

Phase equilibrium constraints on the formation of cratonic mantle

CLAUDE HERZBERG

Department of Geological Sciences, Rutgers University, New Brunswick, NJ 08903, U.S.A. and Center for High Pressure Research and Mineral Physics Institute, State University of New York, Stony Brook, NY 11794, U.S.A.

Abstract—Experimental work at 3 to 8 GPa demonstrates that the melting of fertile mantle peridotite can yield a harzburgite residual mineralogy (L + Ol + Opx) and magmas similar to 2,700 Myr komatiites found on many cratons. The extraction of komatiite magma (46 to 49% SiO₂) from peridotite (45.6 % SiO₂) will yield complementary residues of dunite and olivine-rich harzburgite (41–45 SiO₂) with olivines having Fo_{91–94}. These experimentally-constrained residues are very similar in composition to xenoliths from Greenland, the Slave, Siberian, and Tanzanian cratons, and to many xenoliths from the Kaapvaal craton. Xenoliths possessing olivines with Fo_{92–94} could have formed as residues by about 38 to 50% melt extraction at pressures in the 3 to 6 GPa range, consistent with a plume origin for some cratonic mantle. Other xenoliths with olivines having Fo_{91–92} are similar to residues that had experienced 25 to 38% melt extraction at pressures less than 3 GPa, consistent with the formation of some cratonic mantle in an Archean MORB-like setting. Xenoliths with the lowest pressures of melt extraction typically exhibit the highest pressures of subsolidus equilibration, indicating that they were subducted to the bottom of a cratonic root. It is suggested that thickened cratonic mantle roots formed during accretion of oceanic plateaus and crust. Some peridotite xenoliths from the Kaapvaal craton are unusual in being too enriched in orthopyroxene to have been simple residues, and it is demonstrated that these could have formed as mixtures of olivine-rich residues and orthopyroxene-rich cumulates. Some peridotite xenoliths from Hawaii, Samoa and Tahiti also display Opx enrichment, indicating that Opx-enriched peridotite may be characteristic of plume associations, and that precipitation of cumulates occurs when hot mantle plumes exchange heat with a colder lithosphere.

INTRODUCTION

MANY PERIDOTITE xenoliths from the Kaapvaal, Siberian, Tanzanian, and Slave cratons differ from normal pyrolytic mantle in having smaller but variable amounts of clinopyroxene and garnet (NIXON and BOYD, 1973; BOYD and MERTZMAN, 1987; BOYD, 1987; 1989; BOYD *et al.*, 1997; BOYD and CANIL, 1997; LEE and RUDNICK, 1998). These rocks may be properly termed harzburgites (i.e., olivine + orthopyroxene rocks), and the few percent of clinopyroxene and garnet in them have been interpreted to be of exsolution or metasomatic origin (O'HARA *et al.*, 1975; COX *et al.*, 1987; CANIL, 1992; HERZBERG, 1993; BOYD *et al.*, 1997; GRIFFIN *et al.*, THIS VOLUME). Olivines are typically Fo_{92–94}, higher than olivines from peridotites found in ocean basins and mobile belts (BOYD, 1989; NIU *et al.*, 1997). These mineralogical differences are reflected in the major element geochemistry, which is depleted in Na₂O, CaO, Al₂O₃ and FeO, components that preferentially enter komatiite magmas. There is now substantial agreement that cratonic mantle and komatiites are complementary products of high degrees of melting of pyrolytic mantle (O'HARA *et al.*, 1975; MAALOE and AOKI, 1977; HANSON and LANGMUIR, 1978; BOYD and MERTZMAN, 1987; BOYD, 1987; 1989; TAKAHASHI, 1990; CANIL, 1992; HERZBERG, 1993; BOYD *et al.*, 1997; HERZBERG, THIS WORK; WALTER, 1998; WALTER, THIS VOLUME), a conclusion that has also been ex-

tended to harzburgite xenoliths of unknown age from East Greenland (BERNSTEIN *et al.*, 1997).

Modal orthopyroxene for many xenoliths from the Kaapvaal craton averages 30% (MAALOE and AOKI, 1977; BOYD, 1989), which is higher than would be expected for residues formed by the extraction of basalt or komatiite from pyrolytic (BOYD, 1989; KELEMEN *et al.*, 1992; HERZBERG, 1993). Kaapvaal xenoliths also have higher modal Opx than xenoliths from other cratons (BOYD *et al.*, 1997; BOYD and CANIL, 1997; LEE and RUDNICK, 1998), but additional data are needed to verify this, especially from Laurentia. These SiO₂-enriched Kaapvaal xenoliths could have formed by the unmixing of a residue into olivine- and orthopyroxene-rich domains (BOYD and MERTZMAN, 1987; BOYD, 1989; BOYD *et al.*, 1997), by cumulate processes (HERZBERG, 1993), by the introduction of silicic melts of subduction origin (RUDNICK *et al.*, 1993; KELEMEN and HART, 1996), and by other processes reviewed elsewhere (HERZBERG, 1993). The silica problem and depths and degrees of partial melting required to form komatiite and cratonic mantle have been reexamined in this paper using the most recently available data on experimental petrology at high pressures (HERZBERG and ZHANG, 1997, 1998; HERZBERG and O'HARA, 1998; WALTER, 1998). It is shown that most cratonic mantle xenoliths are simple harzburgite residues left after komatiite extraction, but many Kaapvaal peridotites are similar to mixtures of olivine-rich residues and orthopyroxene-rich cu-

mulates. Both a plume origin (HERZBERG, 1993; PEARSON *et al.*, 1995) and an Archean mid-ocean ridge basalt origin (TAKAHASHI, 1990) are consistent with the spectrum of depths and degrees of partial melting that are inferred for cratonic mantle.

EQUILIBRIUM MELTING OF FERTILE MANTLE PERIDOTITE

The composition of a liquid formed by equilibrium melting of a peridotite will change by the progressive elimination of crystalline phases along cotectics, and the path taken from 1 to 100% melting is critically dependent on both pressure and the composition of the peridotite source region (O'HARA, 1968; 1970; HERZBERG and O'HARA, 1998). There are two basic approaches to evaluating these liquid compositions. The first involves experimental work on a specific peridotite composition, as has been done by WALTER (1998) on a fertile peridotite in the 3 to 7 GPa range. The second approach is a more general one (HERZBERG and O'HARA, 1998) wherein liquids produced by the melting of any peridotite composition are evaluated from liquidus crystallization fields, cotectics and peritectics that have been independently determined for liquid, forsterite, orthopyroxene, clinopyroxene, and garnet phases in the analogue system CaO-MgO-Al₂O₃-SiO₂. Experiments span the 1 to 10 GPa range (PRESNALL *et al.*, 1979; WALTER and PRESNALL, 1994; MILHOLLAND and PRESNALL, 1998; HERZBERG, 1992; HERZBERG and ZHANG, 1997, 1998) and the effects of the components TiO₂, FeO, Na₂O, and K₂O have been evaluated (HERZBERG and ZHANG, 1997; HERZBERG and O'HARA, 1998). Special attention is focused on the harzburgite cotectic equilibrium [L + Ol + Opx] at 3 to 7 GPa, but the large experimental data base that exists at 1 to 3 GPa has been used as an important low pressure reference frame (HERZBERG and O'HARA, 1998); the cotectics shown in Fig. 1 at 3 and 4 GPa differ somewhat from the ones reported by HERZBERG and O'HARA (1998) in conforming to data most recently reported by WALTER (1998). The set of harzburgite saturation curves at 3 to 7 GPa in Fig. 1a were thus constrained by experimental data in CMAS and the Forsterite-Enstatite join (HERZBERG and ZHANG, 1998) together with experimental data of WALTER (1998). This general phase equilibrium method has been applied to a pyrolytic mantle composition listed in Table 1, and the results are shown in Fig. 1.

In this paper, the connection between cratonic mantle and komatiites has been quantitatively evaluated by using experimental petrology to model the melting of a fertile mantle source. This is a forward model, which requires that assumptions be made of the composition

of the peridotite source, and the type of komatiite involved. The peridotite composition shown in Table 1 was extracted as a representative composition of spinel lherzolite xenoliths in the data base of HERZBERG (1993), but differs slightly from that one used by WALTER (1998) in that Al₂O₃ is slightly lower. The composition of WALTER (1998) provides a better source for Gorgona komatiites of Cretaceous age (HERZBERG and O'HARA, 1998), but the peridotite composition chosen here provides a better source for 2,700 Myr Belingwe-type komatiites. There is a wide spectrum of CaO/Al₂O₃ reported for spinel lherzolites and primitive mantle compositions, but only those having ratios of around 0.93 can yield on advanced partial melting CaO/Al₂O₃ that is typical of cratonic mantle and komatiites from the 2,700 Myr Belingwe greenstone belt in Zimbabwe. This leads to the choice of komatiite composition. It may be argued that 3,500 Myr Barberton-type komatiites were more appropriate in the making of cratonic mantle, but there are several lines of evidence that argue against this: 1) most Barberton-type komatiites appear to have separated from a garnet-bearing residue (*e.g.*, HERZBERG, 1993; WALTER, 1998), unlike the low temperature cratonic mantle which is predominantly harzburgite; 2) the few examples of Barberton-type komatiites that exhibit a harzburgite REE signature may have required a peridotite source that was low in Al₂O₃ (CAWTHORN and STRONG, 1975; WALTER, 1998), and it can be demonstrated that complementary harzburgite residues would have had about 50% less Al₂O₃ than those observed for cratonic mantle; 3) Re-Os model ages for cratonic mantle from southern Africa are Archean; minimum estimates of the time of Re-depletion span the 2,400 to 2,900 Myr range (CARLSON *et al.*, 1998), and maximum estimates of mantle extraction ages which are only several hundred million years older (CARLSON *et al.*, 1998); this age bracket is similar to 2,700 Myr komatiites from the Belingwe greenstone belt in Zimbabwe.

Figure 1 illustrates that at pressures between 3.0 and 4.0 GPa, initial melting of mantle peridotite is Opx-bearing [L + Ol + Opx + Cpx + Gt; WALTER, 1998], but becomes Opx-free [L + Ol + Cpx + Gt] when the pressures and temperatures are suitably high for the stabilization of subcalcic Cpx. Liquids formed by advanced melting along the [L + Ol + Cpx + Gt] cotectic become increasingly enriched in SiO₂. Eventually Opx is stabilized by the reaction Ol + Cpx + Gt = L + Opx (O'HARA and YODER, 1967; HERZBERG *et al.*, 1990; LONGHI, 1995; HERZBERG and ZHANG, 1997, 1998; KINZLER, 1997; WALTER, 1998), and when this occurs, melting will once again involve garnet lherzolite [L + Ol + Opx + Cpx + Gt]. It is important to note, however, that Opx can participate during initial melting on the solidus if the source region is more refractory

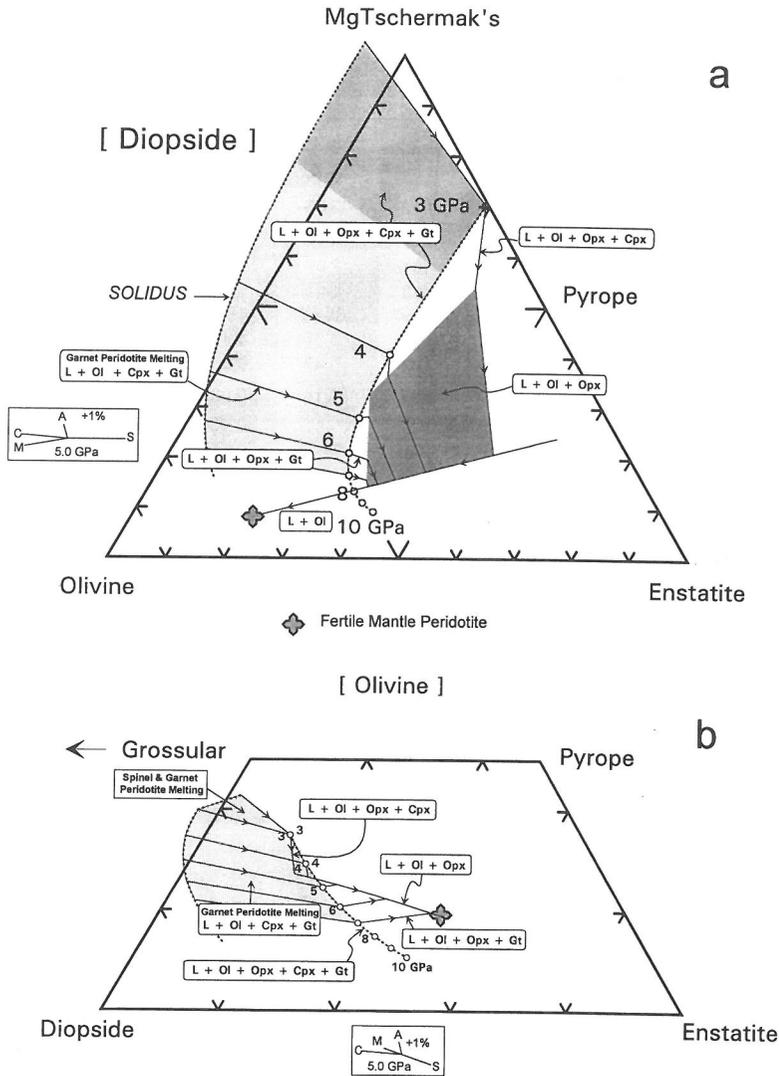


FIG. 1. Projections from Diopside (top) and Olivine (bottom) of model liquids formed by anhydrous equilibrium melting of fertile mantle peridotite (Table 1) from 3 to 10 GPa, modified from HERZBERG and O'HARA (1998). Enstatite-poor shaded region consists of the following low melt fraction assemblages: [L + Ol + Opx + Cpx + Gt], and [L + Ol + Cpx + Gt]. Liquids in equilibrium with [L + Ol + Opx] are most relevant to the understanding of komatiite and cratonic mantle. Small arrows on light lines point to isobaric liquid compositional change owing to increasing temperature. Insets illustrate the effect of adding 1 wt % of C, M, A, or S oxides to the 5 GPa invariant point.

than fertile mantle peridotite (HERZBERG and ZHANG, 1998).

Figure 1b shows that there is a special point at which the polybaric track of liquid compositions in equilibrium with garnet lherzolite [L + Ol + Opx + Cpx + Gt] will pierce the plane of compositions described by harzburgite melting [L + Ol + Opx]. This piercing point will determine the identity of the crystalline phase which is eliminated, and this is governed largely by pressure and CaO/Al₂O₃ of the

bulk composition (O'HARA, 1968; 1970; HERZBERG and O'HARA, 1998). For the peridotite source of interest, garnet is consumed from 3.0 to 4.6 GPa, and liquids are constrained by the cotectic [L + Ol + Opx + Cpx + Gt]. At pressures in excess of 4.6 GPa, clinopyroxene will be eliminated by advanced melting, and liquids will be saturated in garnet harzburgite [L + Ol + Opx + Gt].

When both clinopyroxene and garnet are consumed by progressive melting, the remaining liquid

Table 1. Compositions of Mantle Peridotite and Komatiite (wt%)

	This Study	Kettle River Peridotite ¹	Jagoutz SC1 ²	Griffin ³	Pyrolite ⁴	Hawaiian Xenolith ⁵	Belingwe Komatiite ⁶
SiO ₂	45.70	44.90	45.59	45.19	45.10	48.02	47.78
TiO ₂	0.22	0.15	0.22	0.14	0.20	0.22	0.37
Al ₂ O ₃	3.72	4.30	4.10	3.58	3.30	4.88	6.82
Cr ₂ O ₃	0.38	0.41	0.44	0.39	0.40	0.25	-
FeO	8.1	8.09	7.54	8.46	8.00	9.90	11.13
MnO	0.14	0.13	0.13	0.12	0.15	0.14	0.19
MgO	37.67	37.65	36.84	38.01	38.1	32.35	25.68
CaO	3.46	3.48	3.81	3.37	3.10	2.97	7.00
Na ₂ O	0.35	0.22	0.37	0.42	0.40	0.66	0.65
K ₂ O	0.02	0.09	0.02	0.14	0.03		0.05
NiO	0.24	0.24	0.24	-	0.20	0.20	-
Total	100	99.66	99.30	99.82	98.98	99.59	99.67
CaO/Al ₂ O ₃	0.93	0.81	0.93	0.94	0.94	0.61	1.03

1 WALTER (1998)

2 JAGOUTZ ET AL. (1979)

3 GRIFFIN (1973)

4 RINGWOOD (1979)

5 KUNO AND AOKI (1970)

6 NISBET *et al.* (1993) with Na₂O from HERZBERG and O'HARA (1998)

is saturated in harzburgite [L + Ol + Opx], the phase assemblage that is most relevant for understanding the origin of komatiite and cratonic mantle. Garnet is greatly stabilized at pressures in excess of about 7.5 GPa where the melting sequence [L + Ol + Opx] is replaced by [L + Ol + Gt] (HERZBERG and ZHANG, 1996; WALTER, 1998). This places an upper pressure limit on the formation of cratonic mantle as harzburgite residues [L + Ol + Opx] as discussed in a later section of this paper. For very advanced melting, the remaining liquid will be saturated in only olivine [L + Ol].

FORMATION OF CRATONIC MANTLE AS RESIDUES AND CUMULATES

Figure 2a shows the spectrum of liquid compositions, pressures, and degrees of batch melting of the peridotite source that will yield harzburgite residues [L + Ol + Opx]. At 3 GPa, harzburgite is a stable residuum mineralogy for 24 to 54% batch melting, and at higher pressures this melt fraction can extend to about 65%. Komatiites from the Belingwe greenstone belt in Zimbabwe (NISBET *et al.*, 1977, 1987; BICKLE *et al.*, 1975; 1993) display a wide range of compositions owing to olivine fractionation, but the parental magma composition estimated by NISBET *et al.* (1993) is very similar to a model batch liquid formed at about 4.5 GPa and at 50% melting (Fig. 2a); this is likely to be an aggregate of pooled liquids that formed throughout the melt column (HERZBERG

and O'HARA, 1998). One way to see through the effects of olivine fractionation is by projecting from olivine as seen in Fig. 2b. Now a second trend is observed, but this is described by variable amounts of orthopyroxene. This second trend is almost coincident with the projected compositions of model liquids that are saturated in harzburgite [L + Ol + Opx]; the slight mismatch may be an artifact of alteration (HERZBERG and O'HARA, 1998) or it may be reflecting the addition of a small amount of liquid extracted from garnet peridotite.

Komatiite liquids will separate from harzburgite and dunite residues with the olivine forsterite contents shown in Fig. 2a. These model residues are very similar to peridotite xenoliths from the Siberian craton (BOYD *et al.*, 1997), the Tanzanian craton (LEE and RUDNICK, 1998) and about half the xenolith population from the Kaapvaal craton (BOYD and MERTZMAN, 1987; BOYD, personal communication). Peridotites from the Kaapvaal craton seem to be more Opx-rich than peridotites from other cratons, in agreement with the conclusions reached by LEE and RUDNICK (1998) from work on xenoliths from the Tanzanian craton.

Liquids formed by about 24 to 38% melting will differ from Belingwe komatiites (Fig. 2a), and will yield harzburgite residues [L + Ol + Opx] that contain olivines with Fo_{90.7-92.0}; xenoliths with these olivine compositions are the high temperature types with equilibration temperatures in excess of 1100°C

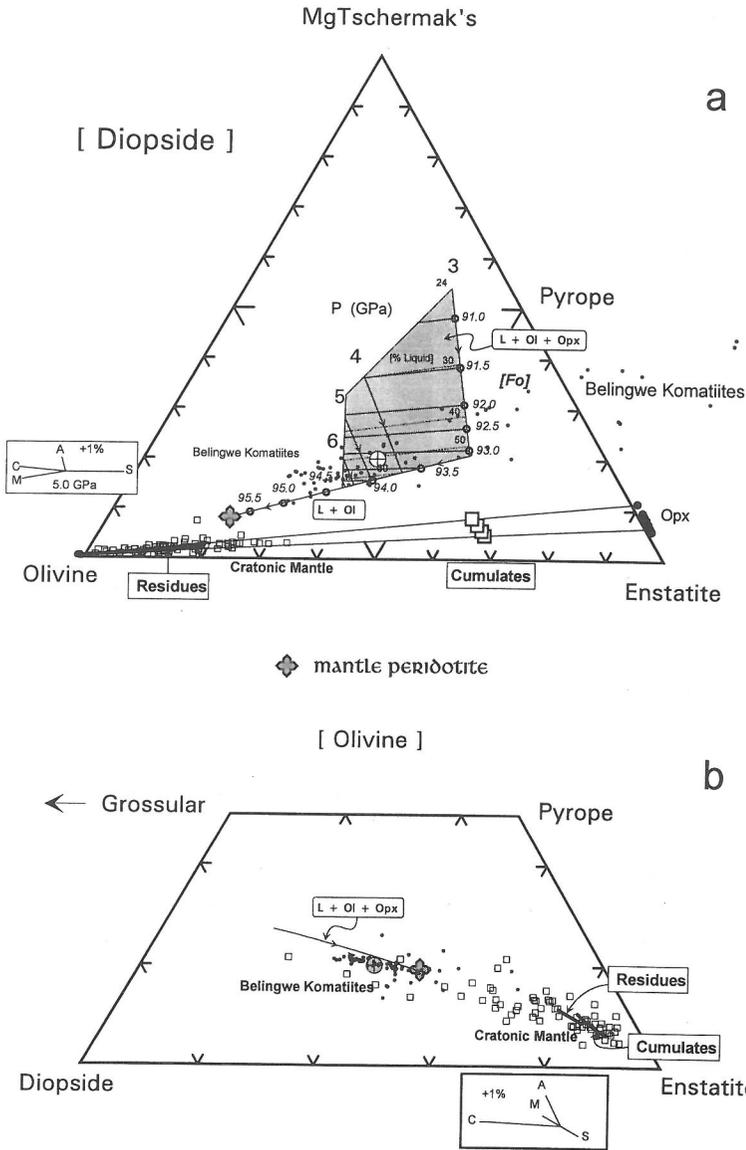


FIG. 2. Projections from Diopside (top) and Olivine (bottom) of model liquids with residual harzburgite [L + Ol + Opx] and dunite [L + Ol] together with Belingwe komatiites and peridotites from the Kaapvaal, Tanzanian, and Siberian cratons. Phase equilibria are from Fig. 1. Large filled cross = fertile mantle peridotite composition (Table 1); dots = Belingwe komatiites; cross-in-circle = estimated Belingwe parental magma (Table 1; NISBET *et al.*, 1993); open squares = cratonic mantle; shaded region = liquids in equilibrium with harzburgite [L + Ol + Opx] at 3 to 7 GPa; solid lines within shaded region = liquids that will crystallize residues with olivines having the Forsterite contents shown; broken lines in shaded region = melt fraction in %. Cumulates are precipitated at 3 GPa.

and pressures in excess of 5 GPa (FINNERTY and BOYD, 1987; BOYD *et al.*, 1997). Liquids formed by 38 to 65% melting are more similar to Belingwe komatiites, and have the potential to form harzburgite residues with olivines having Fo_{92.0-94.0}; xenoliths with these olivine compositions are the low temper-

ature types with equilibration temperatures and pressures that are less than 1100°C and 5 GPa, respectively (FINNERTY and BOYD, 1987; BOYD *et al.*, 1997).

The compositions of model liquids and residues in the projections of Figs. 2a and 2b are shown again as oxide plots in Figs. 3 and 4. Liquid compositions in

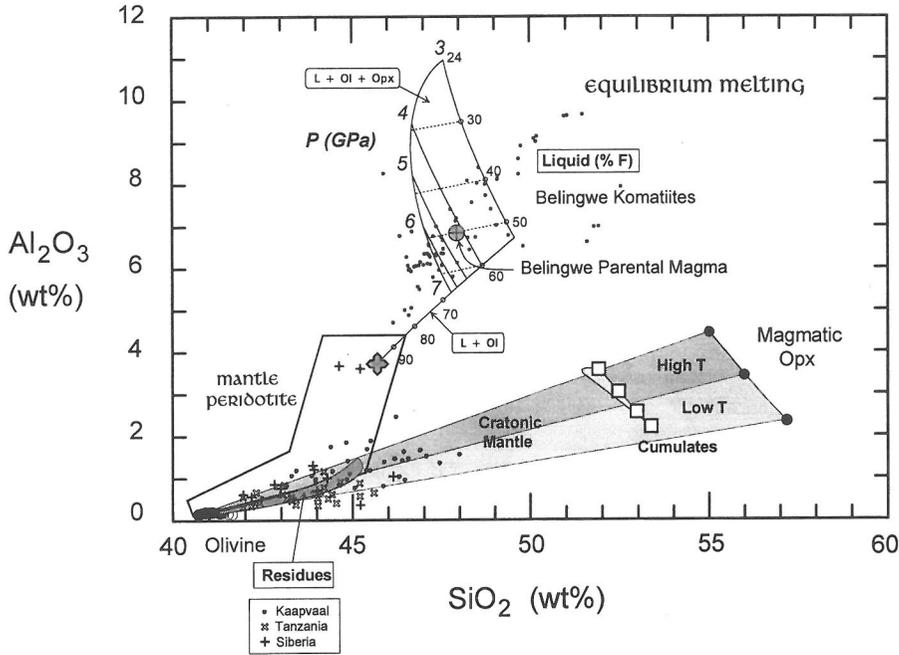


FIG. 3. Al₂O₃ and SiO₂ contents of cratonic mantle compared to model liquids, olivines, orthopyroxenes, residues, and cumulates formed by equilibrium melting and crystallization of mantle peridotite with residual harzburgite [L + Ol + Opx] and dunite [L + Ol] at 3 to 7 GPa. Large filled cross = fertile mantle peridotite composition (Table 1); small dots = Belingwe komatiites; cross-in-circle = estimated Belingwe parental magma (Table 1; NISBET *et al.*, 1993); short broken lines = melt fraction in %; cumulates are precipitated at 3 GPa; coexisting Ol and Opx are indicated by tie lines; Low T = low temperature equilibration for xenoliths having Fo₉₂₋₉₄ olivines. Field for off-craton "mantle peridotite" is from HERZBERG (1993).

equilibrium with olivine (L + Ol) and the experimentally-constrained isobaric harzburgite [L + Ol + Opx] cotectics were computed using an iterative procedure with the batch melting equation:

$$C_L = C_o / [F(1 - D) + D]$$

where C_L = weight% oxide, C_o = initial oxide concentration (Table 1), F = melt fraction, and D is the bulk distribution coefficient. The bulk distribution coefficient is the concentration of the oxide component in the multiphase residue divided by the concentration in the liquid, and takes the form:

$$D = C_s / C_L$$

D is determined from:

$$D = X_{Ol} D_{Ol}^i + X_{Opx} D_{Opx}^i$$

where X_{Ol} and X_{Opx} are the proportions of olivine and orthopyroxene in the residue, and D_i^{Ol/L} and D_i^{Opx/L} are the olivine/liquid and orthopyroxene/liquid partition coefficients, respectively. Mineral/liquid

partition coefficients were computed using a procedure developed by BEATTIE *et al.* (1991) and JONES (1995), which is:

$$D_i^{Ol/L} = A_i^{Ol/L} D_{MgO}^{Ol/L} + B_i^{Ol/L} \text{ and}$$

$$D_i^{Opx/L} = A_i^{Opx/L} D_{MgO}^{Opx/L} + B_i^{Opx/L}$$

A and B parameters listed in Table 2 were from BEATTIE *et al.* (1991) and from regressions of experimental data at pressures in excess of 3 GPa (TAKAHASHI and KUSHIRO, 1983; CANIL, 1992; TRØNNES *et al.*, 1992; HERZBERG and ZHANG, 1996; 1997; WALTER, 1998); the A and B parameters listed in Table 2 are an improved version of the ones used by HERZBERG and O'HARA (1998). For NiO modelling, the Ol/L and Opx/L partition coefficients of BEATTIE *et al.* (1991) were used. Although these are mostly 1 atmosphere determinations, BEATTIE *et al.* (1991) suggested that they provide a good description of experimental data up to 3.0 GPa, and this is seems consistent with preliminary experimental data (WALTER, personal communication).

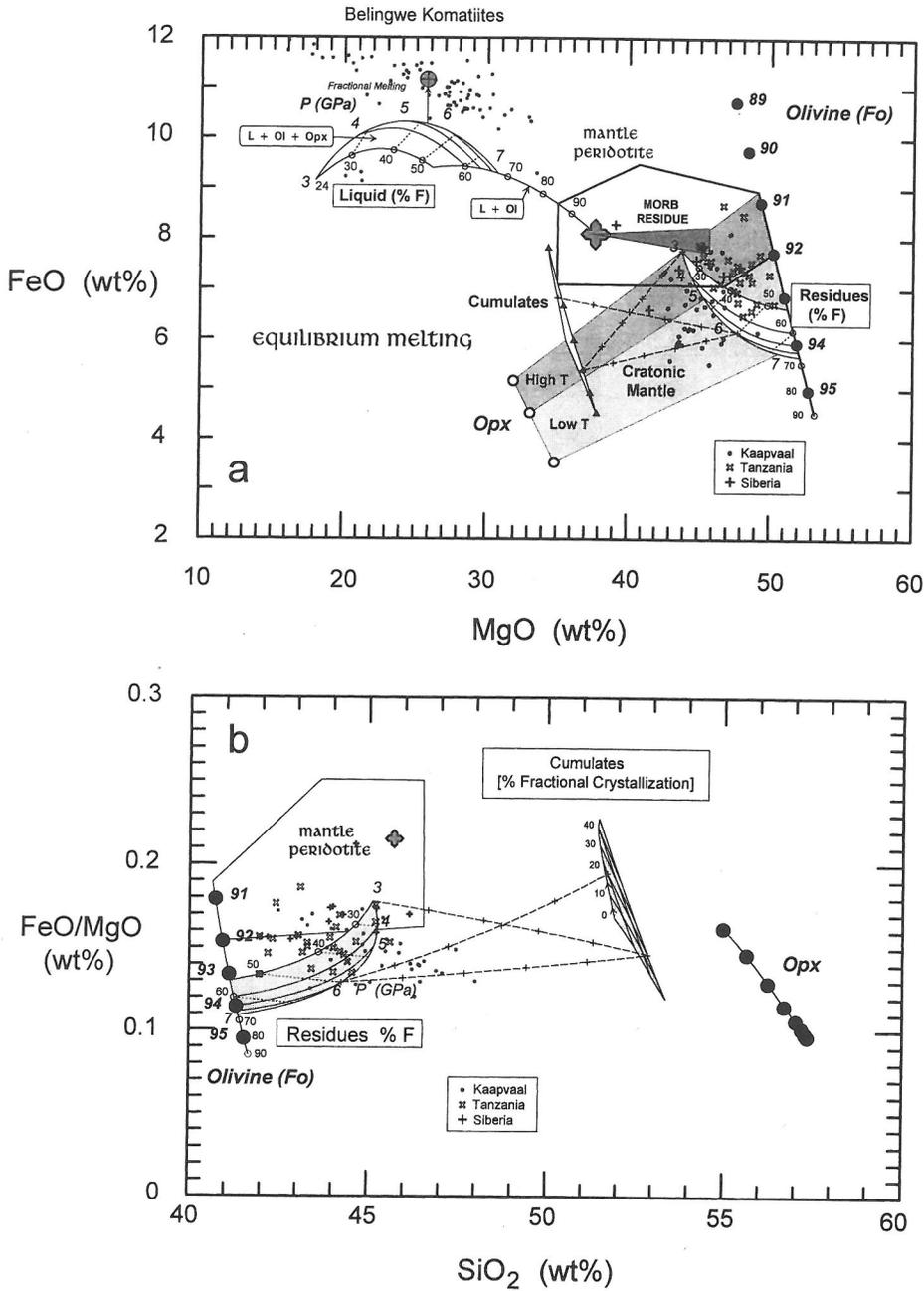


FIG. 4. FeO, MgO, and SiO₂ contents of cratonic mantle compared to model liquids, olivines, orthopyroxenes, residues, and cumulates formed by equilibrium melting and crystallization of mantle peridotite with residual harzburgite [L + Ol + Opx] and dunite [L + Ol] at 3 to 7 GPa. Large filled cross = fertile mantle peridotite composition (Table 1); small dots = Belingwe komatiites; cross-in-circle = estimated Belingwe parental magma (Table 1; NISBET *et al.*, 1993); short broken lines = melt fraction in %; cumulates are precipitated at 3 GPa; coexisting Ol and Opx are indicated by tie lines; long broken lines = mixing lines between cumulates and residues of various compositions, with 20% incremental tick marks. Model MORB residues are from NIU *et al.* (1997). Note that fractional melting provides a better description of the FeO content of komatiites than does equilibrium melting; see text for discussion.

Table 2. Empirical Constants for Computing Partition Coefficients for Ol/L and Opx/L at pressures in the 3 to 7 GPa range

Oxide i	A ^{Ol/L}	B ^{Ol/L}	A ^{Opx/L}	B ^{Opx/L}
TiO ₂	0	0.030	0.148	-0.152
Al ₂ O ₃	-0.022	0.065	0	0.400
Cr ₂ O ₃	0.218	-0.024	2.450	-2.334
MnO	<i>0.259</i>	<i>-0.049</i>	<i>0.352</i>	<i>-0.025</i>
CaO	<i>0.0056</i>	<i>0.0135</i>	-0.133	0.411
Na ₂ O	-0.024	0.075	-0.288	0.587
NiO	3.346	-3.665	1.206	-0.263

$$D_i^{Ol/L} = A_i^{Ol/L} D_{MgO}^{Ol/L} + B_i^{Ol/L}$$

$$D_i^{Opx/L} = A_i^{Opx/L} D_{MgO}^{Opx/L} + B_i^{Opx/L}$$

A & B constants in italics are from BEATTIE *et al.* (1991)

A & B constants for all other oxides from regressions of experimental data listed in text

$$K_{D_{FeO/MgO}}^{Ol/L} = 0.36 \text{ at 3 to 7 GPa (HERZBERG and ZHANG, 1996; 1996)}$$

$$K_{D_{FeO/MgO}}^{Ol/L} = 0.32 \text{ at 1 atmosphere (THY, 1995)}$$

$$K_{D_{FeO/MgO}}^{Opx/L} = 0.32 \text{ at 3 to 7 GPa (HERZBERG and ZHANG, 1997)}$$

Liquids formed by 24 to 50% melting of pyrolitic mantle are komatiites, and these contain 47.0 to 49.0% SiO₂ (Fig. 3). The residues they would generate from peridotite (45.76% SiO₂) are harzburgite and dunite with 41.3 to 45.2% SiO₂, similar to the harzburgite xenoliths from the Siberian craton (BOYD *et al.*, 1997), the Tanzanian craton (LEE and RUDNICK, 1998) and about half the peridotite xenoliths from the Kaapvaal craton (BOYD and MERTZMAN, 1987; BOYD, PERSONAL COMMUNICATION). Data from the Slave Craton (BOYD and CANIL, 1997) and from east Greenland (BERNSTEIN *et al.*, 1997) have not been plotted because they are modal rather than whole rock, but the high olivine contents in these modes are consistent with them being simple residues. In detail, a xenolith with an independently determined olivine Fo content of 92 to 94 will sometimes plot in the model high T fields shown in Figs. 3 and 4a (Fo < 92), although the misfits are minor. This is because whole rock data were variably modified by metasomatism (CANIL, 1992; SMITH and BOYD, 1987; GRIFFIN *et al.*, 1989).

Orthopyroxene-rich harzburgite is mostly characteristic of the Kaapvaal craton. Most xenoliths from the Siberian craton are residues, in agreement with the conclusions reached by WALTER (1998; this volume), and only 1 or 2 samples display Opx-enrichment (Fig. 4). All of the xenoliths from the Tanzanian craton (LEE and RUDNICK, 1998) appear to be simple

residues (Figs. 3 and 4). The average of 372 low and high temperature Kaapvaal peridotites in the data base compiled by HERZBERG (1993) is 45.62% SiO₂; there are 28 samples larger than 0.5 kg in the low temperature data base used by BOYD (1989), and the average is 46.57% SiO₂. Herein lies the silica problem for the Kaapvaal craton. About half the peridotite population is enriched in SiO₂ compared with model residues, and some of the more Opx-rich samples contain about 48% SiO₂ (Figs. 3 & 4b). Residue and cumulate compositions have been modeled, and these are included in Figs. 3 and 4. Cratonic mantle from the Kaapvaal craton can be better described as mixtures of olivine-rich residues and orthopyroxene-rich cumulates. *These mixtures include both low and high temperature peridotites*, although the high temperature peridotites are not as orthopyroxene-rich as the low temperature types.

In order to model the excess Opx observed in peridotite xenoliths from the Kaapvaal craton, the cumulate compositions shown in Figs. 3 and 4 have been computed for 0 to 50% fractional crystallization of olivine-orthopyroxene at 3 GPa. This choice of pressure is appropriate for the base of a preexisting lithosphere where heat may have been extracted from segregated komatiite magmas, and the ratio of Opx/Ol that precipitates is constrained by experiment to be about 3.7:1.0. A number of mixing scenarios are

possible. Partial crystallization in excess about 30% will yield cumulates with FeO/MgO that is elevated above that observed for cratonic mantle, precipitating olivine-orthopyroxenites with Fo contents less than 91. Cumulates with Fo_{90.5} olivines could have mixed with residues having Fo₉₄ olivines, and the resultant mixture could have reequilibrated as an Opx-rich peridotite with Fo₉₃ olivines. A similar mixed rock could have formed as cumulates with Fo₉₄ olivines mixed with residues with Fo₉₂ olivines. Opx-enriched cratonic mantle is most consistent with cumulates formed by about 20% fractional crystallization, although this can be as high as 40% in limited cases.

Mixing lines between candidate residues and cumulates are shown in Figs. 3 and 4. The amount of cumulate component in individual Kaapvaal xenoliths varies considerably from 0%, the most common case, to a maximum of about 40%. It is clear that pure cumulates do not exist, and this is attributed to efficient mixing, possibly by thermal convection, which might have been vigorous in a plume-type scenario with high temperature gradients and low viscosities. Cumulates that formed by about 20% fractional crystallization of a komatiitic magma that itself was generated by about 50% partial melting would contribute about 20% by mass to the mixed rock if mixing was thorough. If mixing was not thorough, preservation of a partially mixed cumulus component might explain the rare xenoliths with Opx-rich bands (e.g., sample LBM17; Cox *et al.*, 1973).

Larger mass fractions of melt extraction yield harzburgite residues with lower FeO contents as modeled in Fig. 4. The FeO content of cratonic mantle is therefore a guide to the extent of melting, although this can be subject to large errors in cases where FeO was metasomatically introduced. The entire xenolith population from the Kaapvaal and Tanzanian cratons is similar to model residues formed by about 25 to 50% melt extraction, and the low temperature types are most consistent with 38 to 50% melting. Xenoliths from the Siberian craton experienced lower degrees of melting, typically the 25 to 38 % range, in good agreement with those of WALTER (1998; this volume). It should be noted that xenoliths from all cratons that display 25 to 38% melting must have formed as residues by the extraction of a liquid that is fundamentally different from Belingwe komatiites (Fig. 2a, 4); these would have been picritic with MgO < 20% and FeO = 9 to 10%.

Pressure information of melt extraction can also be obtained from Fig. 4, although large uncertainties can arise by small cumulate additions of olivine or orthopyroxene to the residue. A MORB residue formed by melt extraction at pressures less than 2.5 GPa will

contain about 8% FeO, essentially the same as the source (NIU *et al.*, 1997). Melt extraction at higher pressures yields residues with lower FeO contents, and the model residues shown at 3 to 7 GPa in Fig. 4 are in good agreement with the model residues calculated by NIU *et al.* (1978) to 2.5 GPa. For xenoliths from all cratons, pressures of melt extraction in excess of 6 GPa are never observed. Low temperature and low FeO xenoliths from the Tanzanian craton record pressures of 3 to 6 GPa, but the more FeO-rich types have the residuum properties of melt extraction at pressures that were less than 3 GPa; this holds true for high temperature xenoliths from the Kaapvaal and Siberian cratons, as originally suggested by WALTER (1998). This gives rise to an interesting paradox that is discussed more fully below; that is, many high temperature xenoliths formed as residues by melt extraction at pressures that were less than 3 GPa, but they reequilibrated in the subsolidus at pressures in excess of 5 GPa.

There is a positive relationship between NiO and MgO in off-craton pyrolytic mantle, reflecting the compatible nature of NiO during partial melting. From 542 peridotites compiled in HERZBERG (1993), this relationship takes the form:

$$\text{NiO (wt\%)} = 0.0096\text{MgO (wt\%)} - 0.125$$

with an uncertainty of ± 0.03 at the 2 σ level. This relationship yields 0.24 ± 0.03 % NiO for the peridotite source considered here. The NiO contents of residues and cumulates are shown in Fig. 5a, and again cratonic mantle can be described as mixtures of these two components. For olivine and orthopyroxene precipitated from komatiite magmas at temperatures in excess of 1600°C, $D_{\text{NiO}}^{\text{Ol/Opx}}$ is computed to be 1.8 to 2.3 from the partition coefficients of BEATTIE *et al.* (1991). Partition coefficients of NiO between olivine and orthopyroxene have also been estimated from xenolith data where temperatures of subsolidus equilibration have been inferred (BOYD, 1997; BODINIER *et al.*, 1987). At about 1000°C the $D_{\text{NiO}}^{\text{Ol/Opx}}$ is typically around 3.8. Cooling of olivine-orthopyroxene assemblages from 1600° to about 1000°C will impose a rotation in the Ol-Opx tie lines as seen in Fig. 5b, resulting in olivines in Opx-rich cratonic mantle with elevated NiO. KELEMEN and HART (1996) and KELEMEN *et al.* (1998) have attributed the elevated NiO in olivine to replacement by orthopyroxene which precipitated from silicic small degree melts in subduction zones. However, elevated NiO in olivines in Opx-rich harzburgite is a basic requirement of equilibrium, and cannot be used to identify the specific geological process from which it originated.

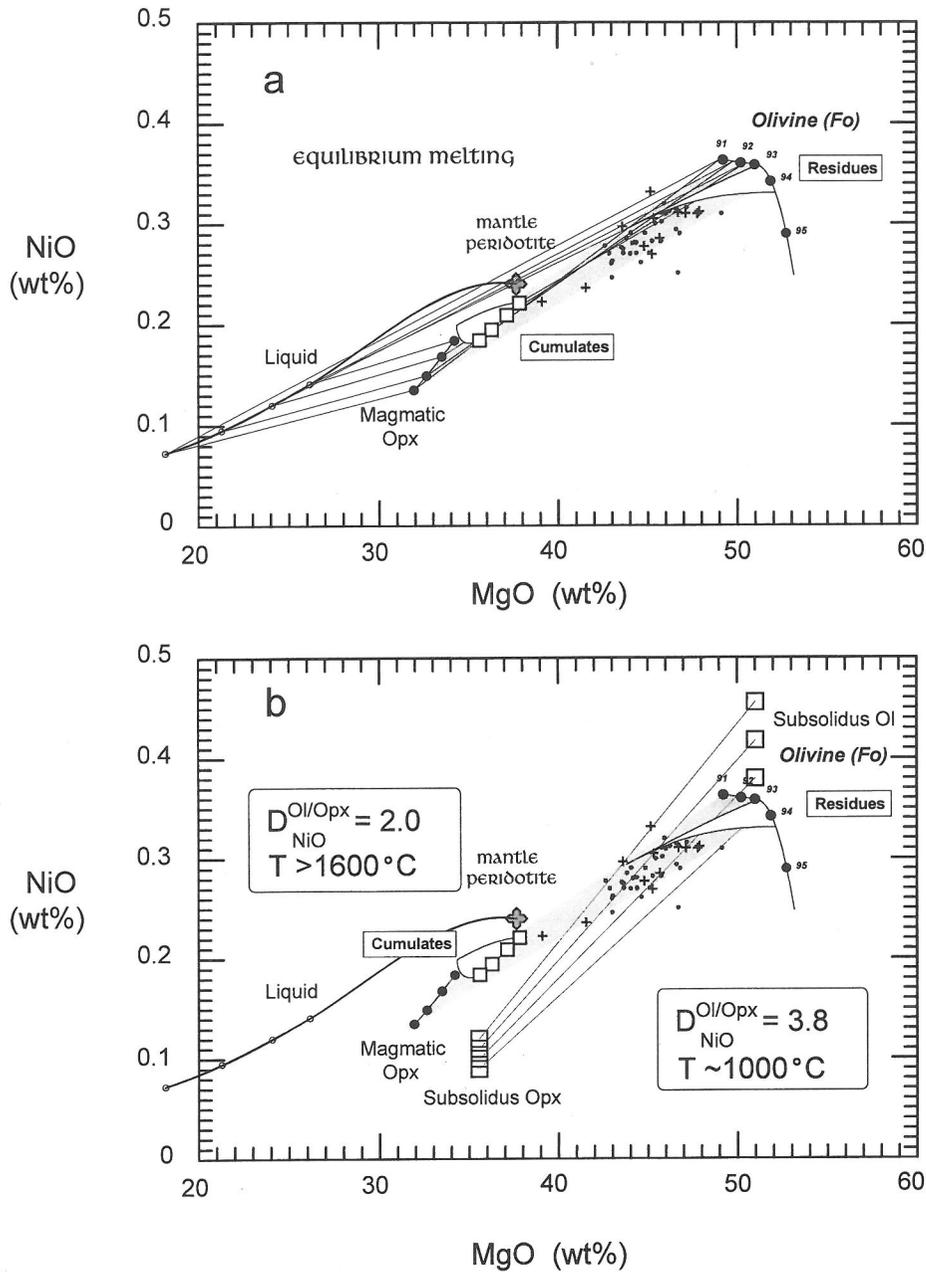


FIG. 5. a) NiO and MgO contents of cratonic mantle compared to liquids, olivines, orthopyroxenes, and residues formed by equilibrium melting of fertile peridotite with residual harzburgite [L + Ol + Opx] and dunite [L + Ol] at 3 to 7 GPa. Cumulates are precipitated at 3 GPa; dots = xenoliths from the Kaapvaal craton (BOYD and MERTZMAN, 1987; BOYD, PERSONAL COMMUNICATION); small crosses = xenoliths from the Siberian craton (BOYD *et al.*, 1997). Many peridotite residues from off-craton occurrences also plot in the shaded region. b) squares = coexisting olivines and orthopyroxenes at low temperatures; filled circles = coexisting olivines and orthopyroxenes at magmatic temperatures.

As discussed above, liquid compositions shown in Figs. 3, 4 and 5 were computed with the batch melting equation. The advantage of this procedure is that

it can be readily modified and applied to more geologically relevant melting processes by replacing the batch melting equation with an accumulated frac-

tional melting equation (i.e., AFM; LANGMUIR *et al.*, 1992). A fractional melting model will be reported separately because this is work in preparation, and its discussion here would greatly expand the scope of this paper. However, readers should note that the following two general conclusions have been obtained:

1) the AFM equation provides a somewhat better description of geochemistry of komatiites than does the batch melting equation, especially for FeO (Fig. 4a) and NiO; however, the differences are not very great, particularly for the other oxides; 2) the AFM equation yields residues that are more refractory than residues depicted in Figs. 3 and 4 using the batch melting equation; however, addition of about 10% komatiite yields residues that are similar to those produced by equilibrium melting. In summary, equilibrium melting provides a good but incomplete description of the geochemistry of komatiites and complementary cratonic mantle.

ARCHEAN PLUME CONTRIBUTIONS TO CRATONIC MANTLE ROOTS

The plume model has become increasingly successful since its inception because it provides the high degrees of melting and high pressures required for the formation of komatiite (FYFE, 1978; JARVIS and CAMPBELL, 1983; ARNDT, 1986; CAMPBELL and GRIFFITHS, 1990, 1992; ARNDT and LESHER, 1992; HERZBERG, 1992; 1995; BICKLE, 1993; NISBET *et al.*, 1993; MCDONOUGH and IRELAND, 1993; ABBOTT *et al.*, 1994; ARNDT *et al.*, 1997; HERZBERG and O'HARA, 1998; WALTER, 1998). The residues of these komatiites are very similar to peridotite xenoliths from the Kaapvaal and Siberian cratons (O'HARA *et al.*, 1975; HANSON and LANGMUIR, 1978; BOYD, 1989; TAKAHASHI, 1990; CANIL, 1992; HERZBERG, 1993; PEARSON *et al.*, 1995; BOYD *et al.*, 1997; WALTER, 1998; this volume), so it is reasonable to infer a plume origin for cratonic mantle (HERZBERG, 1993; PEARSON *et al.*, 1995).

Shown in Fig. 6 is a model cross-section through a plume as it may have appeared during an encounter with a preexisting lithosphere. The petrological structure stems from the phase relations for peridotite as reported by HERZBERG and O'HARA (1998), and consists of an envelope of low degree garnet peridotite melting and an axis of more advanced harzburgite melting. Much of the plume axis and head melts to form komatiite, and this will erupt and differentiate to form thickened crust of basalt and komatiite. Harzburgite is the most common residual mineralogy for komatiites with 2,700 Myr and Cretaceous ages (HERZBERG, 1992; 1995; HERZBERG and ZHANG, 1998;

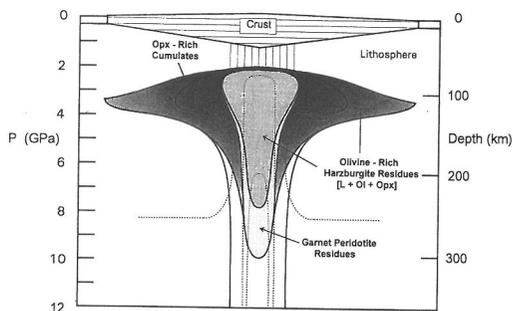


FIG. 6. Thermal and petrological model of a plume that might have contributed to the formation of the Kaapvaal craton, modified slightly from HERZBERG and O'HARA (1998). Initial melting of the plume axis will generate low degree melts with a garnet peridotite residuum; advanced melting during decompression will yield komatiites with a harzburgite residuum [L + Ol + Opx]. Dark bands = orthopyroxene-rich cumulates, which thin and disappear by convective mixing. Continuous tone grey scale = plume residues and cumulates; dark tone = highest Opx/Ol; light tone = lowest Opx/Ol. Dotted lines = isotherms at arbitrary temperatures.

HERZBERG and O'HARA, 1998; WALTER, 1998), and it stayed behind to form cratonic mantle. The advantage of this plume model is that it provides a structural context for generating large volumes of komatiite and complementary harzburgite by minimizing the participation of garnet. Liquids separated from a garnet-bearing residue have a distinctive trace and major element geochemistry which is not observed in the geochemistry of 2,700 Myr komatiites or the primary mineralogy of complementary cratonic mantle (*e.g.*, HERZBERG, 1992; HERZBERG and O'HARA, 1998; HERZBERG and ZHANG, 1998; WALTER, 1998). It is the low FeO (Fig. 4a) of harzburgite that has made cratonic mantle buoyant, and this has kept it isolated from the rest of the convecting mantle over most of the age of the Earth (BOYD and McALLISTER, 1976; GRIFFIN *et al.*, this volume; JORDAN, 1978; BOYD, 1989; HERZBERG, 1993; ABBOTT *et al.*, 1997; O'REILLY *et al.*, 1998; LEE and RUDNICK, 1998). The plume harzburgite would have eventually replaced the original lithosphere by buoyancy-driven diapirism, and this may have occurred early because thermal contributions to the density of harzburgites would have added to compositional buoyancy.

Komatiite magmas must have gravitationally separated from their residues to erupt as surface flows that can now be observed on some Archean cratons. The fundamental question is what happened to these magmas as they moved from the source to the surface. If heat was exchanged by interaction of hot plume magmas with a cold lithosphere, this would have been manifest by partial crystallization of the

komatiite magmas at depth. It is suggested that some xenoliths from the Kaapvaal craton developed Opx-enrichment by magma segregation and precipitation of orthopyroxene cumulates. A parallel process has now been identified for explaining the geochemistry of abyssal peridotite dredged from the world's ocean basins (NIU *et al.*, 1997). These rocks differ from simple MORB residues in being enriched in olivine, and the excess olivine may have been added to the residues by partial crystallization of basaltic magmas as they ascended through cooler lithosphere (NIU *et al.*, 1997).

In addition to vertical ascent, komatiite magmas may have segregated laterally from the plume periphery to core as depicted in Fig. 6 if the plume-lithosphere interface was inclined and if the lithosphere was partially impermeable. Partial crystallization in this situation could have resulted in the distribution of orthopyroxene cumulates closest to the plume axis, resulting in primary Ol/Opx heterogeneities with plume dimensions. In the plume model of Fig. 6, cumulate rocks would exist at the top and core of the plume, and these would be destroyed as discrete entities by convective mixing with residues. But limited or inefficient convective mixing could have preserved primary mineralogical heterogeneities, resulting in banded xenoliths and craton-scale variations in average modal Ol/Opx. Average Kaapvaal-type mantle that is enriched in Opx may originate from a fossil

plume core. Olivine-rich lithospheric mantle from the Slave craton (BOYD and CANIL, 1997), Greenland (BERNSTEIN *et al.*, 1997), Siberia (BOYD *et al.*, 1997), and Tanzania (LEE and RUDNICK, 1998) appear to be simple residues without cumulate contributions, or they may have originated from a fossil plume periphery. The present data base seems to indicate that Opx-rich mantle from the Kaapvaal craton may be unrepresentative of cratons elsewhere, but more work is needed to evaluate this important possibility.

High temperature peridotite xenoliths from the Kaapvaal craton are more diverse mineralogically than the low temperature types. In addition to simple Ol-Opx mixtures, many are garnet lherzolites with a complex metasomatic history (*e.g.*, PHN 1611; SMITH and BOYD, 1987; GRIFFIN *et al.*, 1989). Reference to Fig. 4 illustrates that some garnet lherzolites are similar to residues from melt extraction at pressures that were less than 3 GPa, a suggestion made previously (WALTER, 1998). But some high temperature xenoliths are also too Opx-rich to be simple residues. Orthopyroxene-rich xenoliths have also been described from Hawaii and Samoa, and these are shown in Fig. 7 and Table 1. The Hawaiian peridotites are whole rock data published by JACKSON and WRIGHT (1970) and KUNO and AOKI (1970). These are similar to modal data which show up to 40% Opx in spinel lherzolites from Hawaii (SEN, 1988), as do xenoliths from Samoa (HAURI and HART, 1994) and Tahiti

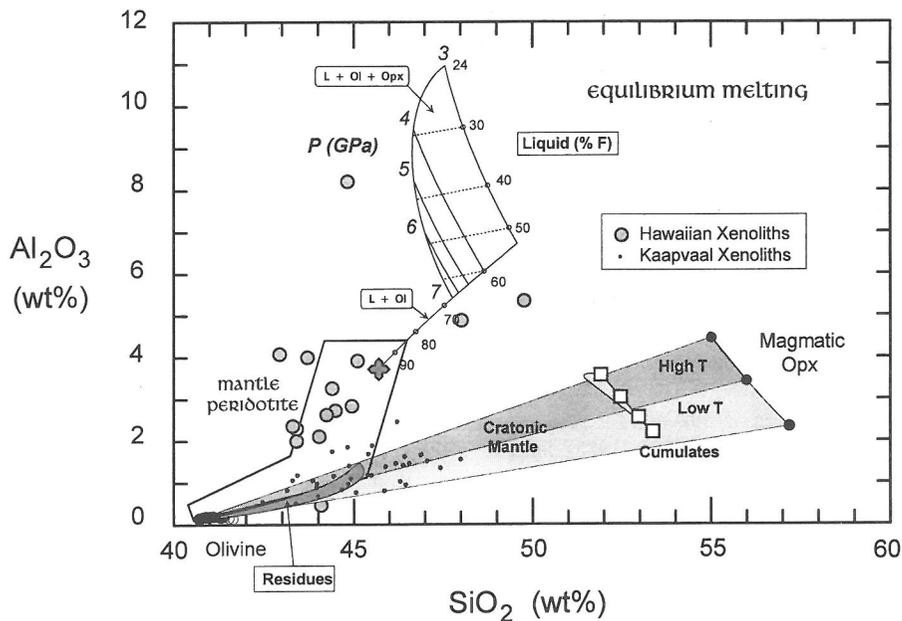


FIG. 7. Compositions of peridotite xenoliths from Hawaii compared to xenoliths from the Kaapvaal craton, showing the wide range of Opx:Ol and elevated Opx, after Fig. 3.

(TRACY, 1980). It is suggested that peridotite xenoliths with high modal Opx are characteristic of magmatism in plumes.

In a previous paper it was suggested that the Kaapvaal peridotite xenoliths may have formed in a gigantic plume that was substantially hotter than most Archean plumes from which komatiites were derived (HERZBERG, 1993), but I no longer believe this to be true. This suggestion was based on the calculations of MILLER *et al.* (1991) which required partial melting to initiate at pressures in excess of 20 GPa in order for melting to reach 50%. These calculations were based on solidus temperatures of TAKAHASHI (1986) which were later shown to be too low by about 200°C at 15 GPa (ZHANG and HERZBERG, 1994; SHIMAZAKI and TAKAHASHI, 1993). It is more likely that melting would have initiated at about 8 to 10 GPa, similar to those which have been estimated for Archean komatiites with 2,700 Myr ages (Fig. 6; HERZBERG and O'HARA, 1998). These komatiites crystallized as lava flows at the surface, producing olivines with maximum forsterite contents in the range $Fo_{93.6-94.5}$ (NISBET *et al.*, 1993). At pressures in excess of 3 GPa, these same komatiites would have separated from harzburgite or dunite residues having olivines with maximum forsterite contents of $Fo_{92.9-93.8}$, based on the effect of pressure on increasing $K_{D}^{OIL/FeO/MgO}$ from 0.32 at 1 atmosphere to 0.36 at higher pressures (THY, 1995; HERZBERG and O'HARA, 1998). Olivines in the source regions of 2,700 Myr Archean komatiites were therefore very similar to those of cratonic mantle, indicating a similarity in the thermal characteristics of these plumes. Gigantic plumes (HERZBERG, 1993), impacts (HERZBERG, 1993), or whole-mantle overturns (BOYD *et al.*, 1997) would be expected to produce near-total melts for which olivines with Fo_{95-96} might precipitate, but these are never observed.

ARCHEAN OCEAN RIDGE CONTRIBUTIONS TO CRATONIC MANTLE ROOTS

Thermobarometry places a minimum thickness of 220 km for the lithosphere below the Kaapvaal craton (FINNERTY and BOYD, 1987; CARLSON *et al.*, 1998), and seismic studies indicate depths of 300 to 400 kilometers (JORDAN, 1978; 1988; NOLET *et al.*, 1994). Phase equilibrium studies indicate that 100 kilometers of residual cratonic mantle could have formed in a plume (Fig. 6), but this estimate is fraught with uncertainties. Thicker roots could have accumulated in a plume head that was continuously fed from below. Conversely, plume heads can detach from their conduit (VAN KELEN, 1997) and result in thinner roots. Fast lithospheric plate motions may also regu-

late maximum root thickness by dispersing the residues downstream, as modeled by PHIPPS MORGAN *et al.* (1995) for Hawaii.

Thick cratonic mantle could also have grown by piggyback underthrusting of thinner units, a mechanism that would be consistent with current models that view the growth of cratons by accretion of oceanic crust and plateaus (STOREY *et al.*, 1991; DE WIT *et al.*, 1992; DESROCHERS *et al.*, 1993; KIMURA *et al.*, 1993; STEIN and GOLDSTEIN, 1996; PUCHTEL *et al.*, 1998; ABBOT *et al.*, 1997). Underthrusting was originally suggested by JORDAN (1988) and is depicted in Fig. 8. Whereas thin lithosphere is subductable, thick lithosphere is not (ABBOTT *et al.*, 1997); between these possibilities may exist lithosphere dimensions that are too buoyant to be subducted into the lower mantle, but are sufficiently dense to be stacked below an existing mantle root. If cratonic mantle extends to depths of 300 to 400 km as inferred from seismological studies (JORDAN, 1978; 1988; NOLET *et al.*, 1994), and if it consists of harzburgite that has not been sampled by kimberlitic volcanism, then underthrusting is essential because harzburgite cannot be a residue of pyrolite by *in situ* magmatic processing at these pressures; instead, residues and cumulates would consist of an assemblage of garnet + olivine (HERZBERG and ZHANG, 1996; WALTER, 1998) and the substantial garnet content would be expected to impart a high velocity signal that is not observed.

The underthrusting model shown in Fig. 8 resolves the pressure paradox of the high temperature xenoliths. Thermobarometry indicates a pressure of subsolidus equilibration in excess of 5 GPa (FINNERTY and BOYD, 1987; BOYD *et al.*, 1997), and yet these have compositions that are more consistent with melt extraction at pressures that were less than 3 GPa (Fig. 4; WALTER, 1998). Subduction of these relatively iron-rich low-pressure residues is a simple way of resolving this paradox. Indeed, these low pressure residues are very similar in composition to those formed by the extraction of ocean ridge basalt at pressures less than 2.5 GPa (Fig. 4a; NIU *et al.*, 1997). TAKAHASHI (1990) proposed that cratonic mantle was formed as residues after extraction of a hot and thickened Archean oceanic crust, and this model might be a successful way of understanding both the high temperature peridotite xenoliths and eclogite xenoliths of crustal origin (SHIREY, *personal communication*). However, pervasive metasomatism presents a challenge to the understanding of the original peridotite protolith (SMITH and BOYD, 1987; GRIFFIN *et al.*, 1989), and this model needs to be examined more carefully.

Thickening Cratonic Roots by Accretion of Oceanic Plateaus & Crust

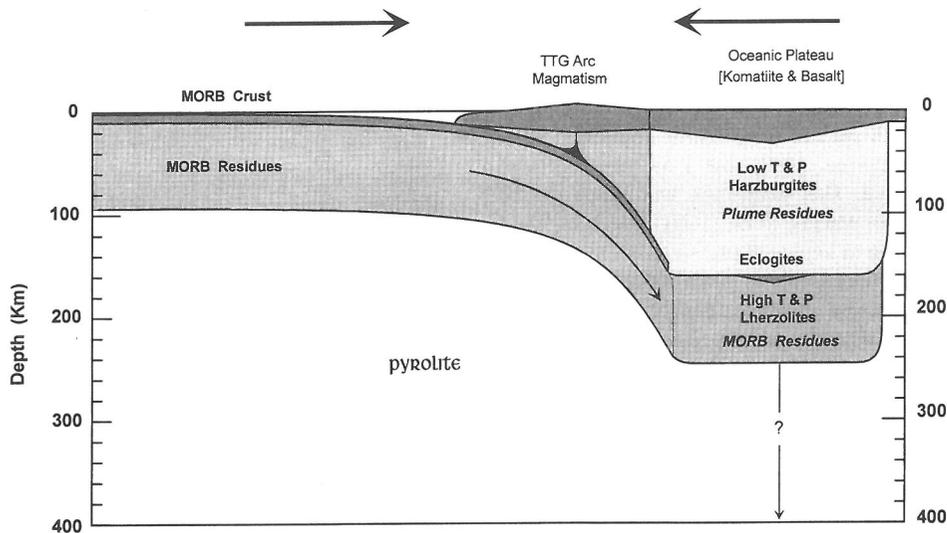


FIG. 8. Archean accretion model for thickening cratonic roots, after JORDAN (1988). Low T & P harzburgites are plume residues; these may have formed by a single plume that was fed continuously from below, or by accretion of multiple roots formed by multiple plumes. High T & P lherzolites are Archean MORB residues, after TAKAHASHI (1990), although extensive metasomatism may complicate this interpretation.

WHERE IS ALL THE KOMATIITE ?

Komatiites are not found in all cratons, and where they occur basalts are a more common lithology. So if komatiites are formed by about 25 to 50% melting of pyrolitic mantle, and the residues occupy cratonic roots with a minimum thickness of roughly 100 km, then about 35 to 100 km of complementary Archean crust should have been produced. Where is all the komatiite? We have a paradox of mass balance requirement and geological observation. The high degrees of melting required for a komatiite—cratonic mantle mass balance with all plausible peridotite source compositions has been recognized for 20 years (HANSON and LANGMUIR, 1978; TAKAHASHI, 1990; HERZBERG, 1993; HERZBERG, this work; WALTER, 1998; WALTER, this volume), and holds true even though controversy remains over such questions as the role of water in komatiite genesis (ARNDT *et al.*, 1998). Therefore, the easiest way to resolve this paradox is by elimination of komatiite from the geological record. Although the destruction of evidence used in the defense of a theory is clearly undesirable, the precedents of missing information in the geological record are numerous and well-known. For every komatiite flow that is observed, it is reasonable to conjecture that greater quantities of komatiite were emplaced at depth in a thickened plateau by under-

plating, and that the crystalline differentiates could have been removed by some form of subduction.

A role for underplating is indicated because the density of FeO-rich komatiite magmas is high and comparable to the density of crystalline gabbro at about 1 GPa (HERZBERG *et al.*, 1983). Komatiites may have underplated the base of a thickened Archean oceanic plateau where they differentiated in sills to olivine cumulates and basalt, similar to models proposed for picritic magmas (COX, 1980; HERZBERG *et al.*, 1983; FARNETANI *et al.*, 1996). Rather than wondering where is all the komatiite, it may be more appropriate to wonder how it was able to erupt at all, or at least to consider the more complex problem of the hydraulic circumstances required for eruption. The physics of this problem has not been modeled.

Having undergone differentiation in thickened oceanic crust, it is not difficult to imagine how the derivative komatiite lithologies could have been vulnerable to some form of subduction. Komatiites have the potential to form dense assemblages of FeO-rich dunite and gabbro because they typically contain 11% FeO; this compares with about 8% FeO in MORB aggregate magmas and MORB residues (Fig. 3a; KINZLER, 1997; NIU *et al.*, 1997). It seems likely that FeO contributions to density and phase transformational contributions to density [i.e., the gabbro to

eclogite transformation; about 1.5 GPa (e.g., HIRSCHMANN and STOLPER, 1996)] could have resulted in a deep komatiitic oceanic crust that was significantly more dense than the cratonic mantle below, although the physics of this important problem has not been numerically modeled as of this writing. In order for cratonic mantle to have stabilized, it seems necessary that overturn, detachment and subduction of dense komatiitic crust from the underlying buoyant mantle occurred at some stage.

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