PHYSICAL, CHEMICAL, AND CHRONOLOGICAL CHARACTERISTICS OF CONTINENTAL MANTLE

Richard W. Carlson
Department of Terrestrial Magnetism
Carnegie Institution of Washington
Washington, D. C., USA

D. Graham Pearson
Department of Geological Sciences
University of Durham
Durham, UK

David E. James
Department of Terrestrial Magnetism
Carnegie Institution of Washington
Washington, D. C., USA

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[1] Unlike in the ocean basins where the shallow mantle eventually contributes to the destruction of the overlying crust, the shallow mantle beneath continents serves as a stiff, buoyant ''root'' whose presence may be essential to the long-term survival of continental crust at Earth's surface. These distinct roles for subcrustal mantle come about because the subcontinental mantle consists of a thick section of material left behind after extensive partial melt extraction, possibly from the wedge of mantle overlying a subducting oceanic plate. Melt removal causes the continental mantle to be cold and strong but also buoyant compared to oceanic mantle. These characteristics allow thick sections of cold mantle to persist beneath continental crust in some cases for over 3 billion years. If the continental mantle becomes gravitationally unstable, however, its detachment from the overlying crust can cause major episodes of intracontinental deformation and volcanism.

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1. INTRODUCTION

[2] Like an iceberg, the main volume of a continent is hidden at depth in ''mantle roots'' that can extend hundreds of kilometers into the upper mantle (Figure 1). Surprisingly, we may know more about the mantle beneath continents than we do of the suboceanic mantle as a result of the occurrence on the continents of deeply derived explosive volcanic rocks, such as kimberlites, that carry fragments of the mantle to the surface. Study of these accidental mantle fragments, called mantle xenoliths [*Nixon*, 1987; *Pearson et al.*, 2003], reveals much about the physical, thermal, compositional, and chronological history of the continental mantle, a subject that has been the topic of several recent volumes [*Fei et al.*, 1999; *van der Hilst and McDonough*, 1999; *Fowler et al.*, 2002; *Jones et al.*, 2004].

[3] Arguably, the most interesting aspect of the continental mantle is that it seems to assist in the long-term survival of continental crust, whereas the shallow oceanic mantle ultimately contributes to the destruction of overlying oceanic crust. On moving away from an ocean ridge, oceanic crust and its relatively thin underlying layer of melt-depleted mantle cool the shallow mantle by conduction of heat into the oceans. This cooling causes the underlying mantle to become mechanically coupled to the overlying plate to form what is known as lithosphere: the rigid conductive thermal boundary layer at the top of the convecting mantle that consists of both crust and that portion of the mantle that move together as a tectonic plate. As the plate ages, oceanic lithosphere cools and thickens, becoming progressively more dense so that at some point it becomes gravitationally unstable and descends back into the mantle in the process known as plate subduction. In marked contrast, continental lithospheric mantle appears to serve as a ''life preserver'' for continental crust. Old sections of continent are underlain by mantle that has experienced very high degrees of partial melting and melt removal [*Boyd and Mertzman*, 1987]. The melt depletion, which includes very efficient removal of water, induces a compositional buoyancy in the residual mantle [*Boyd and McCallister*, 1976] that results in a high-viscosity and buoyant ''raft'' for the overlying crust [*Jordan*, 1975], allowing it to survive at Earth's surface in some cases for as long as 4 Gyr.

[4] Recent advances in our understanding of the petrologic processes involved in creating these highly melt-depleted residues [*Walter*, 2003] and in our ability to date the melting events responsible [*D. G. Pearson et al.*, 2002] allow samples of the continental mantle to provide clues not only to why continents have survived so long at Earth's surface but also to the mechanism(s) of continent formation and modification. In this review we first summarize important observations concerning the physical, chemical, and chronological characteristics of the subcon-

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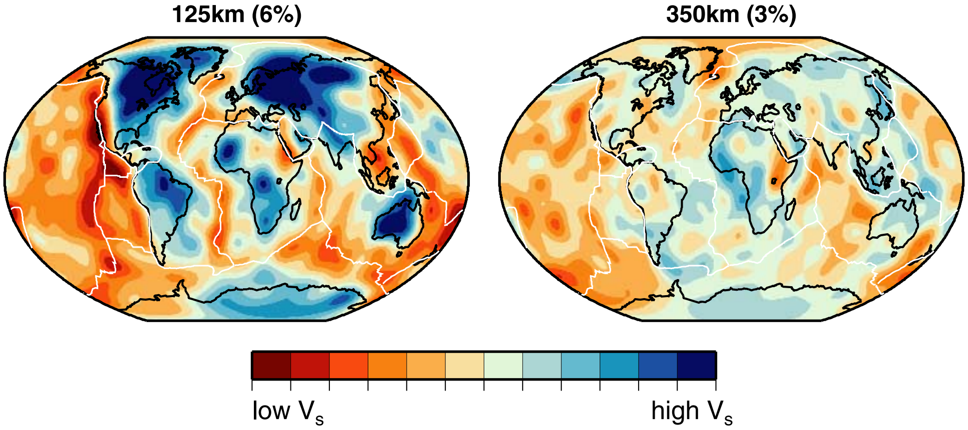

Figure 1. Map of shear wave velocity variation in the mantle at depths of (left) 125 and (right) 350 km. The total range in velocities at 125 km is ±6% and ±3% at 350 km depth. White lines are plate boundaries; continents are outlined in black. At 125 km depth the continent-ocean distinction is clearly visible in the high (blue) seismic velocities characteristic of continental lithospheric mantle. By 350 km depth the correlation between continents and high *Vs* is no longer clear. Figure produced by J. Ritsema with data from the s20rts model of *Ritsema et al.* [2004].

tinental mantle, and then we explore the significance of these observations for understanding the origin and evolution of continental lithosphere.

2. SEISMIC CHARACTERISTICS OF THE CONTINENTAL MANTLE

[5] Figure 1 shows that the clear first-order distinction in seismic velocity between oceans and continents extends well into the shallow upper mantle. Like most of the mantle [*Bina*, 2003], the range of seismic velocities in the continental lithospheric mantle indicates that the most abundant rock type is peridotite, a rock type composed predominantly of the magnesium-rich silicates olivine and orthopyroxene with lesser amounts of calcium-bearing clinopyroxene and an aluminum-rich mineral that changes from plagioclase to spinel to garnet with increasing pressure. Peridotite is also by far the dominant mantle rock recovered as a xenolith [*Nixon*, 1987; *Pearson et al.*, 2003], with eclogite, the high-pressure equivalent of basalt, of secondary abundance. For southern Africa, which has the most comprehensive xenolith data set in the world, *Schulze* [1989] showed that although eclogite abundances in the cratonic mantle may reach 3–15% locally, the overall abundance of eclogite in the cratonic root appears to be 1% or less.

[6] Eclogite that formed as the high-pressure equivalent of oceanic basalt has much higher seismic velocities and density in the upper mantle than does peridotite (Figure 2), so significant quantities of eclogite in the continental mantle should be observed in seismic imaging but only if concentrated into bodies large enough (kilometer scale) to be resolved by the wavelengths of the seismic waves used to image the mantle. In general, high-velocity bodies consistent with large masses of eclogite are not observed in the continental mantle. The exception may be in the Slave craton of northern Canada where seismic reflectors in the shallow mantle may be eclogitic layers too small to be resolved by mantle tomography [*Bostock*, 1997]. Extensive studies in southern Africa, however, have failed to reveal similar reflectors or high-velocity bodies in the cratonic mantle [*James et al.*, 2001; *Gao et al.*, 2002b]. Other old continental areas that have been well studied seismically, such as southeast Brazil, also have turned up no evidence for sizable volumes of eclogitic mantle [*VanDecar et al.*, 1995; *Schimmel et al.*, 2003]. Small regions of eclogitic reflectors in the uppermost mantle, as inferred for the Slave craton, may eventually be revealed through improved high-resolution imaging studies, but at present, eclogite appears to be a relatively minor component of the continental lithospheric mantle.

[7] Seismic velocities in the continental lithospheric mantle vary considerably but generally are higher than in oceanic mantle and, unlike in oceanic mantle, commonly correlate best not with the age of the overlying crust but with the age of last tectonomagmatic activity in the area. In North America, for example, a sharp boundary between a stable eastern region and a tectonically active western region (with demarcation roughly along the Rocky Mountain Front) is marked by a range in shear wave velocity of nearly 10% at 100 km depth [*van der Lee and Nolet*, 1997; *Goes and van der Lee*, 2002; *van der Lee*, 2002; *Godey et al.*, 2004]. Velocity contrasts of that magnitude (10%) correspond to a thermal contrast of ~500°C [*Goes and van der Lee*, 2002], although compositional variations, notably the volumetric fraction of Fe, are commonly cited as a significant contributing factor [*Goes and van der Lee*, 2002; *Godey et al.*, 2004].

[8] Table 1 lists representative seismic velocities and inferred densities for the shallow mantle in a number of different tectonic settings. At the low-velocity end of the spectrum the mantle beneath actively deforming continental areas such as the Basin and Range Province in western North America has both compressional (V_p) and shear (V_s) wave velocities approaching those found in young ocean basins even though much of the crustal section in the Basin

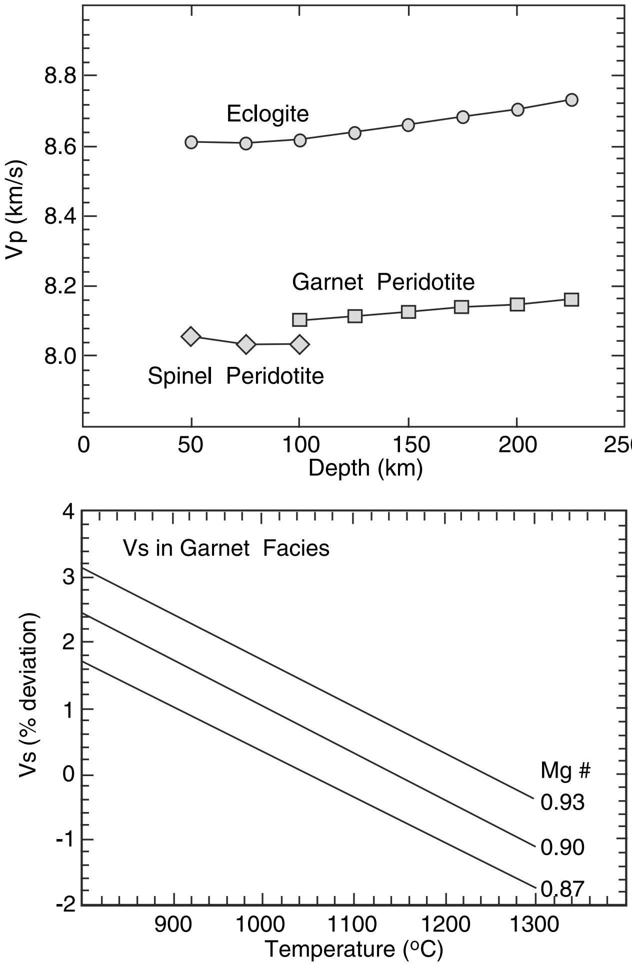

Figure 2. (top) Range of calculated compressional wave velocities (*Vp*) for hypothetical mantle eclogite and fertile peridotite compositions based on the Kaapvaal craton geotherm determined by xenolith thermobarometry [*James et al.*, 2004]. (bottom) Relationship between shear wave velocity (*Vs*) and temperature in garnet peridotite with different magnesium numbers (Mg# = Mg/(Mg + Fe) in the whole rock sample) [after *Lee*, 2003].

and Range is mid-Proterozoic in age. At the other extreme, old, tectonically stable portions of the continents called cratons have V_p and V_s higher than found anywhere in the ocean basins. Figure 3 provides a cross section showing the variation in V_p through the upper mantle of southern Africa, a geologic section that includes two early Archean cratons surrounded by terrains added during Late Proterozoic accretionary events [*Tankard et al.*, 1982]. This tomographic image shows clearly that the older sections of southern Africa are underlain to depths of at least 250 km by seismically fast mantle compared to surrounding Proterozoic terrains [*James et al.*, 2001]. Although most cratons have thick, seismically fast roots in the upper mantle [*Jordan*, 1975, 1978, 1988; *Grand*, 1987; *Polet and Anderson*, 1995; *Bostock*, 1997; *van der Lee and Nolet*, 1997; *Godey et al.*, 2004], there are exceptions; for example, the Wyoming craton of North America where V_s is as low as under much of the actively deforming Basin and Range Province. In this respect, however, the Wyoming craton is anomalous, as the low *S* velocities appear not to be accompanied by similarly low *P* velocities [*van der Lee and Nolet*, 1997; *Goes and van derLee*, 2002], an observation that remains largely unexplained.

[9] In the ocean basins the base of the oceanic lithosphere is marked by a strong decrease in V_s at depths generally less than 100 km beneath the crust. Similar low-velocity zones are seen under tectonically active continental regions, such as the Basin and Range, but available evidence indicates that in these continental areas the mantle low-velocity zone begins immediately beneath the Moho, with a high-velocity "lid" similar to that seen in ocean basins notably absent. Low-velocity zones beneath stable continental regions tend to be either extremely weak or entirely absent. As a result the base of the continental lithosphere is not well defined by seismological data: With increasing depth the high seismic velocities of the stable lithosphere gradually approach those of the convecting mantle across a broad transition region. This zone of velocity transition, in general not well resolved, ranges from 200 to perhaps 400 km deep beneath different cratons. The largest craton, the Superior Province of North America, also has the deepest seismically fast root into the mantle [*Grand*, 1987]. Some seismic results, such as those reported by *VanDecar et al.* [1995] that show evidence for a 130 Ma Paraná "fossil" plume conduit beneath the Brazilian shield extending to at least 600 km depth, can be interpreted to indicate that the entire upper mantle beneath large continents may form a coherent long-lived unit attached to the continental lithosphere.

[10] Both temperature and composition affect seismic velocities of rocks in the mantle, with temperature exerting the dominant control (Figure 2) [*Lee*, 2003; *James et al.*, 2004]. *Goes et al.* [2000] provide estimates for the change in seismic velocity per 100°C temperature of $V_p \sim 0.5-2\%$ and $V_s \sim 0.7-4.5\%$. *Godey et al.* [2004] quantify these estimates more precisely by noting that a 2% increase in velocity can be explained either by a 120°C decrease in temperature, a 7.5% depletion in iron, or a 15% depletion in aluminum.

[11] Cratons are characterized by the lowest surface heat flow of any province on Earth [*Nyblade and Pollack*, 1993; *Artemieva and Mooney*, 2002]. Temperature differences at

TABLE 1. Mantle Shear Wave Velocity (V_s), Density (ρ), and Temperature (*T*) at 100 km Depth for Various Geologic Settings

Geologic Setting[a]	V_s, *km/s*	ρ, *g/cm*^{3}	*T*, °*C*
Kaapvaal craton	4.7	3.33	800
Proterozoic mobile belt	4.6	3.38	850
Midcontinent (United States)	4.5	3.35 (?)	1000
Basin and Range (United States)	3.9	3.32 (?)	1350
Ocean basin (∼60 Ma)	4.43	3.35	1080
Oceanic ridge	<3.7	3.2 (?)	1450
PREM	4.46	3.37	

[a]Midcontinent values are from *Goes and van der Lee* [2002]; ocean basin velocity is based on the oceanic reference model of *Ritsema and Allen* [2003]. Oceanic ridge values are from Mantle Electromagnetic and Tomography Experiment results [*Webb and Forsyth*, 1998]. Basin and Range values are from *van der Lee and Nolet* [1997].

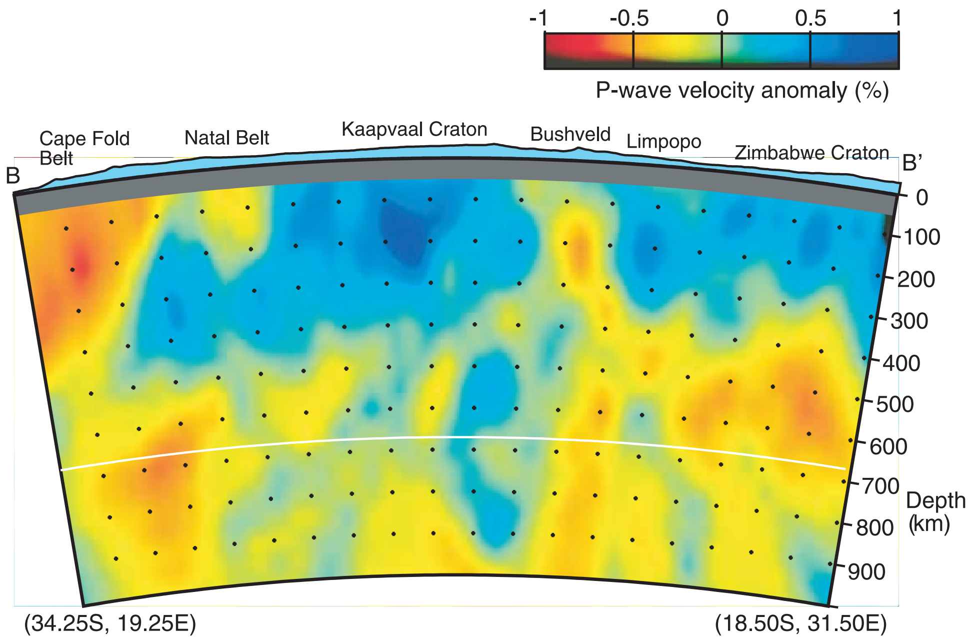

Figure 3. Cross section showing the *P* wave velocity variation in the upper mantle of southern Africa [*James et al.*, 2001]. The cross section runs from SW to NE beginning in Cape Town (B) and extending into central Zimbabwe (B′). The light blue band at the top shows vertically exaggerated surface topography. Along the top of the cross section are listed the major crustal terrains encountered from the circa 600–700 Ma Cape Fold Belt, through the mid-Proterozoic Namaqua-Natal Belt, into the Archean Kaapvaal and Zimbabwe cratons. Also shown are the surface location of the 2.0 Ga Bushveld igneous intrusion into the Archean crust of the Kaapvaal craton and the Limpopo Belt, a zone of late Archean collision between the Kaapvaal and Zimbabwe cratons. Zones of seismically fast mantle are largely restricted to beneath the Archean crust and extend to depths of at least 250–300 km.

depth between cratons and tectonic regions based on heat flow calculations generally are consistent with results based on inversion of seismic velocities to obtain thermal structure [*Goes and van der Lee*, 2002]. Essentially all studies, whether based on seismic velocity, heat flow, gravity, or mantle xenoliths, indicate that cratonic mantle is much colder than oceanic mantle. Using estimates of heat production in Archean crust and mantle and a nominal cratonic surface heat flow of 40 mW/m^2 [*Pollack and Chapman*, 1977], the calculated temperature at 100 km depth beneath a craton is ~750°C [*Nyblade*, 1999]. Adding 500°C to this would put the Basin and Range mantle close to the dry melting temperature of peridotite. While there is active volcanism in the Basin and Range, there is no evidence that the bulk of the shallow mantle is substantially molten. Consequently, the large seismic velocity difference between various regions of the continental mantle or between cratons and oceans [*Forte and Perry*, 2000] is unlikely to be the result only of varying temperature. This observation was the basis for *Jordan*'s [1975] tectosphere hypothesis in which he introduced the principle of a compositionally induced buoyancy compensating for the colder temperature of the cratonic mantle to produce a cratonic root neutrally buoyant with respect to the much hotter oceanic mantle. *Godey et al.* [2004] calculated both temperature and compositional differences from a joint inversion of *S* wave velocities and density perturbations beneath North America and concluded that the maximum temperature range at 100 km depth between craton and tectonic mantle was −440°C and the range in Fe content was ~4%. Curiously, the *S* wave velocity perturbations calculated by *Godey et al.* [2004] do not decrease significantly at depths greater than 100 km, whereas those of *van der Lee and Nolet* [1997] and others do.

3. CONTINENTAL GEOTHERMS

[12] At the high temperatures present in the mantle, diffusion of elements between minerals is relatively rapid, so the constituent minerals of mantle rocks maintain compositions that are consistent with chemical equilibrium between the phases under the extant pressure and temperature conditions. Some of the elemental exchange reactions between minerals are temperature sensitive, and others are pressure sensitive [*Finnerty and Boyd*, 1987; *Brey et al.*, 1990; *Smith*, 1999]. Because the transport mechanism of xenoliths to the surface is so rapid, cooling occurs so quickly that the chemical composition of the minerals as they existed under mantle conditions is preserved. When a number of xenoliths are available from a single volcanic center, the technique of mineral thermobarometry can allow reconstruction of the temperature profile for the whole depth column from which the xenoliths were derived. Typical estimates of the accuracy of pressures derived from thermobarometry are 0.3–0.5 GPa for commonly applied barometers and 30°–180°C for a range of thermometers

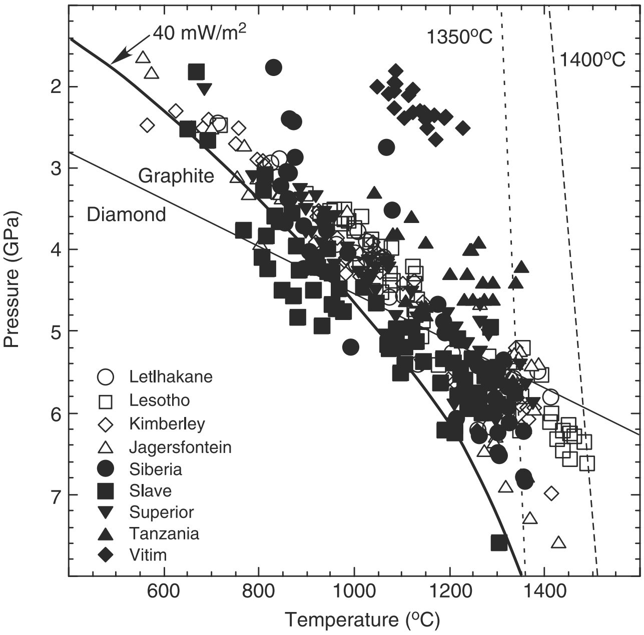

Figure 4. Mineral thermobarometry results for mantle xenoliths from a variety of continental areas [*Ionov et al.*, 1993; *Rudnick and Nyblade*, 1999]. Curve shows a 40 mW/m^2 conductive geotherm assuming that the lithospheric mantle has no internal heat production [*Pollack and Chapman*, 1977]. The straight line is the diamond-graphite phase boundary [*Kennedy and Kennedy*, 1976], and the dotted and dashed lines show the position of adiabats for potential temperatures of 1350°C and 1400°C, respectively [*Rudnick and Nyblade*, 1999].

[*Pearson et al.*, 2003]. Much of the uncertainty in the thermometry arises from the sensitivity of the Fe-Mg cation exchanges to mismatches between Fe^{3+}/Fe^{2+} ratios of samples and experimental calibrations [*Canil and O'Neill*, 1996]. Unfortunately, so far, there has been no detailed xenolith geotherm work done where the effects of Fe^{3+} are fully quantified.

[13] *Boyd* [1973] was the first to use mineral thermobarometry to show that the geotherm recorded by mantle xenoliths from the Kaapvaal craton of South Africa follows a conductive geotherm roughly consistent with the 40 mW/m^2 average surface heat flow of cratons [*Pollack and Chapman*, 1977; *Nyblade and Pollack*, 1993] to depths of the order of 140–180 km depending on location (Figure 4). Note that the exact shape and position of this 40 mW/m^2 geotherm is highly dependent on unverifiable assumptions about the distribution of heat-producing elements in both the crust and conductive mantle lithosphere [*Rudnick and Nyblade*, 1999] as well as uncertainty in the variation of thermal conductivity with depth [e.g., *Hofmeister*, 1999]. In general, the coldest geotherms are found beneath stable cratons where temperatures at 100 km depth are between 750°C and 900°C (Figure 4) [*Finnerty and Boyd*, 1987]. These cold geotherms are common in Archean cratons [*Boyd et al.*, 1997; *Kopylova et al.*, 1999], but geotherms only slightly warmer also can be found beneath long stable sections of Proterozoic crust such as in the State Line district of Colorado [*Eggler et al.*, 1987], the Proterozoic of north Australia [*Jaques et al.*, 1990], and at least in the shallow (<150 km) sections of Proterozoic lithosphere around the cratons of southern Africa [*Finnerty and Boyd*, 1987; *Boyd et al.*, 2004]. Tectonically active areas of the continents tend toward much hotter geotherms even in areas where crustal ages are old. For example, in Tanzania where the Archean crust is currently being rifted by the Central African Rift, xenolith temperatures exceed 1000°C at 100 km depth [*Lee and Rudnick*, 1999]. Even higher mantle temperatures are recorded in tectonomagmatically active areas under younger crustal sections such as at the Vitim volcano in the Baikal Rift where temperatures are estimated to range from 1020° to 1250°C at 80 km depth [*Ionov et al.*, 1993].

[14] For comparison, one of the few areas in the ocean basins that provides a xenolith record covering a sufficient depth range to define a geotherm is in Malaita, a fragment of the Ontong-Java Plateau, where temperatures are near 1000°C at 100 km depth [*Nixon and Boyd*, 1979]. The mantle beneath tectonomagmatically active continental areas thus is at least as hot as, or hotter than, the shallow upper mantle beneath a circa 120 Ma oceanic plateau and is as much as 500°C hotter than cold cratonic mantle at 100 km depth (Table 1). The magnitude of temperature (T) variation in the shallow continental mantle recorded by xenolith thermobarometry therefore is similar to that inferred from the variation in seismic velocity.

[15] One long discussed feature of mantle xenoliths is that they commonly show an inflection in the inferred geotherms at depths ranging from as shallow as 140 km to as deep as 180 km [*Boyd*, 1987]. The geotherm inflection generated the often used terminology of "low temperature" to describe those xenoliths lying on the conductive geotherm and "high temperature" for xenoliths offset from the conductive geotherm to higher temperature [*Boyd*, 1987]. With the addition of new thermobarometers the significance, and even the existence, of the geotherm inflections has been the subject of considerable debate as the inflections appear prominently only with certain combinations of mineral thermometers and barometers [*Smith and Boyd*, 1987; *Brey et al.*, 1990; *Rudnick and Nyblade*, 1999; *Smith*, 1999]. The distinction between low- and high-temperature xenoliths, however, is also generally manifest in terms of their textures. Low-T xenoliths are predominantly coarse-grained rocks with mineral fabrics consistent with continual recrystallization under conditions of low strain. High-T xenoliths are characteristically strongly sheared with matrices of fine-grained recrystallized (neoblastic) olivine flowing around larger grains of the stronger (porphyroclastic) garnet and orthopyroxene [*Harte*, 1977]. Low-T peridotites also tend to be more depleted in the elements that would be removed by partial melting than are the high-T peridotites [*Boyd and Mertzman*, 1987; *Smith and Boyd*, 1987].

[16] The debate over the interpretation of the high-T samples touches on the general question of whether the geotherms recorded by xenoliths indeed are steady state conductive geotherms or instead represent transient condi-

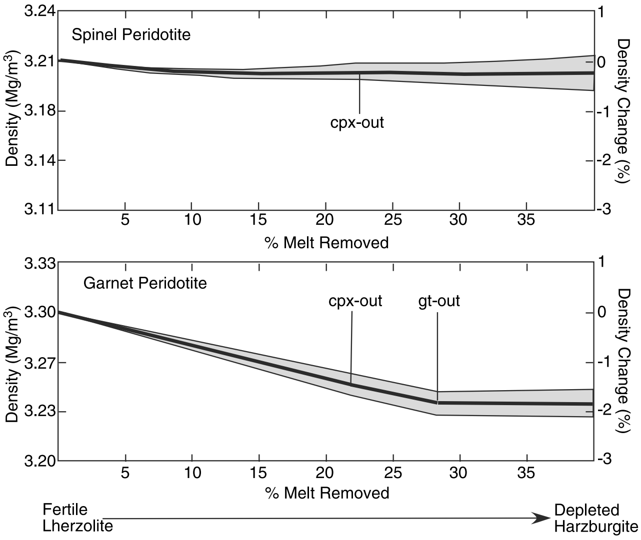

Figure 5. Density change in spinel and garnet peridotite at room temperature and pressure due to melt removal. The bold line in plot shows the density change as a function of the percentage of melt removed, with the shaded band showing the error margin of the density calculation. Also indicated on the plots is the percentage of melting where clinopyroxene (cpx-out) and garnet (gt-out) are completely consumed. Figure modified from Schutt and Lesher (submitted manuscript, 2004).

tions present at the time the xenoliths were brought to the surface [*Harte and Freer*, 1982; *Bell et al.*, 2003]. The sensitivity of xenolith "geotherms" to recent local heating events has long been recognized and should be kept in mind in the interpretation of xenolith thermobarometry. A large change in the basal temperature of the lithosphere requires times on the order of hundreds of millions of years to migrate conductively through a 200 km thick lithosphere [*Nyblade*, 1999]. Given the 70–150 Ma age of many southern African kimberlites [*Smith et al.*, 1985], a clear candidate for a major thermal event that could have modified the geotherm in the southern African lithosphere is the breakup of Gondwana and its associated flood basalt activity in the ~180 Ma Karoo [*Marsh et al.*, 1997] and ~130 Ma Etendeka/Paraná [*Peate*, 1997] provinces. Recent surface wave studies of the mantle beneath the southern Kaapvaal craton, however, show that the present-day *S* velocity structure is consistent with that estimated for xenolith data from 80–90 Ma samples [*Larson et al.*, 2003; *James et al.*, 2004]. These results suggest that to depths of at least 180 km under southern Africa the geotherm has not changed significantly in the past 80–90 Myr.

4. CHEMICAL CHARACTERISTICS OF CONTINENTAL LITHOSPHERIC MANTLE

[17] Peridotites are commonly referred to as "fertile" or "depleted" depending on the degree to which their composition has been modified by extraction of partial melts. A variety of methods have been used to estimate the composition of the mantle before any melting has occurred [*Ringwood*, 1969; *Jagoutz et al.*, 1979; *McDonough and Sun*, 1995; *Palme and O'Neill*, 2003]. When melting occurs in the mantle, certain elements such as calcium and aluminum are concentrated into the melt and are removed with it while other elements, particularly magnesium, selectively remain behind in the solid residue. Those elements that concentrate in the melt are known as "incompatible" elements. For dry melting at low pressure (<3 GPa), iron is nearly equally partitioned between melt and solid, which leads to little change in iron concentration over a wide range of magnesium contents in melt residues [*Kinzler and Grove*, 1992]. This situation changes with increasing pressure where iron concentration drops in the residue with increasing degrees of melt removal [*Walter*, 1999]. A similar result was obtained in water-saturated melting of peridotite [*Kawamoto and Holloway*, 1997] over a wide pressure range in contrast to the results of *Gaetani and Grove* [1998], who showed preferential retention of iron in the residue during water-saturated melting at low pressure.

[18] The mineralogical expression of the transition from fertile to depleted peridotite is the loss first of clinopyroxene and garnet, the main hosts for incompatible elements in the mantle, and then orthopyroxene, resulting in the rock name transition from lherzolite (olivine + orthopyroxene + clinopyroxene + garnet or spinel) to harzburgite (olivine + orthopyroxene) to dunite (just olivine). A critical result of this mineralogical transition is that depleted peridotite is less dense than fertile peridotite at the same temperature. Because garnet is about 15% denser than olivine and orthopyroxene, the reduction of modal garnet, and the increase in Mg to Fe ratio, in a depleted peridotite reduces its density by as much as 2% compared to fertile peridotite at the same temperature [*Boyd and McCallister*, 1976; *Jordan*, 1979; *Poudjom Djomani et al.*, 2001; *Kelly et al.*, 2003; *Lee*, 2003; *James et al.*, 2004]. In a recent comprehensive laboratory study of the effect of progressive depletion on density and seismic velocity, D. L. Schutt and C. E. Lesher (The effects of melt depletion on the density and seismic velocity of garnet and spinel lherzolite, submitted to *Journal of Geophysical Research*, 2004, hereinafter referred to as Schutt and Lesher, submitted manuscript, 2004) have graphed the density as a function of degree of depletion in both spinel and garnet peridotite, the results of which are shown in Figure 5.

[19] Figure 6 and Table 2 separate continental mantle peridotite according to the tectonic setting from which the samples originate. In presenting these averages it is important to recognize that each setting displays a wide range about the median values reported in Table 2, so considerable care in the interpretation of the median composition is warranted. Nevertheless, some clearly distinguishable features are apparent. Both massif peridotites, tectonically obducted fragments of upper mantle [*Bodinier and Godard*, 2003], and mantle xenoliths from off-craton localities range to more fertile compositions than either oceanic peridotites or xenoliths erupted from beneath Archean crustal terrains. This distinction is obvious in the median compositions

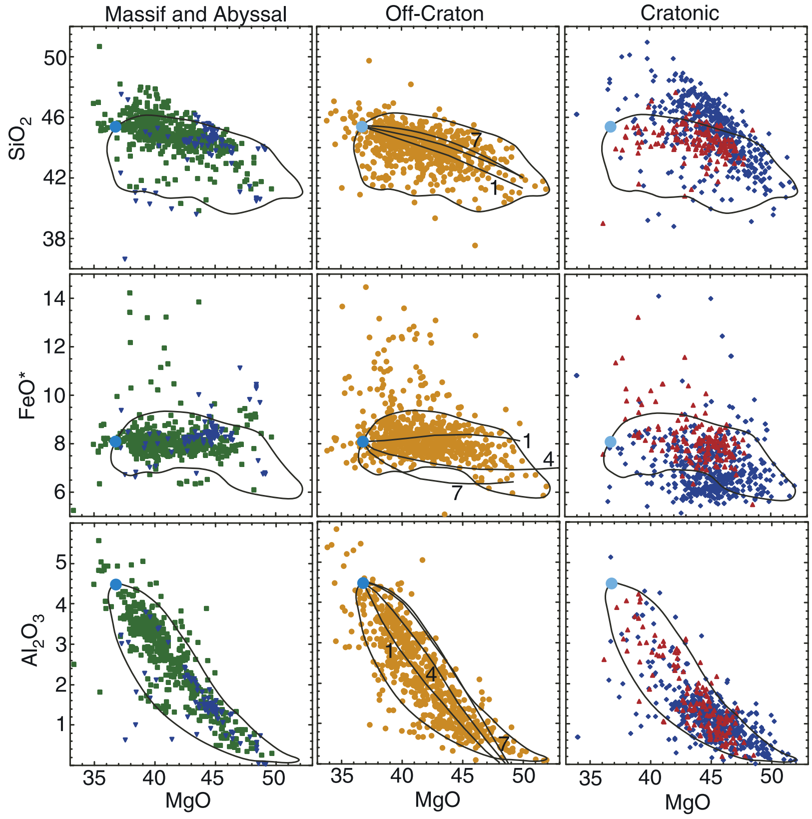

Figure 6. Major element ranges observed for peridotites from a variety of tectonic settings including (left) ocean basins (dark blue points) and mantle sections tectonically obducted onto continental crust (massif peridotites (green points)), (middle) peridotite xenoliths (primarily spinel peridotites) erupted in tectonomagmatically active areas away from cratons, and (right) peridotite xenoliths carried by kimberlites erupting through both Archean and middle to Early Proterozoic crust. In the right column the blue points are for low-T peridotites, and the red points are for high-T samples. The field shown on all plots outlines the majority of the data for off-craton peridotite xenoliths to serve as a reference. The large light blue dot in all plots is an estimate of fertile mantle composition. Lines emanating from fertile mantle in the center column plots show the expected composition of residues of partial melting occurring at 1, 4, and 7 GPa [*Walter*, 1999, 2003]. Peridotite data are from the compilation of *McDonough* [1990].

(Table 2) where the massif peridotites and off-craton peridotite xenoliths have Al_2O_3 and CaO concentrations roughly midway between cratonic samples and estimates of fertile mantle composition. The median composition of low-T cratonic xenoliths is not too different from the highly melt-depleted peridotites characterizing the shallow oceanic mantle with the notable exception of the significantly lower iron concentration of the cratonic samples.

[20] Another feature of the major element characteristics of samples from the continental mantle that has been the subject of much discussion is the tendency of some cratonic peridotites to have SiO_2 concentrations higher than expected for simple partial melt residues (Figure 6). The SiO_2 excess is manifest in anomalously high orthopyroxene contents and was first observed for Kaapvaal low-T cratonic peridotites [*Boyd and Mertzman*, 1987] and then Siberian cratonic peridotites [*Boyd et al.*, 1997] after which it was believed to be a general characteristic of Archean subcontinental mantle. Xenoliths from more recently studied Archean terrains, particularly the Slave [*Kopylova and Russell*, 2000] and North Atlantic cratons (east Greenland) [*Bernstein et al.*, 1998], do not follow this trend but instead plot at the low (~42–43%) SiO_2 concentrations expected for residues of extensive partial melt removal. The high-SiO_2 trend has a number of interesting explanations, including sampling of a more "chondritic" (e.g., higher Si/

TABLE 2. Major Element Composition of Various Mantle Materials[a]

	Fertile	*Oceanic*	*Massif*	*Off-Craton*	*Low-T Craton*	*High-T Craton*
SiO_2	45.40	44.66	44.98	44.47	44.18	44.51
TiO_2	0.22	0.01	0.08	0.09	0.02	0.11
Al_2O_3	4.49	0.98	2.72	2.50	1.04	0.84
FeO*	8.10	8.28	8.02	8.19	6.72	8.08
MgO	36.77	45.13	41.15	41.63	46.12	44.76
CaO	3.65	0.65	2.53	2.44	0.54	1.08
Mg#	0.890	0.907	0.902	0.899	0.925	0.908
Ni	1860	2515	2150	2204	2224	2084
Cr	2520	2809		2736	2470	2588
Number[b]		146	256	171	183	26

[a]All major element concentrations are in weight percent. Ni and Cr concentrations are listed as parts per million. Fertile mantle estimate is from *Palme and O'Neill* [2003]. Remaining compositions are the averages for the data shown in Figure 6, which are derived from the compilation of *McDonough* [1990].

[b]Number of samples included in the average is shown.

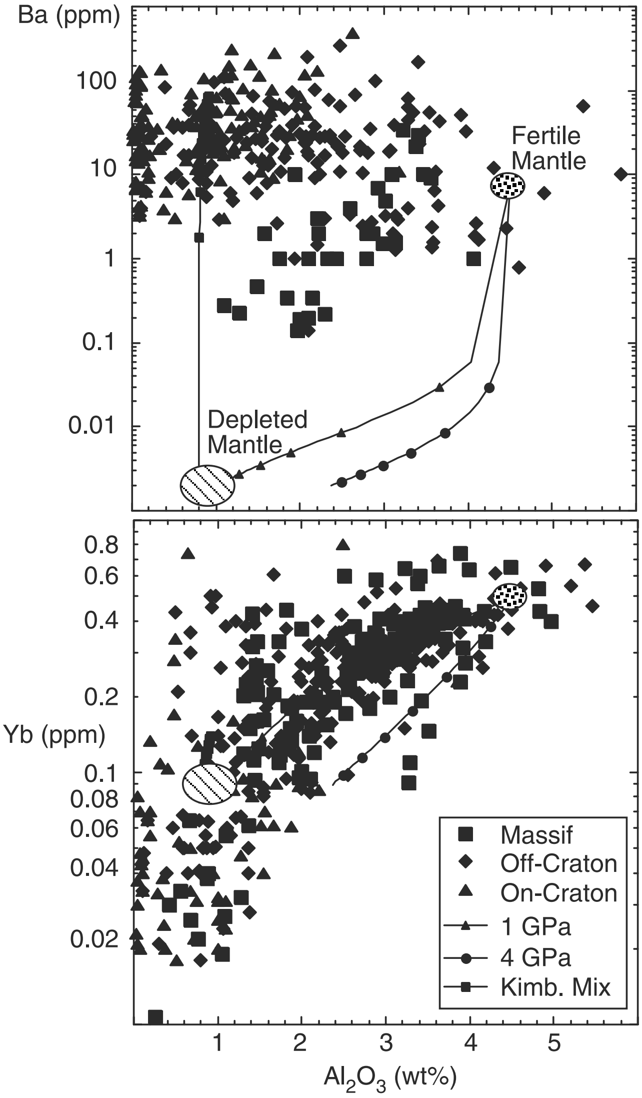

Figure 7. Comparison of the observed concentration of a highly (Ba) and moderately (Yb) incompatible trace element versus Al_2O_3 concentration in mantle peridotites from various tectonic settings. Lines denoted by triangles (1 GPa) and circles (4 GPa) show the compositional variation expected for partial melting residues [*Walter*, 1999] starting from an estimate of fertile mantle shown by the smaller oval. Each point along these lines indicates 10% increments in the amount of partial melt removed. The line marked by the squares shows the mixing trajectory between a residue of high-degree partial melting (large oval) and a primary kimberlitic magma [*Le Roex et al.*, 2003]. Points along these lines reflect 0.5% increments in kimberlite addition to a total of 10% kimberlite, with 90% depleted peridotite. Peridotite data are from the compilation of *McDonough* [1990].

Mg) portion of the primitive mantle [*Herzberg and Zhang*, 1994; *Francis*, 2003], extremely high pressure partial melting and melt interaction [*Herzberg*, 1999], metamorphic concentration of orthopyroxene [*Boyd et al.*, 1997], and melt interaction in the shallow mantle [*Kelemen et al.*, 1998] involving percolation of Si-rich melts derived from subducting slabs [*Rudnick et al.*, 1994]. The debate over these possible explanations continues, but the discovery that the Si enrichment is not a ubiquitous property of Archean subcontinental mantle lends strength to the suggestion that Si enrichment is a secondary (metasomatic) feature imposed upon the subcontinental mantle after its formation as a residue of melting.

[21] The debate over the causes of Si enrichment in mantle peridotites is symptomatic of a general problem in the interpretation of the composition of mantle rocks. All mantle rocks are in essence high-temperature and high-pressure metamorphic rocks. Rapid recrystallization and reequilibration of peridotite at high temperature in the mantle can disguise the textural and chemical effects of the infiltration of lithospheric mantle by melts rising from the underlying mantle. Evidence for secondary melt-rock interaction, called metasomatism, in the continental lithospheric mantle is abundant [*Menzies and Hawkesworth*, 1987]. The majority of oceanic peridotites and a large percentage of massif peridotites are depleted in almost all incompatible elements [*Bodinier and Godard*, 2003] (Figure 7) and thus fit well the model of them being residues of partial melting. In general, the abundance of moderately incompatible trace elements, for example, ytterbium (Figure 7), is also very low in cratonic peridotites, consistent with the major element inferences that cratonic peridotites are the residues of large degree partial melt extraction. In contrast, cratonic peridotite xenoliths commonly are unexpectedly enriched in highly incompatible trace elements, for example, barium (Figure 7). Some of this enrichment in the highly incompatible trace elements simply reflects infiltration of individual xenoliths by the host magma (Figure 7), since kimberlites have among the highest concentration of these elements of any terrestrial magma. However, at least some fraction of this incompatible element enrichment is ancient [*Richardson et al.*, 1984].

[22] The reader may well ask, "Who cares?" about the abundance of some obscure trace elements in the mantle, but some of these elements can strongly influence the physical characteristics of the continental lithospheric mantle. For example, among the highly incompatible trace elements are the radioactive elements potassium, uranium, and thorium. These elements determine the heat-producing capacity of the lithospheric mantle and thus the temperatures present in the conducting lithosphere. Another incompatible "element" in the mantle is water. Water plays two crucial roles in the behavior of the mantle. First, water lowers the melting point of the mantle by as much as 300°C at a depth of 100 km and as much as 1000°C at 330 km depth [*Zhang and Herzberg*, 1994; *Kawamoto and Holloway*, 1997]. Second, water dramatically reduces the strength of olivine, the main constituent of the mantle [e.g., *Karato et al.*, 1986; *Karato and Jung*, 1998, 2003; *Mei and Kohlstedt*, 2000a, 2000b]. As a result the viscosity of dry mantle can be as much as a factor of 20–500 higher than that of water-saturated mantle at a similar temperature [*Hirth and Kohlstedt*, 1996; *Shapiro et al.*, 1999a; *Dixon et al.*, 2004]. Residues of extensive partial melt removal should be completely dry and devoid of heat-producing elements, which should leave them cold, strong, and refractory [*Karato*, 1986; *Pollack*, 1986].

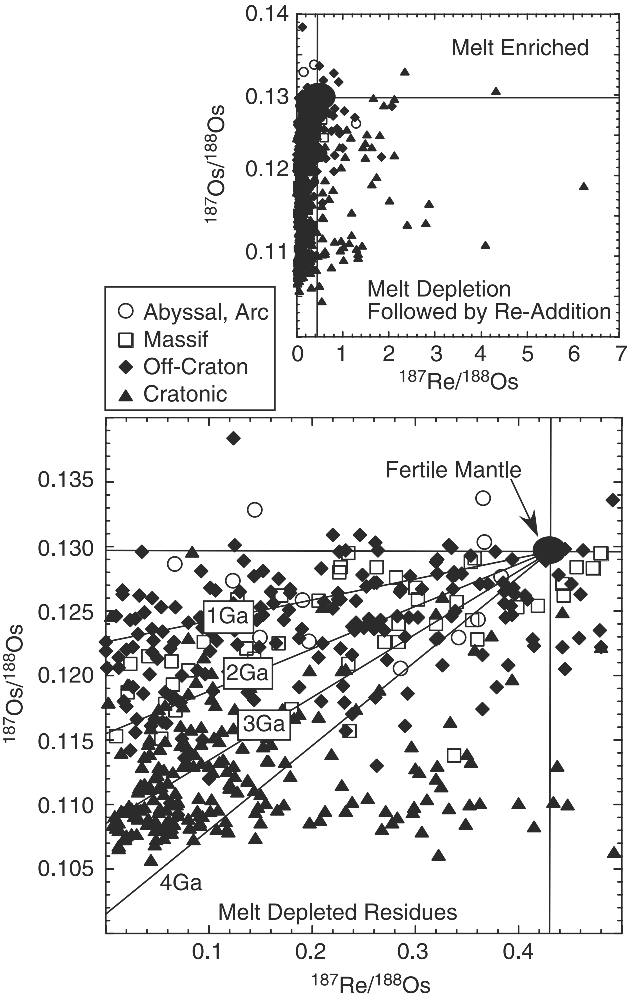

Figure 8. Re-Os isochron diagram showing the data for various types of samples of continental lithospheric mantle. The vertical and horizontal lines in Figure 8 pass through an estimate of the average Re-Os characteristics of the fertile mantle [*Meisel et al.*, 1996] and divide the plot into quadrants that can be reached through single-stage melt depletion (bottom left) and corresponding melts (top right) from average mantle. Data can only reach the top left and bottom right quadrants through an additional chemical fractionation event; for example, the scattering of points into the bottom right quadrant most likely reflects depleted xenoliths that have interacted with their transporting magma thereby raising their Re/Os ratio but not significantly changing their Os isotopic composition. Sloped lines in Figure 8 reflect data arrays expected for single-stage melt residues of 1, 2, 3, and 4 Ga age. Data are from *Walker et al.* [1989], *Reisberg et al.* [1991], *Carlson and Irving* [1994], *D. G. Pearson et al.* [1995a, 1995b, 1995c, 1998b, 1999, 2002, 2004], *N. J. Pearson et al.* [2002], *Reisberg and Lorand* [1995], *McBride et al.* [1996], *Meisel et al.* [1996, 2001], *Handler et al.* [1997, 2003], *Carlson et al.* [1999a, 1999b], *Chesley et al.* [1999], *Menzies et al.* [1999], *Ruiz et al.* [1999], *Burton et al.* [2000], *Lee et al.* [2000], *Peslier et al.* [2000a, 2000b], *Snow et al.* [2000], *Becker et al.* [2001], *Hanghoj et al.* [2001], *Irvine et al.* [2001, 2003], *Lee et al.* [2001], *Richardson et al.* [2001], *Alard et al.* [2002], *Gao et al.* [2002a], *Griffin et al.* [2002], *Schmidt and Snow* [2002], *Wang et al.* [2003], *Westerlund et al.* [2003], *Wu et al.* [2003], and *Carlson and Moore* [2004].

[23] Residues that have been infiltrated by incompatible element–rich melts could well be wet and have moderately high heat production, which would lead them to be relatively fluid and prone to melting. The consequences of metasomatism, however, depend strongly on the nature of the infiltration event. If infiltration is localized in veins, it may only minimally affect the average composition and strength of the lithosphere. If infiltration occurs over wide areas by porous flow, then the consequences could be more dramatic. The general enrichment of the Kaapvaal and Siberian lithospheres in SiO_2 is suggestive of extensive interaction with fluids/melts and raises the possibility that the strength of the lithospheric mantle may be as dependent on its metasomatic history as its formation by extensive melt depletion. On the other hand, perhaps the most sensitive monitors of the average level of metasomatism of the continental mantle are heat flow, and the geotherm since one of the common minerals added in abundance during metasomatism is phlogopite, a potassium-rich mica that would contribute substantially to the heat-producing capability of the mantle. In their study of this issue, *Rudnick et al.* [1998] noted that the average K_2O content (0.15%) measured for cratonic peridotite xenoliths would produce geotherms that are too curved to ever intersect the mantle adiabat. In order to obtain geotherms that match those recorded by the thermobarometry of xenoliths, *Rudnick et al.* [1998] arrived at a preferred, and very low, average K_2O content of the cratonic mantle of 0.03%, which, in turn, implies that the cratonic mantle, on average, has not experienced much metasomatism. This conclusion is also supported by studies of other incompatible elements [*Pearson and Nowell*, 2002] and means that the lithosphere is likely to have low average water content [e.g., *Dixon et al.*, 2004].

5. CHRONOLOGY OF CONTINENTAL LITHOSPHERIC MANTLE FORMATION AND MODIFICATION

[24] Evidence that the mantle beneath old sections of continent is also old first came from the study of silicate and sulfide inclusions contained within diamond [*Kramers*, 1979; *Richardson et al.*, 1984, 1985, 1993]. The first results provided model ages approaching 3.4 Ga for diamonds from South Africa [*Richardson et al.*, 1984], with later studies confirming the old dates but also showing younger diamond formation ages [*Richardson et al.*, 1993; *Pearson and Shirey*, 1999]. Both the old and younger dates for diamond formation have been confirmed with Re-Os age dating of sulfide inclusions in diamonds [*Pearson et al.*, 1998a, 1998b, 1999; *Richardson et al.*, 2001, 2004; *Shirey et al.*, 2002; *Westerlund et al.*, 2003]. However, since both carbon and sulfur are incompatible elements in the mantle, diamond and the sulfides found in both diamonds and mantle peridotites [*Alard et al.*, 2002; *Griffin et al.*, 2002, 2004; *Aulbach et al.*, 2004] likely are metasomatically introduced phases in the melt-depleted peridotites. Consequently, the ages obtained for diamond inclusions and many mantle sulfides likely date this metasomatism and not the event

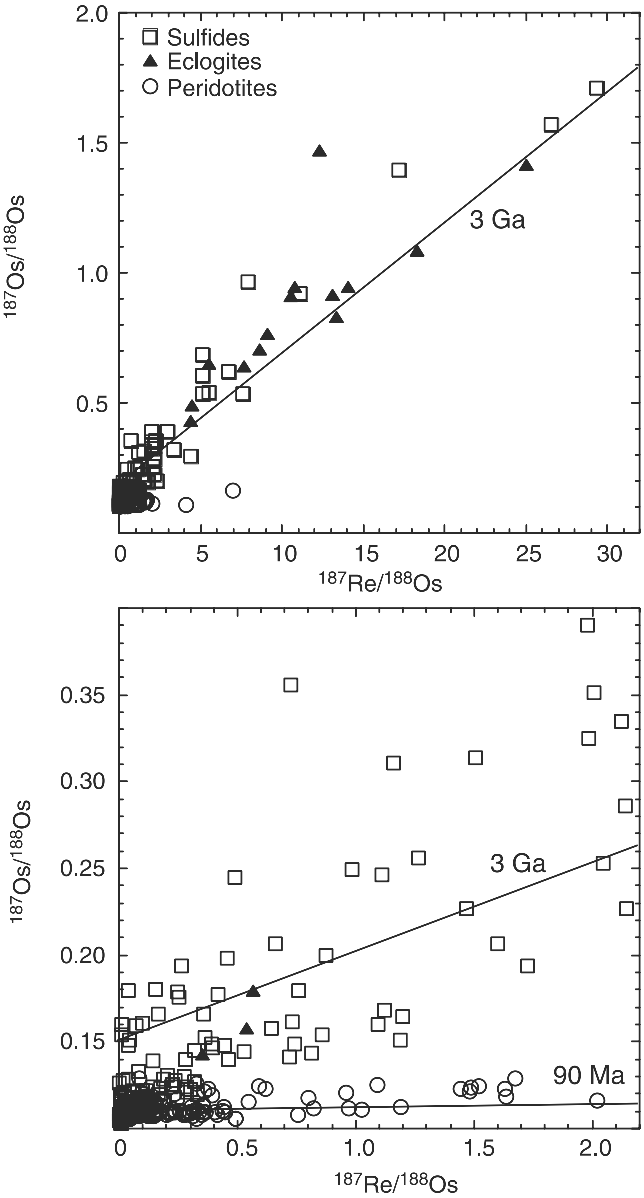

Figure 9. Re-Os isochron diagram for whole rock peridotites (circles), sulfides from Kaapvaal mantle peridotites (squares), and eclogites (triangles) from the cratonic mantle. Lines with slope corresponding to 90 Ma and 3 Ga age are shown for reference. Data for whole rock peridotites are from the sources given in Figure 8; eclogite data are from *Menzies et al.* [2003] and *Pearson et al.* [1995c]; Kaapvaal sulfide data are from *Griffin et al.* [2004].

that created the major element depletion of the continental lithospheric mantle. Indeed, the chemical characteristics of the silicate inclusions contained in diamond usually show strong incompatible element enrichment consistent with their formation from, or interaction with, metasomatic melts/fluids in the deep lithosphere [*Richardson et al.*, 1984; *Stachel et al.*, 1998, 1999; *Shimizu*, 1999].

[25] Because most radiometric systems (Rb-Sr, Sm-Nd, Lu-Hf, and U-Th-Pb) are based on elements that are moderately to highly incompatible during mantle melting, these systems in peridotite are easily overwhelmed by interaction of the peridotites with infiltrating melts. This has, until recently, made it difficult to document the timing of the melt depletion from the continental lithospheric mantle. One radiometric system that has been of considerable use in dating mantle melt extraction events is based on the decay of ^{187}Re to ^{188}Os [*Walker et al.*, 1989; *Shirey and Walker*, 1998; *D. G. Pearson et al.*, 2002; *Carlson*, 2005]. For this application the key feature of the Re-Os system is that Os is compatible during mantle melting, whereas Re is moderately incompatible. Consequently, a residue of melt extraction will have lower Re but higher Os concentration than either the fertile mantle or any mantle melt. This leaves the Re-Os system in residual peridotites less sensitive to contamination by migrating melts in the mantle thereby allowing whole rock xenolith analyses to return useful chronological information on the melt extraction event.

[26] Figure 8 shows the results of Re-Os isotopic analysis of a large number of samples of continental lithospheric mantle. By far, the majority of samples have ^{187}Os/^{188}Os lower than estimated for fertile mantle [*Meisel et al.*, 2001] confirming the expectation of low Re/Os ratio in a residue of melting and that the Re-Os system in whole rock peridotites tends to be dominated by the melt depletion event rather than later melt interaction events. The sloped lines in Figure 8 show the expected trajectory of single-stage melt residues differing in age by 1 Gyr steps. Readily apparent in Figure 8 is that the cratonic xenoliths routinely plot to lower ^{187}Os/^{188}Os and hence older age than the remainder of the samples shown. What is also evident in Figure 8 is that the data show considerable scatter that probably reflects rhenium contamination of the samples by the host magma [*Walker et al.*, 1989; *Carlson et al.*, 1999b] or changes in Re/Os ratio caused by weathering [*Handler and Bennett*, 1999]. Combined Re-Os isotope and platinum group element (Pt, Pd, Ru, Rh, Ir, and Os) analysis shows clearly that most peridotites have more Re than their degree of melt depletion would predict [*Pearson et al.*, 2004].

[27] Traditional isotopic model ages would use the measured parent/daughter ratio and isotopic composition to calculate the time when the sample had the same isotopic composition as the fertile mantle. To avoid the effects of syneruption and posteruption changes to the Re/Os ratio of a xenolith, *Walker et al.* [1989] defined the Re-depletion model age (T_{RD}), a slightly modified way to calculate the model age by assuming that the Re/Os ratio of the undisturbed sample was zero. For residues of low-degree melting where significant Re is left in the melt residue, T_{RD} severely underestimates the true age of melt depletion, but as the degree of melting increases, the Re/Os ratio of the residuum approaches zero, and T_{RD} approaches the true time of melt depletion. In contrast, for samples containing extraneous Re added at the time of xenolith capture, the traditional model age (mantle model age or T_{MA}) calculated using the measured Re/Os ratio of the sample will overestimate the true age of melt depletion. Thus T_{RD} and T_{MA} model ages provide minimum and maximum bounding ages, respectively, to the true time of melt depletion. Because of the clear evidence for recent Re introduction to whole rock

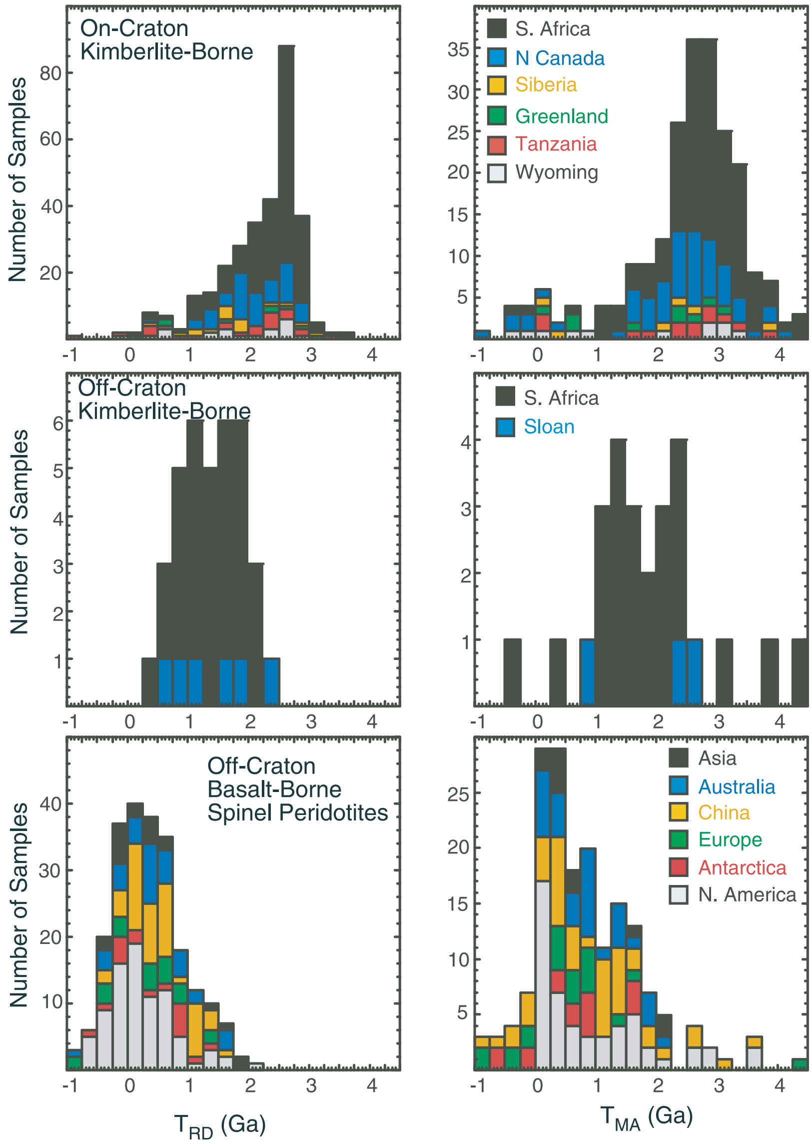

Figure 10. Re-depletion (T_{RD}) and conventional Re-Os mantle (T_{MA}) model ages for peridotite xenoliths from the continental mantle calculated with respect to the primitive mantle Re-Os parameters given by *Meisel et al.* [2001]. The data are grouped according to geologic setting of the eruptive host of the xenoliths. (top) Xenoliths carried by kimberlites erupting through Archean crust. (middle) Kimberlite-borne xenoliths erupting through long stable Proterozoic crustal sections. (bottom) Basalt-borne xenoliths erupting in tectonomagmatically active areas away from cratons. Data sources are as referenced for Figure 8.

peridotites probably caused by kimberlite infiltration of the xenoliths (Figure 9), the T_{RD} parameter can be further modified to subtract the radiogenic Os ingrowth since kimberlite eruption. This adjustment becomes particularly important for old kimberlites such as the Premier kimberlite [*Pearson et al.*, 1995a].

[28] In an attempt to avoid some of the complications involved in whole rock peridotite analysis, several attempts have been made to examine the Re-Os system in sulfides found in mantle peridotites since this phase strongly concentrates Re and Os [*Alard et al.*, 2002; *N. J. Pearson et al.*, 2002; *Griffin et al.*, 2003, 2004; *Aulbach et al.*, 2004]. While offering much complimentary information, there also are complications inherent with this approach. First, the analytical approach used [*N. J. Pearson et al.*, 2002] makes a laser-ablation analysis of only part of a sulfide grain. Exsolution of separate sulfide phases during transport to the surface has been shown to fractionate Re from Os, and hence the whole sulfide must be analyzed to obtain an accurate "whole sulfide" analysis [*Richardson et al.*, 2001]. Second, there is huge complexity in the sulfide data (Figure 9) with most of the data requiring a multiple-stage evolution, and so only very few grains can be simply interpreted in terms of a partial melt extraction event. This is not surprising since at the high degrees of melt extraction experienced by most cratonic peridotites, sulfide should have been completely consumed. Thus most mantle sulfides likely are metasomatically introduced phases. The distinct slopes of the sulfide and whole rock data in Figure 9 show that these introduced sulfides do not significantly influence

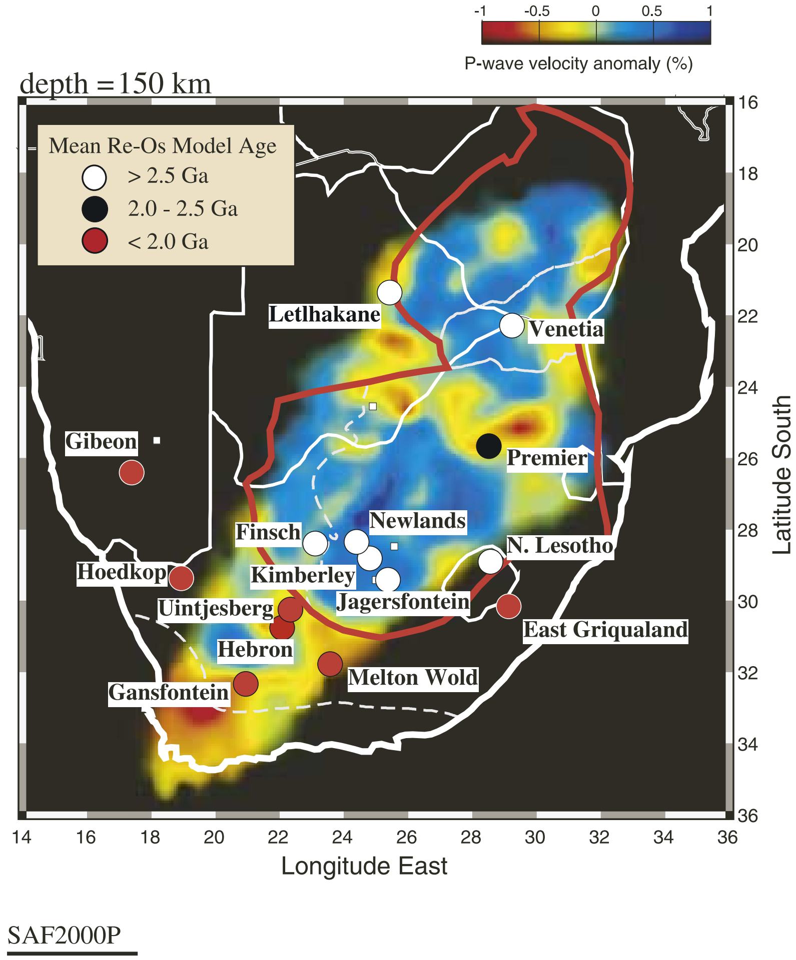

Figure 11. Comparison of *P* wave velocity variations at 150 km depth in the mantle beneath southern Africa [*James et al.*, 2001] with the average Re-depletion model age obtained for xenolith suites from the kimberlite eruption localities shown by the labeled circles [*Carlson et al.*, 1999b]. White circles show localities with mean Re-depletion model ages >2.5 Ga, and these correlate well with the seismically fast roots to the Archean cratons of southern Africa. Red circles show localities with mean xenolith Re-depletion ages <2.0 Ga. Such locations are confined to off-craton eruption sites. The only on-craton kimberlite with mean xenolith ages <2.5 Ga is the Premier locality that sits above a seismically slow portion of the Kaapvaal mantle in an area strongly affected by the 2.0 Ga intrusion of the Bushveld igneous complex.

the whole rock data. Despite the complexity shown by the sulfide data, when the same model age constraints are applied and when the few sulfides are selected that can be interpreted (rightly or wrongly) as single-stage melting products, the first-order information obtained is the same as in whole rock studies [*Pearson et al.*, 2004]. Hence, where systematic whole rock peridotite and sulfide Re-Os isotope studies have been carried out on well-selected samples, the oldest melting events are between late to middle Archean in age. The subjective part of the data analysis is in deciding whether these depletion events record lithosphere-forming events in all cases or whether they record early mantle depletion/heterogeneity prior to craton stabilization, as will be discussed below.

[29] Figure 10 compares the T_{RD} and T_{MA} age distributions obtained for a wide variety of whole rock samples of the continental lithospheric mantle. Combining all the available data onto the few histograms shown in Figure 10 ignores much of the regional complexity in the data but shows a clear first-order conclusion: There is a general correspondence between the age of the crust and the age of its underlying mantle. Mantle xenoliths erupted from beneath Archean crust provide median T_{RD} and T_{MA} ages of 2.4 and 2.8 Ga, respectively; those brought to the surface by kimberlites erupting through Proterozoic crust give median T_{RD} and T_{MA} ages of 1.4 and 2.1 Ga, respectively; and mantle samples from tectonically active areas give median T_{RD} and T_{MA} ages of 0.3 and 0.7 Ga, respectively. This correspondence between crust and mantle age has been particularly well developed for southern Africa where the Re-Os age data for xenoliths has been correlated with crustal terrain boundaries [*D. G. Pearson et al.*, 2002; *Carlson and Moore*, 2004; *Griffin et al.*, 2004], major igneous events in the crustal record [*Carlson et al.*, 1999b], the seismic characteristics of the continental lithospheric mantle (Figure 11), and diamond occurrence [*Shirey*

et al., 2002]. A good temporal connection between crust and lithospheric mantle is also present in the Siberian craton [*Pearson et al.*, 1995b, 1995c, 1998a; *N. J. Pearson et al.*, 2002; *Griffin et al.*, 2002], in the North Atlantic craton as sampled in east Greenland [*Hanghoj et al.*, 2001], and in the Slave craton of northern Canada [*Irvine et al.*, 2003; *Westerlund et al.*, 2003; *Aulbach et al.*, 2004]. No xenolith locality erupting through post-Archean crust brings Archean mantle to the surface, indicating that the cratonic lithospheric mantle is indeed mechanically coupled to the overlying crust. Some areas, however, including Antarctica [*Handler et al.*, 2003], eastern Australia [*McBride et al.*, 1996; *Handler et al.*, 1997] and parts of western North America [*Peslier et al.*, 2000a, 2000b; *Lee et al.*, 2001; *Meisel et al.*, 2001], do provide mantle ages older than expected based on crustal age determinations. This result may reflect the superposition of younger over older terrains through tectonic imbrication during initial assembly of these terrains.

[30] The southern African xenolith data show no resolvable age variation with depth [*Carlson et al.*, 1999b] indicating that at least this portion of continental mantle did not grow downward by accretion as it does for oceanic lithosphere. For example, at Kimberley in the central Kaapvaal craton of South Africa (Figure 11) the crustal basement has zircon core ages of 3.2 Ga with overgrowth ages of 2.9 Ga [*Drennen et al.*, 1990], whole rock peridotite xenoliths give an average Re-Os T_{MA} age of 2.87 ± 0.14 Ga [*Carlson et al.*, 1999b], and sulfide inclusions in Kimberley diamonds give a Re-Os isochron age of 2.89 ± 0.06 Ga [*Richardson et al.*, 2001]. Within the resolution of these dating systems the whole lithospheric mantle section beneath the central Kaapvaal craton appears to have formed, or at least amalgamated, in the late Archean and has remained attached to the overlying crust since that time.

[31] Not all cratons, however, show such a limited range of ages throughout the lithospheric mantle. The Slave craton in northern Canada appears to be compositionally and temporally layered [*Griffin et al.*, 1999; *Kopylova and Russell*, 2000] with an early Archean [*Aulbach et al.*, 2004] highly depleted layer overlying more fertile mantle of significantly younger age [*Irvine et al.*, 2003]. Xenoliths from the northern margin of the Wyoming craton in North America show a step from Early Proterozoic to late Archean T_{RD} ages above 140 km depth to Phanerozoic ages at depths greater than 140 km [*Carlson et al.*, 1999a]. This age step is coincident with a change from depleted peridotites lying on a conductive geotherm to more fertile compositions offset to higher temperature from the conductive geotherm [*Hearn and McGee*, 1984; *Hearn*, 1993]. Thus, in the case of the Wyoming craton, the low-T to high-T transition recorded in the xenoliths may indeed reflect the base of the lithosphere. The Sloan kimberlite that erupted through mid-Proterozoic crust south of the Wyoming craton carried xenoliths that follow a cratonic conductive geotherm to depths of at least 200 km [*Eggler et al.*, 1987]. This could suggest that lithospheric thickness in this area was much greater when the 380 Ma Sloan kimberlite erupted than at 50 Ma when the Wyoming craton kimberlites erupted. This, in turn, suggests the possibility that the lithosphere in this area was thinned by the Laramide event, the major tectonomagmatic event affecting this area of the western United States in the Cretaceous and early Tertiary [*Eggler et al.*, 1988].

[32] Perhaps the best case for removal of old lithospheric mantle comes from the eastern portion of the Sino-Korean craton. The eastern Sino-Korean craton in China currently is characterized by high surface heat flow, abundant Cenozoic volcanism, and low seismic wave velocities in the upper mantle [*Griffin et al.*, 1998]. The mantle beneath this area was sampled at about 460 Ma by magmas carrying xenoliths with the highly depleted compositions, cold conductive geotherm, and ancient Re-Os ages typical of thick Archean lithospheric mantle [*Gao et al.*, 2002a]. This mantle stratigraphy contrasts strongly with the more fertile and much hotter temperatures recorded by mantle xenoliths in Tertiary lavas [*Menzies et al.*, 1993; *Griffin et al.*, 1998]. One Tertiary basalt field that erupted through the central area of the craton that was involved in the circa 1900 Ma Trans-North China Orogen provides xenoliths that define a Re-Os isochron age of 1910 ± 220 Ma [*Gao et al.*, 2002a]. Another Tertiary basalt locality near the eastern margin of the craton has mantle xenoliths with Os isotopic compositions within the range observed for modern oceanic peridotites [*Gao et al.*, 2002a]. These results suggest that the old lithospheric mantle that was present when sampled at 460 Ma has been completely lost from the eastern margin of the Sino-Korean craton. Similar evidence has been presented for the loss of a significant fraction of the lower crust and lithospheric mantle beneath the Sierra Nevada mountain range in California [*Ducea and Saleeby*, 1996; *Lee et al.*, 2000; *Manley et al.*, 2000] and the southern Andes [*Kay et al.*, 1994]. These results indicate that lithospheres are not necessarily forever but that that they can be lost, or severely modified, during major tectonomagmatic events affecting the continents.

6. VIEW FROM THE LITHOSPHERIC MANTLE

6.1. Where and How Does Continental Lithosphere Form?

[33] The very high degree of melt depletion observed in cratonic lithospheric mantle has been interpreted as evidence for formation from plumes of hot material rising from the deep mantle [*Herzberg*, 1999] and has been connected with plume models for the generation of Archean continental crust [*Albarede*, 1998]. In hot plumes, melting would begin at high pressure and, if not impeded by a thick lithosphere, would continue to shallow depths by which point the cumulative extent of melting would be high. Deep melting also has been proposed as an explanation for the relatively high silica contents of cratonic peridotites [*Herzberg*, 1993], but this proposition has been weakened by the failure to produce such high-silica residues in high-pressure melting experiments [*Herzberg*, 1999; *Walter*, 1999], by the recognition that not all cratonic peridotites have high silica concentrations [*Bernstein et al.*, 1998; *Kopylova and Russell*, 2000], and by increasing evidence

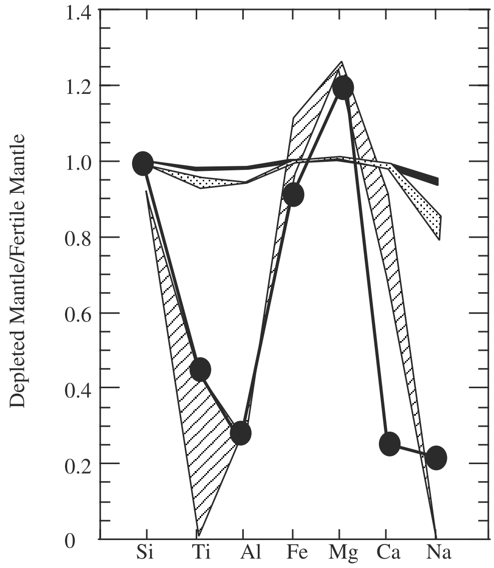

Figure 12. Fertile mantle normalized major element abundances of mantle depleted by the extraction of the current volume of continental crust using extreme estimates for the average composition of the continental crust [*Weaver and Tarney*, 1984; *Taylor and McLennan*, 1985]. The amount of chemical fractionation in the mantle residue depends on the amount of mantle affected by continent extraction as shown by the solid (100% by mass of the mantle affected), dotted (30% by mass of the mantle affected), and hatched (2% by mass of the mantle affected) areas. Dots show the composition of average continental garnet peridotite from *McDonough* [1990].

that the high silica contents of the Kaapvaal and Siberian peridotite xenoliths may be the result of melt infiltration and interaction with low-silica peridotites [*Kelemen et al.*, 1998]. Furthermore, most geochemical characteristics of lithospheric mantle peridotites are most easily reconciled with a relatively low-pressure melting origin [*Canil*, 2004], albeit in the case of cratonic peridotites one taken to very high degrees of melting.

[34] Another problem with the plume model for producing the type of residue seen in cratonic mantle is that a single step of high-degree melting of the mantle produces a high-magnesium magma known as komatiite [*Arndt and Nisbit*, 1982]. Komatiite is much more abundant in Archean than post-Archean crustal sections, but its abundance even in the Archean crustal record is grossly insufficient to balance the amount of highly depleted peridotite found in cratonic mantle. For example, assuming that the cratonic mantle is, on average, a residue of 40% partial melt extraction from fertile mantle, a 150 km thick depleted lithospheric mantle would require on the order of 100 km thickness of komatiite for mass balance. Some have argued that Archean lithospheric mantle forms by the stacking of shallow subducting plates composed of thick komatiitic crustal sections overlying highly depleted mantle [*Helmstaedt and Schulze*, 1989; *deWit et al.*, 1992]. However, the relative amount of eclogite, the high-pressure equivalent of basalt or komatiite, in the cratonic mantle as a whole appears to be about 1% or less, although locally the volume of eclogite may approach as much as 15% [*Schulze*, 1989]. As discussed in section 2, seismic images of the mantle lithosphere of the Kaapvaal craton show no high-velocity bodies that would be consistent with stranded sections of thick oceanic crust [*James et al.*, 2001, 2004; *Gao et al.*, 2002b], but seismic [*Bostock*, 1997; *Snyder et al.*, 2003] and magnetotelluric [*Jones et al.*, 2003] results for the Slave craton do show at least one strongly dipping reflector coincident with a zone of highly depleted mantle [*Griffin et al.*, 1999; *Kopylova and Russell*, 2000] that has been interpreted to be a stranded portion of subducted Archean oceanic lithosphere. While evidence for remnant slabs has only rarely been observed in cratonic mantle, the extent to which improved resolution in deep seismic imaging will reveal (or fail to reveal) more such features remains to be seen.

[35] If one were to ask how much mantle must have been involved in the creation of the existing volume of continental crust, the amount of chemical fractionation of the residual mantle is determined by the mass ratio of continent to mantle affected by continent removal. To achieve the high degree of chemical fractionation observed in the major element composition of average cratonic peridotite, only a very small amount of mantle, on the order of 2% of the mass of the whole mantle, could have contributed major elements to the continental crust (Figure 12). Two percent of the whole mantle is a mass of mantle that if confined beneath continents corresponds to a layer on average 160 km thick. This may be coincidental, but it also could imply that the entire mantle that provided the melts that formed the continental crust now resides in the continental lithospheric mantle. This small volume of mantle, however, cannot have supplied the bulk of the incompatible trace elements in the continental crust. The continental crust contains rubidium or barium, for example, in quantities equivalent to all the rubidium and barium in fertile mantle equivalent to about 30–50% of the mass of the total mantle [*Jacobsen and Wasserburg*, 1979; *Allègre et al.*, 1983]. Mass balance calculations based on major or trace element abundances thus provide estimates of the amount of mantle depleted by continental crust formation that differ by over an order of magnitude.

[36] There is one tectonic setting, however, that effectively decouples the sources of the major and trace elements in the magmas that eventually reach the surface. The major element signatures of convergent margin magmas are largely a function of high-degree, water-aided melting of the mantle wedge above the subducting plate [*Grove et al.*, 2002], whereas most of the incompatible trace elements come from the subducting plate in the form of fluids/melts [*McCulloch and Gamble*, 1991; *Schmidt and Poli*, 1998]. If continents and their deep roots form in convergent margin settings, this melting mechanism provides the means for continents to "scavenge" incompatible trace elements from a much larger volume of mantle than is required to provide the major elements that make up the continental crust. Though it is not clear that convergent margin volcanism alone can produce a crust equal in composition to the

continental crust [*Rudnick*, 1995], a convergent margin setting for continental lithosphere creation provides an attractive explanation for several features of the continental mantle, including the high degrees of melting at relatively low pressure [*Helmstaedt and Schulze*, 1989; *Walter*, 2003; *Canil*, 2004], the silica-enrichment observed in some cratonic peridotites possibly caused by interaction with Si-rich fluids from the subducting slab [*Rudnick et al.*, 1994; *Kelemen et al.*, 1998], the lack of suitable volumes and compositions of komatiite to mass balance the composition of continental mantle, and the evidence from old inclusions in diamonds and garnets in some xenoliths indicating that incompatible element metasomatism accompanied or shortly followed melt depletion [*Richardson et al.*, 1984]. *Pearson et al.* [2003] point out that of modern mantle samples, only those sampled at island or continental arcs approach the degree of depletion of cratonic peridotites. The changing average degree of melt depletion between Archean and younger mantle lithosphere would then primarily reflect a secular decline in mantle temperature; as even in water-fluxed melting, the maximum extent of melting is determined by the temperature in that region of mantle being melted.

[37] Another intriguing piece of evidence suggestive of concurrent melt depletion and metasomatic enrichment consistent with a melting mantle wedge being infiltrated by fluids/melts percolating up from the subducting plate is found in Re-Os data for sulfides in the mantle. The very high Os concentrations of mantle sulfides allow measurement of the Os isotopic composition of single grains thus minimizing the effects of contamination caused by interaction between xenolith and its host magma [*Pearson et al.*, 1998a, 1999; *Richardson et al.*, 2001; *Alard et al.*, 2002; *Griffin et al.*, 2002, 2004; *N. J. Pearson et al.*, 2002; *Shirey et al.*, 2002; *Westerlund et al.*, 2003; *Aulbach et al.*, 2004]. Unlike the Re-Os data for whole rock peridotites that shows little correlation between Re/Os ratio and $^{187}Os/^{188}Os$ (Figure 9), the data for sulfide inclusions in diamond and sulfides in peridotite xenoliths from cratonic mantle show a large range in Re/Os ratio that roughly correlates with $^{187}Os/^{188}Os$ and extends to very radiogenic Os (Figure 9). The sulfide data scatter widely and almost certainly include more than one generation of sulfide, but the slope of the correlation (Figure 9) is consistent with these sulfide grains being formed in the Archean. Their trend to high Re/Os ratio suggests that at least the high Re/Os ratio samples formed from melts. The wide scatter in the sulfide data may reflect variation in the initial Os isotopic composition of these sulfides as a result of interaction between high–$^{187}Os/^{188}Os$ melts from a subducting plate and a depleted mantle wedge with low $^{187}Os/^{188}Os$ [*Ruiz et al.*, 1999; *Westerlund et al.*, 2003] or simply the varying Os isotopic compositions of subducting basalt/sediment mixtures.

6.2. Role of Lithospheric Mantle Buoyancy and Strength in Continent Survival

[38] *Jordan* [1975, 1978, 1988] proposed the "tectosphere" model for continental lithosphere in which horizontal temperature gradients between the continental lithospheric mantle and surrounding oceanic mantle are balanced by chemical gradients so that constant density is maintained at any given depth in the mantle. Neutrally buoyant mantle is at risk of becoming involved in the circulation of the underlying convecting mantle through erosion and/or lateral transport, which explains why simple geodynamic models have difficulty in preserving thick sections of lithospheric mantle for the 3 Gyr time periods observed under some cratons [*Moresi and Lenardic*, 1997; *Lenardic and Moresi*, 1999; *Shapiro et al.*, 1999b; *Lenardic et al.*, 2000]. More recent analyses suggest that at least cratonic mantle lithosphere is intrinsically less dense compared to oceanic mantle to depths of at least 200 km [*Poudjom Djomani et al.*, 2001; *Kelly et al.*, 2003; *James et al.*, 2004]. *Kelly et al.* [2003] suggest that the inferred buoyancy of cratonic peridotite could be balanced by the presence of 7% eclogite to reach neutral buoyancy for the lithosphere. While this amount of eclogite is within the range allowed by concentrate analysis in kimberlites, it is significantly higher than the <1% eclogite abundance inferred for most cratonic mantle regions [*Schulze*, 1989]. *Poudjom Djomani et al.* [2001] suggest that cratons are positively buoyant within the limits of geoid analysis [*Richards and Hager*, 1988; *Forte et al.*, 1995; *Shapiro et al.*, 1999a], and this buoyancy explains both the thickness and longevity of cratonic lithospheric mantle. *Poudjom Djomani et al.* [2001] note that the average degree of depletion, and hence compositional buoyancy, declines for younger lithospheric mantle. They suggest that post-Archean lithospheric mantle therefore becomes negatively buoyant if thicker than 140 km with the result that the deeper lithosphere could become convectively unstable and sink into the general mantle circulation.

[39] Melt depletion not only creates compositional buoyancy in residual mantle but also leaves the mantle lithosphere depleted in radioactive heat-producing elements and in water. The resulting combination of low temperature and low water content of cratonic mantle lithosphere can lead it to have a viscosity as much as 3–4 orders of magnitude higher than warm, wet, asthenospheric mantle [*Dixon et al.*, 2004]. The compositional buoyancy of the lithospheric mantle causes it to resist subduction and/or delamination, while the strength imparted by cold temperatures and lack of water provides the necessary strength to allow thick sections of mantle to remain stable and attached to the overlying crust of similar age.

[40] This relationship between degree of melt depletion, compositional buoyancy, strength, and lithosphere thickness could explain the observation from xenolith geotherms of a general correspondence between lithosphere thickness and age or, more precisely, time since the last major tectonomagmatic event in the overlying crustal section [*Boyd and Gurney*, 1986; *Finnerty and Boyd*, 1987]. The continental lithospheric mantle acts in concert with the overlying crust to resist subduction. It furthermore provides a strong buffer against erosion from below and adds strength to the continent that can explain the continued existence of cratons that

have remained tectonically quiescent over most of Earth history. Should the compositional buoyancy of the continental lithospheric mantle be compromised, from either an insufficient degree of melt removal to compensate for cooling or by pervasive refertilization of the lithosphere through interaction with passing melts, a sudden loss of a thick lithospheric mantle layer through density instability also could explain the sudden onset of magmatism in some continental settings [*Kay et al.*, 1994; *Ducea and Saleeby*, 1996; *Lee et al.*, 2000; *Manley et al.*, 2000].

6.3. What Defines the Bottom of the Continental Lithosphere?

[41] In the ocean basins the base of the lithosphere is defined by a sharp reduction in shear wave velocity. The base of the continental lithosphere, however, is characterized by no such simple pattern. In cratonic regions the transition is evidenced by at most a subtle decrease from the high velocities characteristic of cratonic lithosphere to velocities that with depth gradually become indistinguishable from those of the convecting mantle. In continental areas that exhibit low seismic velocities in the uppermost mantle, such as the Basin and Range, it is not uncommon for there to be no evidence at all for a high-velocity lithospheric "lid" above the very low velocity material that dominates the shallowest mantle. In such circumstances the presence of low-velocity mantle material immediately beneath the continental crust would seemingly suggest that the lithosphere is composed of only crustal material and that if there ever existed a thick mantle lithosphere, it has since been lost. Recent detailed seismic studies in the Rocky Mountains, however, suggest the presence of dipping structures throughout the upper 250 km of the mantle that appear to connect with major terrain boundaries in the overlying crust [*CD-ROM Working Group*, 2002]. Since these crustal terrain boundaries are Proterozoic to Archean in age, if the mantle structures indeed represent a continuation of these boundaries into the upper mantle, then the data imply that a thick lithospheric mantle root still exists beneath the Rocky Mountains, even though the average seismic velocity is very low. The presence of old mantle beneath this area, at least in the shallow mantle (<100 km depth), is supported by Early Proterozoic to late Archean Re-Os model ages for mantle xenoliths from at least one tectonically active area of the Basin and Range [*Lee et al.*, 2001]. These results indicate that if the temperature of a section of lithospheric mantle can be raised sufficiently, it can lose its characteristically high seismic velocities though it may well retain its compositional and rheological distinction from convecting mantle.

[42] *Rudnick et al.* [1998] and *Rudnick and Nyblade* [1999] suggest that the intersection of conductive geotherms with the nominal adiabatic geotherm of the convecting mantle is a reasonable definition of the base of the lithosphere. Unfortunately, conductive geotherms in the lithosphere cannot be calculated with sufficient accuracy just from surface heat flow because the geotherm for a given surface heat flow is a strong function of the distribution of heat-producing elements and thermal conductivity in both the crust and lithospheric mantle [*Rudnick et al.*, 1998]. In addition, the uncertainty in estimating potential temperature of the mantle means that even the adiabat cannot be determined reliably. Given the range of possible values for heat production in both the crust and lithospheric mantle, various geotherm models produce intersections between conductive and adiabatic geotherms that vary in depth from 150 to >400 km [*Rudnick et al.*, 1998]. The "best" (i.e., most widely accepted) estimates for crustal and mantle heat production yield geotherms that suggest lithosphere thicknesses on the order of 200–250 km [*Rudnick and Nyblade*, 1999], considerably thinner than the 300–400 km thicknesses suggested by seismic results in some areas [*Grand*, 1987; *Jordan*, 1988; *James et al.*, 2001].

[43] Another line of evidence comes directly from xenoliths. The deepest low-T xenoliths are found at the Finsch kimberlite in South Africa with equilibration pressures suggesting a depth of origin slightly in excess of 200 km [*Finnerty and Boyd*, 1987]. Other xenoliths have textures that have been interpreted as indicating depths of origin in excess of 300 km [*Haggerty and Sautter*, 1990], although these xenoliths give equilibration pressures indicating reequilibration at depths of <200 km. Evidence for even deeper origins for some lithospheric components comes from the observation of transition zone and lower mantle mineral phases included in diamonds [*Scott-Smith et al.*, 1984; *Harris*, 1992; *Harte et al.*, 1999; *McCammon*, 2001]. It seems likely that these phases were not formed in the lithosphere, however, but were emplaced into the lithosphere by mantle convection after having formed much deeper in the mantle.

[44] The boundary between low-T and high-T xenoliths has been suggested to mark the lithosphere-asthenosphere boundary [*Boyd*, 1987]. Several lines of evidence contradict this interpretation of the high-T peridotite data. First, calculated seismic velocities for high-T peridotites from southern Africa are significantly lower than velocities actually observed at corresponding depths in the mantle [*James et al.*, 2004]. Second, the strongly sheared fabrics [*Harte*, 1977] and the common evidence for mineral zonation [*Smith and Boyd*, 1987; *Shimizu*, 1999] in the high-T peridotites indicate that these features were imposed on the xenoliths only hours to years prior to their transport to the surface [*Ehrenberg*, 1979; *Smith and Boyd*, 1987], implying that the distinguishing features of the high-T samples are not long-term characteristics of the mantle. Third, although more fertile than the low-T peridotites, the great majority of high-T peridotites are not as fertile as estimates of undepleted mantle. Moreover, there is substantial evidence that late stage metasomatism is the cause of increased fertility in many of these samples [e.g., *Smith and Boyd*, 1987]. Finally, osmium isotope compositions of high-T xenoliths are similar to the low-T xenoliths, and both give model ages consistent with them being samples of old continental lithosphere [*Walker et al.*, 1989; *Pearson et al.*, 1995a; *Carlson et al.*, 1999b]. All of these features are consistent with the high-T peridotites forming within the

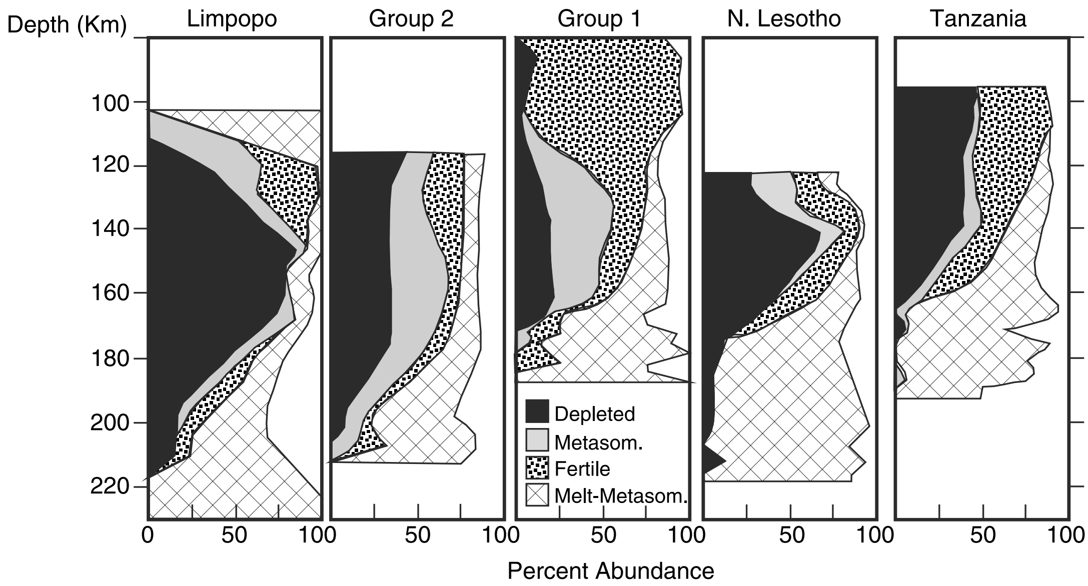

Figure 13. Relative abundance of different chemical types in the African lithospheric mantle as deduced from the analysis of garnet concentrates from the kimberlite groups shown by the labels at the top of each plot. The plots are arranged in age progression of the kimberlite localities from (left) oldest to (right) youngest. Figure modified from that presented by *Griffin et al.* [2003], reprinted with permission from Elsevier.

lower cratonic lithosphere as sheared and metasomatized margins of whatever magmatic body generated their host magma. Some high-T xenoliths have equilibration pressures consistent with depths of origin approaching 250 km [e.g., *Finnerty and Boyd*, 1987]. The uncertainty over whether or not the deepest xenolith reflects the base of the lithosphere, a volatile exsolution horizon for kimberlitic melts where they begin to rise rapidly enough to entrain xenoliths, or simply the random nature of xenolith capture and transport suggests that the maximum depth of xenolith capture should be viewed only as a minimum thickness to the lithospheric mantle.

6.4. Destruction of Continental Lithosphere: Causes and Consequences

[45] Most geodynamic models have difficulty preserving continental lithosphere for even a fraction of the ages recorded by many Archean lithosphere sections [*Lenardic and Moresi*, 1999; *Shapiro et al.*, 1999b], so perhaps the question should instead be, Why does some old lithosphere survive? The small compositional buoyancy imparted by the melt depletion characteristic of lithospheric peridotites is marginal in terms of its ability to keep the lithosphere from being eroded by the convecting mantle flowing across its base. For example, even the large density differences present in oceanic lithosphere between crust and mantle following subduction apparently can be mixed away by mantle convection over billion year time periods [*Christensen*, 1989; *van Keken et al.*, 2002]. We have already noted that the secular decline in the average degree of melt depletion of lithospheric peridotites with age leads to lessened compositional buoyancy, and survivability [*Poudjom Djomani et al.*, 2001], in younger lithospheric mantle. For the degree of depletion observed in cratonic peridotites, the compositional buoyancy coupled with very high viscosity clearly is sufficient to keep this material near the surface, just beneath the continent, for 3 Gyr time periods. More than this, the general first-order correspondence between the age of a crustal section and its underlying mantle (Figure 10) implies that many sections of crust and mantle lithosphere have remained coupled and translated together for their whole history of continental drift, which requires not only buoyancy but high viscosity so that the depleted mantle does not flow horizontally across terrain boundaries to spread out widely beneath the crust. That the boundary between lithospheric mantles of different age can be maintained for long time periods is demonstrated particularly well by data for mantle xenoliths from northern Lesotho [*Carlson et al.*, 1999b; *Irvine et al.*, 2001], erupted on the eastern boundary of the Kaapvaal craton, compared to xenoliths from the Griqualand East kimberlite field that erupted through Proterozoic crust some 70 km east of the northern Lesotho kimberlites [*D. G. Pearson et al.*, 2002]. Mantle xenoliths from Griqualand East give a mean Re-depletion model age of 1.48 Ga with no value >2.22 Ga. In contrast, the mean Re-depletion age for the northern Lesotho samples is 2.42 Ga, and 41 out of 57 samples have T_{RD} greater than the oldest age observed in the Griqualand East section.

[46] Using mineral concentrate (minerals from disaggregated peridotites) data from kimberlites, *Griffin et al.* [2003] suggest that the depleted lithospheric section beneath southern Africa, at a regional scale, shows evidence of increasing refertilization over just the last 500 Myr (Figure 13). For example, at 180 km depth, melt-metasomatized samples account for ∼10% of the mantle sampled by 500 Ma kimberlites but almost 80% of the mantle sampled by the 85 Ma kimberlites of northern Lesotho [*Griffin et al.*, 2003].

Certainly, some of these differences may reflect regional variations in the thickness of the depleted lithospheric mantle, but the data presented by *Griffin et al.* [2003] suggest that the lithospheric mantle of Africa has been significantly infiltrated by melts. A caveat that must be applied to such studies is the possibility that the mineral concentrate sample may not be representative of the entire lithosphere and may be significantly influenced by metasomatic events, some leading to mineral growth, associated with protokimberlite activity. This possibility is reinforced by the fact that the Re-Os data for Kaapvaal peridotites do not show a general decrease in age with depth that would be expected if the trends seen in the concentrate data [*Griffin et al.*, 2003] indeed reflect major chemical modification of the deep lithosphere.

[47] Whether or not melt infiltration into continental mantle is a general phenomenon is critical to the long-term survival of mantle lithosphere because it impacts the whole question of strength and compositional buoyancy. Evidence from eastern China [*Gao et al.*, 2002a] and the Wyoming craton [*Carlson et al.*, 2004] suggests that lithosphere removal can be accomplished only during major tectonomagmatic episodes and not by gradual melt metasomatism of the type documented by *Griffin et al.* [2003]. Nevertheless, if the continental lithospheric mantle can be regionally refertilized by infiltrating melts, when it eventually cools, it may no longer be sufficiently buoyant and viscous to survive as lithosphere. Gentle erosion of the base of the lithospheric mantle may have no expression in the overlying crustal record, but rapid delamination of large sections of lithospheric mantle could result in dramatic uplift and/or magmatism in the overlying crust. This type of lithosphere loss event should be preceded by crustal subsidence as the now dense lithospheric mantle drags down on its overlying crustal section, followed by rapid uplift and volcanism when the lithospheric mantle separates from the overlying crust. This sequence is opposite that expected for the arrival of a deep mantle plume beneath a continent, which should first create a long period of uplift prior to volcanism [*Richards et al.*, 1989]. The fact that subsidence preceded several large-volume igneous events, including the Siberian flood basalts [*Czamanske et al.*, 1998], the Bushveld intrusion of southern Africa [*Jordan*, 2003], the southern Andes [*Kay and Kay*, 1993; *Kay et al.*, 1994], and the southern Sierra Nevada of California [*Saleeby and Foster*, 2004], suggests that lithosphere delamination may be an important mechanism for initiating large-volume continental magmatism.

7. CONCLUSIONS

[48] For centuries most of what we knew about the solid earth came from studies of the continents. With the explosion in ocean studies following World War 2, the dynamic nature of the solid earth that is clearly required to explain the history of ocean basins replaced the concept of a more ‘‘solid’’ Earth derived from the long-term stability of continental cratons. The continental mantle may play the key role in explaining many of the differences between the evolution of continental and oceanic plates. The compositional buoyancy and high viscosity of cratonic mantle is the likely explanation for why cratons seem so able to resist the dynamic motions of surrounding plates and deeper mantle flow. The secular decline in the degree of melt depletion seen in younger sections of continental mantle suggests a reason why continental deformation is mostly accommodated in these younger terrains. It is possible that loss of buoyancy in parts of the deep lithospheric mantle and rapid material transfer from the base of the continental lithospheric mantle into the convecting mantle can explain the sudden onset of episodes of intracontinental tectonism and/or magmatism. This type of process involving gravitational instability is an attractive alternative to the often ad hoc ‘‘plume arrival’’ explanation for such events. With the new, and continually improving, ability to date chemical modification events in mantle rocks it is now possible to explore the third dimension of continent evolution, at least in areas where mantle samples are available. The many recent advances in the ability to investigate the physical and chemical properties of the upper mantle bear much promise for finally being able to understand what role the 70–80% of the continental plate that lies below the Moho plays in the formation and evolution of the continents.

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R. W. Carlson and D. E. James, Department of Terrestrial Magnetism, Carnegie Institution of Washington, 5241 Broad Branch Road, NW, Washington, DC 20015, USA. (carlson@dtm.ciw.edu; james@dtm.ciw.edu)

D. G. Pearson, Department of Geological Sciences, University of Durham, South Road, Durham DH13LE, UK. (d.g.pearson@durham.ac.uk)