., angle of subduction is zero) for making a thick craton is ~500 km. Such distances of lateral transport seem unusually high at face value, but are probably not unreasonable because flat subduction extending >500 km inboard of the trench has been documented (albeit rarely) in the Phanerozoic (Dickinson and Snyder, 1978; Humphreys et al., 2003). In any case, if underthrusting is important for making cratons, the window of opportunity is bounded between 100 and 500 km.

To explore these concepts further we approximate the optimum shear zone thickness H using an analytical formulation. We assume for simplicity that the shear zone has a uniform initial temperature T_o (anywhere between 0 and 200 °C) and is later subjected to heating from its surroundings after it is underthrust. We approximate such a scenario by a uniform temperature boundary condition T_b, which effectively corresponds to the time-averaged boundary temperature that the shear zone experiences during underthrusting (in reality the temperature profile and boundary conditions are asymmetric, so our symmetric boundary condition will result in a maximum estimate of the heating rate). The time it takes for the average temperature of the shear zone to reach the serpentine breakdown temperature T_{sp} (°C) is (Carslaw and Jaeger, 1959):

$$t_{sp} \sim \frac{H^2}{\kappa \pi^2} \ln\left(\frac{8(T_o - T_b)}{\pi^2 (T_{sp} - T_b)}\right) \tag{1}$$

where κ is thermal diffusivity (10^{-6} m^2/s~30 km^2/My) and π is pi. For a given uniform plate velocity V (km/My), the time to underthrust this distance is simply L/V, hence H (km) is simply related to the extent of underthrusting L (km) as follows

$$H \sim \left(\frac{L\kappa\pi^2}{V}\right)^{1/2} \left[\ln\left(\frac{8(T_o - T_b)}{\pi^2 (T_{sp} - T_b)}\right)\right]^{-1/2}. \tag{2}$$

Fig. 4A shows how T in the shear zone increases with time for different shear zone thicknesses H. Serpentine breakdown is assumed to occur at T_{sp}~600 °C (we note that due to a small pressure effect, T_{sp} reaches 700 °C at pressures of 3 GPa; this does not change our results considerably). We also assumed that T_o=0 °C (using up to 200 °C does not change the result significantly) and T_b=800 °C, the latter chosen to represent the average temperature at the interface of the shear zone and its surroundings as this is a reasonable average temperature for lithosphere in the Archean when mantle potential temperatures could have been as high as 1600 °C (Grove and Parman, 2004) (values of 700 and 900 for T_b may be more appropriate for the present and earliest Archean, but do not change our results significantly).

We can also explore how much time it takes to complete one underthrusting event. As discussed above, if ~100–500 km is the favorable window of slip distance (L) to make a craton by underthrusting, the corresponding amount of time favoring craton formation is between 0.5 and 2.5 My, assuming as an example a plate velocity of 20 cm/yr. It can then be seen from Fig. 4A that only a certain range of shear zone thicknesses H is conducive for underthrusting over this duration of time and assumed plate velocity (H~12–24 km).

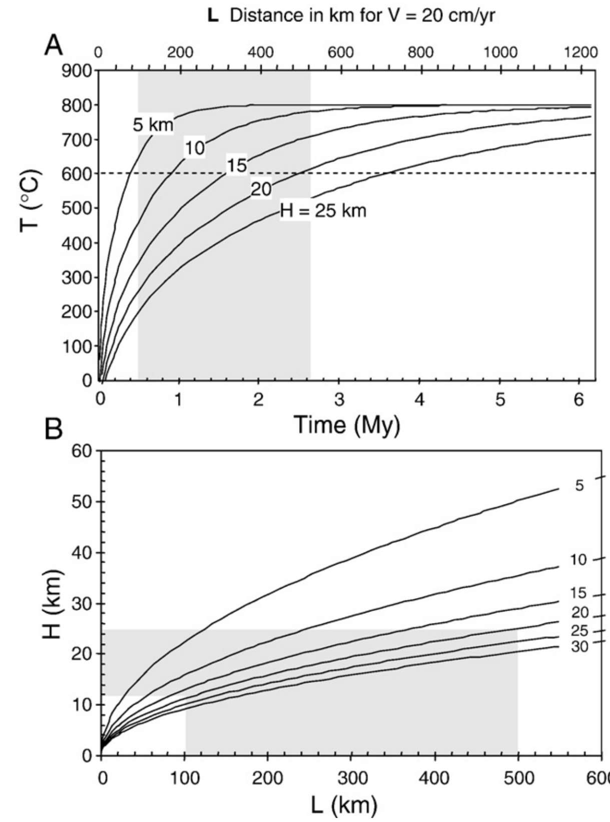

Fig. 4. A. Average temperature of the slab (Eq. (1)) as a function of time for different shear zone thicknesses H and a constant plate velocity of 20 cm/yr (V). Distance underthrusted corresponding to a given time is shown on upper horizontal axis. Gray shaded region represents the displacement L needed to generate a craton. If it is less than this, no underthrusting occurs. If it is greater than this, too much underthrusting occurs and the slab continues to subduct. B. Optimum shear zone thickness versus displacement distance for different plate velocities (Eq. (2)).

The more useful relationship for us is how the optimum shear zone thickness H depends on average plate velocity, which could have changed over Earth's history. This relationship is shown in Fig. 4B via Eq. (2). Here, H is plotted as a function of slip distance L for plate velocities ranging from 5 to 30 cm/yr. The low velocities pertain to average modern plate velocities (Gripp and Gordon, 2002). The higher velocities are considered in order to account for plate velocities in Earth's distant past. Assuming a similar convective regime as the present (e.g., the Archean was characterized by a mobile lid regime), Archean velocities presumably were higher because plate velocities scale with Rayleigh number to the 2/3 power (Turcotte and Oxburgh, 1967) and the Rayleigh number itself was probably higher because mantle temperatures were higher and viscosities were lower. Given that modern plate velocities range up to 11 cm/yr and that the Indian continent may have even reached 20 cm/yr (Jurdy and Gordon, 1984; Gripp and Gordon, 2002), our upper range may not be unreasonably high for Archean plate velocities. Fig. 4B thus shows that optimal shear zone thicknesses H decrease with increasing plate velocity. In the next section, we will discuss the possibility of secular changes in H due to secular changes in the thermal state of the Earth and by inference of the oceanic lithosphere.

One process that we ignored above is shear heating. The effect of shear heating would be to increase the heating rate and healing time of the serpentinized shear zones, thereby shifting optimal shear zone thicknesses for craton formation towards higher values. This effect, however, is probably insignificant. The energy dissipation rate across the shear zone for a constant plate velocity scales with $\eta V^2/H$ where η is viscosity (Pa s). Only when H is small ($\ll 1$ km) is the effect of shear heating important, thus because H is on the order of kilometers, the effect of shear heating is overall small (any effects should be localized in very narrow shear bands). In any case, there is a limit to how hot the shear zone can become because the frictional heat generated will diffuse into its surroundings. The maximum temperature achieved is the steady state temperature, which scales as $\eta V^2/4k$ where k is the thermal conductivity (W/m °C). As the viscosity of serpentine is likely to be at least an order of magnitude less than 10^{21} Pa s, the steady state temperatures for reasonable plate velocities and km-scale shear zone thicknesses are small enough that we can ignore shear heating.

In all of the above it is important to recognize that our model assumes constant plate velocities during underthrusting. This is clearly a simplification as a more realistic approach would deal with underthrusting in a dynamically consistent way by including the coupled effects of lithosphere buoyancy, mantle rheology, and background convective stresses. However, we believe that the first order effects of underthrusting can be drawn out from our simplified approach. A more detailed approach would go far in refining our hypothesis, but is beyond the scope of this paper.

5. When did cratons form?

In the previous section, we showed for a given plate velocity that there is an optimum shear zone thickness that permits craton formation. The question that now arises is what controls the thickness of the shear zone, e.g. the hydrothermally altered zone within oceanic lithosphere? We envision that serpentinization is introduced into the lithosphere via cracks or faults. The maximum thickness of this alteration zone is limited by the temperature at which serpentine breaks down (essentially the 600–700 °C isotherm; Fig. 3B). This does not necessarily mean that alteration will extend to such depths because the extent of alteration depends on the presence of permeable paths, which in turn depend on the presence of faults and fractures. Although the depth extent of faults and fractures may not reach the depth of the 600–700 °C isotherm beneath ocean basins, these fractures are thought to become more extensive and penetrate to greater depths at the bending zone of subduction zones due to bending-related extension (Ranero et al., 2003; Rüpke et al., 2004). Indeed, in modern subduction zones, serpentine is interpreted to have extended well into the slabs and appears to be limited not by fractures but by the isotherm for serpentine breakdown (Brudzinski et al., 2007).

We thus assume that shear zone thickness H is approximately defined by the depth to the 600–700 °C isotherm in the oceanic lithosphere. Recognizing this, H should scale inversely with the average thermal gradient dT/dz of the oceanic lithosphere just before underthrusting commences, that is, $H \propto T_{sp}/(dT/dz)$. Because the average heat flux q out of oceanic lithosphere is given by $q = k(dT/dz)$, H should be inversely proportional to heat flux q. It follows that if the heat flux out of the Earth has decreased with time (Christensen, 1985; Davies, 1993), the average oceanic heat flux dT/dz should have also decreased, implying that H should have increase over the course of Earth's history (Fig. 1B).

In the inset of Fig. 1B, we plot various thermal histories in the form of global average heat flow based on parameterized convection models forced to fit a present day global heat flux of ~80 mW/m^2 (Richter, 1985; McGovern and Schubert, 1989). Because we are interested in the thermal state of oceanic lithosphere just before underthrusting, the relevant thermal state today is that of oceanic lithosphere at the time of subduction, so the oceanic heat flux of interest here is lower than the global average heat flux. For example, today, the average age of subducting lithosphere is ~100 My old and its heat flux is ~40 mW/m^2 (assuming 100 km thick lithosphere and a mantle potential temperature of 1350 °C), so the global average heat flux today (80 mW/m^2) under-estimates the thickness of the serpentinite layer in subducting oceanic lithosphere (Fig. 1B). While we do not know exactly how the oceanic heat flux at the time of subduction/underthrusting scales with global heat flux, we can reasonably assume that in the earliest Archean, when temperatures were much hotter, the heat flux of through oceanic lithosphere just before subduction and the global heat flux would tend to converge to the same high value if the average age of slabs at subduction is very young (and H becomes very small). Thus, we approximate the secular evolution of H by taking the present day average depth to the 600 °C isotherm for 100 My oceanic lithosphere (H~50 km) and extrapolating linearly to zero in the early Archean (red line in Fig. 1B).

The intersection of the range of permissible shear zone thicknesses H (from Fig. 1B) with the thermal history curve gives the time window for craton formation. For reasonable Archean plate velocities between 10 and 30 cm/yr, craton formation is permitted in the interval between ~2 and 3 Gy. Any earlier, shear zones are too thin and heal too fast for sufficient underthrusting to occur. Any later, shear zones are too thick and take too long to heal and terminate. The intersection of modern plate velocities (~5 cm/yr) with the thermal history curve shows that ~1.2 Gy is the absolute latest time for formation of craton-like continents.

We note that our model at face value might permit cratons to be formed in the Phanerozoic if anomalously hot (low H) oceanic lithosphere was being subducted. However, such Archean-like conditions in the Phanerozoic are not likely to be sustained for long. In addition, recall that the other necessary condition for craton formation is that its building blocks are chemically buoyant enough to resist recycling back into the mantle. With progressive cooling of the Earth, the formation rates of chemically buoyant mantle might have decreased. Indeed, modern oceanic lithospheric mantle becomes negatively buoyant within a few million years just by thermal cooling and might be expected to continue subducting instead of being underthrusted to form cratons (Oxburgh and Parmentier, 1977; Lenardic et al., 2003; Sleep, 2003; Lee et al., 2005; Cooper et al., 2006). Thus, our model, in concert with our understanding of the effects of chemical buoyancy, predicts that the early Archean was too hot, the post-Archean too cold, and the late Archean just right for craton formation. In honor of her experimental spirit, we refer to this problem as the "Goldilocks Problem" (Opie, 1974).

6. Further implications

A natural question is what is the fate of the water released from the serpentine breakdown? If such water was redistributed throughout the cratonic mantle, it could re-weaken the mantle and make it prone to subsequent disturbance. Alternatively, the water could rise upwards to flux the lower crust of the overriding plate, causing it to partially melt and generate felsic melts. Such melts would be channelized to the Earth's surface, generating the continental crust and the mafic residues that eventually make up the lower crust (Fig. 1A). Channelizing of these hydrous melts to the surface would provide an efficient means of transporting serpentine-derived water out of the cratonic mantle so that it does not pervasively re-hydrate the lithospheric mantle. This melting process could also be one of the means by which basaltic crusts are subjected to hydrous intracrustal differentiation in shallow "subduction" environments without the need of a modern-type arc, where melts are generated in the mantle wedge (Foley et al., 2003). One attraction of our model is that building thick cratonic mantle roots would be coupled with differentiation of the crust into felsic compositions. Of course, to satisfy the geochemical observations, the mafic residues must be recycled back into the mantle without recycling the peridotitic root, which we know from Re–Os isotopes to have been preserved (Carlson et al., 2005). Exactly how this could happen remains an important area of research.

Finally, our model *does not preclude* the contribution of arcs and plumes to craton formation. In fact, arc magmatism and plumes undoubtedly existed at this time (Watson and Harrison, 2005; Condie and Benn, 2006; Menneken et al., 2007) and contributed to continent formation. Indeed, not all Archean eclogite xenoliths from cratons can be explained as altered oceanic crust; those with higher Mg contents are similar in composition to garnet pyroxenite cumulates and restites found in the roots of Phanerozoic arcs and may have a similar origin (Horodyskyj et al., 2007). We speculate, however, that if lithospheric underthrusting/accretion (the types of lithospheres could include arc and plume-generated lithospheres as well as oceanic lithosphere) was the dominant process by which cratons were assembled, the window of opportunity was in the late Archean to early Proterozoic. Cratonic lithospheric mantle might thus be a mixture of underthrusted oceanic lithospheric segments and sub-arc lithospheric mantle. However, after the Proterozoic, the importance of lithospheric underthrusting as a continent-forming mechanism may have decreased (due to a decrease in plate velocities and plate buoyancy) allowing the relative importance of lateral accretion of arcs to grow. While underthrusting of oceanic lithosphere may result in the generation of unusually thick (>150 km) lithospheres, the thickening associated with lateral arc accretion may not be as large. If so, this may explain why post-Archean lithospheres are generally thinner than in the Archean. More detailed dynamic studies would go far addressing the plausibility of these ideas.

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