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Constraints on the composition of the continental lithospheric mantle

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ABSTRACT

Major and trace element data for 375 spinel lherzolite and harzburgite xenoliths from continental basalts are used to constrain the nature and composition of the continental lithospheric mantle (CLM). The major and compatible trace elements (e.g., Ni, Co) have similar average and median values, whereas median values of the highly incompatible trace elements (e.g., LREE, Rb) are systematically lower than their average values. Both the average and median compositions have LREE-enriched patterns, with La abundances at about 10 and 3 times C1 chondrites, respectively, and a relatively flat HREE pattern at about 2 times C1 chondrites. Relative to the primitive mantle, highly incompatible trace elements are enriched in the average and median, whereas the mildly incompatible elements (e.g., the middle and heavy REE, Hf, Ti, V, Sc, Al, Ca, etc.) are depleted. The enrichment pattern of incompatible elements in these peridotites is comparable to that in ocean island basalts, although at lower absolute abundances. For most elements the average composition provides an estimate of the bulk composition of the CLM, however for those elements with a strong degree of skewedness in their data the median composition may provide a more reasonable estimate.

A comparison of geochemical data for spinel and garnet peridotites reveals many similarities, however, garnet peridotites have, on average, lower concentrations of elements with bulk distribution coefficients close to 1.0 (eg., HREE, Sc, Ca, Mn, Fe, Cr). These differences may be attributed to regional variation or age of the lithosphere sampled. It is proposed that the values reported here may be best viewed as a compositional estimate of the post-Archaean CLM.

The composition of the CLM is consistent with a growth and stabilization model involving the underplating of refractory peridotite diapirs produced during magmatism, and to a lesser extent by advective thickening of the lithosphere. The incompatible element enriched character of the CLM has been developed in intraplate and/or divergent-margin tectonic environments. The absence of a convergent-margin chemical signature in peridotite xenoliths, as characterized by a Nb depletion or enrichment, argues against incompatible element-enrichment of CLM in this environment.

The incompatible element-enriched character of the CLM does not possess the necessary source composition for continental flood basalts, therefore alternative crustal and/or mantle sources need to be considered for these large volume incompatible element-enriched melts. Although the CLM has an incompatible element-enriched character, it does not appear to represent a significant geochemical reservoir for these elements. Based on the median composition, the CLM has less than or equal to 5% of the highly incompatible elements relative to the primitive mantle, and this is about a factor of 5 less than that in the continental crust for such elements.

1. Introduction

The lithospheric mantle, that portion of the lithosphere underlying the crust, can be defined in terms of its chemical, thermal, seismic and/or mechanical properties. Simple petrological and chemical models have been put forth for the bulk composition and mineralogy of the oceanic lithospheric mantle [1–3], but not for the continental lithospheric mantle (CLM) due to its complex and long-term evolution. The CLM has been pos-

tulated to be a mantle reservoir enriched in incompatible elements [4], but its composition has been difficult to define and/or constrain. An incompatible element enriched CLM is commonly invoked as a source for a variety of basalts and some small volume magmatic suites, such as ultrapotassic rocks and kimberlites. Localized regions of the CLM may possess the chemical characteristics necessary for ultrapotassic and kimberlitic magmas, however, is it a reasonable source for larger volumes of basaltic magmas, such as continental

flood basalts? With greater understanding of the composition of the CLM we can better define the distribution of elements in the silicate portion of the earth.

There is much discussion on the nature and composition of the CLM, however, there seems to be little consensus among earth scientists about general aspects of this part of the mantle. The CLM is generally considered to have on average an incompatible element-enriched character, although there is little agreement as to the degree of enrichment and the relative abundances of certain elements (e.g., Nb/La, Ba/Rb, Zr/Sm, Ti/Eu). Related to the question of understanding the geochemical characteristics of the CLM, geochemists are at odds as to whether or not basalts can be used to constrain our compositional models for this region of the mantle. Many would agree that the CLM has developed through a series of tectono-magmatic stages, however, there are marked differences in the existing tectonic models which attempt to explain the growth of the CLM. Likewise, several models exist concerning the average thickness of the CLM in terms of its chemical, mechanical and thermal properties. The oceanic lithosphere presents a simple case, where its thickness is a function of its age, whereas the thickness of the CLM is more difficult to predict.

This paper presents a comprehensive survey of major and trace element data for spinel-bearing peridotite xenoliths carried in alkali basalts and related lavas. The average, standard deviation and median values for 375 samples are given. These data are used to characterize the composition of the CLM and provide some constraints on the source regions of certain basalt types. Furthermore, these data are used to develop models of CLM growth and define the budgetary contribution of the CLM to the global mass balance of elements.

2. The data set

Major and trace element compositions of 375 spinel peridotite xenoliths have been compiled. These xenoliths are generally classified as Group I peridotites [5,6] and they are the most abundant rock type in xenolith suites brought up by alkaline basalts and kimberlites. Group I peridotites [5–7] commonly include lherzolite, harzburgite and

dunite samples, generally with Mg-value > 85 [Mg-value = $100 * \text{Mg}/(\text{Mg} + \text{Fe}_{\text{total}})$]. Seismic models for the uppermost mantle beneath continental regions [2,8] suggest that this portion of the mantle is dominantly composed of these ultramafic lithologies (i.e., lherzolite and harzburgite). Therefore, this material may provide a good estimate of the bulk composition of the CLM beneath these regions.

Xenoliths included in this survey are hosted in basalts from continental regions, not from ocean islands. Most of these basalts are relatively young (i.e., late Tertiary to Recent), however the portions of the lithosphere they have traversed are dominantly Proterozoic and Phanerozoic in age. Peridotite xenoliths from kimberlites, lamproites and related rocks are not included and will be considered in a later study along with garnet-bearing peridotite xenoliths from basalts. Only samples for which major element data were available are included, and samples analyzed before 1970 were excluded to avoid some earlier, less precise data.

Bias of the sample population by any individual study, region or laboratory is minimized by the global coverage and the large number of studies included in this compilation. Some sample types (e.g. dunites) may be under-represented in this compilation because their low trace element concentrations present severe analytical problems that result in them not being analyzed. This may present a potential bias, although its importance is not considered to be great because such samples do not represent a significant population of xenolith suites. Furthermore, their overall effect would be only to lower slightly the absolute abundances of the highly incompatible trace elements.

It is also possible that xenolith sampling processes of magmas are non-random and that the geochemical and isotopic compositions of the peridotites are influenced by their host-basalt and associated magmatism. Several factors, however, argue against these conclusions. First, the host basalts tend to be chemically and isotopically distinct from their entrained xenoliths. Spinel peridotite xenoliths show a wide range of chemical compositions [7], including ones with LREE-enriched and LREE-depleted patterns. Additionally, single xenolith localities from various continents [9–12] have peridotites which display wide variations in major and trace element compositions, in

marked contrast to the basalts which carry them. Such geochemical diversity indicates that the lithospheric mantle sampled by these lavas was produced through a variety of processes and pre-

sumably over a longer time span than host volcanism. Moreover, it has been demonstrated time and again for broadly disparate continental regions (e.g., SE Australia, western U.S., western

TABLE 1
Average and median composition of spinel peridotite xenoliths

	Avg.	$\pm 1\sigma$	<i>n</i>	Med.		Avg.	$\pm 1\sigma$	<i>n</i>	Med.
SiO ₂	44.0	1.26	375	44.1	Nb	4.8	6.3	58	2.7
TiO ₂	0.09	0.09	361	0.09	Mo(b)	50	30	3	35
Al ₂ O ₃	2.27	1.10	375	2.20	Ru(b)	12.4	4.1	3	12
FeO total	8.43	1.14	375	8.19	Pd(b)	3.9	2.1	19	4.1
MnO	0.14	0.03	366	0.14	Ag(b)	6.8	8.3	17	4.8
MgO	41.4	3.00	375	41.2	Cd(b)	41	14	16	37
CaO	2.15	1.11	374	2.20	In(b)	12	4	19	12
Na ₂ O	0.24	0.16	341	0.21	Sn(b)	54	31	6	36
K ₂ O	0.054	0.11	330	0.028	Sb(b)	3.9	3.9	11	3.2
P ₂ O ₅	0.056	0.11	233	0.030	Te(b)	11	4	18	10
Total	98.88			98.43	Cs(b)	10	16	17	1.5
					Ba	33	52	75	17
Mg-value	89.8	1.1	375	90.0	La	2.60	5.70	208	0.77
Ca/Al	1.28	1.6	374	1.35	Ce	6.29	11.7	197	2.08
Al/Ti	22	29	361	22	Pr	0.56	0.87	40	0.21
Fe/Mn	60	10	366	59	Nd	2.67	4.31	162	1.52
					Sm	0.47	0.69	214	0.25
Li	1.5	0.3	6	1.5	Eu	0.16	0.21	201	0.097
B	0.53	0.07	6	0.55	Gd	0.60	0.83	67	0.31
C	110	50	13	93	Tb	0.070	0.064	146	0.056
F	88	71	15	100	Dy	0.51	0.35	58	0.47
S	157	77	22	152	Ho	0.12	0.14	54	0.090
Cl	53	45	15	75	Er	0.30	0.22	52	0.28
Sc	12.2	6.4	220	12.0	Tm	0.038	0.026	40	0.035
V	56	21	132	53	Yb	0.26	0.14	201	0.27
Cr	2690	705	325	2690	Lu	0.043	0.023	172	0.045
Co	112	10	166	111	Hf	0.27	0.30	71	0.17
Ni	2160	304	308	2140	Ta	0.40	0.51	38	0.23
Cu	11	9	94	9	W(b)	7.2	5.2	6	4.0
Zn	65	20	129	60	Re(b)	0.13	0.11	18	0.09
Ga	2.4	1.3	49	2.4	Os(b)	4.0	1.8	18	3.7
Ge	0.96	0.19	19	0.92	Ir(b)	3.7	0.9	34	3.0
As	0.11	0.07	7	0.10	Pt(b)	7	-	1	-
Se	0.041	0.056	18	0.025	Au(b)	0.65	0.53	30	0.5
Br	0.01	0.01	6	0.01	Tl(b)	1.2	1.0	13	0.9
Rb	1.9	4.8	97	0.38	Pb	0.16	0.11	17	0.16
Sr	49	60	110	20	Bi(b)	1.7	0.7	13	1.6
Y	4.4	5.5	86	3.1	Th*	0.71	1.2	71	0.22
Zr	21	42	82	8.0	U	0.12	0.23	48	0.040

Mg-value = 100[MgO/(MgO + FeO)] molecular ratio. Oxide elements are given in wt.%, other elements are given in ppm, unless (b) = ppb. Avg. = average; Med. = median; *n* = number of samples, $\pm 1\sigma$ = 1 standard deviation. Elements for which there were no available data accompanying whole rock compositional data include Be, Rh, I and Hg. Other elements, including H, C, N, and the noble gases were not included in this compilation.

* The abundance of Th may be systematically too high, given a possible analytical problem. Stosch (pers. commun., 1988) in a study of clean mineral separates from peridotite xenoliths found a significant peak around 310 keV only for olivine separates, which usually would be attributed to Th. He interpreted this peak, however, as a double escape peak of the 1332 keV line of ⁶⁰Co resulting from the high Co concentrations in olivines, and pair formation with the subsequent escape of the positron-electron annihilation radiation from the detector. Its effects on the Co 1332 keV peak would have a characteristic energy of: 1332 keV - (2 × 511 keV) (pair formation energy) = 310 keV.

Europe and central Asia) that a range of isotopic compositions (e.g., 15–25 ϵ_{Nd} unit variations) are commonly found in spinel peridotites from a single or adjacent pair of alkali basalt vents [10,13–16]. Such large ranges of isotopic compositions are consistent with the idea that these regions of the CLM have formed over considerable time. Finally, the global distribution of the alkali basalt-hosted peridotite xenolith localities [17] ensures that a broad spectrum of continental regions, which have developed from a diversity of tectonic and magmatic processes, have been sampled. Together these observations support the notion that a fairly representative sampling of the CLM has been carried out by these magmas.

The compositions of these peridotite xenoliths are generally thought to reflect the composition of their source regions in the CLM, however some consideration must be given to effects of weathering. Surface weathering of peridotite xenoliths may represent a significant problem to understanding the abundances of mobile elements in some samples. Many of these elements (e.g., K, Rb, Cs, U) are enriched by more than an order of magnitude in the crust relative to the mantle. Therefore, in the case of anhydrous peridotites, where these highly incompatible elements are concentrated along grain boundaries due to the absence of a host mineral, it may be impossible to determine the initial mantle abundances of these elements.

The average, standard deviation, number of samples and median values for major and trace elements are reported in Table 1. Where average and median values agree the data approximate a normal distribution, whereas when large differences between these values are found these elements show a non-normal distribution. Elements are classified according to their relative bulk distribution coefficients (residue/melt); those enriched in basalts over their source are incompatible elements (e.g., K, Rb, Ba, REE, P, Ti, Ca, Al, Sr, Nb, Ag, Sb, Re, etc.), those enriched in the peridotite residue after melting are compatible elements (e.g., Mg, Cr, Co, Ni, Ir), while a few elements have distribution coefficients close to 1.0 (e.g., Si, Ge, Fe, Mn). The compatible elements and those with a bulk distribution coefficient of about 1 have similar values for their average and median abundances. These elements also show low ($\ll 1$) coefficient of variation values. The

incompatible elements exhibit a wide range in absolute concentration, with average values being higher than their median values. For most (e.g., Ti, Al, Sr, REE) this is generally due to their heterogeneous distribution in the CLM, but in some cases (e.g., P, Nb, Ag, Re) this large relative variation reflects both a heterogeneous distribution and analytical problems. The highly incompatible elements, which are also mobile elements (specifically Cs, Rb, Ba, K, U), show the greatest variation in absolute concentrations and have average values that are much higher than their median values. This must reflect to some extent their heterogeneous distribution in the CLM, but element addition during surface weathering and magma infiltration and, possibly in some cases, analytical problems, may have contributed to the large range of values.

Therefore, the bulk composition of spinel peridotite xenoliths can be estimated by using either the average or median composition for the major elements, the compatible trace elements and only the mildly incompatible trace elements (e.g., V, Sc, Tm, Yb, Lu), given that their average and median values agree. In contrast, for the remaining elements, which includes the highly incompatible elements, the median value is more likely to best represent the bulk composition.

3. Chemical variations in spinel-peridotite xenoliths

The average and median major element compositions of spinel peridotite xenoliths are compared with other average values of spinel and garnet peridotite xenoliths and the primitive mantle in Table 2. There are no significant differences between the average spinel peridotite value reported earlier [18] and the average and median values reported here. Compared with averages for garnet peridotite xenoliths, spinel peridotites are enriched in FeO, Cr₂O₃, CaO, Al₂O₃ and Na₂O. Relative to the primitive mantle composition the average and median compositions of all peridotite xenoliths are enriched in MgO and NiO and depleted in CaO, Al₂O₃, Na₂O and TiO₂. These differences are consistent with the hypothesis that peridotite xenoliths are generally the residues left after partial melting [7,9].

Selected major and minor element variation diagrams are given in Fig. 1. The positive correla-

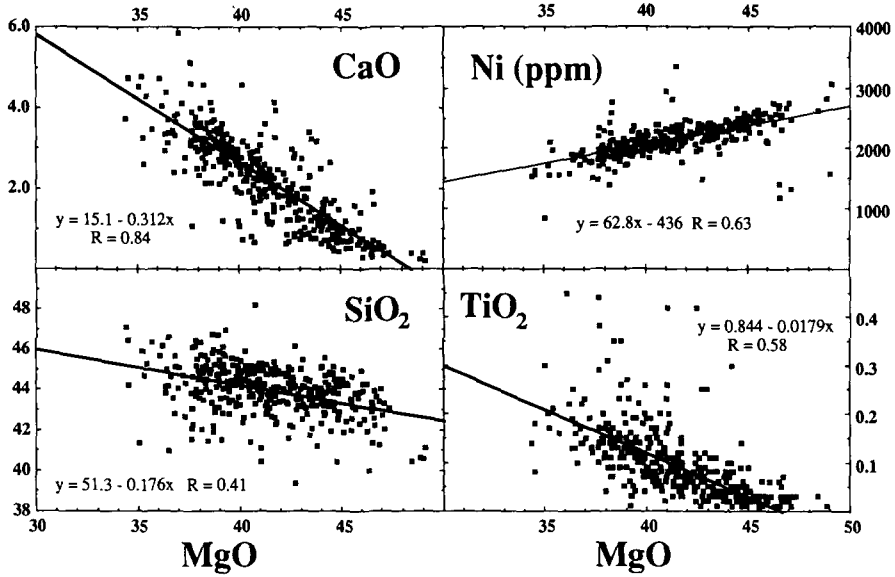


Fig. 1. Major element variation diagrams for spinel lherzolite and harzburgite xenoliths. MgO versus SiO₂, CaO and TiO₂ in wt.% and Ni in ppm. Equations for a linear regression and the correlation coefficient (*R*) are given in each panel.

TABLE 2
Comparison of average and median composition of spinel and garnet peridotite

	Spinel peridotites			Garnet peridotites			Primitive mantle
	Avg.	Med.	M-A sp	M-A gt	B-M	Jordan	
SiO ₂	44.0	44.1	44.15	44.99	45.00	45.55	44.8
TiO ₂	0.09	0.09	0.07	0.06	0.08	0.11	0.21
Al ₂ O ₃	2.27	2.20	1.96	1.40	1.31	1.43	4.45
Cr ₂ O ₃	0.39	0.39	0.44	0.32	0.38	0.34	0.43
FeOtotal	8.43	8.19	8.28	7.89	6.97	7.61	8.40
MnO	0.14	0.14	0.12	0.11	0.13	0.11	0.14
MgO	41.4	41.2	42.25	42.60	44.86	43.55	37.2
NiO	0.27	0.27	0.27	0.26	0.29	-	0.24
CaO	2.15	2.20	2.08	0.82	0.77	1.05	3.60
Na ₂ O	0.24	0.21	0.18	0.11	0.09	0.14	0.34
K ₂ O	0.054	0.028	0.05	0.04	0.10	0.11	0.028
P ₂ O ₅	0.056	0.030	0.02	-	0.01	-	0.022
Total	99.49	99.05	99.87	98.60	100.00	100.00	99.86
Mg-value	89.8	90.0	90.1	90.6	92.0	91.1	88.8
olivine	62	63	67	65	68	66	56 57
opx	24	24	22	28	25	28	22 17
cpx	12	11	9	3	2	3	19 10
spinel	2	2	2	-	-	-	3 -
garnet	-	-	-	4	5	3	- 14

Average (Avg.) and median (Med.) values from Table 1. M-A sp refers to the average continental spinel peridotites from [18] and M-A gt refers to the average continental garnet peridotites from [18]. B-M refers to the average of 24 south African garnet peridotite xenoliths [28]. Jordan refers to the average of continental garnet peridotite xenoliths from [8]. The primitive mantle estimate is from [7,49]. Opx = orthopyroxene, cpx = clinopyroxene. Mg-value is the molar ratios of 100 Mg/(Mg + ΣFe).

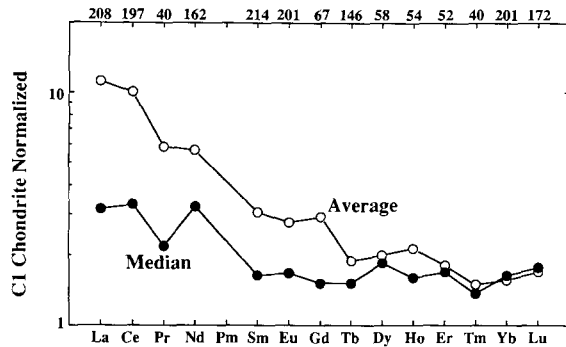


Fig. 2. CI chondrite normalized diagram of the REE for the average and median spinel peridotite. Normalizing values from [24]. Numbers along the top of the figure refer to the number of data entries for each element which went into the average REE pattern.

tion of Ni versus MgO reflects their similar compatibility during melt depletion processes. Likewise, the negative correlations of TiO_2 , CaO and less so SiO_2 versus MgO evidences the contrasting behaviour of these elements during melting. The range of MgO contents reflects the diversity of compositions in xenolith suites and hence the CLM; the more MgO-rich peridotites have generally lost greater amounts of melt.

The REE patterns for average and median compositions display relatively coherent LREE-enriched trends and relatively flat HREE abundances (Fig. 2 and Table 3). The LREE show a non-normal distribution (Fig. 3), which is skewed towards high concentrations. There is no evidence for bimodality of the LREE into LREE-depleted and LREE-enriched end member populations, but instead there appears to be a complete spectrum of LREE patterns. The REE data can also be used to demonstrate that there is no obvious bias in the average and/or median value as a result of a single analyst or sample locality. Data for Pr, Er and Tm come only from a few studies [6,9,11,19–21] and include peridotites from North American, European, Asian and Australian localities. Samples reported in these studies include both LREE-depleted and LREE-enriched varieties. Obviously there are far fewer data for these three REE, however, a REE pattern generated by using only these elements is comparable to those shown by the broader survey REE patterns, indicating that there is no significant bias contributed by the other studies.

In a world wide survey of spinel and garnet peridotite xenoliths McDonough and Frey [7] noted that samples with LREE-enriched patterns

TABLE 3

Selected ratios for average and median spinel peridotite xenoliths, CI chondrite and the primitive mantle

	CI	PM	Avg.	$\pm 1\sigma$	<i>n</i>	Med.
Ni/Cr	4.0	0.64	0.86	(0.31)	308	0.79
Ni/Co	22	17	19.4	(2.4)	165	19.6
FeO/Ni	22	44	39	(11)	308	37
MgO/Ni	14.5	196	190	(33)	308	190
Ca%/Sc	0.15	0.15	0.14	(0.8)	218	0.13
Ca%/Yb	5.4	5.4	7.4	(4.1)	199	6.4
V/Sc	9.4	4.7	5.0	(1.1)	113	4.7
Y/Ho	28	28	26 *	(1.7)	16	25 ^a
Zr/Hf	36	36	35 *	(2)	5 ^a	35 ^a
Nb/Ta	17.6	17.6	17 *	(3)	4 ^a	17 ^a
Ti/Eu	7600	7600	6500	(4000)	193	6520
Ti/Eu ^b	anhydrous peridotites		7700	(2800)	104	7440
Ti/Eu ^b	hydrous peridotites		5200	(4000)	49	4360
(La/Yb) _N	1.0	1.0	7.1	(12)	195	2.7
(La/Sm) _N	1.0	1.0	2.7	(2.8)	200	2.1

Data for CI chondrite (CI) and Primitive mantle (PM) are from [7,24,49].

^a The average and median data for Zr/Hf and Nb/Ta are taken exclusively from the study of Jochum et al. [25], while Y/Ho data are from spark source mass spectrometry and radiochemical neutron activation studies [6,9,25] (see text for details).

^b The average, $\pm 1\sigma$ and median values for Ti/Eu in hydrous and anhydrous peridotites were determined on samples with ≥ 0.3 wt.% TiO_2 . The average, $\pm 1\sigma$ and median ratios were determined from the data population, not from data of the median values in Table 1. *n* = number of samples. Ca%/Sc and Ca%/Yb are given in wt.%/ppm.

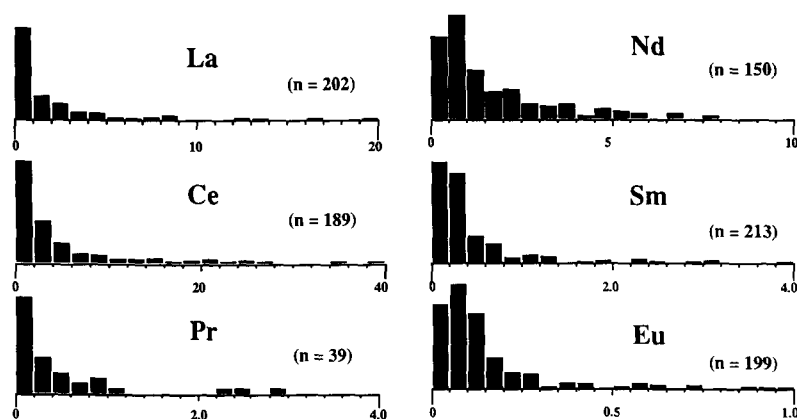


Fig. 3. Histograms showing the LREE concentrations in spinel peridotite xenoliths. These histograms show a non-normal distribution of data for the LREE, with no indication of a bimodal pattern (i.e., LREE-depleted and LREE-enriched grouping). Concentrations are given in ppm. A few data points with very high LREE concentrations were not included in some of these histograms. The number of excluded data points are: La = 6, Ce = 8, Pr = 1, Nd = 12, Sm = 1 and Eu = 2.

commonly have HREE abundances about 1.0–0.1 times C1 chondrite and that HREE abundances decrease with decreasing CaO and Al₂O₃ contents. In contrast, both average and median values show middle to heavy REE (HREE) patterns that are relatively flat at about 2 times C1 chondrite, which is enriched by at least a factor of two over that typically found in individual samples possessing a LREE-enriched pattern and about 2.2 wt.% CaO and Al₂O₃ contents. Such high concentrations of the HREE are often only found in xenoliths with a slight LREE-depletion or a flat LREE pattern and with CaO and Al₂O₃ contents > 2.5 wt.%. This combination of a LREE-enriched and a high, flat HREE pattern (Fig. 2) indicates that the average and median peridotite represents a mixture of features commonly observed in peridotite xenolith suites, but usually not found in individual samples.

The composition of the average and the median peridotite presents a contradiction by having both a LREE-enriched pattern and a major element composition that is depleted in a mafic melt component. This combination of contradictory geochemical signatures is also often found in individual samples [7,9]. The REE are incompatible elements and are therefore partitioned into the liquid and depleted in the residue during melting, whereas MgO and Ni are compatible elements and are enriched in the residue. The fact that these

peridotites have higher Mg-values and Ni contents than primitive mantle, as well as having a LREE-enriched signature, indicates that, on average, these peridotites have evolved through multi-stage processes. Frey and Green [9] originally developed a model which reconciles these apparently contradictory observations. Initially the peridotite body was depleted as a result of basalt extraction, and this residue was later enriched by small amounts of an incompatible element-enriched melt.

The HREE, in particular Yb, are positively correlated with CaO, Al₂O₃, Sc and V contents in peridotite xenoliths [7]. These elements typically show similar degrees of enrichment in basalts over their source composition and thus they would be considered to have relatively similar bulk distribution coefficients. Figure 4 shows the correlations between CaO and Yb and Al₂O₃ and Yb for spinel in peridotite xenoliths. Likewise, Ga abundances show good correlation with Sc and Al₂O₃ abundances suggesting that it also has relatively similar bulk distribution coefficients during melting [22]. It has also been noted that peridotites, as well as primary and near primary basalts from various tectonic environments and Archaean to modern komatiites, have chondritic to near chondritic Ti/Eu ratios [7,23,24]. This indicates that TiO₂ and Eu also have relatively similar incompatibility during melting. Collectively, these

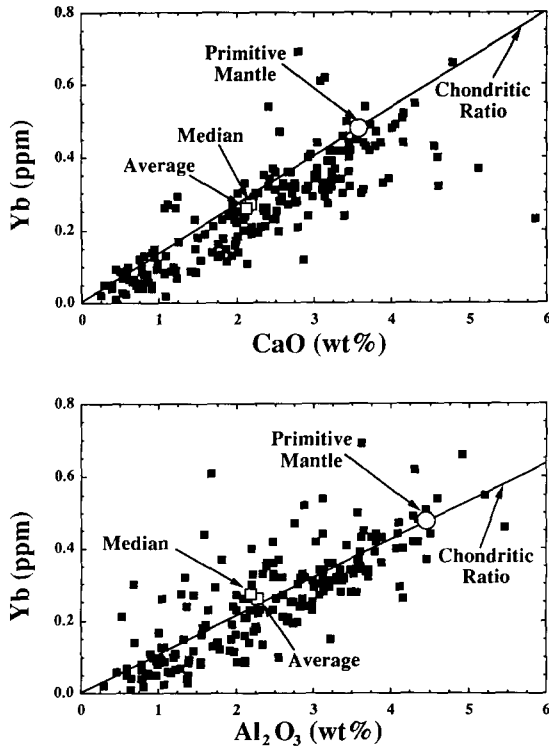


Fig. 4. CaO (in wt.%) and Al₂O₃ (in wt.%) versus Yb (in ppm) for spinel peridotite-xenoliths. The average and median values are from Table 1. Also shown is a line for a chondritic CaO/Yb and Al₂O₃/Yb values and their estimated primitive mantle values [7,24].

features demonstrate that in many instances major and trace elements behave coherently during melting.

There is a general trend for Ca/Yb values in peridotites to be higher than the primitive mantle value (Table 3, Fig. 4), which is consistent with Yb being slightly more incompatible during melting than CaO. In contrast, Al₂O₃/Yb values in peridotites shows considerable scatter around the chondritic ratio (Fig. 4) and the average Al₂O₃/Yb value in spinel peridotite xenoliths is close to the primitive mantle estimate, however, this ratio varies widely in peridotite xenoliths. McDonough and Frey [7] found a good trend for Al₂O₃/Yb versus Yb in anhydrous spinel peridotites, however, the hydrous peridotite xenoliths show greater scatter with many samples having low, sub-chondritic Al₂O₃/Yb ratios. They suggested that the enrichment/metasomatic processes involved in forming the hydrous xenoliths was able to fractionate Al₂O₃ from CaO and Yb.

Selected compatible and incompatible element ratios for the average and median are given in Table 3, together with values for C1 chondrites and the earth's primitive mantle. Ratios of MgO and Co versus Ni are relatively constant as evidenced by their low standard deviations, with only 10–20% overall variation, and excellent agreement between average and median values. Ratios of FeO/Ni and Ni/Cr show larger relative variation, due to the slightly more compatible behaviour of Ni compared to FeO and Cr. The relative differences between average (and/or median) and primitive mantle values for Ni/Cr, Ni/Co, FeO/Ni and MgO/Ni are consistent with these peridotites having lost a melt.

In a detailed study of incompatible trace elements Jochum et al. [25] found peridotites to have Zr/Hf, Y/Ho and Nb/Ta values close to the primitive mantle ratio (Table 3), and concluded that these element ratios do not change significantly during melting. In combination with other high-quality Y/Ho data [6,9] spinel peridotites show only 7% total relative variation in this ratio (Table 3). These observations, together with data for basalts, argue against significant fractionation of these ratios during melting. In contrast, other available data for Zr/Hf and Nb/Ta in peridotite xenoliths show much greater variation (10–320 and 3–100, respectively) and this is more than likely due to poor data quality. It is possible, however, that careful re-analysis of such samples may reveal small differences in these element ratios between peridotites and the primitive mantle. Other incompatible element ratios (e.g. Ca%/Sc, Ca%/Yb, V/Sc, Ti/Eu and Sr/Nd) show somewhat larger relative standard deviations (Table 3), but still having good agreement between average and median values. This suggests that the elements in each pair have similar incompatible behaviour during melting and enrichment processes.

As was found for the REE, the average and median incompatible trace element patterns (Fig. 5) show slight differences at the more incompatible end. Both patterns are relatively smooth, with the median pattern having lower overall concentrations of the more incompatible elements. For the highly incompatible elements [e.g., K to Rb, and other elements (e.g., Cs and Tl) not shown] it is difficult to evaluate fully the reliability of these estimates. This is because most of

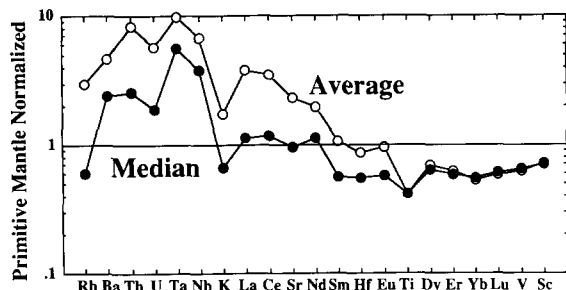


Fig. 5. An incompatible element diagram showing the compositions of the average and median spinel peridotite xenoliths. These patterns are similar to those observed in intraplate basalts [24]. The order of elements and the primitive mantle normalizing values are from [7,24].

these elements are often only found on grain boundaries [11,26,27] which leads to their being readily altered by transport and post-depositional processes. Additionally, for elements such as Nb and Ta, there are just not enough high-quality data at low concentrations (Table 1), which may result in an estimate for their abundance in the CLM that is too high. Given many incompatible trace elements have higher average values relative to their median values and that these elements display a non-random distribution, the median value is chosen over the average value for the best guide to the bulk composition of spinel peridotite xenoliths.

4. Composition of the continental lithospheric mantle (CLM)

Since this data compilation has been restricted to spinel peridotite xenoliths, it may only be representative of the shallow lithospheric mantle and not the deeper garnet peridotite facies region. Additionally, the age of the CLM sampled by the host magma must also be considered, especially if garnet peridotites as sampled by kimberlites are from significantly older portions of the lithosphere than alkali basalt-hosted spinel peridotites. Therefore a brief comparison with garnet peridotite xenoliths is appropriate.

A comparison of spinel and garnet peridotite compositions (Table 2) indicates that both types of peridotites have similar major and minor element compositions. Garnet peridotites, however, do possess lower average values of FeO, MnO, CaO, Al_2O_3 , Cr_2O_3 , and Na_2O and higher values

of SiO_2 and K_2O . The higher K contents may only be due to melt infiltration of their kimberlitic host. Lower CaO, Al_2O_3 and Na_2O contents indicates that these xenoliths are more depleted in a basaltic (or more mafic) component than spinel peridotites. At the same time Ti is more incompatible than these elements and would therefore be expected to be more depleted, although the TiO_2 contents are similar for both garnet and spinel peridotites. This indicates that such comparisons are too simplistic.

Figure 6 shows that there is a considerable range of FeO and MgO contents in spinel and garnet peridotite xenoliths, but that in general garnet peridotites have lower FeO contents than spinel peridotites. The lower Cr_2O_3 and FeO contents in garnet peridotites may be due to there having been greater degrees of partial melt extracted from them. At very high degrees of partial melting Cr_2O_3 and FeO are more depleted in the residual peridotite due to the depletion of the Cr-rich phases and the higher Mg-values in the residual minerals. Spinel and garnet peridotites, in general, follow a common differentiation trend on a Mg/Si versus Al/Si diagram (Fig. 7), with the more refractory samples plotting closer to the olivine-rich end member with very low Al/Si values. Both types of peridotites have experienced melt extraction processes, however, garnet peridotites generally display more depleted characteristics, consistent with them having lost a greater amount of melt.

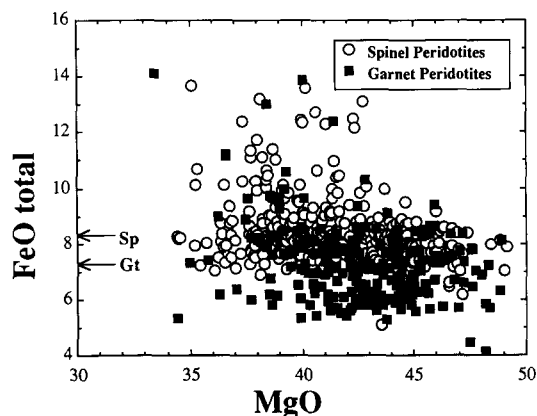


Fig. 6. MgO (wt.%) versus FeO (wt.%) in spinel and garnet peridotite xenoliths. The arrow indicated by *Sp* refers to the average FeO content of spinel peridotites and the arrow indicated by *Gt* refers to the average FeO content of garnet peridotites.

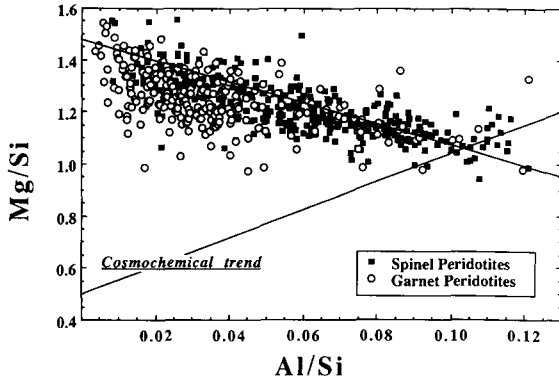


Fig. 7. Mg/Si versus Al/Si (weight ratio) for spinel and garnet peridotite xenoliths. This diagram is adapted from Jagoutz et al. [21], where they suggested the intersection of the cosmochemical and geochemical trends as representative of the upper mantle Mg/Si and Al/Si ratios. Jagoutz et al. referred to the upper line as the geochemical trend and the line shown here is drawn using their coordinates [21]. This line has not been fit to the data shown here.

Further support for the suggestion that garnet peridotites are somewhat more refractory than spinel peridotites comes from a comparison of the average and median REE contents of spinel and garnet peridotites (Fig. 8). (There is substantially less REE data for garnet peridotites, with about 60 analyses each for La, Ce, Nd, Sm and Eu, about 40 for Yb and Lu, 28 for Dy and less than 20 for Gd, Tb, and Er.) The absolute degree of LREE-enrichment in spinel and garnet peridotites

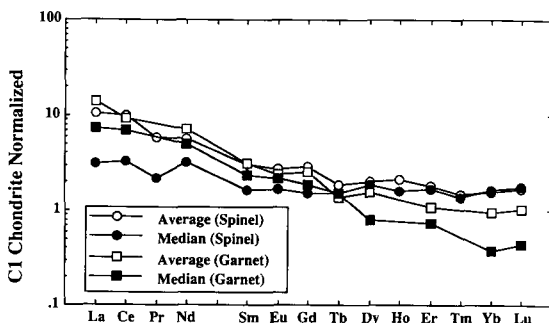


Fig. 8. A comparison of the average and median REE patterns of spinel and garnet peridotite xenoliths. The degree of LREE-enrichments in both suites are relatively similar, however there is a marked depletion of the HREE in the garnet peridotites compared to the spinel peridotites. The greater depletion in the garnet peridotite HREE patterns are consistent with their more depleted major element composition.

CI chondrite normalized values are as in Fig. 2.

are about equal, whereas the HREE are more depleted in the garnet peridotites. The lower HREE contents of garnet peridotites is consistent with the lower CaO and Al₂O₃ contents and with these samples having experienced more melt depletion. Higher (La/Yb)_N in garnet peridotites indicates either they have been more strongly enriched in LREE by open system metasomatism, or the enriching melt/fluid had a higher (La/Yb)_N ratio than that which typically affects spinel peridotites.

The most significant compositional differences between garnet and spinel peridotite xenoliths is in the mildly incompatible element (e.g., Ca, Al, V, Sc and HREE) and mildly compatible element (particularly, Mn, Fe and Cr) concentrations. These elements have bulk distribution coefficients close to 1.0, and are more depleted in garnet peridotites compared with spinel peridotites (Table 1, Figs. 5 and 8), but both types of peridotites are depleted in these elements relative to the primitive mantle. The more incompatible elements (e.g., Ti, Zr, Sr, Nb and middle and LREE) and compatible elements (e.g., Mg, Ni, Co) have comparable abundances in both groups of peridotite xenoliths.

That there are differences in the major and trace element compositions of spinel and garnet peridotites may suggest a vertical zonation in the composition of the CLM, although these differences may also reflect a contrast in the lithosphere traversed by kimberlites and basalts. Basalt-hosted spinel peridotite xenoliths compiled in this study come from Proterozoic and younger terrains and do not include samples from Archaean cratons. Maaløe and Aoki [18] initially observed that there are compositional differences between the dominantly kimberlite-hosted garnet peridotites, the majority of which come from southern Africa, and the basalt-hosted spinel peridotites. They suggested that these differences may be due to regional variations in the composition of the CLM. Boyd and Mertzman [28] and Nixon [17] further emphasized that most garnet peridotite xenoliths are from the Kaapvaal craton of southern Africa, which is Archaean in age, whereas the majority of spinel peridotite xenoliths are from post-Archaean, non-cratonic regions. Moreover, spinel peridotite xenoliths from the Kaapvaal craton are generally considered to be more depleted

than spinel peridotite xenoliths from elsewhere [17].

Geochemical and isotopic data for garnet peridotite xenoliths from the Kaapvaal craton, however, indicate that this region of the CLM has not remained isolated and unaffected by post-Archaean processes. A comparison of epsilon Nd values and average Nd model ages for spinel peridotite and southern African garnet peridotite xenoliths are presented in Fig. 9. This figure shows that there is considerable overlap in the epsilon Nd values of spinel and garnet peridotites and more importantly, that both groups of peridotites have Proterozoic and younger model ages. Single-stage Nd model age calculations for peridotites are generally not applicable because these xenoliths often have geochemical features indicating that they have experienced a melt depletion event followed by a melt enrichment event. The initial melt depletion processes would leave the residue with a LREE-depleted pattern, whereas the subsequent enrichment event leaves the peridotite with a LREE-enriched pattern. Additionally, this enrichment process typically introduces material with a significantly different $^{143}\text{Nd}/^{144}\text{Nd}$ composition than that of the residual peridotite. The fact that southern African garnet peridotites have such young model ages argues for the recent introduction of more radiogenic Nd, and possibly the addition of a LREE-enriched metasomatic melt

into this region of the CLM. Moreover, it would not be unreasonable to assume that large portions of the southern African CLM has experienced multiple enrichment events since its initial stabilization in the Archaean.

In conclusion, the available data do not allow us to fully evaluate whether the compositional differences between spinel and garnet peridotites are a function of age, province or a combination of factors. This is mostly due to the fact that the bulk of the existing major and trace element data for garnet peridotite xenoliths is so heavily biased in favour of samples from the Kaapvaal craton (especially the Kimberley pipes). Thus, it must be emphasized that more data for xenoliths from other Archaean cratons as well as a comparative study of garnet peridotites from off-craton and cratonic localities in southern Africa and elsewhere are needed before one can decide which is the most significant factor influencing the geochemical differences. Given these qualifications, the average or median values presented in Table 1 may be best viewed as the composition of the post-Archaean CLM.

The bulk compositional estimate of the continental lithospheric mantle presented here may represent a lower limit for incompatible elements, since it does not take into account Group II peridotites (Al-Ti augite suite with Mg# < 85; see [5,6]) in the lithospheric mantle. Many Group II

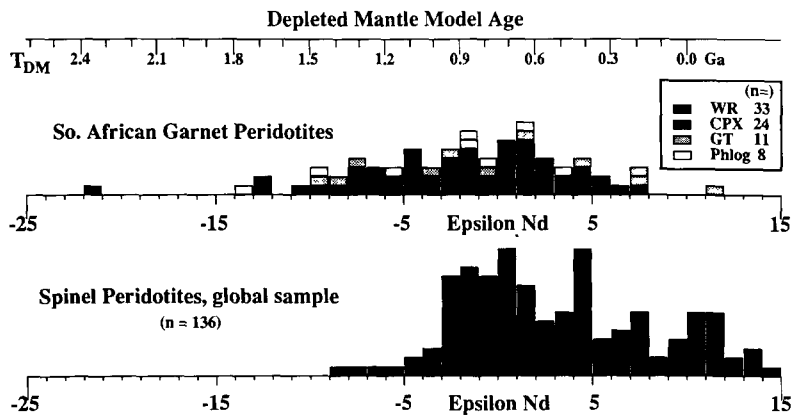


Fig. 9. Epsilon Nd values for spinel and garnet peridotite xenoliths. The upper scale bar presents calculated average Depleted Mantle model ages relative to ϵ_{Nd} values, constructed using an average $^{147}\text{Sm}/^{144}\text{Nd}$ of 0.110 and a depleted mantle value of +10 ϵ_{Nd} (which projects back to ϵ_{Nd} of 0 at 4.5 Ga). The average $^{147}\text{Sm}/^{144}\text{Nd}$ used in this calculation was based on the average Sm/Nd values of whole rocks and clinopyroxenes from spinel and garnet peridotites. These model ages assume a single-stage evolution of the Sm-Nd system, however, for peridotite xenoliths such a model is generally not applicable. Abbreviations used in the inset box are: WR = whole rock, CPX = clinopyroxene, GT = garnet, Phlog = phlogopite. Samples used to construct the southern African garnet peridotites were restricted to those in Cretaceous kimberlites.

peridotite xenoliths from basalts are enriched in incompatible elements (e.g., [29]), as are MARID and other xenoliths (e.g., [30]) from kimberlites. These samples should be considered when calculating the bulk composition of the lithospheric mantle, although obtaining a reliable estimate of the volumetric proportions of Group I to Group II xenoliths and the bulk composition of Group II xenoliths is very difficult. On the other hand peridotites with low incompatible element concentrations may be underrepresented, as mentioned previously, and their contribution must also be considered. Understandably, this CLM composition was chosen as somewhat of a balance between these considerations.

5. Constraints on the growth of the continental lithospheric mantle

Growth of lithospheric mantle is generally attributed to two main processes: conductive cooling and diapiric underplating. Conductive cooling involves the passive incorporation of asthenosphere on to the base of the lithosphere as it cools. This process is suggested to be important for oceanic lithospheric growth away from spreading centers, especially as the oceanic plate cools [1,31,32]. In contrast, diapiric underplating involves the incorporation of Mg-rich, Fe-depleted residues produced by the extraction of a basaltic melt from peridotite [3,33,34]. These peridotitic residues are chemically buoyant relative to the surrounding mantle and can be trapped beneath continents.

The average and median composition of garnet and spinel peridotite xenoliths can be interpreted in terms of a two-stage evolution: an initial melt depletion event followed by the later addition of an incompatible element enriched component. A lithospheric mantle formed by conductive cooling

would be expected to be chemically stratified, becoming more fertile in composition with depth. Ultimately, its composition would be comparable to the source regions of basalts. The depleted composition suggested here for the CLM supports a diapiric underplating growth model. This model is consistent with density arguments of Jordan [35] given an upper mantle of pyrolite composition [2]. Garnet peridotites have lower FeO contents than the primitive mantle and would be more buoyant than the less refractory spinel peridotites. These peridotites, therefore, are more effectively stabilized beneath continents.

The presences of spinel and garnet peridotites with fertile major element compositions does not necessarily indicate that portions of the CLM were accreted through conductive cooling processes. Other geochemical and isotopic evidence reveals that many of these samples have had complex, multistage histories. For example, sheared garnet lherzolites from southern Africa, which have fertile bulk compositions, are pieces of old refractory peridotite which have recently been refertilized [36,37]. Similarly, some spinel peridotites with fertile compositions have also experienced recent refertilization through metasomatic processes [25,26]. It may be that parts of the CLM have been accreted by conductive cooling, however, these portions are believed to be mechanically unstable over geologically long time scales [38].

6. Tectonic environment of incompatible element enrichment of the CLM

Both garnet and spinel peridotites show a second stage enrichment of incompatible elements, which can be related to particular tectonic environments by using incompatible element ratios. Three such ratios are considered important here: Sr/Nd, Ti/Eu and Nb(Ta)/LREE.

TABLE 4
Sr/Nd in primitive basalts and peridotite xenoliths

Primitive basalts	Sr/Nd	Peridotite xenoliths	Sr/Nd
MORB	9 to 15	Spinel Peridotites (Avg.)	15 ± 7
Ocean Island basalts	14 to 22	Spinel Peridotites (Med.)	14
Island Arc Basalts	30 to 35	Garnet Peridotites (Avg.)	19 ± 14
Primitive Mantle	15.6	Garnet Peridotites (Med.)	15

The range of Sr/Nd values for basalts is taken from the literature. Data for the primitive mantle is from [24]. Sr/Nd ratios for average (Avg.) spinel and garnet peridotites are taken from samples where Sr and Nd were both determined, with the variation expressed as ± 1σ.

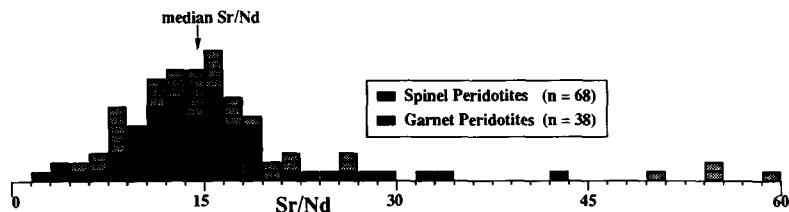


Fig. 10. Histogram of Sr/Nd values in spinel and garnet peridotite xenoliths. The median Sr/Nd value is for both groups of peridotite xenoliths (Table 4). Data for garnet peridotites include samples from southern Africa and elsewhere.

The Sr/Nd ratio of the primitive mantle is 15. Primitive basalts from mid-ocean ridges, continental rifts and intraplate settings (e.g., ocean island type) have Sr/Nd ratios similar to the primitive mantle (Table 4), whereas primitive convergent margin basalts (e.g., island arc basalts) have a Sr/Nd which is a factor of two higher (Table 4). Therefore, a peridotite residue produced by convergent margin magmatism would be expected to have a Sr/Nd value considerably different from the chondritic value. It has been previously noted that spinel peridotite xenoliths from southeast Australia have a relatively constant Sr/Nd of 15 ± 4 [14]. The average and median value for Sr/Nd in garnet and spinel peridotites is also similar to the primitive mantle value (Table 4). A histogram of Sr/Nd in spinel and garnet peridotites (Fig. 10) demonstrates that greater than 90% of peridotite xenoliths have ratios below 30. This range of values is consistent with these samples representing residual peridotites produced in intraplate or divergent-margin tectonic environments.

The fact that spinel and garnet peridotites have Sr/Nd values similar to the primitive mantle value (Table 4) suggests that these elements are not significantly fractionated during partial melting and lithosphere production. Similarly, the continental crust is suggested to have a chondritic or primitive mantle Sr/Nd [39], which also argues against significant fractionation of these elements as a result of crust–mantle differentiation.

The relative abundances of high field strength elements (HFSE: e.g., Nb, Ta, Zr, Hf and Ti) to REE have also been used to characterize basalts from convergent-margin tectonic environments versus basalts from other tectonic settings. Island arc basalts, for example, are often cited as having HFSE-depleted compositions. Therefore, based on the relative order of incompatibility ($Nb = Ta >$

$La > Sm \approx Zr = Hf > Eu \approx Ti$) for these elements in oceanic basalts [24], comparisons of Ti/Eu, La/Nb and Zr/Hf/Sm in peridotites and primitive basalts may be useful in identifying the tectonic environment of incompatible element enrichment of the CLM.

Much of the available data for Zr and Hf in peridotite xenoliths may not be precise enough given the large variation in Zr/Hf. However, most neutron activation analyses data for Sm and Hf may be more reliable because it avoids the problem of comparing results from two different techniques. The average and median Sm/Hf values in spinel (1.7 ± 0.5) and garnet (1.4 ± 0.7) peridotite overlap with one another and the primitive mantle value (1.44), although there is a range of variation in this ratio. This suggests that the CLM does not, on average, appear to possess a significant Hf depletion, relative to Sm, although some peridotite xenoliths do show such depletions. Oceanic basalts and island arc basalts generally have Sm/Hf ratios close to the primitive mantle value [24,40], suggesting that these elements are not significantly fractionated during melting in different tectonic environments.

Salters and Shimizu [41], using an incompatible element diagram, suggested that peridotite xenoliths commonly possess Ti depletions (or a low Ti/Eu) and that this feature is a world-wide character of the upper mantle. They also concluded that this Ti depletion is similar to that found in island arc volcanics. If their observations are correct, then Ti/Eu could also be used to constrain the tectonic environment of CLM growth and enrichment. It has been shown, however, that the initial assumption of Salters and Shimizu (that the incompatible element composition of clinopyroxene reflects that of the whole rock) is generally not valid [7]. This aside, it is still instructive to examine the variation in Ti/Eu.

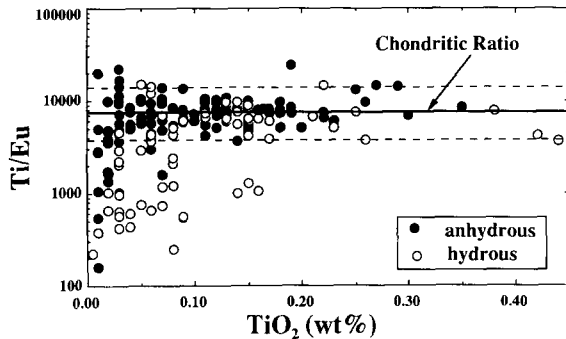


Fig. 11. TiO_2 (wt.%) versus Ti/Eu ratio in anhydrous and hydrous spinel peridotite xenoliths. Hydrous peridotites are distinguished solely on the presence of amphibole and/or mica. The chondritic (or primitive mantle) ratio is shown as a solid bar; a factor of two variation in this ratio is indicated by the upper and lower dashed lines. There is an obvious abundance of hydrous peridotites with low Ti/Eu between 200 and 1500.

Spinel peridotite xenoliths have highly variable Ti/Eu (Table 3), with an average and median (6520) value that is slightly lower than the primitive mantle value (7600). However, the details of this variation in different peridotites reveals some interesting features. The average Ti/Eu ratio is lower in hydrous (amphibole- and/or mica-bearing) peridotites than in anhydrous peridotites (7500 ± 3600), with anhydrous xenoliths only showing 36% total variation (Fig. 11 and Table 3). McDonough and Frey [7] reported similar findings and proposed that hydrous peridotites with low Ti/Eu values (< 3000) were produced by metasomatic processes capable of fractionating Ti from Eu.

Determining whether the CLM has a Nb(Ta)-anomaly is important in constraining the tectonic environment of CLM growth and enrichment. On an incompatible trace element diagram convergent-margin basalts characteristically show Nb(Ta)-depletions relative to La or other LREE, whereas MORBs and intraplate basalts do not show Nb(Ta)-anomalies. The incompatible element signatures of spinel peridotite xenoliths (Fig. 5) are similar to that of intraplate basalts and/or divergent-margin basalts and not convergent-margin basalts. Individual peridotites consistently reveal either patterns comparable to those of MORB, with increasing depletion from Ce through La to Nb and Ta [25] or patterns comparable to those of intraplate basalts. These incompatible

element patterns are not similar to those found in convergent-margin basalts or their residues. It must be remembered, however, that there are few high-quality Nb or Ta data for peridotite xenoliths. This is because these elements are at or near the detection limits for most geochemical laboratories. Thus, conclusions based on the relative abundances of Nb and Ta are made on a limited data set and do not necessarily reflect the bulk of the CLM.

Given that the Sr and Nd isotopic compositions of many of these peridotites differs from that of their host and the associated regional magmatism, most of these xenoliths have not been influenced chemically by the magmatism which has brought them to the surface. Therefore, the incompatible element ratios of these peridotites record earlier, unrelated magmatic events. Certain continental regions where peridotite xenoliths have been sampled (e.g., western U.S., southeast Australia) are areas where extensive convergent margin magmatism has occurred in the past. The chemical enrichment signatures of such magmatism are notably absent. Overall, incompatible element characteristics of peridotite xenoliths show that the enrichment of the CLM has occurred in divergent-margin or intraplate tectonic settings.

7. The continental lithospheric mantle as a source for basalt

It is often suggested that the CLM is either wholly or in part a source for various types of magmas: intraplate alkalic basalts, flood basalts, ultrapotassic lavas and kimberlites. In general, most of these models involve some external force to initiate CLM melting (e.g., upwelling from below causing intrusion of asthenosphere derived melt along with extension, thinning and decompression of the lithosphere). Some localized regions of the CLM may contain the necessary isotopic and incompatible element characteristics needed to be the source of some magma types (e.g., ultrapotassic lavas and kimberlites), although a CLM source for these melts may not be a requirement. Recent Os isotope data [37] support the suggestion that the source of southern Africa Group II kimberlites is in the CLM, whereas such data for the associated Group I kimberlites indicate that these magmas come from sources with

Os isotopic compositions comparable to modern oceanic basalts.

An important question to consider is whether the proposed bulk composition of the CLM is compatible with more general models for the sources of intraplate and flood basalts. Hawkesworth et al. [42] and Carlson [43], for example, have suggested the CLM as a source for large volume flood basalts. Hergt et al. [44] emphasized that some continental flood basalts, in particular basalts from the Gondwanaland breakup period, have incompatible element abundances which mimic the estimated bulk crust compositions (Fig. 8). It is, in fact, this crust-like signature in these basalts which has compelled some to suggest the introduction of a subducted sedimentary component into the source regions of these basalts [43,44]. Furthermore, as noted by Hergt et al. [44], the large negative Eu-anomalies in flood basalts with high Mg-values (~ 60) cannot be produced by feldspar fractionation using reasonable model parameters, and thus forces one to add crust-like material, which possess such an anomaly, to the sources of these basalts. In light of this, it is important to note that Eu-anomalies, negative or positive, are rare in peridotites. Therefore, there are significant differences in the incompatible element pattern of the CLM, particularly for Nb, (Fig. 12) and that of a typical flood basalt. Unless we have not sampled this mantle, then we must consider alternatives to the CLM as a source for these large volume basalts.

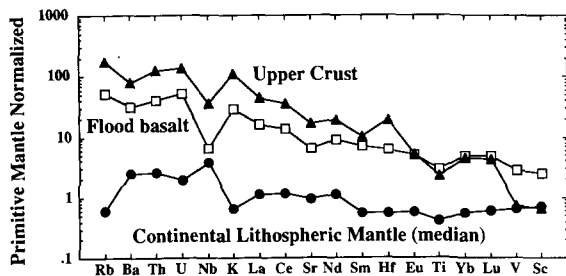


Fig. 12. An incompatible element diagram comparing the median composition of spinel peridotites, as an estimate of the bulk composition of the continental lithospheric mantle, and a typical flood basalt [44]. Also shown for comparison is an estimate of the bulk upper crust [39], which shows many geochemical similarities to continental flood basalts. The normalizing values are the same as those used in Fig. 5.

TABLE 5

Proportional contribution of the continental lithospheric mantle to the primitive mantle for selected elements.

	Concentration			Proportional	
	PM	Avg.	Med.	Avg.	Med.
Si%	20.9	20.6	20.6	1.4	1.4
Mg%	22.5	25.0	24.9	1.6	1.6
Al%	2.36	1.20	1.16	0.74	0.71
Ti	1280	545	545	0.80	0.62
Rb	0.65	1.9	0.38	4.2	0.85
Sr	21	49	20	3.4	1.4
La	0.71	2.6	0.77	5.3	1.6
Eu	0.168	0.16	0.097	1.4	0.85
Yb	0.48	0.26	0.27	0.79	0.82
Hf	0.31	0.27	0.17	1.3	0.80
U	0.024	0.12	0.040	7.3	2.4

Calculation assumes that the continental lithospheric mantle is 1.45% the mass of the primitive mantle. PM = primitive mantle [7,49]. Avg. = average and Med. = median, values from Table 1. Concentrations are expressed in ppm, unless otherwise stated.

It has been suggested the lithospheric mantle contributes to the composition of intraplate alkaline basalts [45–47]. Given the above mentioned similarities between intraplate alkaline basalts and the estimated bulk composition of the CLM, then this would be consistent with the suggestion that the CLM can be a source, in part or wholly, for some intraplate magmas.

8. Mass balance model involving the continental lithospheric mantle

The extraction of continental crust has had a significant effect on the composition of the residual mantle and by implication, the production of the CLM from mantle peridotite has also affected this balance. Therefore, in considering mass balance models for crust-mantle differentiation, it is necessary to also consider the composition of the CLM.

The mass of the continental crust is about 0.57% that of the mantle [39]. Underlying the continental crust is lithospheric mantle with a density of 3.34 g cm⁻³. If this lithospheric mantle is assumed to be about 200 km thick, then the mass of the CLM would be about 8.6 × 10²⁵ g and the mass of the CLM would represent about 1.5% of the mass of the earth's primitive mantle.

Using the average and median composition of these xenoliths (Table 1) as the bulk composition

of the lithospheric mantle, the proportional contribution of the CLM to the total mass of the primitive mantle can be calculated (Table 5). This calculation indicates that the CLM does not represent a significant reservoir for the incompatible elements. A separate mass balance calculation based on a model composition for the crust and various mantle reservoirs [48] is in good agreement with these findings.

Although a few elements (F, C, Cl, Nb, Ru, Ba, Ta, W and Th) appear to be strongly enriched in peridotites, the data suggest that the CLM does not represent a major reservoir for these elements. However, for most of these elements, there are few data available and their estimated primitive mantle values are poorly constrained. Finally, high Th concentrations in the average and median composition may be systematically in error, given most data are from neutron activation analyses (see footnote in Table 1).

9. Summary

The bulk composition of the continental lithospheric mantle (CLM) can be characterized by the average composition of peridotite xenoliths, although for elements that show a strong degree of skewedness in their data the median composition may provide a more reasonable estimate. Given a possible chemical distinction between Archaean and post-Archaean regions of the CLM, the spinel peridotite composition may only be appropriate for the post-Archaean CLM. The compositional estimate of the CLM indicates that it is enriched in the LREE and other incompatible elements, with La being enriched by a factor of 3 above the levels in C1 chondrite. Several element ratios (e.g., Zr/Hf, Y/Ho, Nb/Ta) show relatively minor or no fractionation from their primitive mantle values, indicating similar partition coefficients for such element pairs during melting and metasomatism. The major and trace element characteristics of the CLM is consistent with a two-stage growth history, involving melt depletion followed by incompatible element enrichment.

Individual peridotite xenoliths with LREE-depleted compositions have incompatible element patterns comparable to MORB residues, while those with LREE-enriched compositions have incompatible element patterns similar to intraplate

basalts. Key element ratios indicate that intraplate and/or divergent-margin tectonic environments where important sites of second-stage incompatible element enrichments for these peridotites. In general, peridotite xenoliths do not display any of the characteristic geochemical signatures indicative of convergent-margin (i.e., island arc) magmatism, although regions of CLM sampled by xenolith-bearing lavas should have recorded such magmatism, given its evidence in the overlying crustal rocks of the region.

The geochemical characteristics of the bulk of the CLM are consistent with it being a source for intraplate basalts, however, it does not show the characteristics needed to act as a major source for flood basalts. Using the average and/or median composition of spinel peridotites as best representative of the CLM composition, then this region of the lithosphere does not represent a significant reservoir for the incompatible elements.

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