

Geochemical/Petrologic Constraints on the Origin of Cratonic Mantle

Cin-Ty Aeolus Lee

Department of Earth Science, Rice University, Houston, Texas

Cratons are underlain by thick, cold, and highly melt-depleted mantle roots, the latter imposing a chemical buoyancy that roughly offsets the craton's negative thermal buoyancy associated with its cooler thermal state. Petrologic/geochemical predictions of three endmember scenarios for the origin of cratonic mantle are discussed: (1) high-degree melting in a very hot plume head with a potential temperature $>1650^{\circ}\text{C}$, (2) accretion of oceanic lithosphere, and (3) accretion of arc lithosphere. The hot plume scenario predicts that cratonic peridotites were formed by high degrees of melting at very high pressures (≥ 7 GPa), whereas the two accretion scenarios predict an origin by melting on average at lower pressures (< 4 GPa) followed by subsequent transport of these residual peridotites to the greater depths (3–7 GPa) from which they presently derive. Major-element and mildly incompatible trace-element compositions of cratonic peridotite xenoliths suggest a low pressure origin, favoring the two accretion scenarios. The two accretion scenarios are difficult to distinguish geochemically, but one difference is that garnet pyroxenite xenoliths might be more clinopyroxene- and Si-rich in arc environments than in oceanic environments, which would be more olivine-rich (and Si-poor) due to lower pressures of crystallization in oceanic settings compared to arc settings. High-MgO cratonic garnet pyroxenite xenoliths in fact have major-element systematics similar to high-MgO garnet pyroxenites from Phanerozoic continental arcs. Taken at face value, this suggests that at least some component of cratons may have formed by accretion of arc lithosphere. Low-MgO garnet pyroxenite xenoliths from cratons represent either subducted oceanic crust or arc basalts. Cratonic mantle may be formed by a combination of arc and oceanic lithospheric mantle accretion.

1. INTRODUCTION

The term craton, as originally defined, describes that part of a continent that has been more or less tectonically quiescent for billion year timescales [Stille, 1936; Hoffman, 1988; Hoffman, 1989]. Almost by definition, this means that cratons represent the stable Archean and/or Proterozoic cores of continents. A number of somewhat unique characteristics

are now attributed to cratons. For example, most cratons are characterized by low surface heat flows [Pollack and Chapman, 1977; Pollack, 1986; Pollack et al., 1993] and underlain by thick and cold mantle keels composed of highly melt-depleted peridotitic mantle [Boyd and McAllister, 1976; Jordan, 1978, 1988; Finnerty and Boyd, 1987; Canil, 1992; Herzberg, 1993, 1999; Grand, 1994; Griffin et al., 1999; Pearson et al., 2003; Carlson et al., 2005]. In addition, a large portion of cratonic mantle appears to have remained convectively isolated from the asthenosphere over billion year timescales [Richardson et al., 1984; Walker et al., 1989; Pearson et al., 1995a; Carlson et al., 1999]. From these observations has emerged the view that the long-term stability

of cratons is, in some manner, related to the unique composition and thermal structure of their underlying mantle "keels".

Exactly how these variables conspire to ensure craton longevity is not immediately obvious. However, Jordan proposed an elegant and highly influential hypothesis ("isopycnic hypothesis"), which was successful in relating all of these variables [Jordan, 1975, 1978, 1988]. The motivation for this hypothesis was a simple paradox. Cratons are cold and hence should be negatively buoyant, yet they not only remain convectively stable but there appears to be no significant gravity excess over cold continents. To resolve this paradox, Jordan hypothesized that the negative thermal buoyancy (e.g., densification associated with thermal contraction) imposed by the craton's cooler thermal state is exactly canceled at every depth by intrinsic chemical buoyancy imparted by the low densities of melt-depleted peridotites (relative to fertile peridotites) making up cratonic keels [Jordan, 1988]. In essence, Jordan's isopycnic hypothesis states that cratons are in isostatic equilibrium due to a perfect balance between compositional buoyancy and thermal buoyancy at every depth within the thermal boundary layer. It has now been over two decades since the formulation of the isopycnic hypothesis, and while perturbations and slight modifications to this model have been proposed, overall the basic idea of a *present* balance between compositional and thermal buoyancy beneath cratons still holds.

One might interpret the long lifespan of Jordan's hypothesis to imply that there is not much left to do in cratonic studies. However, the isopycnic hypothesis, while solving one paradox, generates more paradoxes. If continents are presently isopycnic, that is, the thermal and chemical buoyancies are perfectly balanced, the question that arises is whether continents have always been isopycnic. If we allow for the possibility that the average temperature of the continental lithosphere has cooled with time relative to the ambient mantle (and this could have been the case because the peridotite that makes up cratonic mantle melted at some point), then it is not likely that continents have always been isopycnic. It then follows that if, at any point in time, continents were not isopycnic, destabilizing buoyancy forces would have existed. For instance, if the continents started off hotter than they are now, they would have tended to pancake out by gravitational collapse (e.g., in the spirit of Bird [1991]), leaving us without a thick keel. The way out of this is to invoke a cratonic mantle that is more viscous from the outset, perhaps as a consequence of its unique chemical composition, which we will discuss later. But if cratonic mantle is so intrinsically viscous that the pancaking effect is suppressed at higher temperatures, we again return to the conclusion that cratons could not always have been in isostatic balance, unless (1) the thermal states of the cratonic mantle and ambient mantle have remained constant through time,

(2) the elevation of the craton has undergone large adjustments in response to the changing thermal state of the craton, and/or (3) the internal density structure of the craton is continually reorganized in such a manner as to maintain isostasy during thermal evolution of the craton. These are only some of the outstanding questions and apparent paradoxes regarding the origin and evolution of cratons.

These paradoxes stem from the fact that, by itself, the isopycnic hypothesis, which is based on isostatic principles, provides no direct insight into craton origin and evolution, both of which inherently involve the dynamic, e.g., nonstatic, history of cratons. The origin and evolution of cratons thus form the theme of this manuscript. Clearly, we will not be able to answer all of the above questions. We can, however, critically examine some of the endmember hypotheses for craton formation and highlight possible observation-based tests of these hypotheses. Craton formation is discussed in the context of three endmember scenarios (Figure 1): formation from the residue of melting in a single hot plume head, formation by accretion/stacking of oceanic lithosphere, and formation by thickening of arc lithosphere. The hope, of course, is that with a better understanding of how cratons form (and evolve), we may come to a better understanding of why cratons are the way they appear now. This paper focuses primarily on the perspective from the geochemistry and petrology of mantle xenoliths, which represent the only direct samples of cratonic mantle.

2. SALIENT FEATURES OF CRATONIC MANTLE

Comprehensive reviews on the thermal and compositional structure of cratonic mantle are already available [Griffin *et al.*, 1999, 2003; O'Reilly *et al.*, 2001; Pearson *et al.*, 2003; Carlson *et al.*, 2005] and readers are urged to refer to these papers for more details. Our goal here, however, is to briefly review those features of cratonic mantle that are relevant to the foregoing theme of this manuscript. We also outline the assumptions and uncertainties in these thermal and compositional constraints.

2.1. Defining the Thickness of the Cratonic Mantle

The thickness of cratonic mantle is probably the one parameter that leads to the most confusion in scientific discussions, particularly across disciplines. This confusion stems from different definitions and approaches in estimating the thickness of cratonic mantle. Similar confusion arises from the term "lithosphere", which strictly speaking should only be used to describe the strong mechanical boundary layer at the surface of the Earth but has also been used to describe chemical boundary layers and thermal boundary layers without proper specification. For these reasons, the choice has

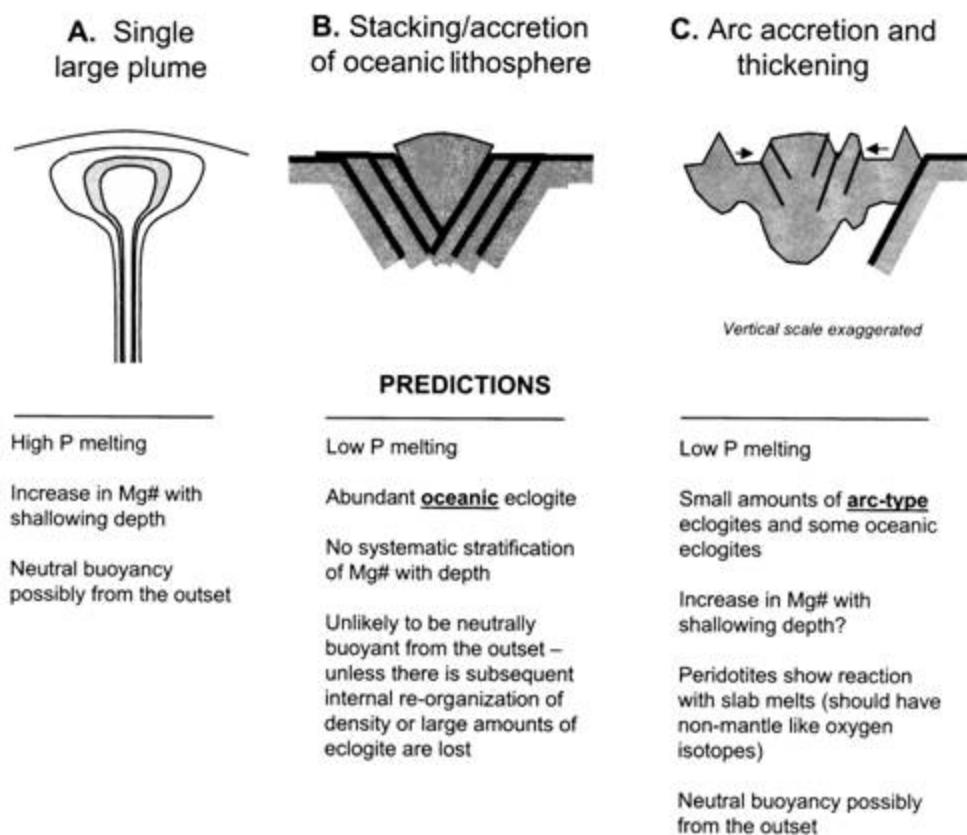


Figure 1. Endmember cartoons for craton formation scenarios and their associated predictions. (A) Plume head, (B) oceanic lithosphere accretion, (C) arc accretion/thickening. The scenarios are not mutually exclusive. Vertical direction is greatly exaggerated.

been made to use the terms, “mechanical”, “chemical” and “thermal boundary layers”, for those instances where we wish to be very specific as to what definition of cratonic mantle thickness is of interest. When speaking in very general terms, “lithosphere” is used as a catch-all phrase for these three boundary layers.

We begin with the most unambiguous definition of the thickness of cratonic mantle, that is, to equate the thickness to that of the thermal boundary layer underlying cratons. The thermal boundary layer at the surface of the Earth is defined to be that part of the uppermost mantle across which the mode of vertical heat transfer transitions from thermal buoyancy-driven advection to conduction. The base of the thermal boundary layer can be uniquely defined with respect to the average potential temperature of the Earth’s mantle by defining the base of the thermal boundary layer to correspond to the depth at which the temperature reaches some pre-defined fraction of the ambient mantle temperature, e.g., 0.9 or 0.95 times that of adiabatically upwelling mantle (Figure 2). Thus, if the temperature of the convecting mantle is known at depth (inferred from extrapolating the mantle potential temperature

to increasing depths along a solid adiabatic gradient), the thickness of the thermal boundary layer is known.

Because of the exponential increase in viscosity with decreasing temperature, the upper part of the thermal boundary layer must be highly viscous and hence strong. If cold enough, the viscosity will be high enough that this part of the thermal boundary layer will behave as a very viscous, and hence strong, layer. This layer will be effectively isolated from secondary convection and corresponds to a viscously defined mechanical boundary layer. The viscous mechanical boundary layer will be thinner than the thermal boundary layer. It is important to note that a continent’s “elastic” thickness, i.e., the effective thickness over which the flexure of a continent under a surface load can be modeled as an elastic plate, does not necessarily equate with the viscously defined mechanical boundary layer, which will, in general, be thicker.

In the next subsection, we will learn that cratonic mantle is compositionally distinct from that of the underlying convecting mantle (asthenosphere). This compositionally distinct layer is referred to as the chemical boundary layer. An important

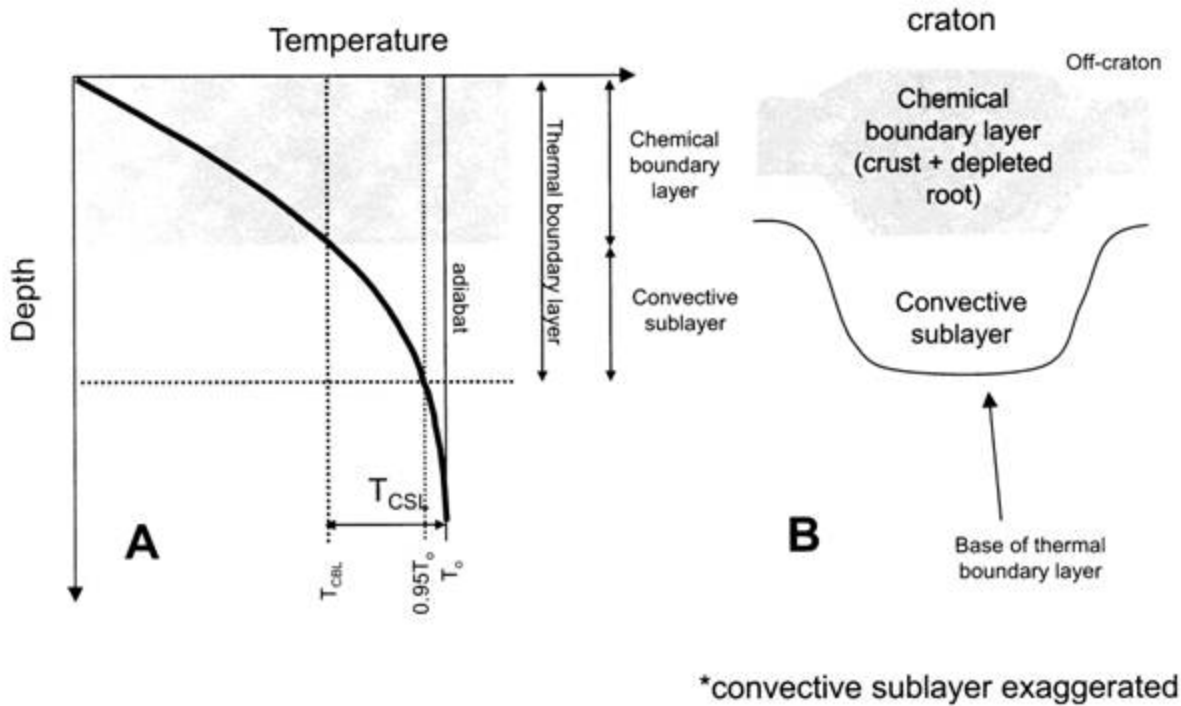


Figure 2. Schematic diagrams illustrating the relationships between the chemical boundary layer, thermal boundary layer, and convective sublayer (CSL) [Cooper *et al.*, 2004]. (A) Depth–temperature diagram showing that the base of the thermal boundary layer is defined by the point at which the temperature reaches 0.95 that of the ambient mantle temperature, T_0 (represented by the mantle adiabat). (B) Schematic cartoon corresponding to the diagram in (A); not to scale.

feature of the chemical boundary layer beneath cratons is that it is characterized by chemically buoyant peridotite as a consequence of high degrees of melt extraction. In the perfectly isopycnic condition proposed by Jordan, the negative thermal buoyancy associated with the craton's cooler thermal state relative to the ambient mantle, is exactly offset by an intrinsic compositional buoyancy at every depth. In this case, the base of the chemical boundary layer will coincide exactly with that of the thermal boundary layer. This equating of the two boundary layers is Jordan's "tectosphere".

If the cratonic mantle is not perfectly isopycnic, the base of the chemical boundary layer need not coincide precisely with the base of the thermal boundary layer. In particular, a possible by-product of extensive melt extraction is that the peridotite residues that make up the chemical boundary layer should have, at least initially, been dehydrated [Pollack, 1986; Hirth and Kohlstedt, 1996; Aubaud *et al.*, 2004]. Dehydration would potentially increase the intrinsic viscosity of the mantle residuum by two orders of magnitude [Pollack, 1986; Hirth and Kohlstedt, 1996]. The increase in viscosity coupled with the compositional buoyancy could help isolate this pre-existing chemical boundary layer from secondary convection [Lenardic *et al.*, 2003; Cooper *et al.*, 2004; Dixon *et al.*, 2004; Lee *et al.*, 2005b]. In this case, the chemical

boundary layer would equate with the viscously defined mechanical boundary layer (although in general the mechanical and chemical boundary layers do not necessarily have to coincide), not the thermal boundary layer, which is thicker. Underlying this strong chemical boundary layer would reside a convectively active sublayer ("convective sublayer") that makes up the base of the thermal boundary layer [Cooper *et al.*, 2004; Lee *et al.*, 2005b].

It is now clear that different approaches in determining the thickness of cratons will lead to different results because each approach is sensitive to different physical properties. For example, gravity and flexural studies of cratons provide constraints on elastic thickness. Seismic tomography will be most sensitive to the thermal boundary layer due to the high sensitivity of seismic velocities to temperature variations. In fact, if we define the thickness of the thermal boundary layer to be 0.95 times that of the ambient convecting mantle, seismic tomography will overestimate the thickness of the thermal boundary layer because it will still be sensitive to the remaining 5% allowable variation in temperature. Likewise, heat flow measurements and the assumption of steady-state conductive heat transfer provide information primarily on the thickness of the thermal boundary layer. Thermobarometric studies of mantle xenoliths may provide the most direct

estimates of chemical and thermal boundary layer thicknesses, provided chemical equilibrium was maintained up until the time these xenoliths were erupted.

Below, we describe different approaches in estimating the thicknesses of cratons. What is important is that although each of these approaches comes with its own uncertainties, a consistent picture of craton thickness is emerging from the collective.

2.1.1. Thickness constraints from heat flow. The heat flow through cratons is inferred by measuring the temperature gradient down a borehole or well and multiplying it by the thermal conductivity of the rock. In general, cratons are characterized by surface heat flows between 30 and 50 mW/m², which is lower than the 50–80 mW/m² typically seen in Phanerozoic regions [Pollack and Chapman, 1977; Nyblade and Pollack, 1990; Pollack et al., 1993; Jaupart and Mareschal, 1999; Nyblade, 1999]. Assuming one-dimensional steady-state and a purely conductive lid, surface heat flow can be used to construct a conductive geotherm for cratonic mantle, provided one has some knowledge (which in many cases is poorly constrained) of the thermal conductivity and concentration of heat-producing elements (U, Th, and K) in the crust and lithospheric mantle [Pollack and Chapman, 1977]. Examples of extrapolating a “skin” measurement (surface heat flow) to depth are shown in Figure 3, where heat production in cratonic mantle has been estimated from the average concentrations of K, U, and Th in mantle xenoliths. It can be seen that the low surface heat flow out of cratons implies that cratons are probably underlain by a thick and cold mantle keel (roughly 200 km), whereas that beneath Phanerozoic regions are much thinner and warmer.

Geotherm calculations, however, should be interpreted carefully because of the great number of assumptions inherent in such calculations. One of the largest uncertainties in extrapolating geotherms is the concentration of heat-producing elements at depth, particularly in the lithospheric mantle. Attempts have been made to use mantle xenoliths as constraints [Rudnick et al., 1998], but this approach is not without uncertainty as the abundances of U, Th, and K in mantle xenoliths vary over an order of magnitude due to melt depletion and re-enrichment processes. Additional uncertainties include the validity of assuming steady-state one-dimensional heat transport and purely conductive heat transfer [Jaupart and Mareschal, 1999]. For example, there is no doubt that cratons drift through the mantle, and because there are significant lateral variations in the temperature of the mantle (e.g., upwellings and downwellings), the transfer of heat through cratons, strictly speaking, cannot be in steady-state. Moreover, the timescales for radioactive decay of U, Th, and K are also on the same order of magnitude as thermal diffusion timescales, such that if there are nonnegligible amounts

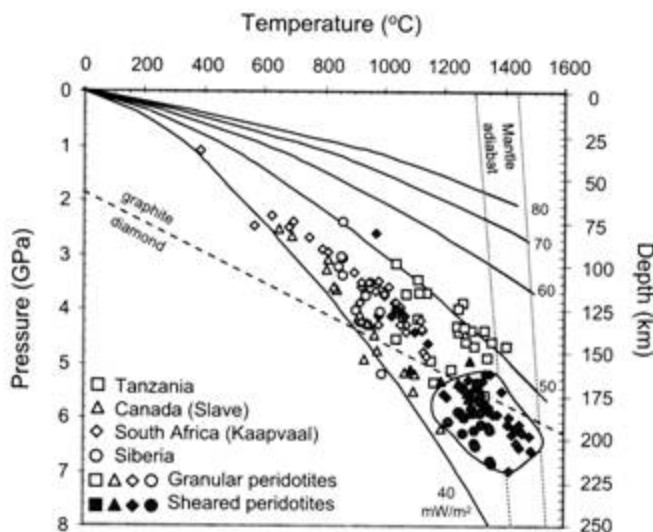


Figure 3. Thermobarometry data of xenoliths (using the Brey and Kohler [1990] Al in orthopyroxene coexisting with garnet barometer and the two-pyroxene thermometer) from different cratons combined with conductive geotherms inferred from surface heat flow (heat productions assumed for upper continental crust, middle continental crust, lower continental crust, and cratonic mantle are 1.1, 0.5, 0.1, and 0.02 $\mu\text{W}/\text{m}^3$, respectively). Open symbols represent low-temperature granular peridotites; colored symbols represent high-temperature sheared peridotites. The field of high-temperature sheared peridotites has been roughly outlined. Shaded vertical region represents range of mantle adiabatic temperatures at present.

of heat-producing elements in the lithospheric mantle (note that in the earlier Pollack and Chapman [1977] family of geotherms, heat production in the lithospheric mantle was assumed to be zero), the current thermal state in cratons cannot be in steady-state [Michaut and Jaupart, 2004]. Additionally, if there are any topographic variations at the base of a craton, heat transfer cannot be one-dimensional [Lenardic and Moresi, 2000]. Finally, the base of a craton is likely to be involved in secondary convection, and as a consequence, the assumption of a steady-state conductive geotherm must break down at great depths [Lenardic et al., 2000].

2.1.2. Thickness constraints from xenolith thermobarometry. The thermal state of cratonic mantle can also be constrained from xenolith thermobarometry, which is based on the principle that the compositions of different mineral phases in a xenolith reflect a frozen-in equilibrium state, generally assumed to be the pressure and temperature at which the xenolith resided in the mantle just prior to eruption of the host kimberlite (although this does not always have to be the case). Thus, if it can be assumed that xenoliths are in thermodynamic equilibrium during their residence in the mantle,

experimentally calibrated relationships between the activities of specific components within mineral phases can be used to estimate temperature and pressure [Ellis and Green, 1979; Harley and Green, 1982; Finnerty and Boyd, 1987; Griffin et al., 1989; Brey and Kohler, 1990]. The results of such calculations on cratonic peridotites are well-known but are reproduced here for completeness. Note that there are no suitable barometers for spinel peridotites. Illustrated in Figure 3 are thermobarometric data for mantle xenoliths from the Tanzanian [Rudnick et al., 1994; Lee and Rudnick, 1999], Slave (Canada; [Kopylova et al., 1999a, b; Kopylova and Russell, 2000]), Kaapvaal (South Africa; [Boyd, 1987; Finnerty and Boyd, 1987; Boyd et al., 1993; Rudnick et al., 1998]), and Siberian cratons [Boyd et al., 1997], all of which are Archean in age. The majority of cratonic mantle xenoliths derive from pressures between 3 and 7 GPa (corresponding to depths between ~90 and 210 km) and temperatures between 600°C and 1500°C. Cratonic peridotites thus define a thermal state which is well below the mantle adiabat and consistent with that determined from cratonic surface heat flow (Figure 3). Tanzanian xenoliths record slightly higher temperatures for a given depth, which may be due to the fact that the Tanzanian xenoliths sample the edge of the Tanzanian craton, which is bound by a Proterozoic mobile belt and is currently being affected by the southward-propagating East African Rift [Nyblade and Brazier, 2002]. The coincidence of the xenolith thermobarometry data with the heat flow-constrained geotherms is remarkable, given the potential uncertainties associated with both approaches.

Xenolith thermobarometric data can be used to reconstruct the thermal state of the cratonic mantle and hence provide information on both the thicknesses of the chemical and thermal boundary layers. This assumes, of course, that the last equilibrium states recorded by the xenoliths were set at the same time. In general, cratonic mantle xenoliths can be subdivided into roughly two textural groups [Boyd, 1987]. One group consists of granular peridotites. These typically have equilibration pressures and temperatures lower than 5.5 GPa and 1200°C. A second group consists of sheared or porphyroblastic textures, and these typically have equilibration pressures and temperatures exceeding 5.5 GPa and 1200°C. The low temperature granular peridotites further differ from the high temperature sheared peridotites in that the former are highly melt-depleted whereas the latter tend to be more fertile, some approaching the fertility of asthenospheric mantle. The transition (150–175 km) between the two groups most likely defines the base of a depleted mantle root, that is, the chemical boundary layer. Because the high-temperature sheared peridotites approach the temperature (~1500°C) of the mantle adiabat at depths of ~200 km, it must be concluded that the thermal boundary layer beneath cratons does not greatly exceed ~200 km (Figure 3), as previously pointed

out by Rudnick et al. [1998]. Thus, cratons are underlain by a ~200-km-thick thermal boundary layer, composed of a 150–175-km-thick chemical boundary layer beneath which lies a thin convective sublayer that makes up the base of the thermal boundary layer (Figure 3).

If, on the other hand, the high-temperature sheared peridotites and low-temperature granular peridotites equilibrated at different times, it is not exactly clear whether the thermobarometric constraints of these two xenolith groups can be pieced together as one continuous geotherm. It has been suggested that the high-temperature sheared peridotites may in fact reflect transient conditions associated with the kimberlite eruption or its precursor. This is based on the fact that many of these xenoliths actually show local textural and chemical disequilibria [Smith and Boyd, 1992; Griffin et al., 1996, 1999]. If the high-temperature sheared peridotites have been affected by very recent heating, their estimated temperatures may overestimate the temperature at the base of the craton prior to heating. This would result in a slight underestimate of the thickness of the thermal boundary layer.

2.1.3. Thickness constraints from seismology. Seismic studies represent a powerful tool for constraining the current structure of cratonic mantle. Because seismic velocities are very sensitive to temperature (e.g., the sensitivity of V_P and V_S to temperature is ~0.7%/100°C and ~1%/100°C, respectively; [Lee, 2003; James et al., 2004]), velocity anomalies are dominated by thermal variations. Seismic tomographic studies of cratons often show the presence of positive anomalies ("blue") extending to depths of 300–400 km beneath cratons [Ritsema et al., 1998; Simons et al., 1999; James et al., 2001]. If these positive velocity anomalies correspond to the depth extent of cratonic mantle, there is a discrepancy with the depth extent of cratons as determined by xenolith thermobarometry.

This discrepancy can be reconciled in two ways. One way is to call upon disequilibrium and transient effects in some xenoliths as discussed above. Another way is to call upon the possibility of smearing effects in seismic tomography and the high sensitivity of seismic tomography to temperature variations. For example, seismic tomography studies of the Kaapvaal craton show "blue" P-wave velocity anomalies extending down to depths between 300 and 400 km [James et al., 2001; Shirey et al., 2002]. But at these depths, the velocity perturbations are only 0.5%. Using our above definition of thermal boundary layer thickness, that is, 0.95 that of the ambient mantle temperature, a temperature difference of roughly 75°C still persists below the defined base of the thermal boundary layer, assuming an ambient mantle temperature at 200 km depth of ~1500°C. This temperature difference corresponds to a ~0.5% variation in P-wave velocity. Without even considering smearing effects, the foregoing suggests

that the “blue” regions in tomographic studies overestimate the true thickness of cratonic mantle.

More recent seismic approaches for estimating craton thicknesses are now appearing to be more consistent with the xenolith constraints. For example, S–P conversions can be used to map out the topography of the transition zone: Beneath Kaapvaal, there appears to be no measurable perturbation to the 410-km seismic discontinuity as would be expected if these cold keels penetrated the transition zone [Niu *et al.*, 2004]. In fact, when this observation is considered in the context of geodynamic constraints, the lack of an elevation of the 410-km discontinuity suggests that, at least for Kaapvaal, an upper bound to the thickness of the thermal boundary layer is ~300 km [Cooper *et al.*, 2004; Niu *et al.*, 2004]. Finally, recent surface-wave anisotropy studies of cratons also seem to suggest thinner keels (~200 km) to cratons [Gung *et al.*, 2003]. Thus, despite the uncertainties and assumptions that go with each of these different approaches in estimating craton thicknesses (heat flow, xenolith

thermobarometry, seismology), a consistent picture seems to be emerging.

2.2. Composition of the Cratonic Mantle and Chemical Buoyancy

One of the unique features of cratonic mantle is that it is dominated by high Mg# (molar $\text{Mg}/(\text{Mg} + \text{Fe}) \times 100$) harzburgitic peridotite (91–94), e.g., olivine- and orthopyroxene-dominated lithologies [Boyd and Mertzman, 1987; Boyd, 1989; Winterburn *et al.*, 1989; Boyd *et al.*, 1993, 1997; Rudnick *et al.*, 1994; Bernstein *et al.*, 1998; Griffin *et al.*, 1999; Lee and Rudnick, 1999; Pearson *et al.*, 2003]. This can be seen in Figures 4–6. Cratonic peridotites are impoverished in clinopyroxene, Ca, Fe, and Al compared to fertile asthenospheric mantle and Phanerozoic peridotites, which are lherzolitic in composition (e.g., clinopyroxene 10–20%), have lower Mg#s (88–89 for fertile asthenosphere and 88–91 for Phanerozoic peridotites), and have higher Ca, Fe, and Al

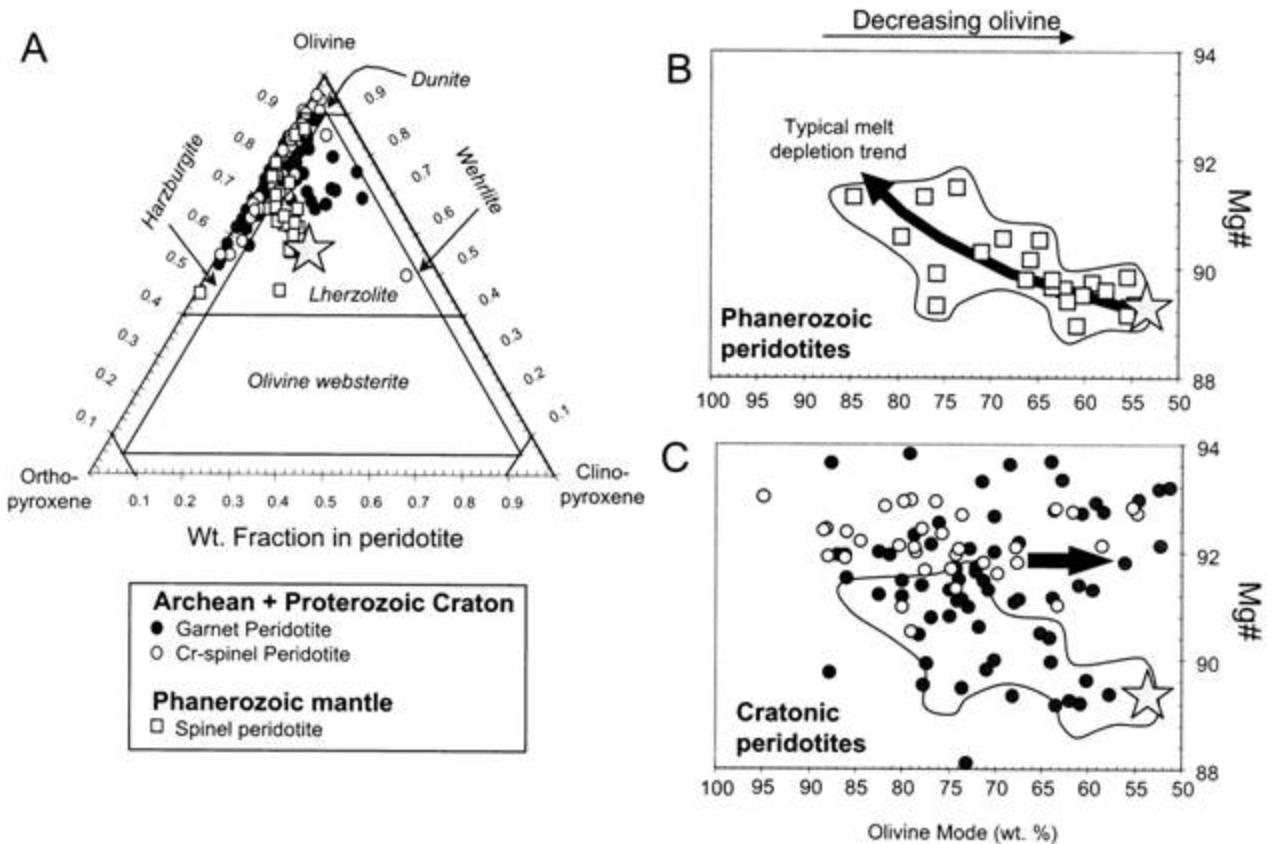


Figure 4. (A) Classification of peridotite lithologies based on weight fraction of olivine, orthopyroxene, and clinopyroxene. Star represents estimated composition of fertile asthenospheric mantle. (B) Mg# (molar $\text{Mg}/(\text{Mg} + \text{Fe}) \times 100$) versus olivine mode in weight percent (note reverse axis) for Phanerozoic peridotites. Arrow denotes the expected trajectory of partial melting residues. (C) Same plot as in B but cratonic peridotites have been superimposed. A number of cratonic peridotites show anomalously high levels of orthopyroxene (low olivine mode) due to Si enrichment (large arrow pointing right).

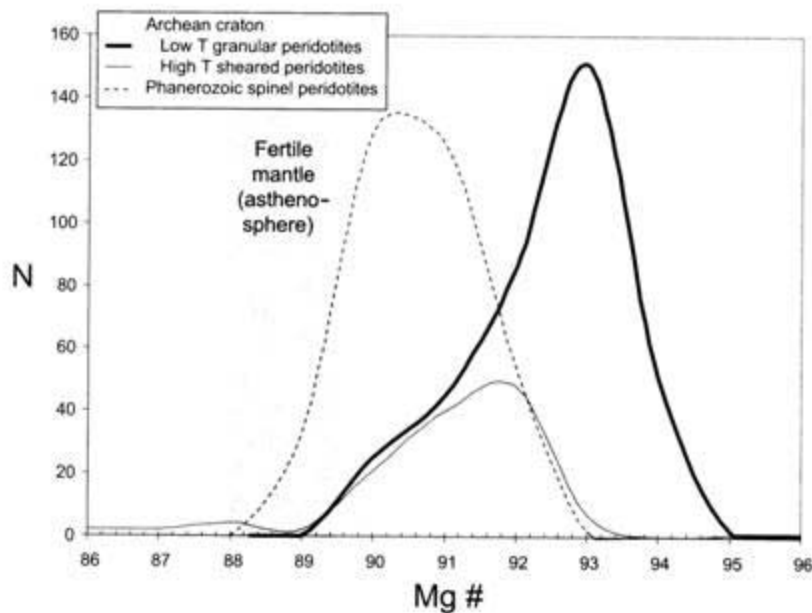


Figure 5. Mg# histogram of cratonic peridotites (low-temperature granular and high-temperature sheared), Phanerozoic mantle (continental lithospheric mantle xenoliths and obducted ophiolite massifs), and estimated fertile asthenospheric mantle (shaded vertical region).

(Figures 4–6). The depletion in clinopyroxene, Ca, Fe, and Al in cratonic peridotites relative to the asthenospheric mantle is most likely due to large degrees of melt extraction. Assuming all cratonic peridotites were originally derived from a more or less homogeneous convecting mantle having a primitive mantle-like composition (at least in terms of major elements and moderately incompatible trace elements), comparisons with partial melting experiments [Walter, 1998, 1999; Herzberg, 1999, 2004] suggest that most cratonic mantle peridotites have had 30–50% melt extracted (Figure 6). Only those cratonic peridotites that derive from the very base of the thermal boundary layer, e.g., the “high-temperature sheared peridotites” have fertile (lherzolitic) compositions (Mg# = 88–91; Figure 5). The higher fertility of the high-temperature sheared peridotites is due either to refertilization at the base of the thermal boundary layer [Boyd, 1987; Smith and Boyd, 1992; Pearson *et al.*, 1995b] or to the possibility that such peridotites have not undergone extensive partial melting.

One subset of cratonic peridotites (Figure 4c) is made up by those that appear to be anomalously Si-rich for their high Mg# and low Ca, Fe, and Al contents [Boyd, 1989; Kelemen *et al.*, 1998]. Melt depletion of a primitive mantle-like starting composition results in a subtle decrease in SiO₂ [Walter, 1998; Herzberg, 1999, 2004]; therefore, the high SiO₂ contents (or equivalently high orthopyroxene mode) for these highly melt-depleted harzburgites suggest that either silica

has been added or their initial starting compositions were Si-rich to begin with. It is largely assumed that the latter is not the case and, as such, most debates have centered around finding a mechanism for the Si-enrichment [Canil, 1992; Herzberg, 1993; Rudnick *et al.*, 1994; Kelemen *et al.*, 1998; Herzberg, 1999; Smith *et al.*, 1999]. One possibility is that harzburgitic residues reacted with silicic melts or fluids, perhaps from partial melting of a subducting slab. This process converts olivine to orthopyroxene, increasing the bulk SiO₂ content [Kelemen *et al.*, 1998; Smith *et al.*, 1999]; Si-enrichment has in fact been observed in a few subduction zone peridotites [Ertan and Leeman, 1996; Smith *et al.*, 1999; McInnes *et al.*, 2001; Arai *et al.*, 2004], possibly providing additional support for this hypothesis. Alternatively, the excess SiO₂ is derived by cumulate addition of orthopyroxene at depth, perhaps associated with melting in a large plume head [Herzberg, 1993, 1999; Lee *et al.*, 2003].

Regardless of how these Si-rich cratonic peridotites formed, an important conclusion is that the highly melt-depleted peridotites (including those with excess SiO₂) have lower intrinsic densities than fertile convecting mantle [Boyd and McAllister, 1976; Lee, 2003]. This is due to the decrease in clinopyroxene and garnet mode as well as the decreasing Fe content in olivine and orthopyroxene. Figure 7a shows how standard temperature and pressure (STP; 25°C and 1 atm) densities calculated for real cratonic peridotites decrease with increasing Mg#. Figure 7b shows the chemical

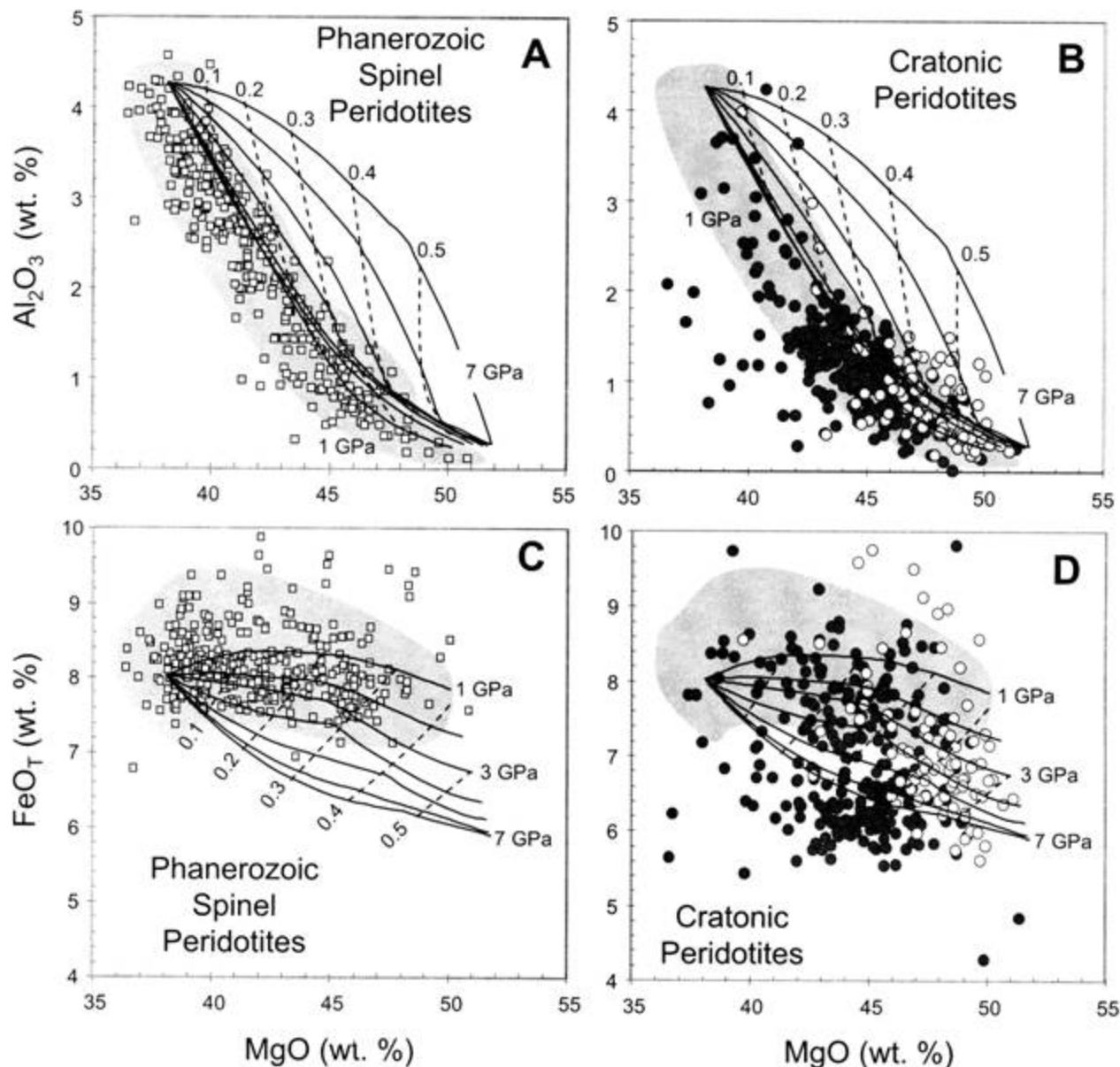


Figure 6. (A) Whole-rock Al_2O_3 versus MgO in Phanerozoic spinel peridotites. Solid curves represent the results of equilibrium melting experiments of a fertile lherzolite starting composition (KR-4003) from 1 to 7 GPa at 1-GPa intervals [Walter, 1998, 1999; Herzberg and O'Hara, 2002; Herzberg, 2004]. Dashed lines represent melt fraction contours at increments of 0.1. (B) Same plot as in (A), but cratonic peridotites have been superimposed. Shaded region represents Phanerozoic spinel peridotites from (A). Dark circles represent garnet-bearing cratonic peridotites. Open symbols represent Cr-spinel-bearing peridotites from the Greenland and Tanzanian cratons. (C) Phanerozoic spinel peridotites plotted in terms of FeO_T (total Fe as FeO) along with equilibrium melting experiments shown in (A). (D) Same as (C) but cratonic peridotites are superimposed.

buoyancy (density deficit) of cratonic mantle that is necessary to exactly offset its negative thermal buoyancy associated with its cooler thermal state (e.g., the isopycnic line). This chemical buoyancy curve can be converted into equivalent $\text{Mg}\#$ (Figure 7c) by using the empirical relationship shown in

Figure 7a. It can then be seen from Figures 7c and 8 that the average $\text{Mg}\#$ of cratonic mantle between 3 and 7 GPa (equilibration pressure range over which most granular xenoliths are sampled) falls within the error range for neutral buoyancy (though perfect isopycnicity is not matched). Phanerozoic

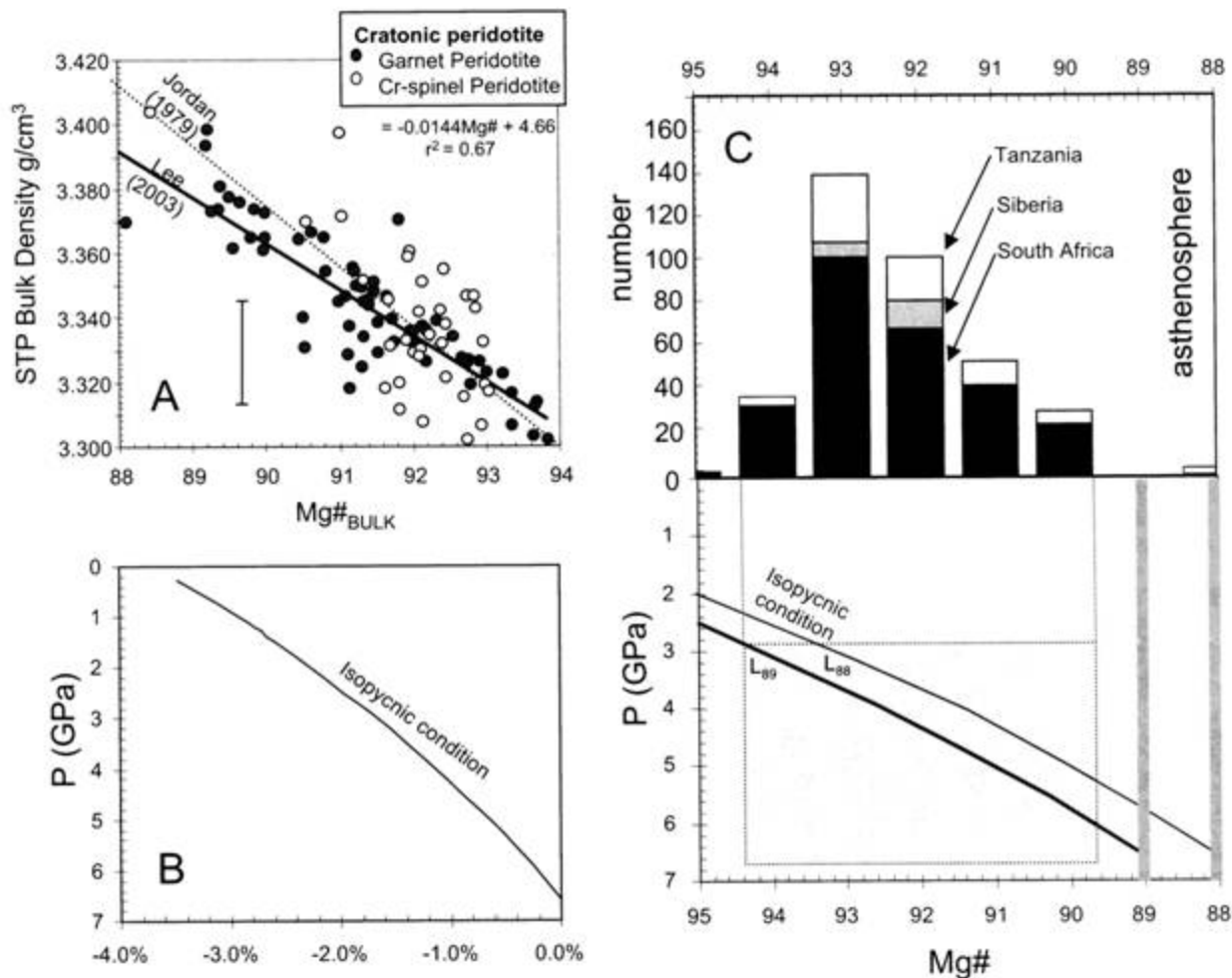


Figure 7. A. Normative (standard temperature and pressure) densities for cratonic peridotites versus Mg# [Lee, 2003]. Vertical bar represents largest estimated error. Line labeled "Lee (2003)" is a regression through the data. Line labeled "Jordan (1979)" represents Jordan's earlier parameterization based on estimated (rather than measured) mineral compositions [Jordan, 1979]. (B) Compositional buoyancy required to maintain neutral buoyancy at every depth for a craton characterized by a thermal state corresponding roughly to the xenolith geotherm shown in Figure 3 (roughly corresponding to the 45 mW/m² geotherm). (C) Top panel shows Mg# histogram of cratonic peridotites. Bottom panel shows the necessary Mg# required to achieve neutral buoyancy for the xenolith geotherm shown in Figure 3 and fertile asthenospheric mantle having Mg#s of 88 and 89 (L88 and L89). Shaded region represents the range of Mg#s and equilibration pressures seen in cratonic peridotites.

oceanic mantle, by contrast, becomes negatively buoyant within 20 Ma due to its less depleted nature (Figure 8).

2.3. Age and Longevity of Cratonic Mantle

One of the most important discoveries was that cratonic mantle is long-lived. The longevity of continental mantle was originally suggested on the basis of highly evolved Sr and Nd isotopic compositions of alkalic basalts and lamproiites derived from the continental lithospheric mantle [Menzies, 1989; Hawkesworth et al., 1990]. These observations suggested that

continental lithospheric mantle resided in an incompatible-element-enriched environment isolated from the convecting mantle for long periods of time. Subsequently, similar isotopic studies of mantle xenoliths from cratons [Menzies, 1983; Cohen et al., 1984; Walker et al., 1989], as well as diamond inclusions [Richardson et al., 1984], also suggested long-term isolation from the convecting mantle. The problem, however, is that Sr and Nd (and Pb) isotopes are based on incompatible element systems. Their isotopic compositions reflect the cumulative effects of metasomatic processes and, in general, do not yield information about the original partial melting

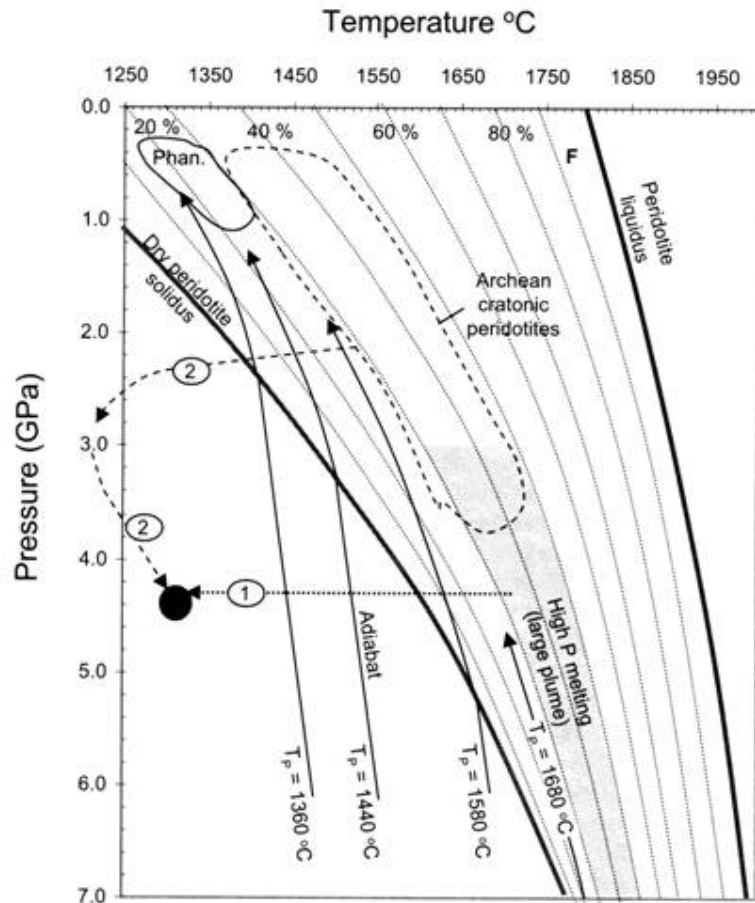


Figure 8. Schematic pressure–temperature diagram showing the dry peridotite solidus and peridotite liquidus along with constant melting degree F contours (in %) shown as dashed lines [Katz *et al.*, 2003]. Arrowed solid curves represent the P – T path of adiabatic decompression melting [McKenzie, 1984; McKenzie and Bickle, 1988]: The subsolidus leg represents the solid–mantle adiabat and the remaining part represents the melting adiabat, assuming the melt stays with the residue. Potential temperatures for each melting adiabat are denoted. Shaded region between 3 and 7 GPa refers to the range in pressures and melting degree F represented by cratonic peridotites. Shaded region at lower pressures (Phan) represents the region characterized by Phanerozoic spinel peridotites from various tectonic settings. Dashed open region for Archean cratonic peridotites reflects the preferred interpretation that cratonic peridotites formed by melting at pressures less than ~4 GPa. Dashed arrowed lines show hypothetical pressure–temperature paths: (1) isobaric cooling of plume head mantle residue, (2) low-pressure melt residues cooled isobarically and then tectonically transported to greater depths. Dark shaded circle at the endpoints of curves 1 and 2 is a schematic representation of present-day equilibration temperatures and pressures of cratonic peridotites.

event. Thus, these isotopes systems, although they suggest long-term isolation, cannot pin down exactly how long cratonic mantle has been convectively isolated.

Re–Os isotopic studies of mantle xenoliths have arguably provided the best (but still crude) constraint on the timing of cratonic mantle formation [Walker *et al.*, 1989; Carlson and Irving, 1994; Pearson *et al.*, 1995a, b, c; Reisberg and Lorand, 1995; Handler *et al.*, 1997; Shirey and Walker, 1998;

Carlson *et al.*, 1999, 2005; Chesley *et al.*, 1999; Lee *et al.*, 2001b; Shirey *et al.*, 2002]. In contrast to Sr, Nd, and Pb, osmium is a highly compatible element and is therefore less easily overprinted by metasomatic processes (although Re can be overprinted). This makes Re–Os isotopic systematics a better recorder of initial melting conditions. Re–Os isotopic studies have shown that cratonic peridotites underwent melt extraction billions of years ago. Moreover, in many cases,

the melt extraction ages (Re–Os model ages) fall within error of the age of the overlying crust, confirming that the present cratons are underlain by long-lived keels dating back to the time of craton formation.

2.4. Garnet Pyroxenites and Eclogites

Although peridotites typically dominate cratonic xenolith suites, garnet pyroxenite xenoliths (a term used to describe rocks composed of garnet and pyroxenes) are occasionally found [MacGregor and Carter, 1970; Helmstaedt and Doig, 1975; Neal *et al.*, 1990; Jacob *et al.*, 1994; Fung and Haggerty, 1995; Beard *et al.*, 1996; Pyle and Haggerty, 1998; Jacob and Foley, 1999; Barth *et al.*, 2001, 2002; Taylor *et al.*, 2003]. In South Africa, garnet pyroxenite xenoliths make up <1% of the xenolith population, peridotites making up the rest [Schulze, 1989]. However, in some rare instances, garnet pyroxenites dominate, such as in the Koidu kimberlite xenolith suite in Sierra Leone [Fung and Haggerty, 1995]. Whether xenolith demographics ever reflect the true distribution of lithologies in the mantle is unclear as it is impossible to prove that kimberlites randomly sample the underlying mantle. However, the high densities of garnet pyroxenites compared to peridotites make it unlikely (at least in the author's opinion) that such lithologies dominate cratonic mantle; if they did, cratons would be negatively buoyant.

Garnet pyroxenite xenoliths in cratons can, in general, be subdivided into high- and low-MgO groups [Fung and Haggerty, 1995; Barth *et al.*, 2001, 2002]. The low-MgO suites are dominated by true eclogites, that is, garnet pyroxenites containing only garnet and omphacitic clinopyroxene and no orthopyroxene. These low-MgO garnet pyroxenites are interpreted to represent fragments of subducted oceanic crust and oceanic gabbroic cumulates. This is based on the following: (1) except for slightly low SiO₂ contents, their major element compositions are very similar to that of mid-ocean ridge basalts [Rollinson, 1997], and (2) they have oxygen isotopic compositions suggestive of hydrothermal alteration [Neal *et al.*, 1990; Jacob and Foley, 1999; Barth *et al.*, 2001], which presumably could only have occurred at or near the surface of the Earth. In addition, many diamonds from kimberlites not only contain mineral inclusions of low-MgO eclogite affinity but also have extremely light carbon isotopic compositions interpreted to have a biogenic origin [Kirkley *et al.*, 1991; Shirey *et al.*, 2002; Pearson *et al.*, 2003]. It has been argued that the low SiO₂ of some of the low-MgO eclogites is due to the fact that they represent the residues of slab melting [Rollinson, 1997; Barth *et al.*, 2001] or of the melting of over-thickened basaltic crust [Foley *et al.*, 2002].

The high-MgO suite, on the other hand, consists of eclogites and garnet websterites. They have MgO contents too high to represent "typical" basaltic oceanic crust. They have

instead been interpreted as primitive cumulates, such as olivine–gabbro cumulates in oceanic settings [Barth *et al.*, 2002; Taylor *et al.*, 2003]. These xenoliths do not have anomalous oxygen isotopic compositions [Barth *et al.*, 2002]. In a later section, we will present an alternative hypothesis that explains these high-MgO eclogites as mafic cumulates associated with basaltic arc volcanism.

2.5. Seismic Heterogeneities in Cratonic Mantle

Recent seismic studies have revealed the presence of internal structure within cratonic mantle. For example, active source seismic experiments in and around the Slave craton have shown the existence of a pronounced dipping reflector on the edge of the craton [Cook *et al.*, 1997; Bostock, 1998, 1999]. This has been interpreted to represent a fossil subduction zone or an accreted sliver of oceanic lithosphere. More recently, teleseismic studies using receiver functions to map out distinct S–P conversions have revealed hints of dipping or possibly irregular interfaces within the Kaapvaal cratonic mantle [Levander *et al.*, 2005]. These seismic discontinuities may turn out to be the remnants of accreted oceanic lithosphere.

2.6. Summary

In summary, cratons are underlain by a thick thermal boundary layer not exceeding 300 km and probably falling within the range of 200–250 km. This thermal boundary layer is composed of a thick chemical boundary layer (150–175 km thick), which is made chemically buoyant due to high degrees of melt depletion. Preservation of ancient isotopic signatures indicates that, except for continent drift, cratons must be largely isolated from the convecting mantle. This implies that the chemical boundary layer underlying cratons must also be inherently strong, and it is speculated that this is the result of dehydration accompanying high degrees of melting. Beneath this strong chemical boundary layer lies the convectively active portion of the craton's thermal boundary layer, across which vertical heat transfer transitions from convection-dominated to conduction-dominated.

3. ENDMEMBER SCENARIOS FOR CRATON FORMATION

Craton formation scenarios can be classified into three endmembers. These include the formation of thick cratonic lithosphere from the mantle residuum of melting in a large mantle plume head (Figure 1a), by the accretion/stacking of oceanic lithosphere (Figure 1b) and/or by collision of arc lithosphere segment (Figure 1c). These endmember scenarios

are not necessarily mutually exclusive. Below, each of these scenarios is described in more detail, emphasizing their specific geochemical and petrologic predictions.

3.1. Formation of Cratonic Mantle in a Single Hot Plume Head

The generation of a thick, highly melt-depleted mantle root requires high degrees of melting. At present, the high degrees of melt extraction characteristic of cratonic mantle (>30%) are represented only by localized hotspot magmatism or by the shallowest portions of the melting column associated with mid-ocean ridge basalt magmatism (Figure 8). Thus, a straightforward hypothesis for the formation of cratonic mantle is to call upon either a hotter and more vigorously convecting Archean mantle or the presence of larger and hotter thermal upwellings, e.g., plumes, in the Archean (Figure 1a). Both scenarios would lead to a thicker residual mantle column in addition to higher average degrees of melting (Figure 8). In such scenarios, the entire craton can be formed in one plume-melting event; that is, cratonic peridotites were formed in situ. The attractive aspect of the plume hypothesis is that a thick, strong, and chemically buoyant mantle keel can be generated by one seemingly simple and elegant process. The posterity of cratonic mantle would be ensured from its inception.

The geochemical/petrologic predictions of this hypothesis are as follows. If cratonic mantle represents the undisturbed residual mantle of a large plume head, that is, the last equilibration pressures of cratonic mantle xenoliths reflect the average pressure at which melt was last extracted, then a mantle with a very high potential temperature is required (Figure 8). For example, to achieve 30–50% melting at pressures of 3–7 GPa (the range in pressures from which cratonic mantle derives), a potential temperature significantly greater than 1650°C is required. This is hotter than the potential temperature of the hottest modern plumes (1400–1600°C) and at least 200°C higher than modern mantle beneath ridges (1250–1450°C) [Herzberg, 2004; Putirka, 2005]. Moreover, at these potential temperatures, partial melting would have actually started at pressures greater than 7 GPa, possibly up to 10 GPa (Figure 8). This is because the dT/dP of the lherzolite solidus decreases with increasing pressure such that at high enough pressures (>7 GPa), the melting adiabat roughly parallels that of the solidus (see Figure 8) and hence melting will occur over a very large depth interval. When we add on the fact that garnet is stable on the solidus over a large temperature range at high pressures [Herzberg and Zhang, 1996; Walter, 1998, 1999; Herzberg and O'Hara, 2002], the hot plume hypothesis predicts that cratonic peridotites should bear a distinct garnet trace-element signature if they indeed derive from a very hot plume head.

A second prediction of the plume hypothesis is that the residual mantle column should be stratified in terms of composition as a consequence of decompression melting. The shallowest portions of the residual column will be most depleted (e.g., highest Mg#) as this is where the temperature difference between the melting adiabat and the solidus is greatest [McKenzie, 1984; Langmuir *et al.*, 1992]. If the high Mg#s (91–94) of cratonic mantle peridotites imparted by partial melting at their current equilibrium pressures (3–7 GPa), one might also expect to find peridotites (e.g., deriving from <3 GPa) with even higher degrees of melt extracted and accordingly higher Mg# (>94); these are not seen, but their absence could be due to biased sampling. The high degrees of melt extraction require the existence of large amounts of a complementary ultramafic magma, such as a komatiite [Takahashi, 1986]. Indeed, komatiites primarily occur in the Archean [Takahashi *et al.*, 1993; Arndt, 2003] but are not as abundant as one might expect (or require) from the plume scenario. However, at high enough pressures (~7–10 GPa), it is possible that komatiitic magmas may be neutrally buoyant with respect to olivine [Stolper *et al.*, 1981; Agee and Walker, 1993] as a consequence, may conveniently sink and dispose of themselves, obviating the need to explain the scarcity of komatiites in the geologic record.

3.2. Formation of Cratonic Mantle by Accretion/Stacking of Oceanic Lithosphere

A second way [Helmstaedt and Schulze, 1989] of generating cratonic mantle is by the accretion and stacking of oceanic lithosphere segments (Figure 1b). The motivation for this hypothesis was largely due to the presence of eclogite (garnet + omphacite lithology) xenoliths in kimberlites, which has been interpreted to have hydrothermally altered oceanic crust protoliths (see section 2.5). This hypothesis has gained additional momentum in light of seismic studies done in the Slave craton, which showed the presence of a dipping seismic reflector on the edge of the craton [Bostock, 1998]. More recently, hints of dipping S–P conversions have been seen in the Kaapvaal craton [Levander *et al.*, 2005].

Unlike the plume-head model, the generation of a cratonic mantle keel by accretion of oceanic lithosphere is a sequential process. Thus, one might expect to see variation with depth in the timing of melt extraction. Attempts have been made using Re–Os isotopic systematics to determine if cratonic mantle is age-stratified [Pearson *et al.*, 1995]. In the Kaapvaal craton, there is no systematic variation of Re–Os model ages with depth, at least to within 0–10 km. Originally, this was argued to provide evidence supporting the plume hypothesis [Pearson *et al.*, 1995c], but it does not rule out the accretion hypothesis if sufficient accretion

stacking of oceanic lithosphere can occur within 0.5 Gy. This time interval is within the noise range of Re-Os model ages. It may also be a reasonable amount of time to form a continent by accretion [Şengör and Natal'in, 1996].

Another distinction between the accretion and plume hypothesis is that cratonic mantle formed by accretion would be predicted to be composed of peridotites that had partially melted at fairly low pressures (<~4–5 GPa, depending on the potential temperature of the mantle) and subsequently transported to the greater pressures (3–7 GPa) from which they presently derive (Figure 8). If so, many cratonic mantle peridotites should have formed by melting in the spinel stability field. The garnets that are now present in cratonic peridotites would then have to represent subsolidus exsolution products from higher temperature (and lower pressure) orthopyroxenes, which are high in aluminum [Cox *et al.*, 1987; Saltzer *et al.*, 2001; Simon *et al.*, 2003]. Thus, in contrast to the plume hypothesis, we would not expect to see trace-element signatures of garnet even though garnet is now present in these xenoliths. The transposition of shallowly melted peridotites to greater depths also relaxes the requirement for extremely hot temperatures to generate high degrees of melting. This is because high degrees of melting can be generated by adiabatically decompressing a parcel of mantle to shallow pressures without calling upon extremely high temperatures (Figure 8). As a consequence, the large amounts of complementary komatiitic magmas required by the plume hypothesis are not required by the oceanic lithosphere accretion hypothesis.

Finally, if cratons are assembled by piecing together wholesale fragments of oceanic lithosphere, a range in fertility of the cratonic mantle might be expected. This is because the fertility of oceanic lithosphere is probably stratified. Although an entire section of oceanic lithosphere has never been sampled (typically, ophiolites sample only the uppermost part of the lithospheric mantle), the stratified nature of oceanic lithosphere is what would be predicted if oceanic lithosphere represents a residual melt column formed by passive upwelling and decompression melting [Langmuir *et al.*, 1992]. In such a scenario, Mg#s would be predicted to vary systematically from ~88 at the base of the melting column to ~91 at the top [Langmuir *et al.*, 1992; Lee *et al.*, 2005b]. Indeed, abyssal peridotites and ophiolitic peridotites, all of which undoubtedly sample the uppermost part of the residual melt column in oceanic lithosphere, are characterized by high degrees of melting and high Mg#s [Dick and Bullen, 1984; Johnson *et al.*, 1990]. Cratonic mantle peridotites, on the other hand, appear to be dominated by highly melt-depleted lithologies (Mg#s >91; Figure 5); only at the base of the cratonic mantle are fertile peridotites seen. Thus, if the oceanic lithosphere accretion hypothesis is correct, the deeper and more fertile portions of the oceanic lithosphere must be

removed during or before the accretion process, so that the resulting cratonic mantle can be biased towards lower fertility as required by observation.

The main problem with the oceanic accretion hypothesis is that presently, oceanic lithosphere becomes negatively buoyant as it cools and hence becomes subductable as early as within 20 Ma of its inception (Figure 9). Thus, it seems unlikely that cratonic mantle could be formed by accretion of oceanic lithosphere segments having a compositional structure similar to today's oceanic lithosphere. On the other hand, if mantle potential temperature was just slightly higher in the Archean (>200°C), a thicker and hence more buoyant oceanic crust might have formed. A hotter mantle might also imply a more vigorously convecting Earth [Davies, 1992]. If so, mantle overturn times might have been shorter and the average age of oceanic lithosphere upon reaching trenches would be younger [Davies, 1992]. Collectively, such buoyancy may permit accretion and stacking of oceanic lithosphere, but there may be only a narrow range in parameter space for this to be a viable process [Cooper *et al.*, 2005].

3.3. Accretion/Thickening of Sub-Arc Mantle

Another scenario proposed for creating a thickened and highly melt-depleted mantle root is by thickening of sub-arc lithospheric mantle, perhaps through accretion or thickening of island arcs (Figure 1c). This was Jordan's [1988] preferred hypothesis for the formation of tectosphere (see also Herzberg [1999]). Dynamically, accretion of sub-arc lithosphere may differ fundamentally from accretion of oceanic lithospheric mantle. For example, arc lithosphere may already be buoyant, unlike oceanic lithosphere, which typically subducts. On the other hand, the petrologic predictions may appear superficially similar to the oceanic lithosphere accretion hypothesis because both accretion scenarios involve constituents formed by melting at shallower pressures and lower temperatures than implied by the hot plume scenario (Figure 8).

There are, however, some subtle differences between the two accretion scenarios. First, if arc lithospheric mantle was originally extracted from the asthenospheric mantle wedge in a subduction zone, one might expect partial melting of the mantle to be greatly influenced by fluids derived from dehydration and/or melting of the subducting slab. Interaction with extensive fluids is unlikely to be involved during the formation of oceanic lithospheric mantle. Of course, subsequent to accretion, it is possible that accreted segments of oceanic lithosphere could interact with later subducting fluids, but the fundamental difference is that such fluids would be interacting with already formed and presumably cold lithosphere whereas in the arc environment, the fluids would be interacting with hot asthenospheric mantle wedge and hence have

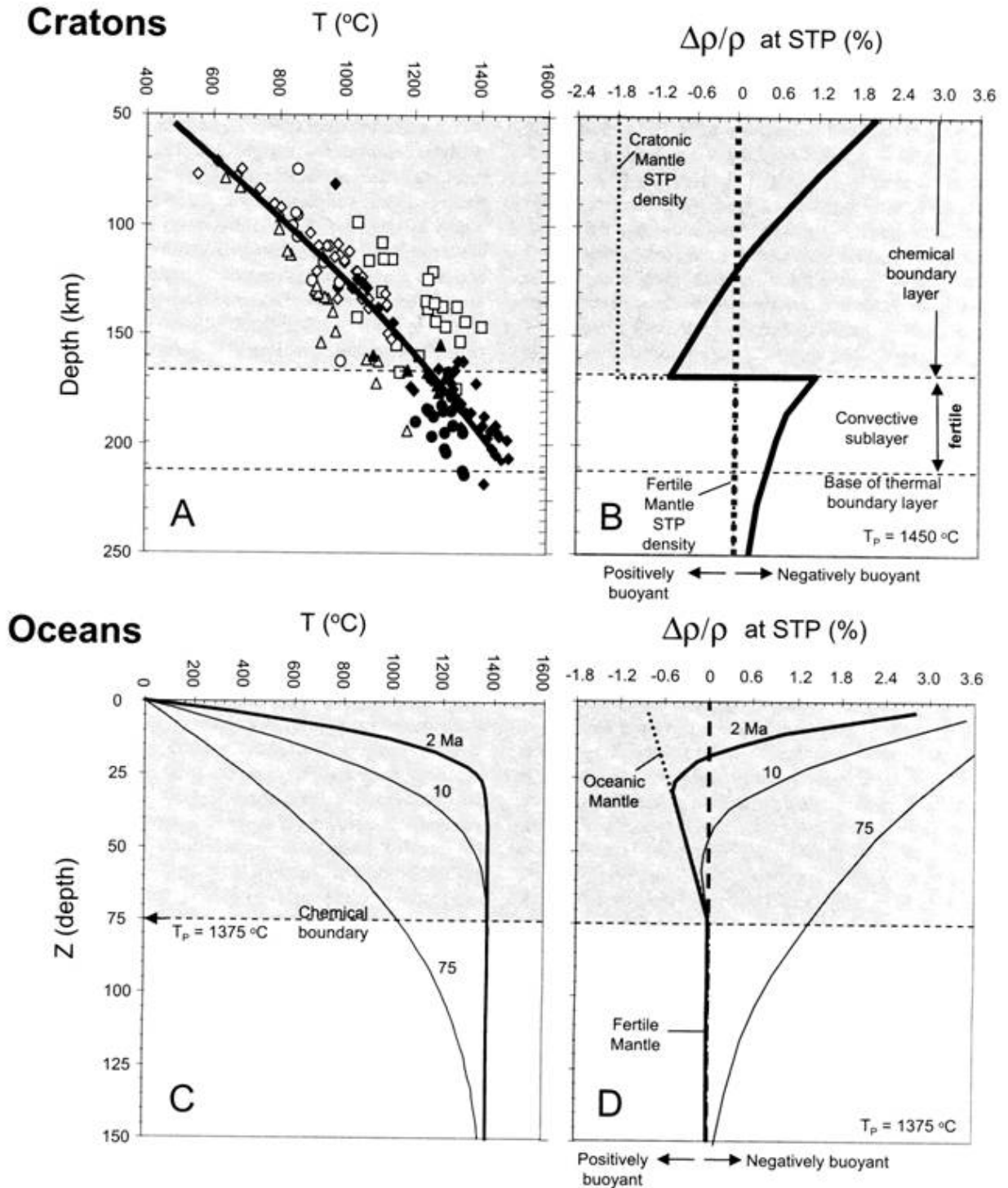


Figure 9. (A) Cratonic geotherm defined by xenolith thermobarometry as shown in Figure 3. (B) Estimated density profile of cratonic thermal boundary layer at roughly steady-state thickness, taking into account the thermal state in (A) and the compositional buoyancy of the chemical boundary layer and the convective sublayer (modified from *Lee et al. [2005b]*). Densities are given in terms of relative deviations from fertile peridotite (all calculations assuming STP conditions). (C) Transient geotherms of cooling oceanic lithosphere based on conductive cooling of an infinite half-space. Age of transient geotherms denoted. (D) Density profile of oceanic lithosphere, taking into account transient thermal states shown in (C) and a predicted compositional buoyancy for oceanic lithosphere. Slanted dashed line represents the standard temperature and pressure density profile of oceanic lithosphere associated with a gradual increase in Mg# with shallowing depth. Note oceanic lithosphere becomes negatively buoyant by 10 Ma.

a direct influence on the melting process. Thus, one useful prediction is that peridotites, originally derived from the mantle wedge, might show reaction with fluids/melts rising from the subducting slab. For example, interaction with silicic slab melts has the propensity to increase the SiO₂ content of highly melt-depleted harzburgites without changing the harzburgite's Mg# significantly [Kelemen *et al.*, 1998]. The anomalously Si-rich harzburgites seen in cratons may in fact have such an origin, although such harzburgites could also be the products of mixing residual peridotites with cumulate orthopyroxenes associated with melting in a large plume head [Herzberg, 1993, 1999]. The test of whether these Si-rich harzburgites represent melt–rock reaction products might lie in their oxygen isotopic compositions, but no such study has yet been systematically done.

Other possible petrologic differences between the two accretion hypotheses are the predicted abundances and compositions of garnet pyroxenite xenoliths in cratonic mantle. The oceanic lithosphere accretion scenario predicts a greater volume proportion of garnet pyroxenite lithologies (specifically low MgO eclogites) in the cratonic mantle, although admittedly, this would be a very difficult prediction to test since it is never clear whether xenolith sampling is representative of the true distribution in the mantle. However, there may be subtle differences in the composition of garnet pyroxenite lithologies found in cratons. For example, eclogite protoliths in the oceanic lithosphere accretion hypothesis would have major element compositions indicative of formation at shallower pressures, e.g., basaltic oceanic crust and cumulate gabbros. In contrast, the garnet pyroxenites formed in arc environments might have formed as higher pressure cumulates, such as at the base of the arc crust itself, which in general is thicker than oceanic crust. In the next section, we discuss whether some cratonic mantle garnet pyroxenite protoliths are arc cumulates.

4. THE PETROLOGIC/GEOCHEMICAL TESTS

4.1. Constraining the Pressures at Which Cratonic Peridotites Formed

As discussed in section 2.1.2, the pressures at which cratonic peridotites last equilibrated are in the range of 3–7 GPa (~90–200 km depth), but the question is whether they melted at these depths or were transposed to these depths since the time of melting. We can estimate melting pressures by examining their bulk compositions and assuming that they originate from a primitive mantle-like or pyrolitic source. Unlike mineral–mineral equilibria, which record conditions of last thermodynamic equilibrium, bulk composition can potentially preserve a record of past partial melting conditions (open system) insofar as the bulk system has remained closed to secondary disturbances since the melting event. In this sense,

major elements may be a natural choice as their concentrations in a peridotite should not change appreciably unless there has been significant mass loss or addition to the system.

Major element systematics, however, give ambiguous results. On the one hand, the low-Fe contents of some cratonic peridotites (Figures 6c, d) suggest, at face value, melting by 30–40% at 7 GPa. This can be seen in Figures 6c and d, where Phanerozoic spinel peridotites and Archean cratonic peridotites are superimposed on isobaric equilibrium melting experiments of a fertile peridotite [Walter, 1998, 1999; Herzberg and O'Hara, 2002; Herzberg, 2004]. We have taken Phanerozoic spinel peridotites as examples of modern-day peridotites formed in oceanic, arc, and continental environments. We do not use abyssal peridotites because they are extensively serpentinized; nevertheless, included in our spinel peridotite list are Phanerozoic ophiolite peridotites, which presumably formed in mid-ocean ridge or back-arc environments (see Lee *et al.* [2003] for references). For simplicity, we have adopted equilibrium isobaric melting rather than polybaric fractional melting. Although more realistic, polybaric fractional melting is highly model-dependent. In any case, the overall conclusions are similar regardless of whether the simplistic or complex approach is taken (for a more comprehensive treatment of polybaric fractional melting, the reader should refer to Herzberg [2004]). This conclusion is that high average pressures of melting (or initial pressures of melting in the polybaric case) results in residues having low-FeO contents due to the greater retention of orthopyroxene and garnet on the solidus at high pressures. It can be seen that Phanerozoic spinel peridotites have FeO contents consistent with melt extraction between 1 and 4 GPa (Figure 6c), whereas Archean cratonic peridotites have FeO contents indicating melting between 1 and 7 GPa, with some suggesting melting at even greater pressures (some of the cratonic peridotites with FeO contents greater than 8 wt.% are represented by the high-temperature sheared peridotites, which show evidence for Fe refertilization; hence the low apparent pressures in some may even be an artifact). In contrast to FeO, a very different picture arises from Al₂O₃–MgO systematics. The extensive involvement of garnet and pyroxene during melting at high pressures results in more aluminous residues for a given MgO content. Interestingly, the Al₂O₃–MgO systematics of cratonic peridotites are largely indistinguishable from those of Phanerozoic spinel peridotites and are easily explained by melting between 1 and 4 GPa (Figures 6a, b). The low Al₂O₃ contents of Archean cratonic peridotites has been used to argue that they actually melted at pressures lower than their current equilibration depths [Bernstein *et al.*, 1998].

We can expand on this debate by assessing the systematics of trace elements that may be sensitive to the presence of garnet. The elements most sensitive to garnet control are V, Y,

Sc, and the heavy rare-earth elements, these elements all being compatible in garnet. A convenient feature of these elements is that their *bulk* partition coefficients in peridotites are such that during partial melting they are only moderately incompatible relative to the bulk rock. This means that they do not become significantly depleted during partial melting and they are only slightly enriched in melts. As a consequence, these elements often remain relatively undisturbed by metasomatic processes, contrasting with the relative ease by which highly incompatible elements are overprinted by metasomatic processes [Canil, 2004]. Such behavior is confirmed by the fact that V, Sc, Y, and the heavy rare-earth elements in peridotites almost always preserve original melting systematics, while highly incompatible elements do not [Lee *et al.*, 2003; Canil, 2004]. Canil [2004] used Sc systematics of cratonic peridotites to show that they could not have formed by high-pressure melting. This is seen in Figure 10, where Sc is plotted against MgO. Partial melting trends of Sc versus MgO are negatively correlated because Sc is incompatible during melting whereas Mg is compatible. Each figure shows the effects of isobaric fractional melting at 1.5, 3, and 7 GPa [Lee *et al.*, 2005a; Li and Lee, 2004]. The extensive involvement of garnet during melting at 7 GPa results in less rapid depletion in Sc than at 1.5 and 3 GPa, where garnet is not present or present only during the very early stages of melting, respectively (Figure 10). Cratonic mantle peridotites (Tanzania, South Africa, and Siberia) have Sc systematics nearly indistinguishable from Phanerozoic spinel peridotites.

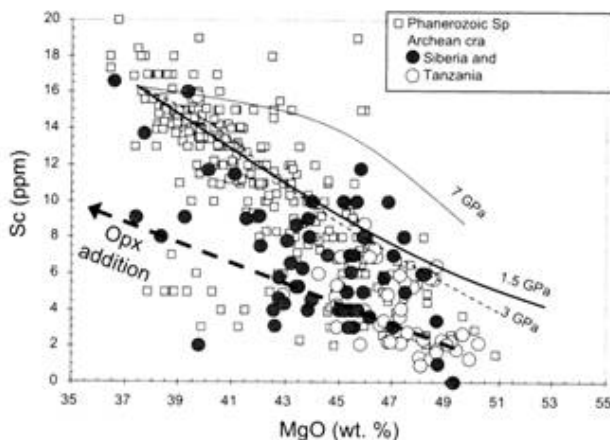


Figure 10. Plot of whole-rock Sc (ppm) versus MgO (wt.%) for Phanerozoic spinel peridotites and Archean cratonic peridotites (Siberia, Tanzania, South Africa). Melting curves for 1.5, 3, and 7 GPa are shown; Sc is elevated during 7-GPa melting due to the greater involvement of garnet on the solidus at high pressures and the high compatibility of Sc in garnet. Dashed arrowed line corresponds to the trend expected for the addition of orthopyroxene through melt-rock reaction or cumulate orthopyroxene addition [Lee *et al.*, 2003].

There thus appears to be no hint from Sc that cratonic peridotites originally melted at 7 GPa. Similar conclusions can be made from V–MgO systematics after the redox-sensitive behavior of V is accounted for [Canil, 2002, 2004; Lee *et al.*, 2003].

It thus seems likely that the majority of cratonic peridotites melted on average anywhere between 1 and 4 GPa, that is, mainly in the spinel stability field (although initial pressure of melting may have been deeper). If so, these peridotites were subsequently transported to greater depths (~7 GPa), after which subsolidus re-equilibration resulted in exsolution of garnet (as well as small amounts of clinopyroxene). An exsolution origin for the small amounts of garnet and clinopyroxene in cratonic harzburgites has been previously suggested on the basis of textural studies [Cox *et al.*, 1987]. These conclusions, if correct, require an explanation for the low FeO contents of some cratonic peridotites (Figure 11), which at face value imply very high pressures of melting [Pearson *et al.*, 1995b]. Herzberg [2004] suggested that the low FeO contents are related to Si-enrichment processes associated with subsequent metasomatism and/or cumulate addition of orthopyroxene (Figure 4b). This can be seen by first identifying those cratonic peridotites that appear to be

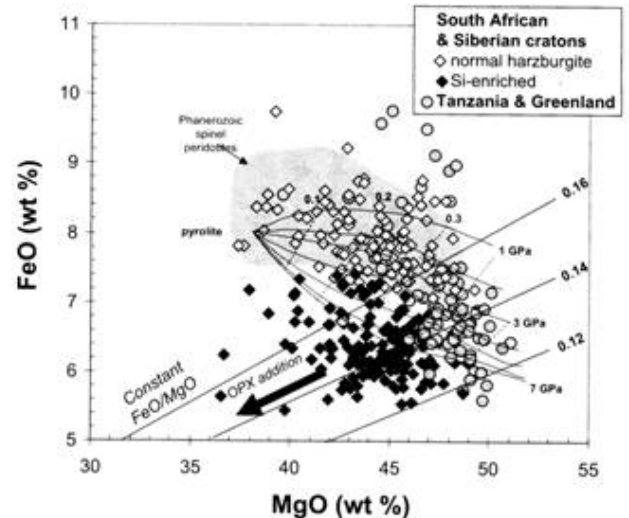


Figure 11. Plot of total FeO versus MgO for cratonic peridotites along with shaded field for Phanerozoic spinel peridotites. Curved lines represent 1–7 GPa equilibrium melting curves as in Figures 6 and 10. Dark-colored diamonds represent those samples identified as Si-enriched, based on their Mg/Si ratios. Arrow shows the direction of Si- and orthopyroxene-enrichment. Diagonal straight lines represent mixing with orthopyroxene having identical FeO/MgO with original peridotite (numbers correspond to FeO/MgO ratios). Note that addition of orthopyroxene can give results in low bulk FeO contents, giving an artifact of a high-pressure melting origin.

anomalously rich in Si (Figure 4b). For example, we can unambiguously identify Si-enriched peridotites as those having a Mg/Si ratio (an inverse proxy for Si-enrichment) less than that of Phanerozoic peridotites at a given Mg#. Using this criterion, we can see that Si-enriched peridotites systematically have lower FeO contents than "normal" cratonic peridotites (Figure 11). It also appears that the lower FeO contents of Si-enriched peridotites are also accompanied by a decrease in MgO content. Effectively, the low FeO and MgO contents of Si-enriched peridotites are consistent with the addition of a Si-rich component (cumulate or melt-rock reaction) having a FeO/MgO (or Mg#) ratio similar to that of the original peridotite. Mantle orthopyroxenes fit the composition of this endmember, consistent with the fact that the Si-enriched peridotites are anomalously rich in orthopyroxene [Herzberg, 2004].

If the Si-enriched peridotites are excluded, the consensus seems to be growing that few cratonic peridotites are residues of melting at average pressures as great as 7 GPa [Bernstein *et al.*, 1998; Canil, 2004; Herzberg, 2004]. The major and moderately incompatible trace element systematics of such cratonic peridotites are very similar to that for Phanerozoic spinel peridotites and can be modeled as 30–50% melt residues formed at pressures \sim 4 GPa. Thus, most cratonic peridotites probably melted at low pressures and have since been transported to the great depths (90–220 km) from which they currently derive. If this is the case, the potential temperature of the mantle in which these cratonic peridotites melted was likely to be between 1400°C and 1650°C (Figure 8). This straddles the upper end of accepted potential temperatures for modern mid-ocean ridge environments (1300–1450°C) [Langmuir *et al.*, 1992; Asimow and Langmuir, 2003; Lee *et al.*, 2005b; Putirka, 2005] and falls within the range of potential temperatures estimated for modern plume settings (1450–1650°C) [Herzberg, 2004; Putirka, 2005]. These conclusions, combined with the fact that there is no gradual decrease in fertility with increasing depth, suggest that cratonic mantle does not represent the residual mantle column of a single, large, and anomalously hot (potential temperature $>$ 1650°C) plume head. Instead, cratonic peridotites could have formed at Archean mid-ocean ridges having a slightly higher potential temperature (by 100–200°C) than in the present. If the Archean mantle potential temperature was identical to the present, then cratonic peridotites must have formed in small thermal plumes, not unlike what is seen today. In either case, cratonic peridotites did not form in situ; thus they have since been transported to greater depths.

4.2. Constraining the Protoliths of Cratonic Eclogites

We now turn to assessing the two accretion hypotheses. To do so, we will focus on the composition of garnet pyroxenite

protoliths with particular emphasis on the high-MgO lithologies (see section 2 for discussion of garnet pyroxenite classification). Garnet pyroxenite protoliths in the oceanic lithosphere accretion hypothesis are likely to be dominated by basaltic oceanic crust and low-pressure gabbroic cumulates, the latter consisting of olivine–pyroxene lithologies (primitive cumulates) and/or pyroxene–plagioclase lithologies (evolved cumulates). The primitive olivine–pyroxene lithologies would give rise to MgO-rich, SiO₂-poor, and Al₂O₃-poor garnet pyroxenites upon subsolidus re-equilibration within the cratonic mantle.

Unlike ridge environments, arc magmas must traverse pre-existing lithosphere such that magmatic differentiation probably commences at greater pressures (e.g. $>$ 1 GPa) than beneath mid-ocean ridges ($<$ 1 GPa). At higher pressures, the olivine phase field diminishes significantly, so instead of precipitating olivine-dominated cumulates, as would occur in mid-ocean ridge settings, the most primitive cumulates in arc settings would be clinopyroxene-rich (this is corroborated by applying the pMELTS thermodynamic algorithm to the crystallization of a primary basaltic magma at 1 GPa). Thus, in the arc case, high-MgO garnet pyroxenites would be represented by clinopyroxene-rich lithologies characterized by high MgO, high SiO₂, and low Al₂O₃ contents.

In Figure 12, we have plotted the bulk compositions of garnet pyroxenite xenoliths derived from the lower crust and lithospheric mantle underlying the Sierra Nevada batholith in California (western North America), an extinct Mesozoic continental arc [cf. Ducea and Saleeby, 1996, 1998; Lee, 2005; Lee *et al.*, 2006]. These pyroxenites can be subdivided into a high-MgO ($>$ 8 wt.%) and a low-MgO group [Lee *et al.*, 2006], the former being more primitive and the latter more evolved. Both groups are represented by garnet clinopyroxenites, but the high-MgO group has more than 50% modal clinopyroxene and the latter has more than 50% garnet. This is consistent with the fact that the high-MgO group has SiO₂ contents of \sim 50–52 wt.% (Figure 12A) and Al₂O₃ contents of 6–10 wt.%, while the low-MgO group has lower SiO₂ (Figure 12) and higher Al₂O₃ (Figure 12B). Also shown in Figure 12 are plutonic rocks from the Sierras, volcanic rocks from the Cascades volcanic arc (GEOROC database), and gabbroic cumulates (plagioclase and hornblende-rich) associated with the Sierra Nevada batholith [Sisson *et al.*, 1996]. The high SiO₂ contents of the high-MgO group eliminate an olivine-bearing protolith and are instead more consistent with a clinopyroxene-rich protolith. This clinopyroxene-rich protolith could also have had original cumulate garnet, but moderate enrichment in heavy rare earths (not shown) suggests that most of the garnet is subsolidus. The low-MgO group, on the other hand, is similar to the more evolved Sierran cumulates in terms of major elements. Thus, one possible interpretation is that the high-MgO Sierran garnet pyroxenites

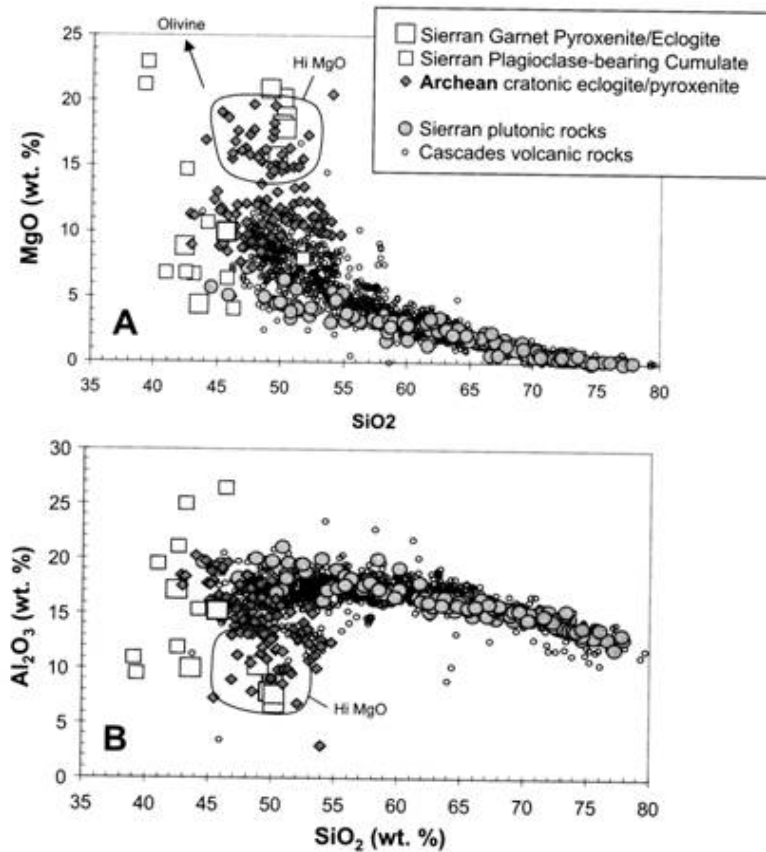


Figure 12. Whole-rock MgO (A) and Al₂O₃ (B) versus SiO₂, showing cratonic eclogites (Siberia, South Africa, West Africa) compared to Phanerozoic eclogites/garnet pyroxenites from an arc environment (Sierra Nevada batholith). References are given in text. Also shown are Cascades magmatic arc series and Sierra Nevada plutonic rocks. It can be seen that some of the high-MgO cratonic eclogites/pyroxenites have compositions similar to those of the high-MgO Sierran garnet pyroxenites.

represent primitive high-pressure cumulates, whereas the low-MgO pyroxenites represent more evolved cumulates. Indeed, initial crystallization of the high-MgO pyroxenites is consistent with the fact that the primitive Sierran plutons have anomalously low MgO and high Al₂O₃ contents for a given SiO₂. Subsequent crystallization of the more evolved low-MgO pyroxenites could then drive the initial increase in SiO₂ seen in the Sierran magmatic differentiation array.

In Figure 12, Archean cratonic garnet pyroxenite (mostly bimineralic eclogites) xenoliths from Siberia [Jacob *et al.*, 1994; Beard *et al.*, 1996; Jacob and Foley, 1999; Taylor *et al.*, 2003], South Africa [Pyle and Haggerty, 1998], and Sierra Leone in west Africa [Fung and Haggerty, 1995] have been superimposed. As discussed previously, the rough subdivision of the cratonic garnet pyroxenites into high- and low-MgO groups is apparent. The low-MgO group appears to have compositions similar to modern basaltic oceanic crust,

although Al₂O₃ and SiO₂ contents are slightly lower. The low-MgO group has been argued to have an oceanic crust protolith coupled with subsequent partial melting in the eclogite or amphibolite stability field [Barth *et al.*, 2001]. The high-MgO group has been interpreted to be either low-pressure olivine-plagioclase-pyroxene cumulates or high-pressure (2–3 GPa) garnet pyroxenite cumulates [Barth *et al.*, 2002]. However, as shown in Figure 12, the high-MgO Archean pyroxenites are remarkably similar to the high-MgO Sierran pyroxenites, interpreted here to be high-pressure arc cumulates.

If these compositional similarities are not a matter of coincidence, the easiest interpretation is that at least some of the high-MgO Archean garnet pyroxenites represent primitive cumulates associated with Sierran-like continental arc magmatism. If so, the high-MgO Archean garnet pyroxenites are evidence that at least some Archean cratons may have formed

by accretion/thickening of arcs. The low-MgO Archean eclogites are generally interpreted to represent former subducted oceanic crust. Collectively, this suggests that craton formation may be a combination of arc and oceanic lithosphere accretion. Alternatively, the low-MgO Archean eclogites may represent arc basalts rather than oceanic crust basalts.

5. AFTERTHOUGHTS

In light of the foregoing discussions, it is worthwhile returning to the questions of why cratons are so long-lived and why are they presently neutrally buoyant? Any successful model for the origin of cratons must satisfy these conditions. In this context, the plume-head scenario seems highly attractive because it provides a thick chemically depleted and possibly dehydrated root that is strong from the outset, ensuring its longevity and approximate neutral buoyancy. This model, however, predicts that cratonic peridotites have a high pressure origin, which appears not to be the case in general.

The oceanic lithosphere accretion hypothesis is attractive in that it can explain the origin of low-MgO eclogite xenoliths in cratons as well as the presence of dipping reflectors in the margins of cratonic mantle. As pointed out above, however, for neutral buoyancy to be maintained, this hypothesis requires a very delicate balance between the amounts of eclogitized oceanic crust and depleted oceanic mantle: The more depleted (and hence chemically buoyant) the oceanic mantle, the more complementary eclogitized oceanic crust that must exist. Too much eclogite or garnet pyroxenite, regardless of how depleted the oceanic mantle is, will prevent the craton from ever being neutrally buoyant.

The arc accretion/thickening hypothesis is also attractive. In this hypothesis, highly depleted mantle would be expected throughout the cratonic mantle, unlike accretion/stacking of oceanic lithosphere, which would yield a periodic variation in fertility with depth. In addition, the amounts of eclogite/pyroxenite in the cratonic mantle would probably be less, especially if most arc-generated pyroxenites are delaminated or removed during or shortly after arc formation. By the time arc accretion occurs, only small amounts of garnet pyroxenite remain. Thus, the range of conditions, which permit neutral buoyancy in this hypothesis, seems (at face value) more easily achieved and maintained, but the dynamic feasibility of this process needs to be tested.

Clearly, craton formation is a complicated process and the true answer could very well be a mixture of formation scenarios. Identifying these formation processes undoubtedly will help us understand to what extent the dynamics of the Earth's mantle have changed through time. It has been shown that, on average, there may have been a rough secular change in the composition and structure of continental lithospheric mantle

[Griffin *et al.*, 1999]; that is, the thickest and most melt-depleted lithospheric mantles are associated with Archean cratons, while the thinner and more fertile ones are Phanerozoic (Proterozoic lithospheres are intermediate). In light of the endmember formation scenarios, these secular changes could imply any of the following: the occurrence of larger and hotter plumes in the Archean, a hotter Archean mantle, or faster spreading rates in the Archean. Faster spreading rates might increase the probability of generating thickened arc lithosphere or stacked oceanic lithosphere sections.

In some cases, however, there appears to be no correlation between thickness and composition versus age. For example, late Archean to Paleo-Proterozoic lithospheric mantle underlies the Mojave block in southwestern United States [Lee *et al.*, 2001b]. Although this mantle is fertile, the adjacent middle Proterozoic Colorado Plateau is underlain by a thick, highly melt-depleted cratonic root, not unlike "archetypal" Archean cratons. Another example is in Australia, where it has been shown seismically that some of the thickest cratons underlie Proterozoic regions rather than Archean regions [Simons *et al.*, 1999]. These "exceptions to the rule" show that although there may very well be a secular change in the way or rate at which cratons are formed, one must conclude that at least some of the cratons seen today are simply a result of biased preservation. If so, this begs the question of how much crust and lithospheric mantle was generated in the past but just didn't have the necessary compositional, viscosity, and thermal structure to survive.

While we have focused primarily on the origin of cratons, it could be argued that an equally important perspective might come from studying the eventual fate of cratons and of continental lithosphere in general. Once a stable continental lithosphere is formed, does it last forever [O'Reilly *et al.*, 2001]? If, as suggested here, cratonic mantle might have started out dehydrated due to the high degrees of melt extraction, cratonic mantle could be inherently strong and hence stable with respect to the convecting mantle. However, this outcome seems at face value inconsistent with the fact that many cratonic mantle xenoliths bear evidence for hydrous metasomatism (see Pearson *et al.* [2003] for a review). It has even been suggested that continental lithospheric mantle is weak, perhaps as a consequence of hydrous metasomatism [Maggi *et al.*, 2000a, 2000b; Jackson, 2002]. If so, this challenges the view of a strong lithospheric mantle. The paradox, of course, is that rehydration might lead to weakening of the continental lithosphere, destroying all hope of a long life. Indeed, there are some lines of observational evidence that suggest continental lithospheric mantle, under some circumstances, may in fact be destabilized and recycled back into the convecting mantle [Kay and Kay, 1993; Ducea and Saleeby, 1996; Lee *et al.*, 2000, 2001a; Gao *et al.*, 2002]. Understanding the mechanisms of lithospheric destabilization will thus remain

an area of open debate. A number of geodynamic studies have shown that lithospheric mantle can be destabilized during compressional orogenies [Houseman *et al.*, 1981; Conrad and Molnar, 1997; Houseman and Molnar, 1997; Molnar *et al.*, 1998; Jull, 2001], but none of these studies have incorporated the effects of dehydration (or rehydration); hence, the obvious question is what role does water play in the stabilization and potential destabilization of continental lithospheric mantle? The issue of water, that is, how it got there (or left), how it influences rheology, and how much is present, may ultimately be one of the most important foci for future research in the dynamics of continents.

6. CONCLUSIONS

The salient features of cratonic mantle were reviewed in this paper and discussed in the context of geochemical and petrologic predictions associated with three endmember formation scenarios: (1) formation by large degrees of melting in a single, large plume head, (2) formation by accretion/stacking of oceanic lithosphere, and (3) formation by accretion/thickening of arc lithosphere. The following conclusions were made:

1. Cratons are underlain by a thick thermal boundary layer composed of a highly melt-depleted and compositionally buoyant chemical boundary layer. This chemical boundary layer must be strong in order to ensure long-term preservation. The chemical boundary layer is underlain by a thin convectively active sublayer that makes up the base of the thermal boundary layer.
2. Cratonic peridotites appear to have melted on average at lower pressures (<4 GPa) than those from which they presently derive; hence, they have been transposed to greater depths after their formation.
3. The protoliths of Archean low-MgO eclogites are related to subducted basaltic crust of either oceanic or arc origin.
4. High-MgO Archean eclogites are similar to high-pressure cumulates found in Phanerozoic arc environments.

The low pressures from which cratonic peridotites originally melted suggest that cratonic mantle does not represent the residual melt column of a single, anomalously hot plume head. The low pressures are instead more consistent with formation by arc or oceanic lithosphere accretion. These latter scenarios are unfortunately difficult to distinguish petrologically. However, an arc component seems to be required in at least some cratons.

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- C.-T. A. Lee, Department of Earth Science, MS-126, Rice University, 6100 Main Street, Houston, Texas 77005, USA. (ctlee@rice.edu)