Chemical and temporal variations in the Earth's lithosphere

C.J. Hawkesworth, D.G. Pearson and S.P. Turner

Email alerting service	Receive free email alerts when new articles cite this article - sign up in the box at the top right-hand corner of the article or click **here**

THE ROYAL
SOCIETY

Chemical and temporal variations in the Earth's lithosphere

By C. J. Hawkesworth[1], D. G. Pearson[2] and S. P. Turner[1]

[1]Department of Earth Sciences, The Open University,
Milton Keynes MK7 6AA, UK
[2]Department of Geological Sciences, University of Durham,
Durham DH1 3LE, UK

The lithosphere is the rigid outer layer of the Earth, but its composition, and by implication its origins, are different in terrains of different ages. The continental lithosphere is typically weaker, although often much thicker, than the oceanic lithosphere. Oceanic mantle lithosphere accretes new material at its base as it cools (thermal accretion), and so it is compositionally similar to the underlying asthenosphere. Much of the continental mantle lithosphere, and particularly that beneath Archaean cratons, is relatively depleted in major elements. Thus, continental mantle lithosphere is less dense than the oceanic lithosphere, and its thickness depends on the compositional difference between it and the underlying asthenosphere, as well as on the geothermal gradient. The errors and accuracy of the mineral thermobarometry of peridotite samples are briefly reviewed, and both mineral thermobarometry and heat-flow studies consistently suggest minimum thicknesses of 150–200 km for the lithosphere beneath Archaean cratons. Os isotopes reflect the age of major-element depletion, and, hence, the stabilization of the mantle lithosphere. They confirm that in many cases the continental crust and uppermost mantle stabilized at about the same time, and have remained as a coherent unit ever since. In general, areas of Archaean lithosphere appear not to have thickened significantly since they were formed, whereas the lithosphere beneath Proterozoic and younger areas has been more prone to subsequent magmatic and tectonic events.

Keywords: lithosphere definitions; mineral thermobarometry;
Os isotopes; mantle compositions; subcratonic mantle

1. The lithosphere and its rheology

The lithosphere is the rigid outer layer of the Earth and it can be divided into a number of plates. It consists of crust and mantle portions that are separated by the Moho. In the simplest model (but see below), the principal, and most easily observable, distinction between the oceanic and continental lithosphere is the composition of the crust, which is basaltic and (roughly) tonalitic in the oceanic and continental lithospheres, respectively.

In detail, there has been much discussion on how the lithosphere is best defined, not least because significant changes in different physical properties take place at different depths, and so definitions of the lithosphere vary depending on the particular properties under investigation. A conventional approach considers the lithosphere in terms of the upper *thermal boundary layer*, within which heat is transferred primarily by conduction, to a deeper convecting system in the asthenosphere (see figure 1).

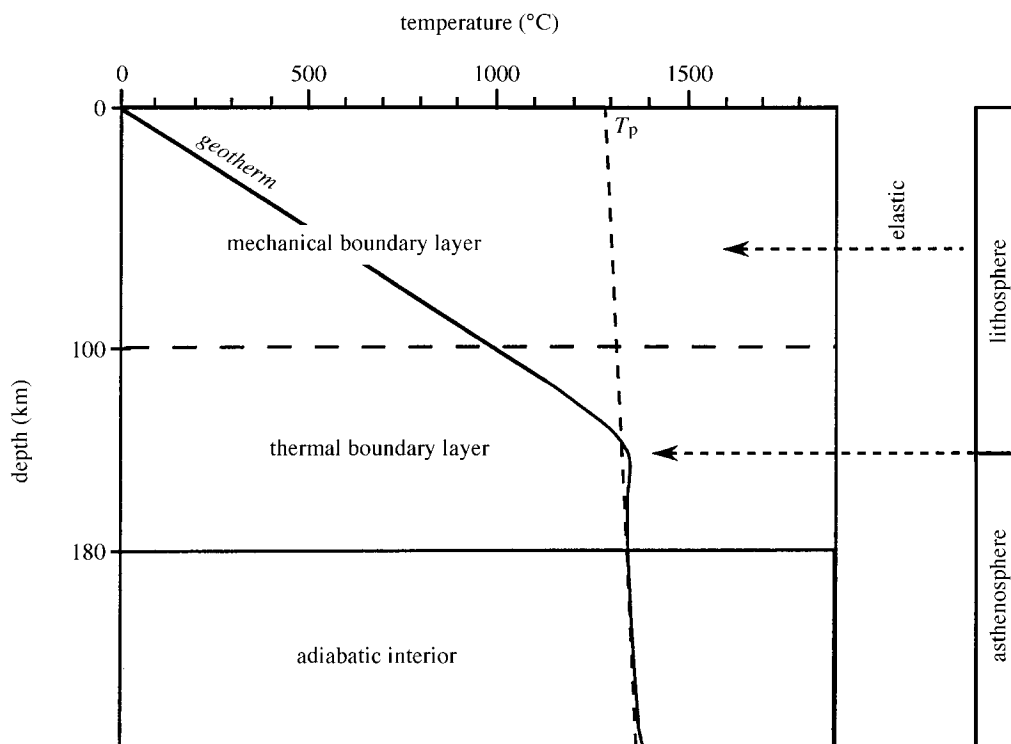© 1999 The Royal Society
T$_{\text{E}}$X Paper

Figure 1. Definitions of the mechanical and thermal boundary layers, and lithosphere (after White 1988). The solid curve is the horizontally averaged equilibrium thermal structure of old oceanic lithosphere over normal temperature asthenosphere (potential temperature $T_p = 1280\ ^\circ C$).

However, the upper part of the thermal boundary layer may also be regarded as a *mechanical boundary layer*, as a consequence of the strong temperature dependence of rock strength or viscosity. Thus, on geological time-scales, the mechanical boundary layer is effectively isolated from the relatively rapid turnover of the convecting asthenosphere below. In this context, the accepted definition of the lithosphere is in terms of a mechanical boundary layer, which takes no part in convection, underlain by the lower part of the thermal boundary layer which does convect, albeit more sluggishly than the underlying asthenosphere (see, for example, Parsons & McKenzie 1978; McKenzie & Bickle 1988).

The *thermal lithosphere* is defined somewhat differently and in the oceans it is constrained by heat flow and bathymetry. The thermal lithosphere tends to be slightly thicker than the mechanical boundary layer definition, but its base is still within the thermal boundary layer (White 1988). Another commonly cited 'definition' of lithospheric thickness is in terms of its effective elastic thickness (EET), determined by modelling the flexural response to loading, e.g. volcanic islands, mountain belts, sedimentary basins (Turcotte 1979). However, neither the mechanical nor thermal boundary-layer lithospheres need coincide with the *elastic lithosphere*, because they are measures of a different physical response. The EET is the thickness of the layer that supports elastic stresses on geological time-scales, and it depends on the magni-

tude of applied stress, strain rates and also temperature (Davies 1988). The mantle below the elastic lithosphere does not support elastic stresses, but it is still sufficiently viscous to move as a coherent part of the plate, and, therefore, it is part of the mechanical and thermal lithospheres.

The essential point is that the definition of the term lithosphere varies depending on the purpose of a particular investigation. The debate is particularly acute in continental areas, but, in the context of recognizable geochemical signatures, the continental lithosphere is most usefully defined in terms of a mechanical boundary layer isolated from the homogenizing tendencies of convection. The continental lithosphere so defined can, therefore, accumulate and maintain distinct geochemical and isotopic anomalies for much longer than the underlying asthenosphere. An alternative view is put forward by Anderson (1995) who argues that there would be much less confusion in the literature if the terms lithosphere and asthenosphere were restricted to rheological matters. In this definition, mantle hotter than $ca.\,650\,°C$ does not qualify as lithosphere, and Anderson (1995) prefers the term perisphere to describe the shallow trace-element-enriched mantle source of certain intraplate magmas. The perisphere is envisaged to be present in both oceanic and continental areas, but, since the age of such material can be shown to be older beneath the continents, it does appear to be attached to the overlying rheologically defined lithosphere over geological time-scales.

The temperature distribution in the lithosphere controls both the density and the strength. Thus, the depth of the oceans increases with the age of the lithosphere because, as the oceanic lithosphere cools and thickens, it becomes more dense and, to maintain isostatic equilibrium, it sinks. A concurrent increase in the strength of the oceanic lithosphere with age is inferred from analysis of its response to volcanic island loads, where the vertical deflection associated with a volcano emplaced on old oceanic lithosphere is demonstrably less than that for an equivalent age volcano on young lithosphere (Turcotte 1979). An apparent anomaly between the behaviour of continental and oceanic lithospheres is that it appears that most continental lithosphere is weaker than the oceanic lithosphere, even though it is at least as thick and often much thicker (Turcotte & McAdoo 1979; Sclater *et al.* 1980; Boyd & Gurney 1986; Jordan 1975; Richter 1988). One reason for this is a difference in bulk rheological properties, and, in particular, the presence of a quartz-dominated (weak) crust in the continents and olivine-dominated (strong) crust in the oceans.

The rheology of the lithosphere and its response to extensional deformation is most strongly controlled by its temperature and (to a lesser extent) its mineralogy. For example, materials can only maintain stresses over geological time if τ, the ratio of the temperature (in K) of a material to its melting point, is less than about 0.4 (Ashby & Verrall 1977; Weertman 1978). This corresponds to $ca.\,100\,°C$ for wet granite and 280 and $390\,°C$ for hydrous and anhydrous peridotite, respectively, at 2 GPa (70 km depth), illustrating that it is only the upper parts of the crust and lithospheric mantle that can maintain stresses over long periods of time. The mineralogy of the lithosphere is a function of bulk composition, although most models usually use the simplifying assumption that the rheology will be dictated by the mineral abundant enough ($ca.\,60\%$) to form a matrix of inter-touching grain boundaries. Thus, most numerical models of the response of the continental lithosphere to extension (see, for example, Houseman & England 1986; Sonder & England 1989) assume a highly simplified rheology controlled by failure of quartz $\pm$ plagioclase in the crust and olivine in the mantle (figure 2).
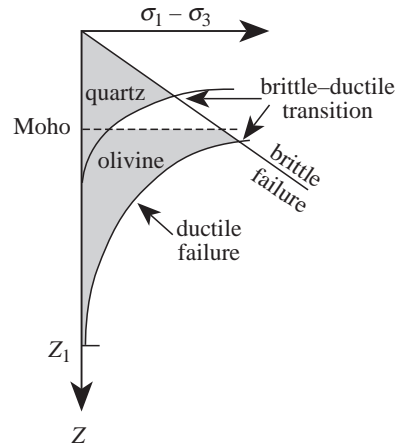
Figure 2. A plot of the stress at which failure will occur (σ_1–σ_3) plotted against depth (z) for a simple model of the continental lithosphere. The strength is given by the grey area under the curves and it is a function of the vertical compositional structure of the lithosphere, and the geotherm.

The two- or three-phase rheological parametrization of the lithosphere is clearly an extreme oversimplification. Moreover, there is abundant evidence that natural deformation mechanisms are complex, and that pre-existing fabrics, grain size changes, metamorphic reactions and heterogeneous strain partitioning are all important (see, for example, Knipe 1989; Handy 1994). However, while these mechanisms clearly control deformation style at the outcrop scale, it is impressive that, on the scale of the lithosphere, the simple rheological models have been successful in predicting first-order observations, such as the behaviour of extensional sedimentary basins (see, for example, McKenzie 1978; White & McKenzie 1988; Newman & White 1997). Here, we review information on chemical and temporal variations in the Earth's lithosphere as a step towards appraising the need to modify these assumptions in future models.

2. Oceanic lithosphere

In the oceanic basins, the bathymetry and inferred lithospheric thickness are a simple function of age (Parsons & Sclater 1977). The basaltic oceanic crust (*ca.* 7 km thick) is produced by decompression melting of the asthenosphere at mid-ocean ridges, following which the subjacent lithospheric mantle accretes new material at its base as it cools (thermal accretion, see McKenzie & Bickle (1988)). Again, in the simplest model, this lithospheric mantle is simply asthenospheric peridotite that has become too cool to be able to be separated from the overlying plate and convect. After *ca.* 70–80 Ma, when the lithosphere reaches a thickness of *ca.* 100 km, the heat flow into the bottom is balanced by that flowing out through the top and, subsequently, small-scale convective instabilities at the base of the thermal boundary layer maintain a constant lithospheric thickness and bathymetry (Parsons & McKenzie 1978).

The description of the oceanic lithosphere just outlined conforms to the plate model and, indeed, it is the oceanic lithosphere that is best described by plate tectonics: the oceans are composed of a small number of large rigid plates with deformation being concentrated along narrow zones between the plates. Although there is increasing

evidence that the oceanic lithospheric mantle often has a composition that differs from that expected by simple thermal accretion of asthenosphere (e.g. enrichment by small partial melt fractions; Class & Goldstein (1997)), for the most part, its composition is well predicted by simple geodynamic models and relatively invariant with age. The corollary is that deformation is localized along the oceanic plate margins. The lack of internal deformation of the oceanic lithosphere reflects its strength. The mineral that controls rheological behaviour of the oceanic lithosphere is olivine, and this will not change even with quite big shifts in bulk composition (excluding for the moment the effects of dissolved volatiles within olivine).

3. Continental lithosphere

The continental lithosphere is demonstrably different from oceanic lithosphere. First, production of a crust that is broadly tonalitic in composition clearly requires processes in addition to, or instead of, single-stage decompression melting of the asthenosphere (see, for example, O'Nions & McKenzie 1988). Second, there may be a very crude relation between the age of the lithosphere and its thickness in continental areas that include old Proterozoic and Archaean cratons, but in this case, it is at least in part due to differences in composition. In the oceans, no lithosphere older than Cretaceous is preserved, principally because oceanic lithosphere becomes negatively buoyant (i.e. susceptible to subduction) after *ca.* 80 Ma. In contrast, the longevity, stability and composition of portions of the continents are interlinked: survival of continental lithosphere is facilitated by its buoyancy and/or the resistance of continental lithospheric mantle to convective removal (Jordan 1988; Richter 1988). The implication is that there is a link between age (stability), composition and rheology. Buoyancy is a function of density, which is dependent on composition and, to a lesser extent, temperature.

In the remaining sections we concentrate our discussion on the composition of the mantle portion of the continental lithosphere for several reasons. Firstly, the composition of the continental crust (broadly tonalitic) has been assessed in detail in a number of recent reviews (Taylor & McLennan 1995; Rudnick & Fountain 1995; Plank & Langmuir 1998). Secondly, the strength of the continental lithosphere is likely to be largely dictated by the mantle part of the lithosphere rather than the crust. The continental crust generally has a low density, and its rheology is usually approximated by quartz and/or plagioclase, although this is clearly an oversimplification. Nevertheless, even if the lower crust is quite mafic in composition (e.g. Rudnick & Fountain 1995), any compositional increases in strength will be mediated by thermal weakening due to increasing temperatures at depth.

(*a*) *Thermal accretion or compositional stabilization*

In the simplest model, the continental mantle lithosphere grows downwards by thermal accretion in much the same way as that inferred to occur beneath the oceanic crust, and it has the same composition, and thus rheological behaviour, as the oceanic mantle lithosphere (i.e. it is indistinguishable from the asthenosphere except by its temperature). However, there is now abundant evidence from xenoliths and studies of the composition of continental magmatism that large portions of the continental lithospheric mantle have a different and often highly heterogeneous composition

reflecting a protracted and complex multistage history (see, for example, Maaloe & Aoki 1977; Boyd & Mertzman 1987; Boyd 1989; Hawkesworth *et al*. 1990; Herzberg 1993). One consequence is that the thickness of the lithosphere depends on the difference in composition between it and the underlying asthenosphere, as well as on the geothermal gradient (i.e. it is compositionally stabilized).

Xenoliths brought to the surface by continental lavas frequently contain evidence for a multistage history involving depletion in a basaltic melt component followed by enrichment in incompatible trace elements at some later time (see, for example, reviews in Menzies & Hawkesworth 1987; Menzies 1990). It is the melt depletion event that is most critical here, because it leaves a residue that is relatively depleted in Fe and, therefore, has a lower density. The increased buoyancy of the mantle increases its likelihood of survival by making it more resistant to convection, even at asthenospheric temperatures. Moreover, if clinopyroxene has been exhausted during the basalt extraction melting event, such lithospheric mantle will also have a raised solidus and increased resistance to removal by further partial melting unless it is subsequently enriched by incompatible trace elements and volatiles.

The significance of the geochemical data would be greatly enhanced if there were clear criteria for distinguishing xenoliths scavenged from within the mantle lithosphere from those originating in the underlying asthenosphere. However, this is a difficult issue which has been the subject of much debate (Nixon & Boyd 1973; Harte 1983; Nixon & Davies 1987; Harte & Hawkesworth 1989; Haggerty & Sautter 1990). The criteria may be textural, in that most coarse annealed samples are thought to represent fragments of the mantle lithosphere, and they may be based on the calculated pressures and temperatures and the extent to which the latter reflect likely conditions within or below the continental lithosphere.

Many kimberlites yield two largely distinct populations of peridotite. One has mineral compositions that yield temperature estimates typically below 1100 °C, whereas the other population yields temperature estimates in the range 1200–1500 °C (Boyd & Nixon 1975; Harte 1983; Finnerty & Boyd 1987; Carswell & Gibb 1987). The majority of lower-temperature (cold) peridotites have coarse textures and relatively Mg-rich (Fe-poor) compositions, whereas the higher-temperature (hot) peridotites are usually deformed, and they often have more fertile, particularly more Fe- and Ti-rich, compositions than the coarse cold peridotites. Xenoliths with intermediate characteristics are observed at a few localities that have more Fe-rich, or deformed, cold peridotites-pyroxenites (Harte 1983; Harte & Hawkesworth 1989), and others that have low-Ca garnet harzburgites apparently restricted to a limited depth range (Boyd *et al*. 1993).

The two groups of low- and high-temperature peridotites also tend to define arrays with significantly different slopes on *P–T* diagrams. The slope for the low-temperature suite approximates to that expected for a cool conductive shield geotherm consistent with a surface heat flow of 41 mW m^{-2}, whereas the slope for the high-temperature suite corresponds to that of a higher heat flow geotherm greater than, or equal to, those of oceanic areas (Boyd 1973; Finnerty & Boyd 1984, 1987). Thus it was tempting to suggest that the junction between the two suites marked the junction between cool rigid lithospheric mantle, and the hot convecting asthenosphere below. However, it now appears that (1) the so-called high-*T* rocks have suffered significant metasomatism and (2) they preserve old Os isotope signatures, indicating that they too should be considered part of the mantle lithosphere (Pearson 1999).

(*b*) *Pressures, temperatures and lithosphere thickness*

A key question in assessing the thickness of the continental lithosphere is the accuracy of the mineral thermobarometry on peridotite xenoliths. This is difficult to evaluate due to a scarcity of calibrations against natural systems and uncertainties in the extrapolation of results between different compositional systems (Smith 1999). One test has been to compare the results of thermobarometry on natural graphite- and diamond-bearing peridotites against their position in *P–T* space relative to the experimentally determined graphite–diamond stability curve (Pearson *et al.* 1994*a*). This showed general agreement between the natural and experimentally determined graphite–diamond stability fields, to within 0.7–0.8 GPa in most cases, for a number of thermometer–barometer combinations. The currently preferred means of calculating equilibration conditions for garnet peridotites is by using the pyroxene–solvus thermometer and the Al-solubility in orthopyroxene barometer formulations of Brey *et al.* (1990). This combination gives similar results for several independent thermometers and barometers in the range between 2.5 and 5 GPa and between 900 and 1400 °C (Brey & Kohler 1990; Smith & Boyd 1992; Smith 1999). Large errors are likely at low temperatures due to steepening of the pyroxene solvus limb, and Smith (1999) noted scatter in the estimates from natural xenolith suites at around 1400 °C that could be due to inaccuracies at high temperatures.

Despite the apparent general success of mineral thermobarometry in garnet lherzolite systems, a number of likely sources of error remain that may result in worst-case confidence limits of estimated temperatures and pressures being no better than ± 100 °C, and possibly as much as 1 GPa, respectively. These sources of error include insufficient understanding and accountance of the effects of minor elements such as Na and Cr in natural versus synthetic systems, and possible large effects associated with the presence of Fe^{3+} (Canil *et al.* 1994; Canil & O'Neill 1996). Perhaps more realistic errors, for the currently favoured Brey *et al.* (1990) thermobarometer for rocks having equilibrated in the temperature and pressure range of the experiments used to construct the thermobarometers, are ± 60 °C and 0.45 GPa (2 s.d.) (Brey 1990).

Testing thermobarometry by observed concordance with the diamond–graphite stability field can be looked at from the more general perspective of the success in explaining the presence or absence of diamonds in cratonic and circum-cratonic areas. Peridotite xenolith suites erupted by kimberlites within craton boundaries give equilibration conditions ranging well into the experimentally determined diamond stability field, consistent with the presence of diamonds in many of these kimberlites. In contrast, xenolith suites erupted by kimberlites in off-craton areas, e.g. Namibia and East Griqualand, give equilibration conditions entirely within the graphite stability field, consistent with the absence of diamonds in these kimberlites (Boyd & Gurney 1986). Similar conclusions have been reached using trace-element thermobarometry and successfully applied to diamond exploration strategies (Griffin & Ryan 1995). In addition, thermobarometry on silicate inclusions within diamonds almost invariably gives pressures within the diamond stability field and often of 60 kbar or more (Harris 1992). Hence, at this level, we seem to be able to place reasonable limits on the depth of the mechanically stable lithospheric mantle via mineral thermobarometry.

Recent studies on mantle xenoliths give maximum depths at different cratonic localities of 150–220 km (see, for example, Boyd *et al.* 1998), and these can be compared with the constraints available from the low surface heat flows that charac-
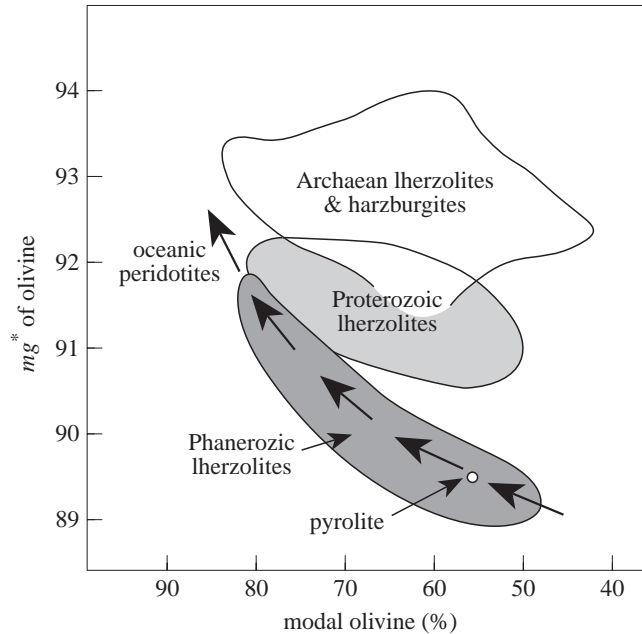
Figure 3. Variation in the mg^* of olivines with the modal olivine content for mantle-derived
xenoliths from terrains of different ages (after Boyd *et al.* 1993; Griffin *et al.* 1998*a*).

terize Archaean areas. Rudnick *et al.* (1998) showed that for a surface heat flow of
$41 \, \mathrm{mW \, m^{-2}}$ for Archaean cratons, the mantle root must have U, Th and K contents
that are significantly less than the average values calculated from xenolith analyses.
Using an average K_2O content of $0.03 \, \mathrm{wt\%}$ for the mantle lithosphere, they noted
that if the xenolith P–T points were assumed to reflect the present-day thermal
regime beneath the Archaean cratons, the lithosphere thickness would not exceed
ca. 150–200 km. However, that requires an unexpectedly low crustal heat production,
and it may be that the xenolith P–T data represent geotherms perturbed by even
the kimberlite magmatism. In that case, more U, Th and K can be accommodated
within the crust, and the lithosphere extends to depths of over 200 km (Rudnick *et
al.* 1998). In summary, mineral thermobarometry and heat-flow studies consistently
suggest minimum thicknesses of 150–200 km for the lithosphere beneath Archaean
cratons.

(*c*) *Major elements*

Depleted (Fe-poor) mantle xenoliths dominate those erupted through Archaean
crust, suggesting that the preservation of such ancient crustal remnants is linked to
the intrinsic buoyancy and survivability of the underlying lithospheric mantle. This is
the compositionally stabilized chemical boundary layer, long championed by Jordan
(1988). Boyd (1989) used a diagram of the *mg* number of olivine against its modal
abundance to highlight the differences in composition between oceanic and cratonic
peridotites (see also table 1). The oceanic peridotites plot close to the calculated array
of residual mantle compositions formed by progressive melt extraction from fertile
peridotite (figure 3). In contrast, the low-T coarse peridotites from Archaean areas—

Table 1. *Average compositions of mantle from different domains compared with the composition of pyrolite (Ringwood 1979) and primitive mantle (Hart & Zindler 1986)*

((*a*) Spinel lherzolites from Herzberg (1993), (*b*) abyssal peridotite from Dick & Fisher (1984), (*e*) Kaapvaal (Boyd 1989), and the (*f*) Archaean, (*g*) Proterozoic and (*h*) Phanerozoic Gnt SCLMs (subcontinental lithospheric mantles) are the bulk compositions estimated from garnet compositions from Griffin *et al*. (1998*b*). Estimated densities in, for example, the Archaean and Phanerozoic Gnt SCLMs are 3.31 and 3.37 $g\,cm^{-3}$, respectively (Griffin *et al*. 1998*b*). (*c*) Pyrolite; (*d*) primitive mantle.)

	(*a*)	(*b*)	(*c*)	(*d*)	(*e*)	(*f*)	(*g*)	(*h*)
SiO_2	44.3	43.6	45.6	46.0	46.6	45.7	44.7	44.5
TiO_2	0.09	0.02	0.2	0.18	0.04	0.04	0.09	0.14
Al_2O_3	2.36	1.18	3.3	4.06	1.46	0.99	2.1	3.5
Cr_2O_3	0.43	0.22	0.4	0.47	0.4	0.28	0.42	0.40
FeO	8.31	8.21	8.1	7.54	6.2	6.4	7.9	8.0
MnO	0.13	0.14	0.15	0.13	0.12	0.11	0.13	0.13
MgO	41.6	45.2	38.5	37.8	44.1	45.5	42.4	39.8
CaO	2.2	1.13	3.1	3.21	0.79	0.59	1.9	3.1
Na_2O	0.23	0.02	0.4	0.33	0.09	0.07	0.15	0.24
K_2O	0.04	—	0.03	0.03	—	—	—	—
NiO	0.27	0.22	0.2	0.28	0.27	0.3	0.29	0.26
mg	89.9	90.8	89.4	89.9	92.6	92.7	90.5	89.9

which are primarily from the Kaapvaal in southern Africa (Boyd 1989), but also from Siberia (Boyd *et al*. 1997)—are displaced to high *mg* numbers at relatively low modal olivine abundances. The high *mg* numbers in the olivines reflect high whole-rock *mg* numbers, and, hence, the degree of major-element depletion. The relatively low modal olivine values for the Archaean samples are primarily due to their relatively high enstatite, and, hence, elevated silica contents. In contrast, Pearson *et al*. (1994*b*) have shown that the mantle lithosphere underlying Proterozoic areas has higher modal olivine (higher Mg/Si) and lower olivine *mg* numbers, reflecting a secular change in bulk lithosphere composition between on- and off-craton environments. Recently, Griffin *et al*. (1998*a*) have shown that similar secular changes in peridotite compositions can also be documented from analyses of Cr-pyrope garnets included in volcanic rocks that have been scavenged from the peridotite wall rocks en route to the surface. They argue that while the major change in composition is between Archaean and post-Archaean peridotites, significant differences can also be recognized between Proterozoic and Phanerozoic peridotites (figure 3; table 1).

The origin of the distinctive low-FeO, high-silica, low-Ca/Al Archaean peridotites remains conjectural. Since they appear to be confined to the Archaean, they presumably reflect conditions that have not prevailed since, and popular interpretations have included residues after komatiite melt extraction (see, for example, Boyd 1989; Takahashi 1990). However, Herzberg (1993) emphasized that these peridotites were too silica-rich to be simple residues of komatiite melt extraction, and this has been confirmed by recent experiments that demonstrate that opx-rich residues cannot be due to komatiite melt generation from anhydrous garnet peridotite (Walter 1997). Such residues might occur if melting was both shallow and took place in the presence of water (Gaetani & Grove 1998), but other studies have interpreted these

Archaean xenoliths either as cumulates, possibly modified by later metamorphic segregation processes (Boyd 1989), or as the products of progressive melt–rock inter-action (Kelemen *et al.* 1992). Of significance here is the fact that the Archaean sam-ples have densities that are on average *ca.* 2% less than, for example, recent oceanic peridotites (Jordan 1988; Hawkesworth *et al.* 1990; Griffin *et al.* 1998*a*). Moreover, Boyd *et al.* (1998) presented new data on spinel peridotites from the Kaapvaal Cra-ton and showed that, overall, there is an increase in calculated densities of *ca.* 3–4% as depth increases from *ca.* 50 to 200 km. They cautioned that the highest densities are recorded in poorly understood metasomatized rocks, and so they may not be representative of the mantle depth range from which they were derived.

Overall, it is inferred from the study of xenoliths, and from the observation that primary economic concentrations of diamonds are very largely confined to Archaean areas (see, for example, Boyd & Gurney 1986), that there are deep lithospheric roots down to at least 200 km beneath Archaean cratons (cf. 125 km thick lithosphere beneath oceanic areas). Whole-rock and mineral analyses confirm that the low-*T* xenoliths, in particular, are compositionally different from peridotites from younger areas, and these differences are presumably, at least in part, responsible for the presence of such deep roots. This is consistent with suggestions that lithospheric mantle at least 200 km thick must be compositionally stabilized in order to avoid thermal removal (Davies 1979), and with geoid anomalies and numerical convection models that both support the existence of deep depleted lithospheric mantle roots beneath cratons (Doin *et al.* 1996, 1997). Beneath post-Archaean areas, the mantle is less depleted, in that the olivines have lower *mg* numbers (figure 3), and the bulk rock samples have higher Ca/Al and Mg/Si ratios (see, for example, Herzberg 1993; Griffin *et al.* 1998*a*, *b*).

(*d*) *Radiogenic isotopes*

It has proved difficult to determine the age of mantle lithosphere directly, in part because of the complex history of most peridotite samples. More fundamentally, how-ever, the formation of mantle lithosphere is typically linked to melt extraction pro-cesses, and to the stabilization of major-element-depleted refractory mantle; yet most widely used radioactive decay schemes involve incompatible trace elements (Rb–Sr, Sm–Nd, U–Th–Pb). Thus, Sr, Nd and Pb isotope ratios tend to provide information on the age of minor and trace-element enrichment events, and not about major-element depletion episodes. Old ages have been recorded from such decay schemes, most notably from Pb and Sm–Nd isotopes in sulphide and silicate inclusions in diamond (Kramers 1979; Richardson *et al.* 1984), but, presumably, even they reflect an enrichment event in mantle lithosphere that had already stabilized.

In another geologically useful decay scheme, ^{187}Re decays to ^{187}Os with a half life of 4.23×10^{10} years. Re is moderately incompatible, and Os is compatible in the upper mantle. Consequently, residual mantle has lower Re/Os ratios, and, hence, with time it has lower ^{187}Os/^{188}Os than Os, even in the source of MORB (Walker *et al.* 1989; Pearson *et al.* 1995*a*; Snow & Reisberg 1995). Also, the Os isotope ratio of depleted peridotite is much less readily modified in response to subsequent melt and/or fluid-related enrichment processes than, for example, the ^{87}Sr/^{86}Sr or ^{143}Nd/^{144}Nd ratios, simply because residual peridotite has much higher Os content than most metaso-matic fluids and melts. Thus, Os isotopes are increasingly used to investigate the
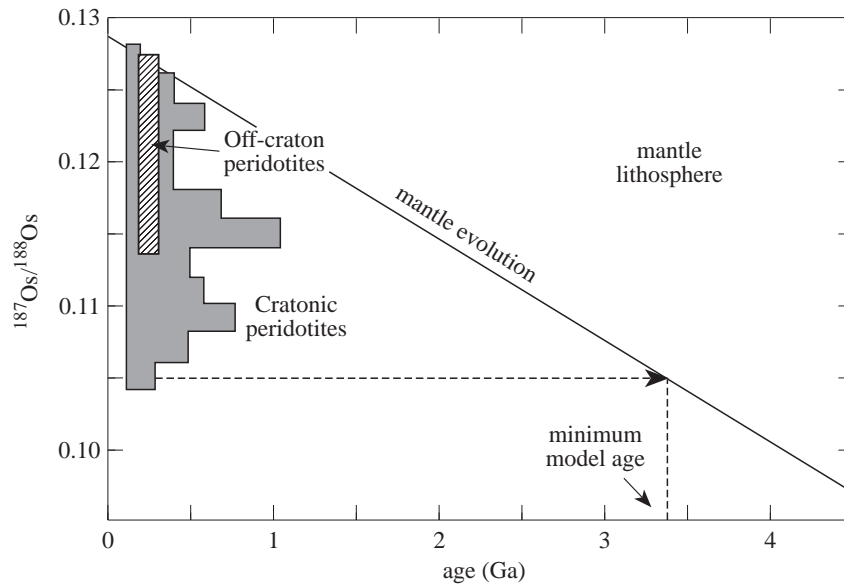
Figure 4. Os isotope evolution diagram illustrating the range of present-day Os isotope ratios of cratonic mantle peridotites from Siberia and southern Africa (grey area). The bar with diagonal hachures for the off-craton peridotites illustrates the range obtained from Namibia and East Griqualand. Data from Walker *et al.* (1989), Pearson *et al.* (1994*b*, 1995*a*, *b*) and Pearson (1999; D. G Pearson, unpublished work). Also illustrated is a Re-depletion model age (T_{RD}), which is calculated by extrapolating the present-day Os isotope ratio back to the mantle evolution line, assuming that $Re/Os = 0$ in the peridotite sample.

depletion history of lithospheric mantle, and, in particular, lithospheric mantle that has lower Re/Os ratios than the source of MORB (see, for example, Walker *et al.* 1989; Carlson & Irving 1994; Pearson *et al.* 1995*a*, *b*; Reisberg & Lorand 1995; Nägler *et al.* 1997; Burnham *et al.* 1998). More recently, they have been used to document unexpectedly old melt depletion events in fore-arc peridotites (Parkinson *et al.* 1998).

In practice, it is often difficult to determine when the measured Re/Os ratios of mantle samples were last modified. Thus, rather than calculate Os model ages using the parent/daughter isotope ratio in the same way as for Nd model ages, widespread use has been made of Re-depletion model ages (T_{RD}), particularly in xenolith studies (Walker *et al.* 1989). The underlying assumptions are that at large degrees of melting most of the Re in peridotite is removed and that this effectively 'freezes' the Os isotope composition of the peridotite at that time. Re-depletion model ages are therefore calculated by extrapolating the present-day Os isotope ratios back to the mantle evolution line, assuming that $Re/Os = 0$ in the peridotite samples (figure 4). A key point is that these model ages provide only minimum estimates of the ages of the depletion events since, if the residual peridotites contained significant Re over a period of time, the calculated T_{RD} age will be less than the true depletion age.

The oldest T_{RD} ages calculated from the available Os isotope data on cratonic peridotite xenoliths from the Kaapvaal, Wyoming and Siberia Cratons are similar at *ca.* 3–3.5 Ga (Pearson *et al.* 1995*a*, *b*; Pearson 1999). These oldest T_{RD} ages are, therefore, taken to approximate the time of the depletion events that marked craton

stabilization. In detail, samples from the Premier kimberlite in southern Africa consistently give post-Archaean T_{RD} ages of *ca.* 2 Ga (Pearson *et al.* 1995*a*; Carlson *et al.* 1998), even though this kimberlite was also emplaced through Archaean cratonic basement. However, it is in the area of the large Bushveld intrusion that formed close to 2 Ga, and which may have been associated with the generation of a sizeable volume of new lithosphere. In the circum-cratonic xenolith suites, such as those from Namibia and East Griqualand, surrounding the Kaapvaal Craton, the oldest T_{RD} ages are *ca.* 2.1–2.3 Ga (Pearson *et al.* 1998), much younger than the oldest T_{RD} ages on the Craton (figure 4). Thus, the Os isotope studies confirm suggestions, made on the basis of Sr, Nd and Pb isotope variations, that the oldest mantle ages from any particular area are similar to the oldest ages in the overlying crust. The new insight available from Os isotopes is that they reflect the age of major-element depletion. Thus, they can be more confidently linked to the stabilization of the bulk lithosphere than to subsequent trace-element enrichment processes. It is concluded that the crust and uppermost mantle stabilized at about the same time, and in many cases have remained a coherent unit ever since (see, for example, Pearson *et al.* 1995*a*, *b*; McBride *et al.* 1996; Handler *et al.* 1997). In addition, Os isotope studies on more mafic samples (eclogites and pyroxenites) of the mantle lithosphere suggest that many samples from cratonic areas are also Archaean in age and may reflect subducted Archaean basaltic crust (Pearson *et al.* 1995*c*).

(*e*) *Volatiles*

As noted above, there are various lines of evidence for enrichment of depleted mantle lithosphere involving the addition of volatiles. Volatile-bearing olivine is significantly weaker than anhydrous olivine (Mackwell *et al.* 1990). However, xenolith studies suggest that the added volatiles are usually taken up in amphibole overgrowths on clinopyroxene, rather than being incorporated into olivine, and, thus, will lower the solidus of the peridotite by several hundred degrees Celsius relative to that of anhydrous peridotite (Olafsson & Eggler 1983). Although appropriate experimental data are limited, it seems likely that amphibole (Riecker & Rooney 1969) will be just as strong as clinopyroxene (Kirby & Kronenberg 1984), and so there is as yet little reason to think that volatile-enriched major-element-depleted peridotite should be any weaker than normal anhydrous fertile peridotite (Turner & Hawkesworth 1995). On the contrary, the increased thickness and resultingly cooler geotherm, and the opx-rich mineralogy, combine to make Archaean lithosphere significantly stronger than Proterozoic and younger lithosphere; a feature presumably linked to its longer survival. This notion is supported by the observation that few extensional basins or compressional orogens are developed within Archaean cratons. What is less clear is the extent to which these models can be applied to younger Proterozoic and Phanerozoic continental lithospheres. For example, the rapidly thickened lithosphere beneath Tibet was presumably not compositionally stabilized, and so it underwent convective thinning soon after collision of India with Asia (see, for example, Houseman *et al.* 1981; Molnar *et al.* 1993; Turner *et al.* 1996).

(*f*) *Lithosphere-derived magmatic rocks*

In principle, the composition of the lithosphere can be inferred from magmas derived either entirely, or in large part, from within the mantle lithosphere. Mantle

domains of different age and incompatible-element characteristics have been recognized by mapping out regional variations in volcanic rocks. For example, the so-called 0.706 line in the western USA demarks the boundary between the accreted Mesozoic terrains in the west and both Precambrian crust and mantle lithosphere to the east. Similarly, provinces of high- and low-Ti basalts have long been recognized in the Mesozoic flood basalts of southern Africa and South America, and attributed to different provinces in the underlying mantle lithosphere (Erlank *et al.* 1988; Piccirillo *et al.* 1988).

Mantle-derived magmas that have erupted through continental crust are susceptible to crustal contamination. Thus, it is rare to get good evidence of the age and nature of their mantle source regions. However, lamproites have high incompatible-element abundances, and, in some areas, they have been shown to have been derived from source regions that are probably depleted in major elements, and range in age from the Mid-Proterozoic to the Late Archaean (see, for example, Fraser *et al.* 1985; Nelson *et al.* 1986). Continental flood basalts (CFBs) are more prone to crustal contamination, but some suites whose compositions have been corrected for the effects of contamination still have major- and trace-element contents that differ significantly from those of oceanic basalts derived from the sub-lithospheric mantle (Turner & Hawkesworth 1995). Figure 5 summarizes the total Fe and Na_2O contents of selected CFB suites corrected back to 8% MgO, to minimize the effects of fractional crystallization and any associated crustal contamination. A number of the CFB fields extend to low $Fe_{8.0}^*$ at low $Na_{8.0}$ relative to the fields for MORB, and for the Deccan, which has $Fe_{8.0}^*$ and $Na_{8.0}$ values consistent with decompression melting of an upwelling mantle plume at a potential temperature of 1530 °C (Hawkesworth *et al.* 1999). It is argued that these low $Fe_{8.0}^*$ CFBs were derived from source regions that are not sampled by oceanic basalts, and they are inferred to be slightly melt-depleted portions of the mantle lithosphere (Turner & Hawkesworth 1995). Such CFBs, which appear to have been derived from within the mantle lithosphere, tend to have Proterozoic or younger source ages (Hawkesworth *et al.* 1993). There is no evidence that such tholeiitic magmas were derived from Archaean mantle lithosphere, and this is consistent with the more extreme major-element-depleted composition of such lithosphere inferred from the study of mantle xenoliths (figure 3; table 1). Nevertheless, at least some portions of post-Archaean mantle lithosphere under continents may be compositionally stabilized, and the Os isotope data suggest that, typically, the lithospheric mantle was stabilized at about the same time as the overlying crust (see, for example, Pearson *et al.* 1995*a*; McBride *et al.* 1996; Handler *et al.* 1997).

4. Discussion

Kimberlite pipes contain xenoliths of crustal as well as mantle lithologies. In southern Africa, mafic granulite xenoliths appear to be a feature of off-cratonic, and not cratonic, kimberlites. The latter observation may be linked with the characteristic bimodal granite–greenstone association of many Archaean areas, and with models in which Archaean granitic rocks reflect remelting of mafic rocks at depths great enough to stabilize residual garnet. In contrast, the post-Archaean lower crust may contain greater thicknesses of mafic material marking both a change in crust generation processes and a greater role for magmatic underplating. The available radiogenic isotope data indicate that such underplating took place at about the same time as
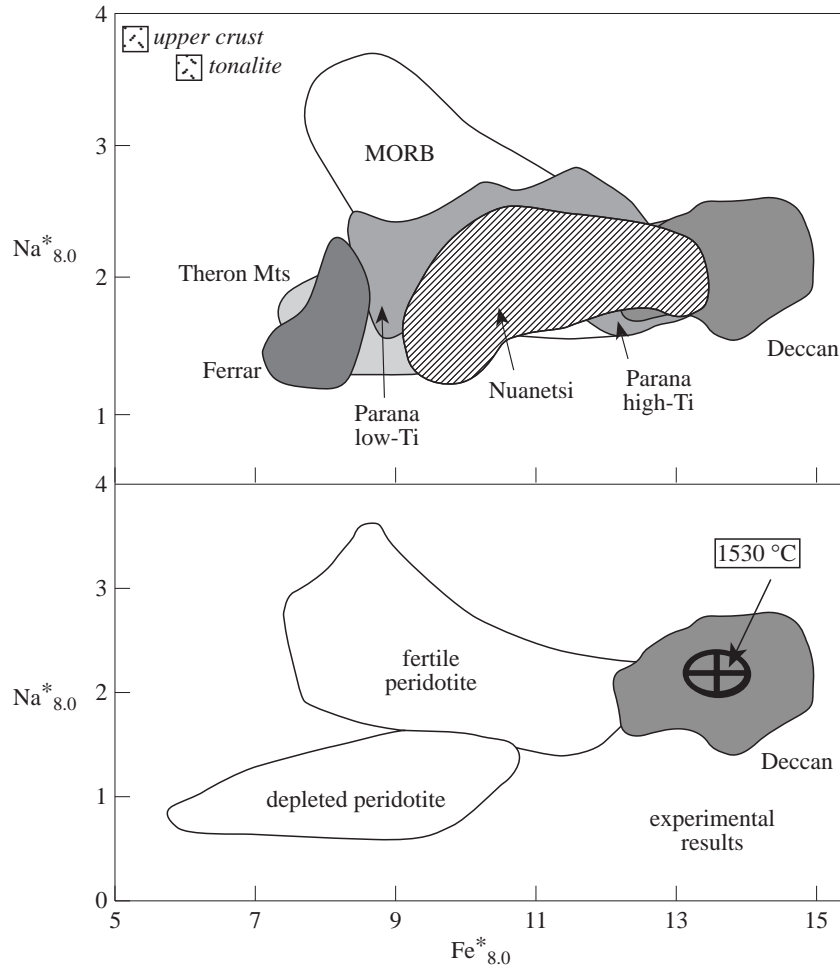
Figure 5. A plot of the fields for the Na and Fe contents of selected CFBs extrapolated to 8% MgO (after Klein & Langmuir 1987; Turner & Hawkesworth 1995; Hawkesworth *et al*. 1999). In sharp contrast to the MORB array, many of the continental flood basalts define overlapping sub-horizontal arrays that extend to low $Fe^*_{8.0}$ and $Na_{8.0}$. These are compared with experimental data on melts from fertile and depleted peridotite from Takahashi & Kushiro (1983) and Kushiro (1990). The model composition labelled 1530 °C is that for decompression melting of fertile anhydrous peridotite at that potential temperature following the parametrization of McKenzie & Bickle (1988). It is concluded that the uncontaminated Deccan CFB were generated by melting within a mantle plume, and that many of the other CFBs were derived from melt-depleted peridotite sources located within the continental mantle lithosphere. Details and data sources in Turner & Hawkesworth (1995).

the upper crust was stabilized, rather than in subsequent magmatic episodes (Rogers & Hawkesworth 1982; Huang *et al*. 1995). Durrheim & Mooney (1991) used seismic-velocity models to conclude that the thickness of the crust is generally thinner in Archaean (*ca*. 35 km) than in Proterozic areas (*ca*. 45 km), and that the latter has a markedly thicker high-velocity layer (greater than 7.0 km s^{-1}) at its base. However,

Rudnick & Fountain (1995) could find no such differences in either crustal thickness or velocity structure between Archaean and Proterozoic crust.

Mineral and whole-rock analyses have documented a clear secular variation in the composition of the uppermost mantle beneath continental areas (Maaloe & Aoki 1977; Boyd 1989, 1996; Griffin *et al.* 1998*a*). Mantle xenoliths from beneath crust generated before *ca.* 2.5 Ga consist of highly depleted lherzolites and harzburgites with low Mg/Si at high *mg*. Since such rocks are effectively restricted to cratonic areas, they reflect processes of Archaean age, and their Archaean age has been confirmed by both Nd and Os isotope studies (see Pearson (1998) for a review). In general, the age of the mantle lithosphere is similar to that of the overlying crust, and so stabilization was linked to thermal accretion and/or the processes of crust formation. Strikingly, these data require that lithospheric roots that have been analysed directly in samples derived from pressures corresponding to depths of at least 200 km have been linked to the overlying crust since the Archaean. Most models concur that such deep roots must have been compositionally stabilized to have survived (see, for example, Davies 1979; Jordan 1988; Richter 1988), and recent estimates suggest that the Archaean mantle lithosphere may be 2–3% less dense than the convecting upper mantle (Griffin *et al.* 1998*a, b*). Richter (1988) has further suggested that the Archaean lithosphere may have been stabilized to temperatures of almost 1400 °C, which would require that it had different rheological properties to the surrounding mantle, in addition to density.

Post-Archaean mantle lithosphere appears to be both thinner (not a source of significant amounts of diamond), and less depleted in major elements than Archaean lithosphere (figure 3). Its density is therefore closer to that of the surrounding mantle, and so it will be both thinner and will undergo convective removal more readily. Thus, not all of the continental mantle lithosphere may be compositionally stabilized. To begin with, at least some post-Archaean lithospheric mantle must be thermally accreted, as evidenced by the thermal sag stage of sedimentary basin formation following extension (see, for example, McKenzie 1978; White & McKenzie 1988). Furthermore, convective thinning of thickened lithospheric mantle beneath orogenic mountain belts (Houseman *et al.* 1981) would be unlikely to remove low-density depleted mantle peridotite. The high-K magmas produced during such convective thinning events provide clear evidence for the presence of mantle lithosphere beneath Mesozoic crust that has had a long multistage history of depletion and enrichment (Turner *et al.* 1996, 1999). Indeed, it is likely that many parts of the continental mantle lithosphere are vertically stratified with older, more depleted and buoyant material at shallower levels, giving way to more fertile peridotite at depth. The temporal change in magma chemistry in several areas lends support to this concept (see, for example, Garland *et al.* 1996; Turner *et al.* 1999).

References

Anderson, D. L. 1995 Lithosphere, asthenosphere, and perisphere. *Rev. Geophys.* **33**, 125–149.

Ashby, M. F. & Verrall, R. A. 1977 Micromechanisms of flow and fracture, and their relevance to the rheology of the upper mantle. *Phil. Trans. R. Soc. Lond.* A **288**, 59–95.

Boyd, F. R. 1973 A pyroxene geotherm. *Geochim. Cosmochim. Acta* **37**, 2533–2546.

Boyd, F. R. 1989 Compositional distinction between oceanic and cratonic lithosphere. *Earth Planet. Sci. Lett.* **96**, 15–26.

Boyd, F. R. 1996 Origins of peridotite xenoliths: major element considerations. In *High pressure and high temperature research on lithosphere and mantle materials: Proc. Int. Sch. Earth Planet. Sci, Sienna*, pp. 89–106. University of Sienna.

Boyd, F. R. & Gurney, J. J. 1986 Diamonds and the African lithosphere. *Science* **232**, 472–477.

Boyd, F. R. & Mertzman, S. A. 1987 Composition and structure of the Kaapvaal lithosphere, Southern Africa. In *Magmatic processes: physiochemical principles* (ed. B. O. Mysen), pp. 3–12. The Geochemical Society, Univeristy Park, PA.

Boyd, F. R. & Nixon P. H. 1975 Origins of ultramafic nodules from some kimberlites of northern Lesotho and the Monastery Mine, South Africa. *Phys. Chem. Earth* **9**, 431–451.

Boyd, F. R., Pearson, D. G., Nixon, P. H. & Mertzman, S. A. 1993 Low-calcium garnet harzburgites from Southern Africa; their relations to craton structure and diamond crystallization. *Contrib. Mineral. Petrol.* **113**, 352–366.

Boyd, F. R., Pokhilenko, N. P., Pearson, D. G., Mertzman, S. A., Sobolev, N. B. & Finger, L. W. 1997 Composition of the Siberian cratonic mantle: evidence from Udachnaya peridotite xenoliths. *Contrib. Mineral. Petrol.* **128**, 228–246.

Boyd, F. R., Pearson, D. G. & Mertzman, S. A. 1998 Spinel-facies peridotites from the Kaapvaal root. In *Abst. 7th Int. Kimb. Conf., Cape Town.*

Brey, G. P. 1990 Geothermobarometry for lherzolites: experiments from 10 to 60 kb, new thermobasometers and application to natural rocks. Habil. thesis, University of Aarmstadt, Germany.

Brey, G. P. & Kohler, T. 1990 Geothermobarometry in four-phase lherzolites. II. New thermobarometers, and practical assessment of existing thermobarometers. *J. Petrol.* **31**, 1353–1378.

Brey, G. P., Kohler, T. & Nickel, K. G. 1990 Geothermobarometry in four-phase lherzolites. I. Experimental results from 10 to 60 kb. *J. Petrol.* **31**, 1313–1352.

Burnham, O. M., Rogers, N. W., Pearson, D. G., van Calsteren, P. W. & Hawkesworth, C. J. 1998 The petrogenesis of the Eastern Pyrenean peridotites: an integrated study of their whole-rock geochemistry and Re–Os isotope composition. *Geochim. Cosmochim. Acta.* **62**, 2293–3210.

Canil, D. & O'Neill, H. S. C. 1996 Distribution of ferric iron in some upper mantle assemblages. *J. Petrol.* **37**, 609–635.

Canil, D., O'Neill, H. S. C., Pearson, D. G., Rudnick, R. L., McDonough, W. F. & Carswell, D. A. 1994 Ferric iron in peridotites and mantle oxidation states. *Earth Planet. Sci. Lett.* **123**, 205–220.

Carlson, R. W. & Irving, A. J. 1994 Depletion and enrichment history of subcontinental lithospheric mantle: Os, Sr, Nd and Pb evidence for xenoliths from the Wyoming Craton. *Earth Planet. Sci. Lett.* **126**, 457–472.

Carlson, R. W., Pearson, D. G., Boyd, F. R., Shirey, S. B., Irvine, G., Menzies, A. H. & Gurney, J. J. 1998 Regional age variation in the southern African mantle: significance for models of lithospheric mantle formation. In *Ext. Abs. 7th Int. Kimb. Conf., Cape Town.*

Carswell, D. A. & Gibb, I. G. F. 1987 Evaluation of mineral thermometers and barometers applicable to garnet litherzolite assemblages. *Contrib. Mineral. Petrol.* **94**, 403–416.

Class, C. & Goldstein, S. L. 1997 Plume-lithosphere interactions in the ocean basins: constraints from source mineralogy. *Earth Planet. Sci. Lett.* **150**, 245–260.

Davies, G. F. 1979 Thickness and thermal history of continental crust and root zones. *Earth Planet. Sci. Lett.* **44**, 231–238.

Davies, G. F. 1988 Ocean bathymetry and mantle convection. 1. Large scale flow and hotspots. *J. Geophys Res.* **93**, 10 467–10 480.

Dick, H. J. B. & Fisher, R. L. 1984 Mineralogic studies of the residues of mantle melting: abyssal and alpine peridotites. In *Kimberlites II: the mantle and crust–mantle relationships* (ed. J. Kornprobst), pp. 295–308. New York: Elsevier.

Doin, M. P., Fleitout, L. & McKenzie, D. 1996 Geoid anomalies and the structure of continental and oceanic lithospheres. *J. Geophys. Res.* **101**, 16 119–16 135.

Doin, M. P., Fleitout, L. & Christensen, U. 1997 Mantle convection and stability of depleted and undepleted continental lithosphere. *J. Geophys. Res.* **102**, 2771–2787.

Durrheim, R. J. & Mooney, W. D. 1991 Archean and Proterozoic crustal evolution; evidence from crustal seismology. *Geology* **19**, 606–609.

Erlank, A. J., Duncan, A. R., Marsh, J. S., Sweeney, R. J., Hawkesworth, C. J., Milner, S. C., Miller, R. M. & Rogers, N. W. 1988 A laterally extensive geochemical discontinuity in the sub-continental Gondwana lithosphere. In *Geochemical evolution of the continental crust*, pp. 1–10. Brazil: Poços de Caldas.

Finnerty, A. A. & Boyd, F. R. 1984 Evaluation of thermobarometers for garnet peridotites. *Geochim. Cosmochim. Acta* **48**, 15–27.

Finnerty, A. A. & Boyd, F. R. 1987 Thermobarometry for garnet peridotite xenoliths: a basis for upper mantle stratigraphy. In *Mantle xenoliths* (ed. P. H. Nixon), pp. 381–402. New York: Wiley.

Fraser, K. J., Hawkesworth, C. J., Erlank, A. J., Mitchell, R. H. & Scott Smith, B. H. 1985 Sr-, Nd- and Pb-isotope and minor element geochemistry of lamproites and kimberlites. *Earth Planet. Sci. Lett.* **76**, 57–70.

Gaetani, G. A. & Grove, T. L. 1998 The influence of water on melting of mantle peridotite. *Contrib. Mineral. Petrol.* **131**, 323–346.

Garland, F., Turner, S. & Hawkesworth, C. 1996 Shifts in the source of the Paraná basalts through time. *Lithos* **37**, 223–243.

Griffin, W. L. & Ryan, C. G. 1995 Trace elements in indicator minerals: area selection and target evaluation in diamond exploration. *J. Geochem. Explor.* **53**, 311–337.

Griffin, W. L., O'Reilly, S. Y., Ryan, C. G., Gaul, O. & Ionov, D. A. 1998*a* Secular variation in the composition of subcontinental lithospheric mantle: geophysical and geodynamic implications. *AGU Spec. Monogr.*, composition **26**, 1–26.

Griffin, W. L., O'Reilly, S. Y. & Ryan, C. G. 1998*b* The composition of sub-continental lithospheric mantle: garnet-based estimates. *Ext. Abs. Goldschimdt Conf. Toulouse 1998, Min. Mag.* **62A**, 541–542.

Haggerty, S. E. & Sautter, V. 1990 Ultradeep (greater than 300 kilometers), ultramafic upper mantle xenoliths. *Science* **248**, 993–996.

Handler, M. R., Bennett, V. C. & Esat, T. M. 1997 The persistence of off-cratonic lithospheric mantle: Os isotopic systematics of variably metasomatised southeast Australian xenoliths. *Earth Planet. Sci. Lett.* **151**, 61–75.

Handy, M. R. 1994 Flow laws for rocks containing two non-linear viscous phases: a phenomonological approach. *J. Struct. Geol.* **16**, 287–301.

Harris, J. W. 1992 Diamond geology. In *The properties of natural and synthetic diamond* (ed. J. E. Field), pp. 345–393. Academic.

Hart, S. R. & Zindler, A. 1986 In search of a bulk-earth composition. *Chem. Geol.* **57**, 247–267.

Harte, B. 1983 Mantle peridotites and processes—the kimberlite sample. In *Continental basalts and mantle xenoliths* (ed. C. J. Hawkesworth & M. Norry), pp. 46–91. Nantwich, UK: Shiva.

Harte, B. & Hawkesworth, C. J. 1989 Mantle domains and mantle xenoliths. In *Kimberlites and related rocks, vol. 2—their mantle/crust setting, diamonds and diamond exploration*, pp. 649–686. *Spec. Publ. Geol. Soc. Aust.*, no. 14. Blackwell.

Hawkesworth, C. J., Kempton, P. D., Rogers, N. W., Ellam, R. M. & van Calsteren, P. W. 1990 Continental mantle lithosphere, and shallow level enrichment processes in the Earth's mantle. *Earth Planet. Sci. Lett.* **96**, 256–268.

Hawkesworth, C. J., Gallagher, K., Pearson, D. G., Turner, S. P. & van Calsteren, P. 1993 The continental lithosphere: a geochemical perspective. *An. Acad. Bras. Ci.* **65**, 199–225.

Hawkesworth, C., Kelley, S., Turner, S., Le Roex, A. & Storey, B. 1999 Mantle processes during Gondwana break-up and dispersal. In *Proc. Alex du Toit Special Gondwana, 10* (ed. M. J. de Wit). *J. African Earth Sci.* **28**, 5–27.

Herzberg, C. T. 1993 Lithosphere peridotites of the Kaapvaal craton. *Earth Planet. Sci. Lett.* **120**, 13–29.

Houseman, G. & England, P. 1986 A dynamical model of lithosphere extension and sedimentary basin formation. *J. Geophys. Res.* **91**, 719–729.

Houseman, G. A., McKenzie, D. P. & Molnar, P. J. 1981 Convective instability of a thickened boundary layer and its relevance for the thermal evolution of continental convergent belts. *J. Geophys. Res.* **86**, 6115–6132.

Huang, Y., van Calsteren, P. & Hawkesworth, C. J. 1995 The evolution of the lithosphere in Southern Africa: a perspective on the basic granulite xenoliths from kimberlites from Southern Africa. *Geochim. Cosmochim. Acta* **59**, 4905–4920.

Jordan, T. H. 1975 The continental tectosphere. *Geophys. Space Phys.* **13**, 1–12.

Jordan, T. H. 1988 Structure and formation of the continental tectosphere. In *Oceanic and continental lithosphere: similarities and differences* (ed. M. A. Menzies & K. G. Cox). *J. Petrol.* (Spec. Vol.), pp. 11–37.

Kelemen, P. B., Dick, H. J. B. & Quick, J. E. 1992 Formation of harzburgite by pervasive melt/rock reaction in the upper mantle. *Nature* **358**, 635–641.

Kirby, S. H. & Kronenberg, A. K. 1984 Deformation of clinopyroxenite: evidence for a transition in flow mechanisms and semibrittle behaviour. *J. Geophys. Res.* **89**, 3177–3192.

Klein, E. M. & Langmuir, C. H. 1987 Global correlations of ocean ridge basalt chemistry with axial depth and crustal thickness. *J. Geophys. Res.* **92**, 8089–8115.

Knipe, R. J. 1989 Deformation mechanisms—recognition from natural tectonites. *J. Struct. Geol.* **11**, 127–146.

Kramers, J. D. 1979 Lead, uranium, strontium, potassium and rubidium in inclusion-bearing diamonds and mantle derived xenoliths from southern Africa. *Earth Planet Sci. Lett.* **42**, 58–70.

Kushiro, I. 1990 Partial melting of mantle wedge and evolution of island arc crust. *J. Geophys. Res.* **95**, 15 929–15 939.

Maaloe, S. & Aoki, K. 1977 The major element composition of the upper mantle estimated from the composition of lherzolites. *Contrib. Mineral. Petrol.* **63**, 161–173.

McBride, J. S., Lambert, D. D., Greig, A. & Nicholls, I. A. 1996 Multistage evolution of Australian subcontinental mantle: Re–Os isotopic constraints from Victorian mantle xenoliths. *Geology* **24**, 631–634.

McKenzie, D. P. 1978 Some remarks on the development of sedimentary basins. *Earth Planet. Sci. Lett.* **40**, 25–32.

McKenzie, D. & Bickle, M. J. 1988 The volume and composition of melt generated by extension of the lithosphere. *J. Petrol.* **29**, 625–679.

Mackwell, S. K., Bai, Q. & Kohlstedt, D. L. 1990 Rheology of olivine and the strength of the lithosphere. *Geophys. Res. Lett.* **17**, 9–12.

Menzies, M. A. 1990 *Petrology and geochemistry of the continental mantle*, pp. 1–184. Oxford: Clarendon.

Menzies, M. A. & Hawkesworth, C. J. (eds) 1987 *Mantle metasomatism*. Academic.

Molnar, P., England, P. & Martinod, J. 1993 Mantle dynamics, uplift of the Tibetan plateau and the Indian Monsoon. *Rev. Geophys.* **31**, 357–396.

Nägler, Th. F., Kramers, J. D., Kamber, B. S., Frei, R. & Prendergast, M. D. A. 1997 Growth of subcontinental lithospheric mantle beneath Zimbabwe started at or before 3.8 Ga: Re–Os study on chromites. *Geology* **25**, 983–986.

Nelson, D. R., McCulloch, M. T. & Sun, S.-S. 1986 The origins of ultrapotassic rocks as inferred from Sr, Nd and Pb isotopes. *Geochim. Cosmochim. Acta.* **50**, 231–245.

Newman, R. & White, N. 1997 Rheology of the continental lithosphere inferred from sedimentary basins. *Nature* **385**, 621–624.

Nixon, P. J. & Boyd, F. R. 1973 Petrogenesis of the granular and sheared ultrabasic nodule suite in kimberlite. In *Lesotho kimberlites: Cape Town, Cape and Transvaal* (ed. P. H. Nixon), pp. 48–56. Lesotho National Development Corporation, Maseru.

Nixon, P. J. & Davies, G. R. 1987 Mantle xenolith perspectives. In *Mantle xenoliths* (ed. P. J. Nixon), pp. 741–756. Wiley.

Olafsson, M. & Eggler, D. H. 1983 Phase relations of amphibole, amphibole-carbonate, and phlogopite-carbonate peridotite: petrologic constraints on the asthenosphere. *Earth Planet. Sci. Lett.* **64**, 305–315.

O'Nions, R. K. & McKenzie, D. P. 1988 Melting and continent generation. *Earth Planet. Sci. Lett.* **90**, 449–456.

Parkinson, I. J., Hawkesworth, C. J. & Cohen, A. S. 1998 Ancient mantle in a modern arc: osmium isotopes in Izu–Bonin–Mariana Forearc peridotites. *Science* **281**, 2011–2013.

Parsons, B. & McKenzie, D. 1978 Mantle convection and the thermal structure of the plates. *J. Geophys. Res.* **83**, 4485–4496.

Parsons, B. & Sclater, J. G. 1977 An analysis of the variation of ocean floor bathymetry and heat flow with age. *J. Geophys. Res.* **82**, 803–827.

Pearson, D. G. 1999 Evolution of cratonic lithospheric mantle: an isotopic perspective. In *Mantle petrology: field observations and high-pressure experimentation: Spec. Publ. in Honour of Francis R. (Joe) Boyd* (ed. Y. Fei & C. Bertka). Cambridge University Press.

Pearson, D. G., Boyd, F. R., Haggerty, S. E., Pasteris, J. D., Field, S. W., Nixon, P. H. & Pokhilenko, N. P. 1994a The characterisation and origin of graphite in cratonic lithospheric mantle: a petrological carbon isotope and Raman spectroscopic study. *Contrib. Minerl. Petrol.* **115**, 449–466.

Pearson, D. G., Boyd, F. R., Hoal, K. E. O., Hoal, B. G., Nixon, P. H. & Rogers, N. W. 1994b A Re–Os isotopic and petrological study of Namibian peridotites: contrasting petrogenesis and composition of on- and off-craton lithospheric mantle. *Min. Mag.* A **58**, 703–704.

Pearson, D. G., Carlson, R. W., Boyd, F. R., Shirey, S. B. & Nixon, P. H. 1995a The stabilisation of Archaean lithospheric mantle: a Re–Os isotope study of peridotite xenoliths from the Kaapvaal craton. *Earth Planet. Sci. Lett.* **134**, 341–357.

Pearson, D. G., Shirey, S. B., Carlson, R. W., Boyd, F. R., Pokhilenko, N. P. & Shimizu, N. 1995b Re–Os, Sm–Nd and Rb–Sr isotope evidence for thick Archaean lithospheric mantle beneath the Siberia craton modified by multi-stage metasomatism. *Geochim. Cosmochim. Acta* **59**, 959–977.

Pearson, D. G., Snyder, G. A., Shirey, S. B., Taylor, L. A., Carlson, R. W. & Sobolev, N. V. 1995c Archaean Re–Os age for Siberian eclogites and constraints on Archaean tectonics. *Nature* **374**, 711–713.

Pearson, D. G., Carlson, R. W., Boyd, F. R., Shirey, S. B. & Nixon, P. H. 1998 Lithospheric mantle growth around cartons: a Re–Os isotope study of peridotite xenoliths from East Griqualand. In *Ext. Abs. 7th Int. Kimb. Conf., Cape Town.*

Piccirillo, E. M., Melfi, A. J., Comin-Chiaramonti, P., Bellieni, G., Ernesto, M., Marques, L. S., Nardy, A. J. R., Pacca, I. G., Roisenberg, A. & Stolfa, D. 1988 Continental flood volcanism from the Paraná Basin (Brazil). In *Continental flood basalts* (ed. J. D. Macdougall), pp. 195–238. Dordrecht: Kluwer.

Plank, T. & Langmuir, C. H. 1998 The geochemical composition of subducting sediment and its consequences for the crust and mantle. *Chem. Geol.* **145**, 325–394.

Reisberg, L. & Lorand, J. P. 1995 Longevity of sub-continental mantle lithosphere from osmium isotope systematics in orogenic peridotite massifs. *Nature* **376**, 159–162.

Richardson, S. H., Gurney, J. J., Erlank, A. J. & Harris, J. W. 1984 Origin of diamonds in old enriched mantle. *Nature* **310**, 198–202.

Richter, F. 1988 A major change in the thermal state of the Earth at the Archean–Proterozoic boundary: consequences for the nature and preservation of continental lithosphere. In *Oceanic and continental lithosphere: similarities and differences* (ed. M. A. Menzies & K. G. Cox). *J. Petrol.* (Spec. Vol.), pp. 39–52.

Riecker, R. E. & Rooney, T. P. 1969 Water-induced weakening of hornblende and amphibole. *Nature* **224**, 1299–1300.

Ringwood, A. E. 1979 Origin of the Earth and the Moon. New York: Springer.

Rogers, N. W. & Hawkesworth, C. J. 1982 Proterozoic age and cumulate origin for granulite xenoliths, Lesotho. *Nature* **299**, 409–413.

Rudnick, R. L. & Fountain, D. M. 1995 Nature and composition of the continental crust; a lower crustal perspective. *Rev. Geophys.* **33**, 267–309.

Rudnick, R. L., McDonough, W. F. & O'Connell, R. J. 1998 Thermal structure, thickness and composition of continental lithosphere. *Chem. Geol.* **145**, 395–412.

Sclater, J. G., Jaupart, C. & Galson, D. 1980 The heat flow through oceanic and continental crust and the heat loss of the Earth. *Rev. Geophys. Sp. Phys.* **18**, 269–311.

Smith, D. 1999 Temperatures and pressures of mineral equilibration in peridotite xenoliths: review, discussion and implications. In *Mantle petrology: field observations and high-pressure experimentation. Spec. Publ. in honour of Francis R. (Joe) Boyd* (ed. Y. Fei & C. Bertka). Cambridge University Press.

Smith, D. & Boyd, F. R. 1992 Compositional zonation in garnets in peridotite xenoliths. *Contrib. Mineral. Petrol.* **112**, 134–147.

Snow, J. E. & Reisberg, L. 1995 Erratum of Os isotope systematics of the MORB mantle: results from altered abyssal peridotites. *Earth Planet. Sci. Lett.* **133**, 411–421.

Sonder, L. J. & England, P. C. 1989 Effects of a temperature dependent rheology on large-scale continental extension. *J. Geophys. Res.* **94**, 7603–7619.

Takahashi, E. 1990 Reconstruction of the Archean Earth. *Eos* **71**, 967.

Takahashi, E. & Kushiro, I. 1983 Melting of a dry peridotite at high-pressures and basalt magma genesis. *Am. Min.* **68**, 859–879.

Taylor, S. R. & McLennan, S. M. 1995 The geochemical evolution of the continental crust. *Rev. Geophys.* **33**, 241–265.

Turcotte, D. L. 1979 Convection. *Rev. Geophys. Space Phys.* **17**, 1090–1098.

Turcotte, D. L. & McAdoo, D. C. 1979 Geoid anomalies and the thickness of the lithosphere. *J. Geophys. Res.* B **84**, 2381–2387.

Turner, S. & Hawkesworth, C. 1995 The nature of the sub-continental mantle: constraints from the major element composition of continental flood basalts. *Chem. Geol.* **120**, 295–314.

Turner, S., Arnaud, N., Liu, J., Rogers, N., Hawkesworth, C., Harris, N., Kelley, S., van Calsteren, P. & Deng, W. 1996 Post-collision, shoshonitic volcanism on the Tibetan Plateau: implications for convective thinning of the lithosphere and the source of ocean island basalts. *J. Petrol.* **37**, 45–71.

Turner, S. P., Platt, J. P., George, R. M. M., Kelley, S. P., Pearson, D. G. & Nowell, G. M. 1999 Magmatism associated with orogenic collapse of the Betic-Alboran Domain, SE Spain. *J. Petrol.* (In the press.)

Walker, R. J., Carlson, R. W., Shirey, S. B. & Boyd, F. R. 1989 Os, Sr, Nd and Pb isotopic systematics of southern African peridotite xenoliths: implications for the chemical evolution of the subcontinental mantle. *Geochim. Cosmochim. Acta* **53**, 1583–1595.

Walter, M. J. 1997 Origin of cratonic peridotite; a test for sorting of olivine and orthopyroxene. *Eos* **78**, 747.

Weertman, J. 1978 Creep laws for the mantle of the earth. *Phil. Trans. R. Soc. Lond.* A **288**, 9–26.

White, R. S. 1988 The Earth's crust and lithosphere. In *Oceanic and continental lithosphere: similarities and differences* (ed. M. A. Menzies & K. G. Cox). *J. Petrol.* (Spec. Vol.), pp. 1–10.
White, N. & McKenzie, D. 1988 Formation of the 'steers head' geometry of sedimentary basins by differential stretching of the crust and mantle. *Geology* **16**, 250–253.

Discussion

N. WHITE (*Bullard Laboratories, University of Cambridge, UK*). The thickness of the lithospheric plate beneath cratons is critically dependent upon the location of the diamond–graphite phase boundary. I understand that the key experiments were carried out some time ago. How well determined are the pressure–temperature conditions of this boundary, and what is the scope for error?

C. J. HAWKESWORTH. It is true that the experiments were carried out a long time ago, but they are still held in high regard. Tests by Joe Boyd and Graham Pearson using various thermobarometers in diamond- and graphite-bearing xenoliths suggest that there is good agreement with the experimentally determined diamond–graphite stability fields, and an overlap into the wrong stability fields of not more than 7–8 kbar, or *ca.* 25 km. Also, thermobarometry on diamond inclusions almost invariably gives pressures in the diamond stability field and often 60 kbar or more.

N. WHITE. How thin is the lithosphere beneath cratons?

C. J. HAWKESWORTH. There is considerable variation in the pressure estimates from different kimberlite localities, but they range from *ca.* 150 km to 200 km for xenoliths from the South African craton, with an error of perhaps 25 km.

D. MCKENZIE (*Bullard Laboratories, University of Cambridge, UK*). I want Professor Hawkesworth to say a little bit more about the pressures and temperatures, because the standard method of calculating the pressures gives values that are higher than recent estimates from laboratory experiments. But the calculated pressures are consistent with Kennedy and Kennedy's diamond–graphite boundary, which the more recent expressions based on laboratory experiments are not.

C. J. HAWKESWORTH. The currently preferred means of calculating equilibration conditions for garnet peridotites is by using the pyroxene-solvus thermometer and the Al-solubility in orthopyroxene barometer formulations of Brey and colleagues. This combination gives comparable results to several independent thermometers and barometers in the range between 2.5 and 5 GPa and 900 and 1400 °C, although large errors are likely at low temperatures due to steepening of the pyroxene-solvus limb, and there is significant scatter at higher temperatures. As indicated earlier, there is good agreement with the experimentally determined stability fields for diamond and graphite, and the overlap into the wrong stability fields appears to be not more than 7–8 kbar, or *ca.* 25 km.

D. MCKENZIE. The problem with the laboratory experiments has been that they are not compatible with the diamond–graphite stability boundary. There have always been two possible explanations: that something's wrong with the more recent experiments, or that there's something wrong with the diamond–graphite boundary, which has been measured, to my knowledge once, by Kennedy and Kennedy something like 20 years ago and has never been repeated. Basically, Kennedy's reputation is so great that no one thought it worthwhile to redo it, but 20 years is a very long time ago.

If you take the laboratory experiments, then the depths could be shallower, perhaps by as much as 50 km.

M. MENZIES (*Royal Holloway College, University of London, UK*). It is important to distinguish between Archaean regions of the Earth that are tectonically and magmatically inactive (e.g. Kaapvaal craton, South Africa) and Archaean regions that have been reactivated (e.g. North China craton, eastern China). In the former, crust older than 2.5 Ga is underlain by a stable mantle keel that reaches to *ca.* 150–200 km, whereas in the latter the total lithosphere (i.e. crust and lithospheric mantle) thickness is *ca.* 80 km. Given that the Kaapvaal is the most studied craton can we assume that its lithostratigraphy can be used as a reference against which to compare other cratons (e.g. Slave Canada, Siberia)?

C. J. HAWKESWORTH. Historically, the problem has been that the Kaapvaal has been the archetypal or almost only craton that's been well studied and that perhaps too much of our arguments have hinged on it. It appears to be exceptionally well preserved, and so it may provide some of the most reliable information about the nature of the Archaean lithosphere at the time of its formation. In areas where the Archaean lithosphere is now thinner, it is hard to unravel what it was like before the recent tectonics. Is that right?

M. MENZIES. Yes, but the Kaapvaal might have been thicker in the past and thinner now.

C. J. HAWKESWORTH. Well, certainly Kerry Gallagher and Tony Erlank, among others, have made a couple of points: one is that there was a Cretaceous metasomatic event in the mantle beneath the Kaapvaal. Second, it has been argued that the high-temperature peridotites are restricted to xenoliths that are in kimberlites which are less than 90 million years old, and that they therefore reflect relatively recent remobilization, which may even be linked to changes in the erosion history. However, the osmium isotope data on the high-temperature peridotites do indicate that they have been within the lithosphere for billions of years.

M. MENZIES. But the metasomatism might have affected the pressures and temperatures in some of those deformed peridotites.

C. J. HAWKESWORTH. It would make them younger not older, assuming equilibrium in the metasomatism.

E. NISBET (*Royal Holloway College, University of London, UK*). Can I raise a problem which has been puzzling me for many years, which is the presence of detrital diamonds in the late Archaean Witwatersrand sediments? This of course fits with the sort of ages which Jan Kramers was getting, but the presence of diamond in sediment shows very clearly that by the later Archaean there were diamonds not only forming down there but actually being erupted, eroded and redeposited: in other words the lithosphere was cool, thick and of fairly modern aspect. Yet the regional geology (e.g. in Barberton) is consistent with what you were saying that the nearby lithosphere dates back to about 3.3 Ga or 3.4 Ga. If so, and if the two areas were contiguous, that really leaves you with a very short time-scale to make cool enough conditions for diamonds. Conductively it is possible, but difficult, to cool things down that much in the time-period, though the tectonic history could have induced cooling by introducing cooler material at depth, for example by subduction.

C. J. HAWKESWORTH. Can you clarify for us just how good the time constraints are between the deposition of those diamonds: what is the age?

E. NISBET. Well, the diamonds were deposited in the sediment before about 2.7 Ga.

C. J. HAWKESWORTH. So you have *ca.* 700 million years to cool the lithosphere by conduction, which should be alright.

R. S. WHITE (*Bullard Laboratories, University of Cambridge, UK*). Is the lower Archaean lithosphere full of diamonds everywhere? The market in gem quality diamonds and their high prices is built largely on the restricted number available for purchase. Is there in fact an unlimited supply of diamonds beneath our feet?

C. J. HAWKESWORTH. I think it is regarded as pretty unlimited, and in Australia diamondiferous lamproites are found in basement with ages down to *ca.* 1.7 or 1.8 Ga. So provided you stay with basement older than that, there should be lots of diamonds down there.

M. OSMASTON (*Woking, UK*). Can Professor Hawkesworth be completely sure that the nodule compositions from which he is inferring so much have not suffered any syn-magmatic depletion? Is the rhenium/osmium system any use here?

C. J. HAWKESWORTH. Syn-magmatic depletion is unlikely to affect the osmium isotope ratio of the peridotite, because there is so little osmium in the magma and so much in the peridotite. However, interaction with the kimberlite could change the rhenium/osmium ratio, which is why we don't use it in the calculation of the model ages.