

The thickness and heat production of Archean lithosphere: constraints from xenolith thermobarometry and surface heat flow

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Abstract—The pressure-temperature arrays derived from thermobarometry of peridotite xenoliths are compared for four Archean cratons: Kalahari, Slave, Superior and Siberia. Except for Siberia, which shows significant scatter in the pressure-temperature data, the arrays for these cratons are relatively coherent and do not show large differences in temperature ($<200^{\circ}\text{C}$ at 5 GPa), or temperature gradient, between cratons. This observation suggests that the heat flowing through the mantle root is similar for each region. If we assume that the P-T arrays reflect the present-day thermal regime of the cratons, inferences can be made regarding the heat production and thickness of cratonic lithosphere in regions where sufficient P-T data exist to define the conductive geotherm well. The Kalahari craton is one such region, where over 100 P-T data points exist for mantle xenoliths. We employ a grid search technique to generate some 18,000 geotherms by varying the thermal, structural and chemical properties of the lithosphere within reasonable limits. Successful models, defined as those producing geotherms falling within the 95% confidence limits of a linear least-squares fit to the P-T array, yield estimates of mantle heat production between 0 and $0.07 \mu\text{W}/\text{m}^3$ (with most models $\leq 0.03 \mu\text{W}/\text{m}^3$) and lithospheric thickness of 200–250 km. Using the average surface heat flow for the Kalahari craton ($47 \pm 2 \text{ mW}/\text{m}^2$), the modeling results also limit crustal heat production to between 0.5 and $0.8 \mu\text{W}/\text{m}^3$ and require a significant heat flux through the Moho (17 to $25 \text{ mW}/\text{m}^2$), and the base of the lithosphere (8 to $22 \text{ mW}/\text{m}^2$).

INTRODUCTION

STUDY OF the structure, composition and thermal evolution of Archean-aged continental crust and underlying mantle lithosphere is fundamental to understanding the processes of continental growth and evaluating the long-term stability of continents. Archean cratons are stable regions where surface heat flow is observed to be relatively low and uniform (MORGAN, 1984; NYBLADE and POLLACK, 1993). These conditions imply a steady state heat flux, which is controlled by conduction of heat from two sources: 1) heat produced by radioactive decay of K, Th and U in the lithosphere and 2) heat conducted from the underlying convective mantle (POLLACK, 1980). Depending upon the concentration and distribution of heat producing elements in the lithosphere and the thermal conductivity, a large range of conductive geotherms can be produced for any given surface heat flow, spanning temperature differences of up to several hundreds of degrees Celsius at depths of 100–200 km (DAVIES and STREBECK, 1982; RUDNICK *et al.*, 1998). Thus surface heat flow alone does not tightly constrain the thermal conditions within mantle lithosphere.

Xenoliths, foreign rock fragments picked up and carried rapidly to the Earth's surface in magmas such as kimberlites and alkali basalts, provide some of the few petrological and geochemical constraints on the thermal properties and state of the deep lithosphere (*e.g.*, BOYD, 1973). Mineral chemical data appropri-

ate for thermobarometry are now available from mantle xenoliths from several Archean cratons [*e.g.*, Kalahari (Kaaopvaal and Zimbabwe), Siberia, Slave, Superior and Tanzania – see below] and, although differences exist between xenoliths from different cratons, there are several important similarities:

- 1) Both coarse-granular and deformed xenoliths occur in each craton, with the deformed or porphyroclastic xenoliths generally recording higher equilibration temperatures than the coarse-granular xenoliths.

- 2) The P-T array defined by coarse, low temperature xenoliths from each craton are similar, falling close to a $40 \text{ mW}/\text{m}^2$ reference geotherm of POLLACK and CHAPMAN (1977).

- 3) The maximum equilibration depth of cratonic peridotite xenoliths is ≤ 7 GPa (230 km depth). [Some rare garnet lherzolites from the Jagersfontein kimberlite have inclusions in garnets that are interpreted as exsolution of pyroxene from an original majorite garnet (SAUTTER *et al.*, 1991). Likewise, ultra high pressure phases such as majorite (MOORE and GURNEY, 1985) and possible lower mantle minerals (HARRIS *et al.*, 1997; HARTE and HARRIS, 1994), have been reported as inclusions in diamonds. These rare minerals and rocks may indicate formation of the kimberlite very deep within the earth (*e.g.*, HAGGERTY, 1994; RINGWOOD *et al.*, 1992), but cannot be taken as evidence for the thickness of the lithosphere

unless one is prepared to argue that the lithosphere extends into the lower mantle.]

In this paper we assume that the P-T arrays defined by cratonic xenoliths represent equilibration to a conductive, steady-state geotherm at the time of kimberlite eruption (*i.e.*, that the kimberlite magmatism has not thermally affected the xenoliths or the lithosphere). Furthermore, because of the long time constant for thermal diffusion [100's of Ma for a heat flow anomaly on the order of 5 mW/m² to propagate through a 200 km thick lithosphere (NYBLADE, 1999)], we assume that these P-T arrays are representative of present-day conditions beneath Phanerozoic kimberlite pipes; the kimberlites included in this study range from Mesozoic (Kalahari, Slave, Superior) to Permian (Siberia). Given this, a "geotherm window" can be defined for cratons with sufficient data coverage (*i.e.*, Kalahari craton) by using the P-T estimates to place bounds on the range of permissible temperatures at specified depths in the mantle. By varying lithospheric thermal and structural parameters, families of geotherms that satisfy this geotherm window can be calculated. The range of acceptable geotherms provides limits on the amount of heat production in the lithospheric mantle. Given surface heat flow observations, this approach can also constrain the amount of heat production in the crust. Finally, the intersection of the conductive geotherms with the mantle adiabat is used to define the thickness of the Archean thermal boundary layer, and this thickness, when combined with the heat production of the crust and lithospheric mantle, yields estimates of the proportion of surface heat flow coming from the lithospheric mantle and convecting mantle.

P-T ARRAYS OF CRATONIC XENOLITHS

Thanks to the pioneering work of F. R. BOYD (*e.g.*, BOYD, 1973; BOYD, 1984; FINNERTY and BOYD, 1987) the past several decades have witnessed large advances in our understanding of the thermal structure of Archean cratons through the application of thermobarometry to cratonic mantle xenoliths. An excellent summary of thermobarometry as applied to cratonic xenoliths is found in SMITH (1999), in which he reviews the evidence for equilibration in these samples. Zoning is common in the deformed, high temperature samples, leading to uncertainty regarding the veracity of their P-T arrays (*e.g.*, LEE, 1998; SMITH, 1999). Evidence for trace element and isotopic disequilibria has been documented in some coarse granular peridotites (GÜNTHER and JAGOUTZ, 1994; SHIMIZU *et al.*, 1997), but the general lack of zoning of major elements in these minerals is cited as evidence for their equilibration to ambient conditions

(SMITH, 1999). This conclusion is consistent with the evidence for near isotopic equilibration at the time of eruption for the Sm-Nd and Rb-Sr systems in rocks that are clearly much older (RICHARDSON *et al.*, 1985). For Siberian samples, however, some evidence for major element disequilibria has been documented (BOYD *et al.*, 1997). Thus, one must bear in mind that any use of xenolith P-T arrays to infer thermal structure of cratonic lithosphere is subject to the underlying assumption that equilibrium has been attained. If this is shown in the future to be incorrect, then the conclusions drawn here must be revised accordingly.

The most recent and thorough experimental calibrations of the Fe-Mg exchange thermometer between clinopyroxene and orthopyroxene and the Al in orthopyroxene barometer of Brey and colleagues (BREY and KÖHLER, 1990 and BREY *et al.*, 1990) (hereafter termed the "BKN" thermobarometers) are used to compare equilibration conditions of cratonic xenoliths from four Archean cratons: Kalahari (BOYD and MERTZMAN, 1987; SHEE, 1978; STIEFENHOFER *et al.*, 1997), Slave (BOYD and CANIL, 1997; KOPYLOVA *et al.*, 1998; PEARSON *et al.*, 1998), Siberia (BOYD *et al.*, 1997; GRIFFIN *et al.*, 1996; POKHILENKO *et al.*, 1993; ZHURAVLEV *et al.*, 1991) and Superior (MEYER *et al.*, 1994; SCHULZE, 1996) [Data for peridotite xenoliths from the Tanzanian craton have not been considered here since evidence exists for incipient heating by rift-related magmas (DAWSON *et al.*, 1997; LEE and RUDNICK 1999)]. Pressure and temperature have been calculated directly from mineral chemical data from the literature or supplied from the authors; in all cases Fe³⁺ is assumed to be zero.

Figure 1 shows the P-T arrays for peridotite xenoliths from the four cratons; coarse-granular peridotites are shown as solid symbols, porphyroclastic or deformed peridotites are shown as open symbols. The data are plotted relative to the best fit conductive geotherm for the Kalahari craton (described below). Also shown in the Kalahari plot is the 40 mW/m² geotherm of POLLACK and CHAPMAN (1977). The thermobarometry data for xenoliths from each craton generally form a coherent trend. The exception is Siberia, for which evidence of mineralogical disequilibria has been found (BOYD *et al.*, 1997; GRIFFIN *et al.*, 1996). More limited scatter exists in the P-T arrays for the other cratons, which is probably due to 1) analytical errors in mineral compositions, propagated through the thermobarometry, 2) uncertainties in the thermobarometry calibration (estimated at $\pm 20^\circ\text{C}$ and ± 0.3 GPa for the BKN calibration, BREY and KÖHLER, 1990), and possibly 3) subtle variations in temperature (beyond the resolution of thermobarometry) within a given craton.

Deformed, high temperature peridotites were originally interpreted as deriving from the asthenosphere

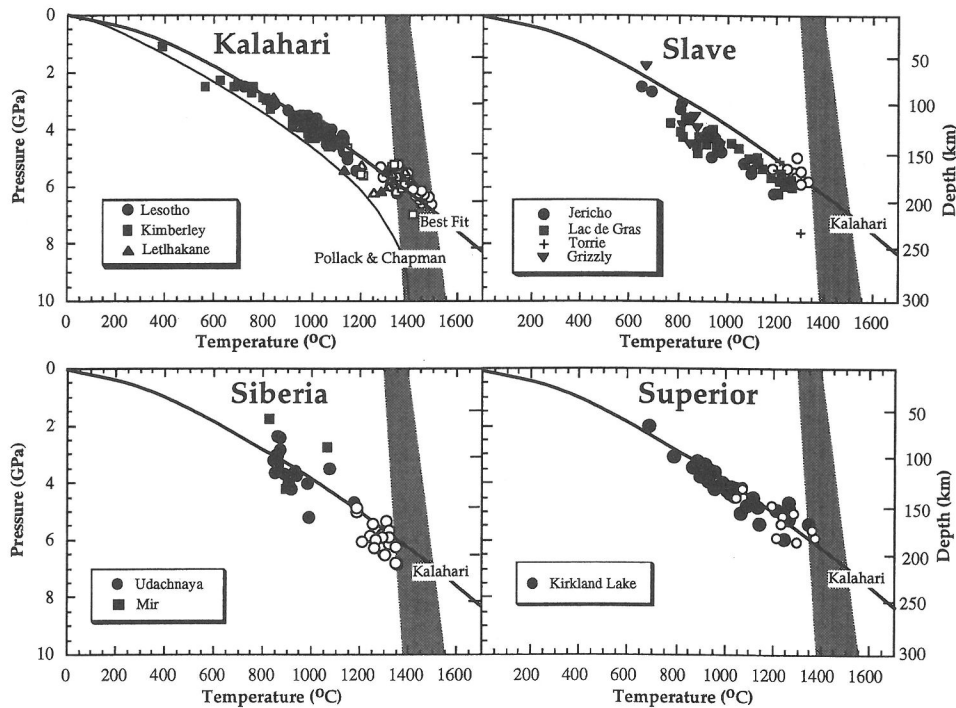


FIG. 1. Pressure and temperature of equilibration for peridotite xenoliths in kimberlites from different Archean cratons, calculated using the BREY and KÖHLER (1990) formulations of the 2 pyroxene thermometer and Al in orthopyroxene barometer. Solid symbols are coarse-granular peridotites, open symbols are deformed peridotites. The best fit geotherm for the Kalahari data is plotted for reference in each diagram. (a) Kalahari craton. Data sources: Lesotho (FINNERTY and BOYD, 1984; NIXON and BOYD, 1973) and Kimberly (unpublished data from F. R. BOYD), Kaapvaal craton; Lethakane kimberlites, Zimbabwe craton (SHEE, 1978; STIEFENHOFER *et al.*, 1997). Two geotherms are plotted: the Pollack and Chapman 40 mW/m² conductive geotherm and the best fit geotherm for the P-T data (see Table 1 for parameters). (b) Slave craton. Data sources: Jericho (KOPYLOVA *et al.*, 1998), Lac de Gras (PEARSON *et al.*, 1998), Torrie (CANIL, unpublished data), Grizzly (BOYD and CANIL, 1997). (c) Siberian craton. Data sources: Udachnaya (BOYD *et al.*, 1997; GRIFFIN *et al.*, 1996; SPETSIOUS and SERENKO, 1990), Mir (ZHURAVLEV *et al.*, 1991) (d) Superior craton. Data sources: (MEYER *et al.*, 1994; SCHULZE, 1996; VICKER, 1997).

(NIXON and BOYD, 1973) because (1) they have porphyroclastic textures, indicating flow, (2) they were thought to be fertile, *i.e.*, close in composition to a model primitive mantle composition, such as the famous sample PHN1611, used in several experimental investigations of mantle melting (*e.g.*, SCARFE and TAKAHASHI, 1986) and (3) they appeared to define a “kink” in the xenolith P-T array, which might have marked a thermal boundary layer at the base of the lithosphere. However, this interpretation has not held up in light of more recent investigations. A compilation of major element chemistry for Kaapvaal peridotites shows that the average Al₂O₃ and CaO content of the deformed peridotites is similar to that of coarse-granular peridotites (MCDONOUGH and RUDNICK, 1998), indicating that they are equally refractory. However, Fe contents are higher (and Mg# lower) than coarse-granular peridotites, and together

with the zoning seen in garnets in many of these peridotites (GRIFFIN *et al.*, 1989; SMITH and BOYD, 1987), is interpreted to reflect metasomatic enrichment of Fe, Ti and Zr shortly before the xenolith’s entrainment in the kimberlites. The “kink” in the geotherm appears to be an artifact of choice of thermometer and geotherm. The kink is present when using the FINNERTY and BOYD (1987) plus MACGREGOR (1974) combinations of the two-pyroxene thermometer and Al in orthopyroxene barometer, but is not apparent when using the BKN thermobarometers. In addition, the POLLACK and CHAPMAN (1977) geotherm curves strongly away from the xenolith P-T array below depths of 150 km, giving rise to an apparent discrepancy between the P-T array and the conductive geotherm (this is discussed more fully below). Finally, Os isotopic measurements indicate that many of these high temperature peridotites

experienced ancient melt depletion, and must therefore be pieces of the ancient lithosphere that were recently overprinted (PEARSON *et al.*, 1995). For these reasons, and the fact that they form a continuum with the coarse granular peridotites in a temperature vs. pressure plot (*e.g.*, Fig. 1), we include them in our thermal modeling.

Although still widely used within the petrological community today, the POLLACK and CHAPMAN (1977) family of geotherms is not particularly applicable to Archean cratons. They were modeling the thermal structure of relatively thin continental lithosphere (120 km). Because of this, the heat production in their mantle jumps from $0.01 \mu\text{W}/\text{m}^3$ between 40 and 120 km depth (their model lithosphere), to $0.084 \mu\text{W}/\text{m}^3$ below these depths – the pyrolite model of the day. This high heat production at depths greater than 120 km gives rise to the pronounced curvature of their geotherm, causing it to fall below the high temperature end of the xenolith P-T array in every craton. Although the heat production of lithospheric mantle cannot be well constrained from the K, Th and U of cratonic xenoliths (see RUDNICK *et al.*, 1998, for a more detailed discussion), the P-T array itself suggests that the heat production of the lithosphere is relatively low, $\leq 0.03 \mu\text{W}/\text{m}^3$, as discussed below.

The well-defined Kalahari geotherm serves as a marker by which to compare the absolute temperature differences between different cratons. Variations in equilibration conditions of the coarse granular peridotites are minimal within individual cratons (except the Siberian craton where significant scatter exists even in one pipe), but significant differences exist between cratons. Peridotites from the Kalahari and Superior cratons plot at higher temperatures for a given pressure than peridotites from the Siberian and Slave cratons. However, the maximum temperature difference between different cratons ($<200^\circ\text{C}$ at 5 GPa) is smaller than the difference between xenoliths from Archean cratons and those from adjacent Proterozoic crust (*e.g.*, BOYD, 1979; FRANZ *et al.*, 1996; MITCHELL, 1984). The overall uniformity of the xenolith P-T array between Archean cratons suggests a similar amount of heat flowing through the mantle lithosphere at each locality. We will now use these observations to make inferences about crustal and mantle heat production and overall craton thickness. We focus our efforts on the Kalahari craton, for which a large data density exists.

INFERENCES FROM THE KALAHARI P-T ARRAY

The xenolith thermobarometry data from the Kalahari Craton reasonably constrain temperatures be-

tween depths of about 70 and 200 km (Fig. 1). We therefore use these data and the approach outlined in the introduction to obtain estimates of crust and mantle heat production and lithosphere thickness.

To find the range of geotherms that satisfy the Kalahari P-T array, we first determine the 95% confidence limits (2 standard deviations) of the P-T array by fitting the data using linear and second-order least squares regressions. Both regressions yield similar fits to the data (*r* values of 0.969 and 0.970 for the linear and second-order fit, respectively), and for simplicity we use the confidence limits for the linear regression in our modeling, which are shown in Fig. 2.

Next we perform a grid search to obtain a family of geotherms that fall within the 95% confidence limits of the P-T array. In this grid search, geotherms are calculated for a range of plausible lithospheric parameters using the method outlined by CHAPMAN (1986). The parameter range is shown in Table 1. The surface heat flow from the Kaapvaal and Zimbabwe Cratons is similar (JONES, 1988; JONES, 1992; NYBLADE *et al.*, 1990), giving an average heat flow for the Kalahari Craton of 47 ± 2 (st. error) mW/m^2 (NYBLADE *et al.*, 1990). The heat flow range in Table 1 is based on the $\pm 2 \text{ mW}/\text{m}^2$ uncertainty. Values of mean crustal heat production were chosen to span a range wider than the range reported in the literature

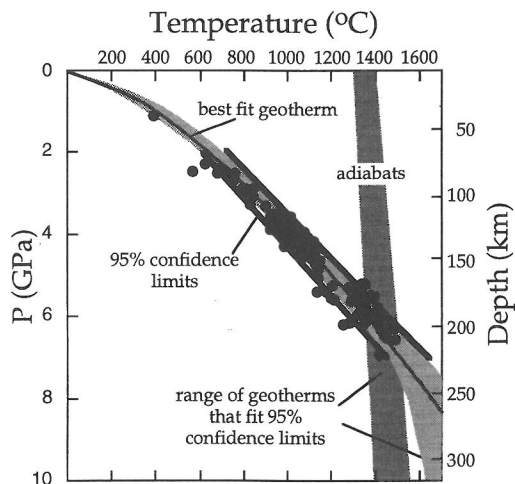


FIG. 2. P-T data array for Kalahari craton with 95% confidence limits for the linear regression of the data. Also shown are the upper and lower bounds (gray field) for the family of geotherms ($n = 535$) that fall within the 95% confidence limits and the best fit geotherm to the regressed data (Table 1). Adiabats correspond to potential temperatures of 1300 and 1400°C. Note that the high temperature xenoliths fall below the highest adiabat and there is no need to appeal to a special origin via diapiric upwelling to explain these data (*cf.* JORDAN, 1988).

Table 1. Input and output values of parameters used in geotherm equation in the grid search test. The best fit geotherm is the one having the lowest RMS error with the regressed P-T data

Kalahari model parameters	Permissible Models			
	Input Range	Minimum	Maximum	Best Estimate
Surface Q (mW/m ²)	45-49	45	49	45
Crustal HP (μ W/m ³)	0.3-0.9	0.5	0.8	0.7
K Crust (W/m.K)	2.4-3.0	2.4	3	2.4
Mantle HP (μ W/m ³)	0-0.1	0	0.07	0
Crustal thickness (km)	35-45	35	45	38
Q from lithospheric mantle (mW/m ²)		0	15	0
Q from crust (mW/m ²)		20	32	27
Q Moho (mW/m ²)		17	25	18
Q Base of Lithosphere (mW/m ²)		8	21	18
Lithosphere Thickness (km)		203	253	224

for Archean crust (see RUDNICK *et al.*, 1998, and references therein). Distribution of heat producing elements in the crust follows the three layer model of RUDNICK and FOUNTAIN (1995), with 60% in the upper crust, 34% in the middle crust and 6% in the lower crust. The range of mean thermal conductivity for the crust was chosen to span a range of values typical of crustal rocks (CERMAK and RYBACH, 1982; ROY *et al.*, 1981), including the effects of pressure and temperature on conductivity (CHAPMAN, 1986; SASS *et al.*, 1992; ZOTH and HAENEL, 1988). In the lithospheric mantle, we used the thermal conductivity model of SCHATZ and SIMMONS (1972), which accounts for the effects of pressure and temperature. Since heat production in the lithospheric mantle is poorly constrained, a very large range of values was chosen so as not to influence the results of the modeling. Lastly, the range of crustal thickness was chosen to span the range of average crustal thicknesses for Archean crust (40 ± 5 km, CHRISTENSEN and MOONEY, 1995; RUDNICK and FOUNTAIN, 1995). The few published estimates of crustal thickness for the Kalahari Craton fall within the 40 ± 5 km range (DURRHEIM *et al.*, 1992; DURRHEIM and GREEN, 1992; GANE *et al.*, 1956; STUART and ZENGENI, 1987), suggesting that the Kalahari crust is not significantly different from Archean crust elsewhere.

Some 18,000 geotherms were generated in the grid search by incrementally changing each model parameter individually. Five hundred and thirty-five geotherms fell within the range of acceptable temperatures defined by the 95% confidence limits (Fig. 2). Table 1 gives the range (minimum and maximum) of values for each parameter represented in the ensemble

of successful models. In the case of crust and mantle heat production, the range of values in the successful models is a subset of the range of input values, demonstrating that the range of input values had little effect on the outcome of the grid search. In addition, results from sensitivity analyses indicate that the results of the grid search are not influenced by the choice of surface temperature used in calculating the geotherms, nor by the details of how heat production is distributed in the crust (in agreement with DAVIES and STREBECK, 1982), as long as a large percentage of the heat production resides in the upper crust. Table 1 also shows the parameters of the "best fit" geotherm – the geotherm giving the lowest RMS error when compared to the regressed P-T data. This geotherm is plotted for reference against all the P-T data arrays in Fig. 1.

Figure 3 shows histograms of the values of the different parameters obtained from the successful models. In most cases, the histograms approximate normal distributions (*e.g.*, lithospheric thickness, heat flow into lithosphere, crustal heat production), indicating that most successful models yield values in the middle of the output range. However, this is not the case for mantle heat production, which shows a strongly skewed distribution, indicating that most of the successful models have low lithospheric mantle heat production (Fig. 3).

From the above analysis, the bounds on average crustal heat production in the Kalahari craton is 0.5 to 0.8 μ W/m³ and lithospheric mantle heat production ranges from 0 to 0.07 μ W/m³. However, only a very small number of models have mantle heat production as high as 0.07 μ W/m³, and the vast majority of the

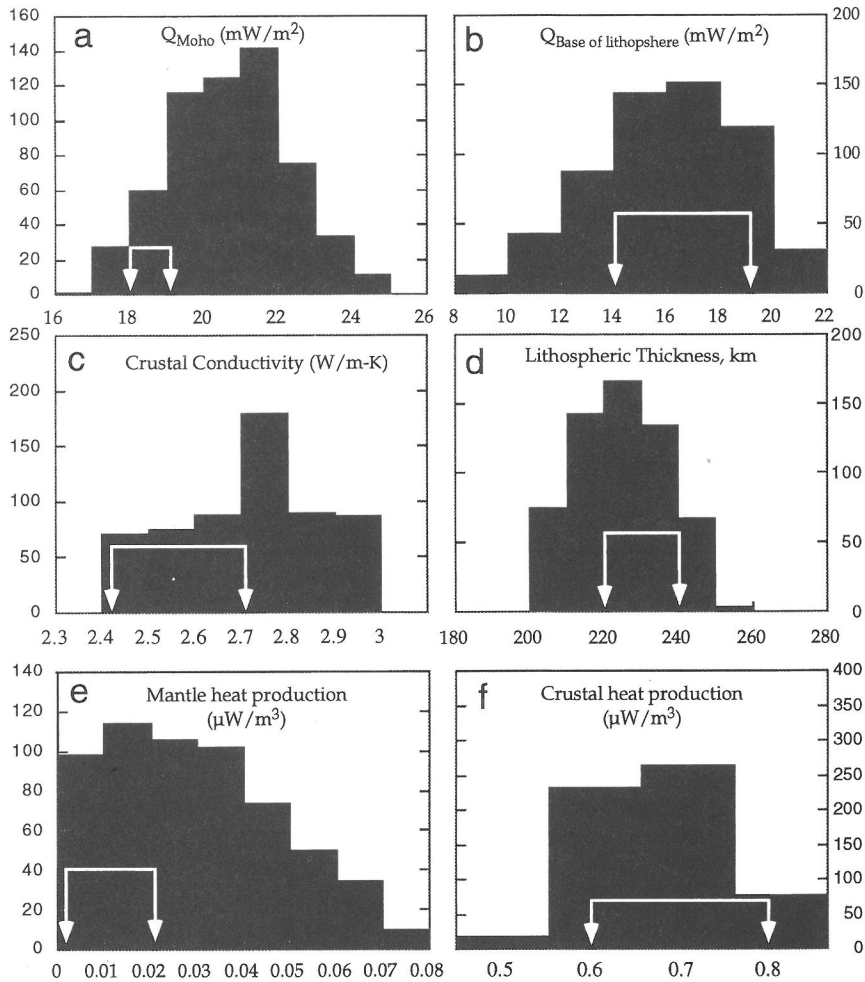


FIG. 3. Histograms of the parameters derived from the grid search of geotherms that fall within the 95% confidence limits of the P-T data for the Kalahari craton. Arrows mark range of values for the 10 geotherms that yield the closest fit (lowest rms) to the regressed P-T data. (a) heat flux across the Moho, (b) heat flux through base of lithosphere, (c) crustal conductivity, (d) lithospheric thickness, (e) lithospheric mantle heat production, (f) crustal heat production.

models have heat production $< 0.03 \mu\text{W}/\text{m}^3$ (Fig. 3). Thus, although we cannot exclude a relatively high heat production in the lithospheric mantle on the basis of this modeling, we consider it unlikely. The best fitting geotherm is for a surface heat flow of $45 \text{ mW}/\text{m}^2$, a mean crustal heat production of $0.70 \mu\text{W}/\text{m}^3$, a crustal thickness of 38 km, and no heat production in the lithospheric mantle.

The thickness of the conductive mantle root beneath the Kalahari craton can also be constrained from the analysis of the xenolith P-T array. The thickness is determined from the intersection of the family of conductive geotherms obtained from the grid search with the mantle adiabat. We adopt a range of possible mantle adiabats with potential tempera-

tures between 1300 and 1400°C and thermal gradients between 0.3 and $0.5^\circ\text{C}/\text{km}$ (FEI, 1995; NAVROTSKY, 1995). This yields a range of thicknesses between 200 and 250 km (Table 1), with the best fit geotherm giving a maximum thickness of 224 km (using the hotter adiabat).

OTHER CRATONS

Although the data density from thermobarometry of xenoliths in other cratons is not high enough to allow a statistical treatment like the one provided above for the Kalahari craton, several qualitative interpretations can nonetheless be made. The thermobarometry data from the Superior Province (Fig. 1)

plot on top of the Kalahari data, suggesting that the conclusions regarding lithospheric mantle heat production and thickness for the Kalahari craton can also be applied to the Superior craton. These interpretations are consistent with a recent analysis of the thermal structure of that craton (JAUPART *et al.*, 1998). In the Superior craton, however, surface heat flow is lower than in the Kalahari and the crust appears to be thicker. Assuming that the P-T array observed for the Kalahari craton is also representative of the Superior craton, we re-ran our models changing the surface heat flow bounds to 35–39 mW/m² and crustal thickness bounds to 40–45 km. As expected, this yielded lower crustal heat production than for the Kalahari, but similar mantle heat production and lithospheric thickness. Our modeling suggests that, in contrast to the conclusions of JAUPART *et al.* (1998), the amount of heat flowing across the Moho in the Superior Province is relatively large (18–25 mW/m², *cf.* their estimate of 12–13 mW/m²). This discrepancy implies that either the estimated heat production of cratonic crust of JAUPART *et al.* (1998) is too high, or that the xenolith P-T array is hotter than the present geotherm beneath the craton.

In comparison to the Kalahari and Superior cratons, both the Siberian and Slave cratons appear to have somewhat cooler lithospheres, suggesting a lower amount of heat flowing from the underlying mantle. Interestingly, surface heat flow data for these two cratons lie at the extremes of observed surface heat flow in Archean cratons. Average surface heat flow in the Siberian craton is very low – less than 35 mW/m² (DUCHKOV, 1991), whereas the limited data for the Slave (one measurement near Yellowknife) suggest that heat flow is high –50–53 mW/m² (LEWIS and WANG, 1992). If these values are confirmed by additional measurements, then they would imply large differences in the crustal composition between these cratons. Moreover, the slightly cooler geotherms could allow a thicker lithospheric root, but probably not much deeper than 250 km, based on extrapolation of the xenolith P-T arrays and the mantle adiabat in Fig. 1 (b and c).

COMPARISON WITH PREVIOUS ESTIMATES

In this section we compare our results with previous estimates of crustal and mantle heat production and lithospheric thickness in Archean cratons. The mean crustal heat production determined from the Kalahari data of 0.5–0.8 $\mu\text{W}/\text{m}^3$ falls within the range of most other estimates based on a number of techniques and assumptions (see RUDNICK *et al.*, 1998). The lower bound (0.5 $\mu\text{W}/\text{m}^3$) matches estimates of average Archean crustal heat production of

TAYLOR and MCLENNAN (1985) and RUDNICK and FOUNTAIN (1995). The upper bound (0.8 $\mu\text{W}/\text{m}^3$) is higher than most estimates of Archean crustal heat production (which are generally $\leq 0.7 \mu\text{W}/\text{m}^3$, MCLENNAN and TAYLOR, 1996; WEAVER and TARNNEY, 1984) but is lower than that estimated for the Archean crust of eastern China (0.9 $\mu\text{W}/\text{m}^3$, GAO *et al.*, 1998). As discussed previously (RUDNICK *et al.*, 1998), this very high heat production inferred for the Chinese crust is anomalous and may be a contributing factor in the high surface heat flow ($\sim 60 \text{ mW}/\text{m}^2$) observed there.

Heat production in the mantle lithosphere is a poorly constrained quantity. Difficulties associated with inferring heat production in the mantle lithosphere from xenolith compositions have been reviewed by McDONOUGH (1990) and RUDNICK *et al.* (1998) and include the enhancement of heat producing element contents due to host infiltration, alteration and weathering, as well as the persistent question regarding the representativeness of the xenoliths for highly incompatible elements such as these. The simple analysis of P-T data presented here may provide a more robust approach to estimating mantle heat production than xenolith compositions, since the amount of curvature in the mantle geotherm directly reflects the amount of heat production.

Is the 200–250 km thick Archean lithosphere estimated here in accord with the results of numerous seismic studies of Archean cratons over the past two to three decades? Estimates of Archean lithosphere thickness from many seismological studies are in the depth range of 300–400 km (*e.g.*, GRAND, 1994; JORDAN, 1975; JORDAN, 1988; SU *et al.*, 1994), significantly deeper than our estimate based on thermal structure. However, more recent interpretations of seismological data are consistent with a ≤ 250 km thick lithosphere (RICARD *et al.*, 1996; VAN DER LEE and NOLET, 1997) and preliminary results from broadband studies seeking to image the lithosphere beneath the Australian (VAN DER HILST *et al.*, 1998), Tanzanian (RITSEMA *et al.*, 1998), and Kalahari (JAMES *et al.*, 1998) cratons estimate minimum thicknesses of 200–250 km, overlapping with our observations. It remains to be seen whether the apparent discrepancies between the earlier seismological models and the thermal modeling persist in light of newer seismological data. If they do, it implies either that the xenolith P-T arrays are too hot and are not representative of present-day conditions beneath the cratons or that cratons may be sites of cold convective downwelling in the mantle (*e.g.*, KING and ANDERSON, 1998).

CONCLUSIONS

Assuming that cratonic peridotite xenoliths record the P-T conditions of present-day Archean mantle roots, the following conclusions can be drawn:

1) The average crustal heat production in the Kalahari craton lies between 0.5 and 0.8 $\mu\text{W}/\text{m}^3$, with a best estimate of 0.70 $\mu\text{W}/\text{m}^3$.

2) The average lithospheric mantle heat production is $<0.07 \mu\text{W}/\text{m}^3$, and, for most models less than 0.03 $\mu\text{W}/\text{m}^3$, with a best estimate of 0 $\mu\text{W}/\text{m}^3$.

3) The intersection of the family of geotherms that fit the xenolith data with a 1400°C mantle adiabat suggests maximum lithospheric thicknesses of 250 km with a best estimate of 220 km for the Kalahari craton.

4) The heat flux across the Moho derived from the xenolith P-T data is 17–25 mW/m^2 , which is higher than recent estimates from the Superior Province (12–13 mW/m^2). The differences are not easily reconcilable and imply either that the estimates of crustal heat production in the Superior Province are too high or that the xenolith P-T arrays are too hot.

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REFERENCES

- BOYD F. R. (1973) A pyroxene geotherm. *Geochim. Cosmochim. Acta* **37**, 2533–2546.
- BOYD F. R. (1979) Garnet lherzolite xenoliths from the kimberlites of East Griqualand, South Africa. *Carnegie Inst. Washington Yearbook* **79**, 296–302.
- BOYD F. R. (1984) Siberian geotherm based on lherzolite xenoliths from the Udachnaya kimberlite, U.S.S.R. *Geology* **12**, 528–530.
- BOYD F. R. and CANIL D. (1997) Peridotite xenoliths from the Slave craton, Northwest territories. In *Seventh Annual V.M. Goldschmidt Conference*, LPI Contribution No. 921, pp. 34. Lunar and Planetary Institute.
- BOYD F. R. and MERTZMAN S. A. (1987) Composition and structure of the Kaapvaal lithosphere, southern Africa. In *Magmatic Processes: Physicochemical principles*, Special Pub. No. 1 (ed. B. O. MYSEN), pp. 13–24. The Geochemical Society.
- BOYD F. R., POKHILENKO N. P., PEARSON D. G., MERTZMAN S. A., SOBOLEV N. V., and FINGER L. W. (1997) Composition of the Siberian cratonic mantle: evidence from Udachnaya peridotite xenoliths. *Contrib. Mineral. Petrol.* **128**, 228–246.
- BREY G. P. and KÖHLER T. (1990) Geothermobarometry in four-phase lherzolites II. New thermobarometers, and practical assessment of existing thermobarometers. *J. Petrol.* **31**, 1353–1378.
- BREY G. P., KÖHLER T., and NICKEL K. G. (1990) Geothermobarometry in four-phase lherzolites I. Experimental results from 10 to 60 kb. *J. Petrol.* **31**, 1313–1352.
- CERMAK A. and RYBACH L. (1982) Thermal conductivity and specific heat of minerals and rocks. In *Landolt-Bornstein Numerical Data and Functional Relationships in Science and Technology*, Vol. 16 (ed. G. ANGENHEISTER). Springer-Verlag.
- CHAPMAN D. S. (1986) Thermal gradients in the continental crust. In *The nature of the lower continental crust*, Vol. 24 (eds. J. B. DAWSON, D. A. CARSWELL, J. HALL, and K. H. WEDEPOHL), pp. 63–70. Geol. Soc. Spec. Publ.
- CHRISTENSEN N. I. and MOONEY W. D. (1995) Seismic velocity structure and composition of the continental crust: a global view. *J. Geophys. Res.* **100**(B7), 9761–9788.
- DAVIES G. F. and STREBECK J. W. (1982) Old continental geotherms: constraints on heat production and thickness of continental plates. *Geophys. J. R. astr. Soc.* **69**, 623–634.
- DAWSON J. B., JAMES D., PASLICK C., and HALLIDAY A. M. (1997) Ultrabasic potassic low-volume magmatism and continental rifting in north-central Tanzania: association with enhanced heat flow. *Proc. Sixth Int. Kimberlite Conf., Russian Geology and Geophysics* **38**, 69–81.
- DUCHKOV A. D. (1991) Review of Siberian heat flow data. In *Terrestrial heat flow and the lithosphere structure* (eds. V. CERMÁK and L. RYBACH), pp. 426–443. Springer-Verlag.
- DURRHEIM R. J., BARKER W. H., and GREEN R. W. E. (1992) Seismic studies in the Limpopo Belt. *Precam. Res.* **55**, 187–200.
- DURRHEIM R. J. and GREEN R. W. E. (1992) A seismic refraction investigation of the Archaean Kaapvaal Craton, South Africa, using mine tremors as the energy source. *Geophys. J. Int.* **108**, 812–832.
- FEI Y. (1995) Thermal expansion. In *Mineral Physics & Crystallography: A handbook of physical constants* (ed. T. J. AHRENS), pp. 29–44. AGU.
- FINNERTY A. A. and BOYD F. R. (1984) Evaluation of thermobarometers for garnet peridotites. *Geochim. Cosmochim. Acta* **48**, 15–27.
- FINNERTY A. A. and BOYD F. R. (1987) Thermobarometry for garnet peridotites: basis for the determination of thermal and compositional structure of the upper mantle. In *Mantle Xenoliths* (ed. P. H. NIXON), pp. 381–402. John Wiley & Sons.
- FRANZ L., BREY G. P., and OKRUSCH M. (1996) Steady state geotherm, thermal disturbances, and tectonic development of the lower lithosphere underneath the Gibeon kimberlite province, Namibia. *Contrib. Mineral. Petrol.* **126**, 181–198.
- GANE P. G., ATKINS A. R., SELLSCHOP J. P. F., and SELIGMAN P. (1956) Crustal structure in the Transvaal. *Bull. Seis. Soc. Am.* **46**, 293–316.
- GAO S., LUO T.-C., ZHANG B.-R., ZHANG H.-F., HAN Y.-W., HU Y.-K., and ZHAO Z.-D. (1998) Chemical composition of the continental crust as revealed by studies in East China. *Geochim. Cosmochim. Acta*, **62**, 1959–1975.
- GRAND S. P. (1994) Mantle shear structure beneath the Americas and surrounding oceans. *J. Geophys. Res.* **99**, 11,591–11,621.
- GRIFFIN W. L., KAMINSKY F. V., RYAN C. G., O'REILLY S. Y., WIN T. T., and ILUPIN I. P. (1996) Thermal state

- and composition of the lithospheric mantle beneath the Daldyn kimberlite field, Yakutia. *Tectonophysics* **262**, 19–33.
- GRIFFIN W. L., SMITH D., BOYD F. R., COUSENS D. R., RYAN C. G., SIE S. H., and SUTER G. F. (1989) Trace-element zoning in garnets from sheared mantle xenoliths. *Geochim. Cosmochim. Acta* **53**, 561–567.
- GÜNTHER M. and JAGOUTZ E. (1994) Isotopic disequilibria (Sm/Nd, Rb/Sr) between mineral phases of coarse grained, low temperature garnet peridotites from Kimberley Floors, southern Africa. In *Kimberlites, Related Rocks and Mantle Xenoliths, (Proceedings Fifth Int. Kimb. Conf.)*, Vol. 1 (eds. H. O. A. MEYER and O. H. LEONARDO), pp. 354–364. C.P.R.M.
- HAGGERTY S. E. (1994) Superkimberlites: a geodynamic diamond window to the Earth's core. *Earth Planet. Sci. Lett.* **122**, 57–69.
- HARRIS J., HUTCHINSON M. T., HURSTHOUSE M., LIGHT M., and HARTE B. (1997) A new tetragonal silicate mineral occurring as inclusions in lower-mantle diamonds. *Nature* **387**, 486–488.
- HARTE B. and HARRIS J. W. (1994) Lower mantle mineral associations preserved in diamonds. *Mineralogical Magazine* **58**, 384–385.
- JAMES D. E., VAN DER LEE S., GAO S., SILVER P., VANDECAR J., KUEHNEL R., JORDAN T. H., SALTZER R., GAHERTY J., GORE J., ZENGENI T., NGUURI T., WRIGHT C., WEBB S., BURFORD D., DOUCOURE M., MOLISANA M., GREEN R., ROBEY J., HARVEY J., KOSTLIN E., and REICHHARDT F. (1998) Southern Africa seismic experiment. *EOS, Spring Meeting Abstracts*, S228.
- JAUPART C., MARESCHAL J. C., GUILLOU-FROTTIER L., and DAVAILLE A. (1998) Heat flow and thickness of the lithosphere in the Canadian shield. *J. Geophys. Res.* **103**, 15,269–15,286.
- JONES M. Q. W. (1988) Heat flow in the Witwatersrand Basin and environs and its significance for the South African shield geotherm and lithosphere thickness. *J. Geophys. Res.* **93**, 3243–3260.
- JONES M. Q. W. (1992) Heat flow anomaly in Lesotho: implications for the southern boundary of the Kaapvaal craton. *Geophys. Res. Lett.* **19**, 2031–2034.
- JORDAN T. H. (1975) The continental tectosphere. *Rev. Geophys. Space Phys.* **13**(3), 1–12.
- JORDAN T. H. (1988) Structure and formation of the continental lithosphere. In *Oceanic and Continental Lithosphere; similarities and differences* (eds. M. A. MENZIES and K. COX), pp. 11–37. *J. Petrol.*, Special Lithosphere Issue.
- KING S. D. and ANDERSON D. L. (1998) Edge-driven convection. *Earth Planet. Sci. Lett.* **160**, 289–296.
- KOPYLOVA M. G., RUSSELL J. K., and COOKENBOO H. (1998) Upper-mantle stratigraphy of the Slave craton, Canada: insights into a new kimberlite province. *Geology* **26**, 315–319.
- LEE C.-T. (1998) Are inflected geotherms real? In *7th Internat. Kimberlite Conf. Abstracts*, pp. 489–491. Univ. Cape Town.
- LEE, C.-T. and RUDNICK, R. L. (1999) Compositionally stratified cratonic lithosphere: petrology and geochemistry of peridotite xenoliths from the Labait Volcano, Tanzania. In *Proceedings of the Seventh Kimberlite Conference* (eds. J. J. GURNEY and S. H. RICHARDSON) in press.
- LEWIS T. J. and WANG K. (1992) Influence of terrain on bedrock temperatures. *Global Planetary Change* **98**, 87–100.
- MACGREGOR I. D. (1974) The system MgO-Al₂O₃-SiO₂: Solubility of Al₂O₃ in enstatite for spinel and garnet peridotite. *Am. Mineral.* **59**, 110–119.
- MCDONOUGH W. F. (1990) Constraints on the composition of the continental lithospheric mantle. *Earth Planet. Sci. Lett.* **101**, 1–18.
- MCDONOUGH W. F. and RUDNICK R. L. (1998) Mineralogy and composition of the upper mantle. In *Ultra High Pressure Mineralogy, Reviews in Mineralogy*, Vol. 37 (ed. R. J. HEMLEY), pp. 139–164. Mineral. Soc. Am.
- MCLENNAN S. M. and TAYLOR S. R. (1996) Heat flow and the chemical composition of continental crust. *J. Geol.* **104**, 396–377.
- MEYER H. O. A., WALDMAN M. A., and GARWOOD B. L. (1994) Mantle xenoliths from kimberlite near Kirkland Lake, Ontario. *Can. Mineral.* **32**, 295–306.
- MITCHELL R. H. (1984) Garnet lherzolites from the Hanaus-I and Louwrensia kimberlites of Namibia. *Contrib. Mineral. Petrol.* **86**, 178–188.
- MOORE R. O. and GURNEY J. J. (1985) Pyroxene solid solution in garnets included in diamond. *Nature* **335**, 784–789.
- MORGAN P. (1984) The thermal structure and thermal evolution of the continental lithosphere. In *Structure and Evolution of the Continental Lithosphere*, Vol. 15 (eds. H. N. POLLACK and V. R. MURPHY), pp. 107–193. Phys. and Chem. Earth.
- NAVROTSKY A. (1995) Thermodynamic properties of minerals. In *Mineral Physics & Crystallography: A handbook of physical constants* (ed. T. J. AHRENS), pp. 18–28. AGU.
- NIXON P. H. and BOYD F. R. (1973) Petrogenesis of the granular and sheared ultrabasic nodule suite in kimberlites. In *Lesotho Kimberlites* (ed. P. H. NIXON), pp. 48–56. Lesotho National Development Corporation.
- NYBLADE A. A. (1999) Heat flow and the structure of Precambrian lithosphere. *Lithos*, (in press).
- NYBLADE A. A. and POLLACK H. N. (1993) A global analysis of heat flow from Precambrian terrains: implications for the thermal structure of Archean and Proterozoic lithosphere. *J. Geophys. Res.* **98**, 12207–12218.
- NYBLADE A. A., POLLACK H. N., JONES D. L., PODMORE F., and MUSHAYANDEBU M. (1990) Terrestrial heat flow in East and Southern Africa. *J. Geophys. Res.* **95**, 17,371–17,384.
- PEARSON D. G., CARLSON R. W., SHIREY S. B., BOYD F. R., and NIXON P. H. (1995) The stabilisation of Archean lithospheric mantle: A Re-Os isotope study of peridotite xenoliths from the Kaapvaal craton. *Earth Planet. Sci. Lett.* **134**, 341–357.
- PEARSON N. J., GRIFFIN W. L., DOYLE B. J., O'REILLY S. Y., VAN ACHTERBERGH E., and KIVI K. (1998) Xenoliths from kimberlite pipes of the Lac de Gras area, Slave craton, Canada. In *7th Internat. Kimberlite Conf. Extended Abst.*, pp. 670–672. Univ. Cape Town.
- POKHILENKO N. P., SOBOLEV N. V., BOYD F. R., PEARSON D. G., and SHIMIZU N. (1993) Megacrystalline pyrope peridotites in the lithosphere of the Siberian platform: mineralogy, geochemical peculiarities and the problem of their origin. *Russ. Geol. Geophys.* **34**, 56–67.
- POLLACK H. N. (1980) The heat flow from the earth: a review. In *Mechanisms of Continental Drift and Plate Tectonics* (eds. P. A. DAVIES and S. K. RUNCORN), pp. 183–192.
- POLLACK H. N. and CHAPMAN D. S. (1977) On the regional variation of heat flow, geotherms, and the thickness of the lithosphere. *Tectonophysics* **38**, 279–296.
- RICARD Y., NATAF H.-C., and MONTAGNER J.-P. (1996) The

- three-dimensional seismological model a priori constrained: Confrontation with seismic data. *J. Geophys. Res.* **101**, 8457–8472.
- RICHARDSON S. H., ERLANK A. J., and HART S. R. (1985) Kimberlite-borne garnet peridotite xenoliths from old enriched subcontinental lithosphere. *Earth Planet. Sci. Lett.* **75**, 116–128.
- RINGWOOD A. E., KESSON S. E., HIBBERSON W., and WARE N. (1992) Origin of kimberlites and related magmas. *Earth Planet. Sci. Lett.* **113**, 521–538.
- RITSEMA J., NYBLADE A. A., OWENS T. J., LANGSTON C. A., and VANDECAR J. C. (1998) Upper mantle seismic velocity structure beneath Tanzania, East Africa: implications for the stability of cratonic lithosphere. *J. Geophys. Res.* **103**, 21,201–21,214.
- ROY R. F., BECK A. W., and TOULOUKIAN Y. S. (1981) Thermo-physical properties of rocks. In *Physical Properties of Rocks and Minerals* (eds. Y. S. TOULOUKIAN, W. R. JUDD, and R. F. ROY), pp. 409–502. McGraw-Hill.
- RUDNICK R. L. and FOUNTAIN D. M. (1995) Nature and composition of the continental crust: a lower crustal perspective. *Rev. Geophys.* **33**(3), 267–309.
- RUDNICK R. L., McDONOUGH W. F., and O'CONNELL R. J. (1998) Thermal structure, thickness and composition of continental lithosphere. *Chem. Geol.* **145**, 399–415.
- SASS J. H., LACHENBRUCH A. H., MOSES T. H., and MORGAN P. (1992) Heat flow from a scientific research well at Cajon Pass, California. *J. Geophys. Res.* **97**, 5017–5030.
- SAUTTER V., HAGGERTY S. E., and FIELD S. (1991) Ultra-deep (300 kilometers) ultramafic xenoliths: petrological evidence from the transition zone. *Science* **252**, 827–830.
- SCARFE C. M. and TAKAHASHI E. (1986) Melting of garnet peridotite to 13 GPa and the early history of the upper mantle. *Nature* **322**, 354–356.
- SCHATZ J. F. and SIMMONS G. (1972) Thermal conductivity of earth materials at high temperatures. *J. Geophys. Res.* **77**, 6966–6983.
- SCHULZE D. J. (1996) Kimberlites in the vicinity of Kirkland lake and Lake Timiskaming, Ontario and Québec. In *Searching for Diamonds in Canada*, Vol. 3228 (eds. A. N. LECHEMINANT, D. G. RICHARDSON, R. N. W. DILABIO, and K. A. RICHARDSON), pp. 73–78. Geological Survey Canada, Open File Rept.
- SHEE S. R. (1978) *The mineral chemistry of xenoliths from the Orapa kimberlite pipe, Botswana*. PhD Thesis, Univ. Cape Town.
- SHIMIZU N., POKHILENKO N. P., BOYD F. R., and PEARSON D. G. (1997) Geochemical characteristics of mantle xenoliths from the Udachnaya kimberlite pipe. *Proc. Sixth Int. Kimberlite Conf., Russian Geology and Geophysics* **38**, 205–217.
- 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: A Tribute to Francis R. (Joe) Boyd*, Vol. (this volume) (eds. Y. FEI, C. M. BERTKA, and B. O. MYSEN). The Geochemical Society.
- SMITH D. AND BOYD F. R. (1987) Compositional heterogeneities in a high-temperature lherzolite nodule and implications for mantle processes. In *Mantle Xenoliths* (ed. P. H. NIXON), pp. 551–561. John Wiley & Sons.
- SPETSISUS Z. V. and SERENKO V. P. (1990) *Composition of the continental upper mantle and lower crust beneath the Siberian platform*. Moscow "Nauka". (in Russian)
- STIEFENHOFER J., VILJOEN K. S., and MARSH J. S. (1997) Petrology and geochemistry of peridotite xenoliths from the Letlhakane kimberlites, Botswana. *Contrib. Mineral. Petrol.* **127**, 147–158.
- STUART G. W. and ZENGENI T. G. (1987) Seismic crustal structure of the Limpopo mobile belt, Zimbabwe. *Tectonophysics* **144**, 323–335.
- SU W.-J., WOODWARD R. L., and DZIEWONSKI A. M. (1994) Degree 12 model of shear velocity heterogeneity in the mantle. *J. Geophys. Res.* **99**, 6945–6980.
- TAYLOR S. R. and MCLENNAN S. M. (1985) *The Continental Crust: its Composition and Evolution*. Blackwell.
- VAN DER HILST R. D., SIMMONS F. J., and KENNETT B. L. N. (1998) Constraints on the structure of the upper mantle beneath Australia from waveform tomography. *EOS, Spring Meeting Abstracts*, S228.
- VAN DER LEE S. and NOLET G. (1997) Upper mantle S velocity structure of North America. *J. Geophys. Res.* **102**, 22815–22838.
- VICKER P. A. (1997) Garnet peridotite xenoliths from kimberlite near Kirkland Lake, Canada. M.S. Thesis, Univ. Toronto.
- WEAVER B. L. and TARNEY J. (1984) Empirical approach to estimating the composition of the continental crust. *Nature* **310**, 575–577.
- ZHURAVLEV A. Z., LAZ'KO Y. Y., and PONOMARENKO A. I. (1991) Radiogenic isotopes and REE in garnet peridotite xenoliths from the Mir kimberlite pipe, Yakutia. *Geokhimiya* **7**, 982–994.
- ZOTH G. and HAENEL R. (1988) Thermal conductivity. In *Handbook of Terrestrial Heat Flow Density Determination* (eds. R. HAENEL, L. RYBACH, and L. STEGENA), pp. 449–463. Kluwer Academic.