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Chemical and isotopic relationships between peridotite xenoliths and mafic–ultrapotassic rocks from Southern Brazil

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Abstract

Peridotite xenoliths from the late-Cretaceous Alto Paranaíba and Goiás mafic–alkalic magmatic provinces of central and southeast Brazil reveal the existence of compositionally and temporally distinct lithospheric mantle beneath these areas. Garnet and spinel–lherzolites and spinel–harzburgites from the Alto Paranaíba province are generally depleted in Ca, Al and Re, which indicates that they are residues of melt extraction. Old Re-depletion model ages (average 2.4 Ga) for these peridotites show that this area is underlain by the early-Proterozoic to late-Archean melt-depleted lithospheric mantle of the São Francisco Craton. In contrast, spinel–lherzolites from the Goiás alkalic province, located 500–600 km northwest of the Alto Paranaiba province, have major- and trace-element compositions similar to modern fertile mantle. Most Goiás peridotites have fertile mantle $^{187}Os/^{188}Os$ (0.1261–0.1292), but two samples with lower $^{187}Os/^{188}Os$ define Re-depletion model ages of ~1.2 Ga.

The compositional distinction between the lithospheric mantle samples is mirrored in the kamafugitic rocks of these two provinces, which suggests that the composition of these magmas was strongly influenced, or determined entirely, by the lithospheric mantle. Most of the kamafugitic rocks have $^{187}Os/^{188}Os$ higher than observed for peridotite, which suggests that the source of these magmas is a mixture of lithospheric peridotite of varying age, veined and/or interlayered with various olivine-poor components. The relatively limited range in Sr, Nd, and Hf isotopic composition of the mafic–alkalic magmas indicates that the olivine-poor component was added regionally, overprinting the incompatible-element characteristics of the lithospheric mantle beneath these provinces. The age of metasomatism in the lithospheric mantle of this area is poorly constrained, but the isotopic characteristics of the mafic–alkalic magmatism suggest that this event may have occurred during the mid- to late-Proterozoic assembly of the Brasília mobile belt.

Keywords: Radiogenic isotopes; Xenoliths; Lithospheric mantle; Alkalic volcanism; Brazil; Igneous geochemistry

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1. Introduction

The largest kamafugite (Wooley et al., 1996) province in the world occurs in Minas Gerais and Goiás States in southern Brazil (Gaspar and Danni, 1981; Danni, 1985; Danni and Gaspar, 1992, 1994; Gibson et al., 1995a,b; Sgarbi and Valenca, 1995; Carlson et al., 1996; Sgarbi et al., 1998; Araujo et al., 2001a,b; Sgarbi and Gaspar, 2001). The kamafugites, along with other silica under-saturated rock types, such as kimberlite and carbonatite, are found in a number of provinces (Fig. 1) on the northern and western margins of the Paraná Basin of southern Brazil, the eruption sight of ~800,000 km^3 of flood-basalt lavas between 138 and 129 Ma (Peate, 1997). Subsequent alkalic magmatism, with an age range between 90 and 55 Ma (Bizzi, 1993; Gibson et al., 1995a; Thompson et al., 1998) post-dates the flood-basalt activity, but shares some important chemical and isotope characteristics with the flood basalts. Among these are: (i) the presence of low- and high-Ti compositional groups in both the flood basalts (Mantovani et al., 1985; Piccirillo et al., 1989; Peate et al., 1992) and the mafic–alkalic magmas (Gibson et al., 1995b, 1997) and (ii) similar Sr, Nd and Pb isotopic compositions between the high-Ti Paraná basalts (Mantovani et al., 1985; Piccirillo et al., 1989; Peate et al., 1992), high-Ti mafic–alkalic magmas (Gibson et al., 1995b; Carlson et al., 1996; Gibson et al., 1997; Araujo et al., 2001a), and the lavas of the Rio Grande Rise, Walvis Ridge (Richardson et al., 1982; Gibson et al., 2005) and the central Atlantic island Tristan da Cunha (O'Nions et al., 1977), the current location of the mantle plume proposed as the source of the Paraná flood basalts.

Whether or not flood-basalt magmas originate from melting of continental lithospheric mantle (Hawkesworth et al., 1992) or upwelling mantle plumes (White and

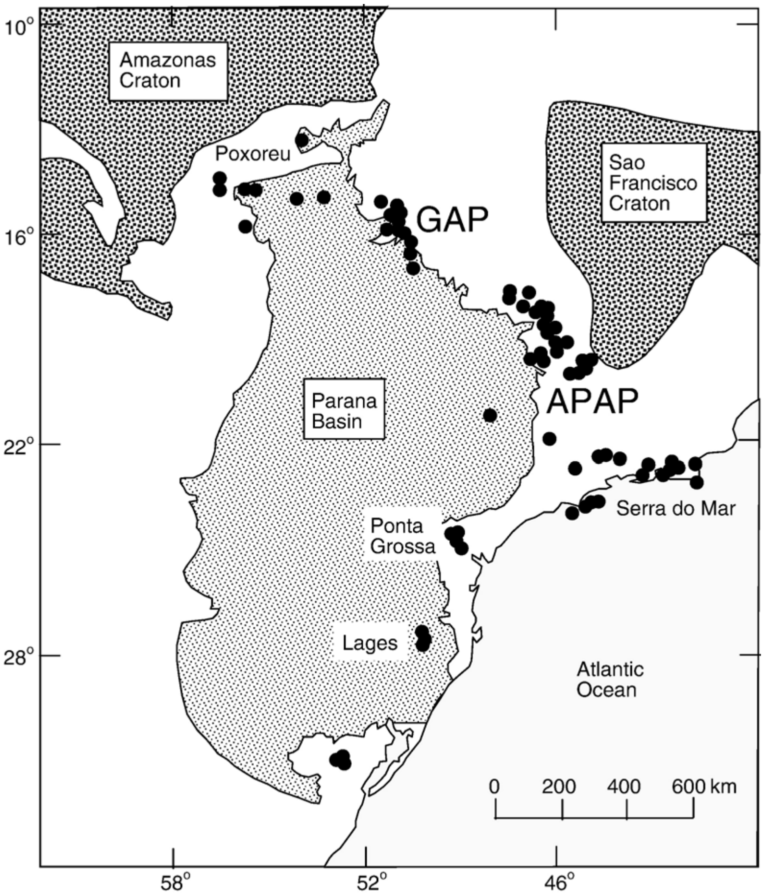

Fig. 1. Geological map of central and southern Brazil showing the location of the many late-Cretaceous mafic–alkalic magmatic provinces (large black circles) in relation to the two nearby Archean cratons and the Paraná basin. Both the Alto Paranaíba alkalic province (APAP) and Goías alkalic province (GAP) intrude the Mesoproterozoic Braziliano belt basement at the margin of the Paraná basin. Map modified from Gibson et al. (1997).

McKenzie, 1989) has been the topic of considerable debate. Because the majority of the mafic–alkalic magmatism in central and southeast Brazil post-dates Paraná flood-basalt activity by some 45 Myr (Amaral et al., 1967; Hasui and Cordani, 1968; Bizzi, 1993; Gibson et al., 1995a; Sonoki and Garda, 1998), this magmatism clearly cannot be related to the same thermal event that caused flood-basalt volcanism. Similarly, there is no good reason to assume that the very distinctive isotopic composition typical of Tristan da Cunha and high-Ti Paraná flood basalts was resupplied by a new plume over a wide area of southern Brazil at 85 Ma when the center of the Tristan plume was far off the east coast of Brazil following opening of the South Atlantic Ocean. Nevertheless, this isotopic signature is present in many of the mafic–alkalic magmas, which suggests that there may be some connection with the flood-basalt activity. Because the mafic–alkalic magmas are much smaller degree mantle melts than flood basalts, the composition of the mafic–alkalic magmas may more likely indicate the role of small-volume, but incompatible-element-rich, components in the lithospheric mantle. Some of the mafic–alkalic intrusions contain xenoliths from the shallow lithospheric mantle that can be examined to determine if the magmas contain compositional signatures consistent with those shown by xenoliths from the different mantle basement terranes through which they were emplaced.

In order to address these issues, we report new major- and trace-element and Sr and Nd isotopic data for kamafugites from the Alto Paranaíba and Goiás alkalic provinces. We add Hf and Os isotope data for these rocks in order to address whether the sources of these magmas were dominantly peridotite or contained significant amounts of olivine-poor (i.e. metasomatic) components. In addition, we report compositional and Os isotopic data for peridotite xenoliths contained in the mafic–alkalic lavas that allow us to directly address the role of lithospheric mantle as a source for these magmas.

2. Geologic setting

Radiometric ages for the Goiás alkalic province cluster between 80 and 85 Ma and for the Alto Paranaíba between 80 and 90 Ma (Amaral et al., 1967; Hasui and Cordani, 1968; Bizzi, 1993; Gibson et al., 1995a; Sonoki and Garda, 1998). The Goiás and Alto Paranaíba alkalic provinces are members of an NW–SE lineament (Fig. 1) of late-Cretaceous mafic–alkalic magmatism in Brazil that has been postulated to represent a mantle plume track (Gibson et al., 1995a,b; Thompson et al.,

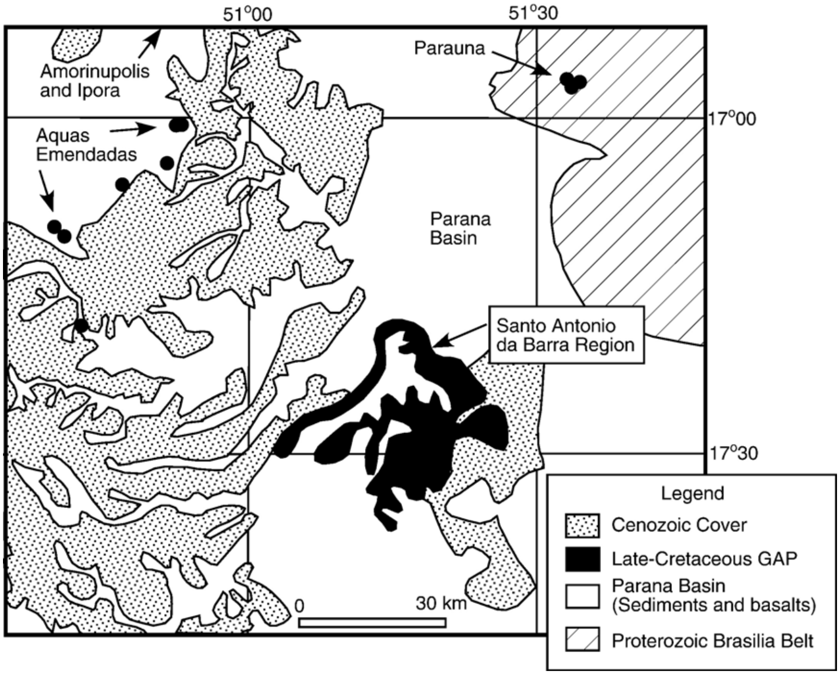

Fig. 2. Geological map of the southern portion of the late-Cretaceous Goías Alkalic Province, showing the Santo Antônio da Barra, Paraúna, and Águas Emendadas regions. Modified from Junqueira-Brod (1998).

1998; Gibson et al., 1999) although the temporal progression of this trend is not obvious.

In the Alto Paranaíba alkalic province, the kamafugites are temporally and spatially associated with kimberlites and carbonatites (Almeida and Svisero, 1991; Sgarbi and Valenca, 1993; Gibson et al., 1995a; Sgarbi and Valenca, 1995; Carlson et al., 1996; Brod et al., 2000; Araujo et al., 2001a). The Alto Paranaíba alkalic province rocks occur as small intrusions, carbonatitic complexes and the voluminous kamafugitic lava flows of the Mata da Corda Formation (Herz, 1977; Almeida, 1986; Almeida and Svisero, 1991; Leonardos et al., 1991; Sgarbi and Valenca, 1993; Gibson et al., 1995a; Sgarbi and Valenca, 1995; Carlson et al., 1996; Brod et al., 2000; Araujo et al., 2001a). In the Goiás alkalic province (Fig. 2), rocks with kamafugitic affinities occur in several localities including Santo Antônio da Barra, Aguás Emendadas, Paraúna, Amorinópolis and Iporá (Gaspar and Danni, 1981; Danni, 1985; Danni et al., 1990; Danni and Gaspar, 1994; Gibson et al., 1995a; Junqueira-Brod et al., 2000; Sgarbi and Gaspar, 2001). The Goiás alkalic province rocks are mostly small intrusions, but include lava flows and pyroclastic deposits in the Santo Antônio da Barra area (Gaspar and Danni, 1981; Danni, 1985; Danni et al., 1990; Danni and Gaspar, 1994; Sgarbi et al., 1998; Junqueira-Brod et al., 2000; Sgarbi and Gaspar, 2001). These crop out in an area of at least 370 km^2, with a minimum calculated volume of 23 km^2 (Junqueira-Brod et al., 2000). The Alto Paranaíba magmas intrude the Neoproterozoic Brasília Mobile Belt near the southwestern tip of the Archean São Francisco Craton (Fig. 1). The Goiás alkalic province rocks also penetrate Brasília Belt basement, but intrude and overlie basalts and sedimentary rocks from the Paraná Basin (Figs. 1 and 2).

3. Sample petrography

Kamafugites from the Paraúna region in the Goiás province are very fine grained, but olivine, clinopyroxene and phlogopite can be recognized as their major constituents. Kamafugites from the Santo Antônio da Barra flows and Aguás Emendadas intrusions are aphanitic porphyritic or glomeroporphyritic in texture and contain clinopyroxene, olivine, kalsilite, leucite, analcime, and magnetite (Danni et al., 1990; Sgarbi et al., 2000; Sgarbi and Gaspar, 2001). The alkalic rocks from the Mata da Corda Formation (Alto Paranaíba kamafugitic flows) are aphanitic porphyritic to fine/medium-grained in texture and contain clinopyroxene, perovskite, magnetite, olivine, phlogopite, apatite, kalsilite, leucite and melilite pseudomorphs (Sgarbi and Valenca, 1991, 1995). The kamafugitic intrusions from the Alto Paranaíba alkalic province, though mineralogically and texturally similar to the Mata da Corda Formation and Goiás alkalic rocks, have higher olivine contents (Araujo et al., 2001a).

The mantle xenoliths from the Alto Paranaiba province derive from small kimberlite intrusions whereas those from the Goias province are from small kamafugite intrusions (Paraúna Region — Fig. 2). The Alto Paranaíba xenoliths are garnet- and spinel-bearing lherzolites and spinel–harzburgites. They are rounded, small (4–6 cm diameter), medium- to coarse-grained and show a slightly strained porphyroclastic to granular texture. The garnet lherzolites contain kink-banded olivine, pyroxene (diopside and enstatite), idiomorphic garnet and serpentine. Mineral contacts are irregular, with rare straight grain boundaries. The spinel–lherzolites and spinel–harzburgites are porphyroclastic to granular in texture and consist of kink-banded irregular to euhedral olivine grains, pyroxene, irregular and/or interstitial red, brown and black spinels, serpentine, phlogopite and carbonate. The phlogopite occurs as continuous interstitial aggregates, and in kelyphitic borders to garnets, probably associated with later metasomatic introduction. Kelyphitic borders around garnet and pyroxenes, and veins filled with carbonate and serpentine, also are common features in these rocks.

The spinel–lherzolites from the Goiás alkalic province are rounded, up to 7 cm diameter, coarse-grained, predominantly granular in texture, although strained textures also have been observed. They are mineralogically similar to spinel–peridotites from the Alto Paranaíba province, except for their lesser abundance of spinel and by the absence of phlogopite. Kelyphitic borders around pyroxene and spinel grains, serpentine fill in fractures in olivine, and veins filled with serpentine and carbonate, also are present. Mineral thermobarometry gives equilibration temperatures of 850 ± 80 °C and pressures of 1.4 ± 0.3 GPa indicating that these are shallow mantle samples from an area with a hot geotherm (Danni et al., 1994).

4. Analytical methods

The xenoliths were manually separated from the host rock with a hammer and then crushed to powder with an alumina disk mill. Only interior fragments of the xenoliths were crushed in order to avoid including any adhering host rock. Kamafugite samples from the Alto Paranaíba province were powdered in a similar manner. Aliquots of the powdered samples were analyzed for major- and trace-element composition by XRF at

Franklin and Marshall University by Stanley Mertzman. Procedures for these analyses are described in Boyd and Mertzman (1987). Kamafugite samples from the Goiás Province were analyzed at the Universities of Durham and Brasilia (Junqueira-Brod, 1998), using a combination of XRF for major elements and ICP-MS for trace elements, with additional XRF analyses carried out at Universidad Nacional de Salta.

Isotopic analyses were carried out at the Department of Terrestrial Magnetism (DTM) except for the Sr and Nd isotopic measurements of the Goiás alkalic rocks (see below). At DTM, Sr, Nd and Hf isotope analyses of the kamafugites were performed on ~50 mg samples that were digested in HF and HNO_3 in closed beakers. Samples were spiked with isotopically enriched tracers of Rb, Sr, Sm, Nd, Lu and Hf prior to dissolution. Chemical separations were performed using the procedures described by Carlson et al. (2004). Sr isotopic compositions were determined by thermal ionization mass spectrometry on the DTM VG-354. Sr is fractionation corrected to $^{86}Sr/^{88}Sr = 0.1194$ and reported relative to a value of $^{87}Sr/^{86}Sr = 0.71025$ for the NIST 987 Sr standard. Rb, Sm, Lu, Nd and Hf isotopic compositions were measured on the DTM VG-P54 multi-collector ICP-MS. Nd is fractionation corrected to $^{146}Nd/^{144}Nd = 0.7219$ and reported relative to a value of $^{143}Nd/^{144}Nd = 0.511860$ for the La Jolla Nd standard. Hf is fractionation corrected to $^{179}Hf/^{177}Hf = 0.7325$ and reported relative to a value of $^{176}Hf/^{177}Hf = 0.282160$ for the JMC-475 Hf standard. External reproducibility for Sr, Nd and Hf isotopic compositions are better than 0.0025% using these procedures.

For the samples from Goiás, Sm–Nd isotopic analysis followed the method described by Gioia and Pimentel (2000) and were carried out at the Geochronology Laboratory of the University of Brasília. Whole-rock

Table 1

Major and trace elements and Os isotopic composition of peridotite xenoliths from the Três Ranchos and Coromandel kimberlites, Alto Paranaíba Province

Sample	X721	APC7-1	AP03-4	APC7-2	X219a	X219	AP03-2	CT-2
Rock type	Sp–Hz	Gt–Lhz	Gt–Lhz	Gt–Lhz	Sp–Lhz	Sp–Lhz	Gt–Lhz	Gt–Lhz
SiO_2 (wt.%)		43.23	38.89			39.75		41.03
TiO_2		0.09	0.12			0.24		0.20
Al_2O_3		1.88	0.34			0.91		0.90
Fe_2O_3		1.30	1.04			1.64		1.67
FeO		5.54	7.18			10.25		6.45
MnO		0.12	0.12			0.19		0.15
MgO		40.55	46.21			40.00		40.74
CaO		0.76	0.28			3.12		1.03
Na_2O		0.24	0.14			0.22		0.19
K_2O		1.19	0.13			0.13		0.89
P_2O_5		0.03	0.05			0.18		0.05
LOI		4.23	4.94			2.67		5.82
Total		99.16	99.44			99.30		99.12
Mg#		0.92	0.91			0.86		0.90
Sr (ppm)		149	130			204		159
Zr		31	37			46		53
V		95	69			81		94
Ni		2390	3115			2685		2545
Cr		6200	202			1445		2745
Co		102	118			109		97
Re (ppb)	0.23	0.10	0.33	0.20	0.10	2.06	0.11	0.16
Os (ppb)	6.36	8.13	2.61	2.65	3.84	4.09	3.00	2.98
$^{187}Re/^{188}Os$	0.1740	0.0595	0.6046	0.3694	0.1267	2.423	0.1840	0.2524
$^{187}Os/^{188}Os$	0.10990	0.11157	0.11228	0.11447	0.11203	0.10915	0.11517	0.11411
$^{187}Os/^{188}Os$ (i)	0.10965	0.11149	0.11142	0.11395	0.11185	0.10572	0.11491	0.11375
T_{RD} (Ga)	2.64	2.39	2.40	2.05	2.34	3.17	1.92	2.08
T_{MA} (Ga)	4.40	2.77	−6.05	14.29	3.31	−0.59	3.35	5.00

Rock type abbreviations: Sp — spinel, Gt — garnet, Hz — harzburgite, Lhz — lherzolite. Mg# is the molar ratio of (Mg/(Mg+Fe)). $^{187}Os/^{188}Os$ (i) is the initial Os isotopic composition calculated for an age of 85 Ma. T_{RD} and T_{MA} are the Re-depletion and Re–Os model ages (Walker et al., 1989), respectively in Ga.

powders (ca. 50 mg) were mixed with ^{149}Sm–^{150}Nd spike solution and dissolved in Savillex capsules. Sm and Nd extraction from whole-rock samples followed conventional cation-exchange techniques followed by teflon columns containing LN-Spec resin (HDEHP — di-ethylhexil phosphoric acid supported on PTFE powder). Sm and Nd samples were loaded on Re evaporation filaments of double filament assemblies and the isotopic measurements were carried out on a multi-collector Finnigan MAT 262 mass spectrometer in static mode. Uncertainties for Sm/Nd and ^{143}Nd/^{144}Nd ratios are better than ±0.4% (2σ) and ±0.005% (2σ) respectively, based on repeated analyses of international rock standards BHVO-1 and BCR-1. ^{143}Nd/^{144}Nd ratios were normalized to ^{146}Nd/^{144}Nd of 0.7219. Sr isotopic analysis for these samples followed conventional dissolution and cation-exchange procedures similar to those described by Pankhurst and O'Nions (1973). Samples were loaded onto double filament arrangements and the analyses were carried out in static mode in a multi-collector Finnigan MAT-262 mass spectrometer. Mass fractionation corrections were made using ^{88}Sr/^{86}Sr equal to 8.3752. 2σ uncertainties on the measured ^{87}Sr/^{86}Sr are better than 0.01%.

For the xenolith and kamafugite Re–Os analyses, 2 g of sample, and respective spikes, were digested in HCl–HNO_3 in sealed Pyrex tubes held at 240 °C overnight. Os was extracted from the aqua-regia into CCl_4, and then back-extracted into HBr. The Os was further purified by a microdistillation step from an H_2SO_4–CrO_3 mixture. Re was separated by anion-exchange chromatography. Re and Os isotopic compositions were determined by negative thermal-ionization mass spectrometry. Procedures for Re–Os separation and isotopic measurement are described in more detail in Carlson et al. (1999).

5. Results

Major- and trace-element analyses of the mantle xenoliths are given in Tables 1 and 2 with major element variation diagrams (renormalized to anhydrous) shown

Table 2
Major and trace elements and Os isotopic composition of peridotite xenoliths from Paraúna, Iporá Province, Goiás

Sample	IP03	IP2.5	IP-01	MBO-01	IP2.7e	IP2.7b	IP-02	IP2.6	IP2.7c
Rock type	Sp–Lhz	Sp–Lhz	Sp–Lhz	Sp–Lhz	Sp–Lhz	Sp–Lhz	Sp–Lhz	Sp–Lhz	Sp–Lhz
SiO_2 (wt.%)	44.18	43.54	43.02	44.13	43.07	42.65	43.30	43.42	43.98
TiO_2	0.18	0.19	0.20	0.16	0.15	0.20	0.23	0.25	0.20
Al_2O_3	4.40	2.74	3.33	4.06	3.49	5.98	4.26	3.29	4.10
Fe_2O_3	1.81	1.82	1.37	1.61	1.61	1.91	1.66	1.89	1.73
FeO	6.64	6.76	7.31	6.78	7.08	6.48	6.27	6.85	6.36
MnO	0.15	0.14	0.15	0.15	0.15	0.17	0.14	0.14	0.15
MgO	37.05	39.00	38.89	37.41	39.26	35.27	35.52	36.60	36.10
CaO	3.46	3.30	2.58	2.76	2.64	3.47	4.92	3.41	4.05
Na_2O	0.43	0.39	0.44	0.42	0.39	0.47	0.57	0.52	0.48
K_2O	0.07	0.15	0.29	0.15	0.17	0.22	0.21	0.26	0.16
P_2O_5	0.03	0.04	0.06	0.05	0.05	0.05	0.04	0.06	0.04
LOI	1.23	1.87	2.01	1.74	1.87	2.00	2.10	2.37	2.40
Total	99.63	99.94	99.65	99.42	99.54	98.87	99.22	99.06	99.75
Mg#	0.89	0.89	0.89	0.89	0.89	0.88	0.89	0.88	0.89
Sr (ppm)	57	105	119	103	85	96	108	120	94
Zr	32	38	46	36	35	39	37	42	35
V	106	100	90	98	92	114	116	109	116
Ni	2295	2600	2440	2215	2365	2240	2290	2525	2025
Cr	2465	1470	1695	2425	1710	4050	2570	1850	2405
Co	98	106	110	103	110	103	97	105	96
Re (ppb)	0.10	0.34	0.19	0.15	0.31	0.23	0.28	0.18	0.32
Os (ppb)	1.24	2.02	2.34	3.16	2.08	2.93	1.95	2.58	3.01
^{187}Re/^{188}Os	0.4031	0.8013	0.3804	0.2276	0.7136	0.3774	0.6972	0.3336	0.5101
^{187}Os/^{188}Os	0.12866	0.12728	0.12097	0.12646	0.12097	0.12881	0.12972	0.12834	0.12993
^{187}Os/^{188}Os (i)	0.12809	0.12614	0.12043	0.12614	0.11996	0.12828	0.12873	0.12787	0.12921
T_{RD} (Ga)	0.09	0.36	1.16	0.36	1.22	0.06	0.00	0.12	−0.07
T_{MA} (Ga)	1.71	−0.41	10.36	0.78	−1.84	0.54	0.01	0.55	0.35

^{187}Os/^{188}Os(i) is the initial Os isotopic composition calculated for an age of 85 Ma.

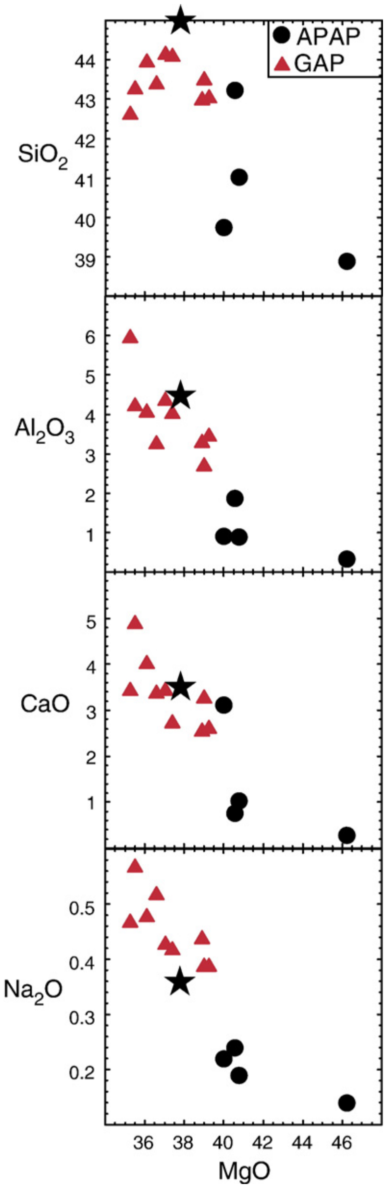

Fig. 3. MgO versus SiO$_2$, Al$_2$O$_3$, CaO, FeO(t), TiO$_2$, and Na$_2$O concentrations for xenoliths from the Alto Paranaíba and Goiás alkalic provinces. The large star shows the composition of fertile mantle from McDonough and Sun (1995). The xenolith data have been renormalized on an anhydrous basis.

in Fig. 3. The Alto Paranaíba peridotite xenoliths have high MgO content and Mg# (mostly >0.90) with low Al$_2$O$_3$ (0.34–1.88 wt.%), CaO (0.28–3.12 wt.%) and Na$_2$O (0.14–0.24 wt.%) contents (Table 1) with no clear compositional distinction between garnet and spinel facies. One Alto Paranaíba xenolith (X219) has a high CaO content for its very low Al$_2$O$_3$ concentration, most likely reflecting metasomatic addition to this sample. This sample also is enriched in Fe, Sr, and particularly Re compared to the other Alto Paranaíba samples, consistent with a metasomatic addition. The scatter in SiO$_2$ concentration might reflect some input of SiO$_2$ to these samples through alteration, but the routinely low Na$_2$O concentrations and the good Na$_2$O vrs MgO correlation of these samples suggest that alteration has not dramatically modified the major element concentration of these xenoliths.

The Goiás (Paraúna) spinel peridotites have much more fertile compositions than the Alto Paranaíba xenoliths with lower Mg# (0.88–0.89) and higher Al$_2$O$_3$ (2.74–5.98 wt.%), CaO (2.58–4.92 wt.%) and Na$_2$O (0.39–0.57) contents (Table 2) that overlap estimates of fertile-mantle composition in these elements (McDonough and Sun, 1995). The Na$_2$O contents of the Goias samples are high compared to fertile-mantle estimates (Fig. 3), and we have no obvious explanation for this. Na can be added by both metasomatism and alteration. Metasomatism also usually results in additions in Ti, Fe and Ca, yet most of the Goias samples have concentrations of these elements either similar to, or lower than, primitive mantle estimates (Fig. 3). In support of their more refractory nature, the Alto Paranaíba samples have slightly higher average Ni (2684 vrs 2368 ppm) and Cr (2648 vrs 2293 ppm) concentrations compared to the Goiás samples, but the Alto Paranaíba samples also have slightly higher Sr concentrations (average 161 vrs 99 ppm) compared to the Goiás samples suggestive of some metasomatic addition to the Alto Paranaiba samples.

Re and Os concentrations overlap between the two groups of xenoliths although the Alto Paranaíba xenoliths tend to have higher Os concentrations (mean 4.21 vrs 2.37 ppb) compared to the Goiás peridotites. All but two of the Alto Paranaíba xenoliths have ^{187}Re/^{188}Os less than estimated for the fertile mantle (^{187}Re/^{188}Os = 0.435, Meisel et al., 2001) consistent with the melt-depleted character of these peridotites. Re/Os ratios of the Goiás samples generally are higher than those in the Alto Paranaíba xenoliths and provide an average ^{187}Re/^{188}Os (0.49) similar to that estimated for fertile mantle. The Re/Os ratios of the Goias peridotites provide another line of evidence that the composition of these xenoliths does not reflect metasomatic refertilization. Metasomatism can drastically raise the Re/Os ratio of mantle peridotite, as shown, for example, by the Alto Paranaiba sample X219.

The Alto Paranaíba xenoliths have low ^{187}Os/^{188}Os values (initial ^{187}Os/^{188}Os between 0.10572 and 0.11491) that are significantly below those estimated for modern fertile mantle (0.1296, Meisel et al., 2001). As a result, these samples have old Re-depletion model ages ranging from 1.92 to 3.17 Ga (mean = 2.4 ± 0.4 Ga). Two of the Goiás peridotites have distinctly lower $^{187}Os/^{188}Os$ values (IP-01 = 0.12043 and IP2.7e = 0.11996) then do the remainder of the Goiás samples that have initial $^{187}Os/^{188}Os$ >0.126. The two low $^{187}Os/^{188}Os$ Goiás samples have Re-depletion ages of 1.16 and 1.22 Ga, but most of the Goiás samples have Os isotopic compositions ($^{187}Os/^{188}Os$ = 0.1261 to 0.1292) within the range of estimates for modern fertile mantle, and hence give no reliable model age information.

The kamafugites from the Alto Paranaíba (Table 3) and Goiás (Tables 4 and 5) alkalic provinces show similar ranges in MgO, FeO_t, and CaO, but are distinct in other major element abundances, particularly SiO_2, TiO_2, Al_2O_3, and Na_2O (Fig. 4). The Alto Paranaíba kamafugites studied here represent only a small fraction of the chemical variability in this province where rock compositions span the range from kimberlite to carbonatite (Carlson et al., 1996; Araujo et al., 2001a).

Table 3
Major, trace element, and isotopic results for Alto Paranaíba kamafugites

Sample	QP-37-980	QP-191	QP221-6	QP-197-2	QP-229-5
SiO_2	36.50	38.87	34.84	38.03	36.91
TiO_2	7.48	5.46	7.18	4.43	7.24
Al_2O_3	4.72	7.32	4.98	5.33	4.73
Fe_2O_3	11.89	11.21	10.36	8.12	10.62
FeO	2.72	1.56	4.45	5.79	4.29
MnO	0.24	0.17	0.20	0.22	0.22
MgO	9.24	6.48	9.78	16.43	9.08
CaO	15.06	11.36	16.20	11.02	17.79
Na_2O	0.57	1.76	0.43	1.01	0.81
K_2O	1.12	1.61	1.17	1.57	2.06
P_2O_5	1.21	1.20	1.78	0.92	1.80
LOI	7.09	6.82	7.79	6.31	2.54
Total	97.84	93.82	99.16	99.18	98.09
Mg#	0.55	0.50	0.56	0.69	0.54
Rb (XRF)	119	594	149	174	97
Ba	15900	50100	5210	3560	12100
Th	29.9	21.9	24.6	22.4	26.1
U	6.9	7.3	8.1	4.8	6.1
Nb	242	253	264	207	237
La	180	170	162	177	213
Ce	344	374	279	324	379
Sr (XRF)	1698	1960	1702	1445	2300
Zr	776	708	732	650	262
Y	35.7	29.7	34.7	27.3	38.1
Ga	10.4	3.6	14.6	10.8	7.8
Zn	139	112	177	105	101
Co	54	35	54	77	53
Cr	140	13	94	870	58
Cu	174	134	180	100	44
Ni	264	45	109	696	99
Sc	40	17	20	21	41
V	349	228	450	181	303
Rb (ID)	123	583	149	185	95
Sr (ID)	1747	1984	1728		
$^{87}Rb/^{86}Sr$	0.2032	0.8500	0.2499		
$^{87}Sr/^{86}Sr(m)$	0.704556	0.706080	0.705684		
$^{87}Sr/^{86}Sr(i)$	0.704311	0.705053	0.705382		
Sm	35.8	22.9	28.7	23.3	31.2
Nd	268	160	206	179	215
$^{147}Sm/^{144}Nd$	0.08060	0.00867	0.08432	0.07895	0.08764
$^{143}Nd/^{144}Nd$	0.512496	0.512316	0.512320	0.512298	0.512313
εNd(i)	−1.5	−4.2	−4.9	−5.3	−5.1
T_{DM} (Ga)	0.80	0.65	1.02	1.01	1.06

Major elements are in wt.%, trace elements in ppm. Initial Nd isotopic composition calculated for an age of 85 Ma and the εNd is relative to modern bulk-earth $^{147}Sm/^{144}Nd$ = 0.1966 and $^{143}Nd/^{144}Nd$ = 0.512636.

Table 4
Major- and trace-element results for mafic–alkalic rocks from the Santo Antônio da Barra Region, Goiás alkalic province

Sample	DG-10A	DG-11A	DG-11C	DG-14A	DG-16C	DG-18A	LM-17	01SB	02SB	03SB	05SB
Lat.								17.507	17.507	17.507	17.501
Long.								50.656	50.656	50.656	50.683
SiO_2	41.68	42.67	41.79	38.89	39.68	40.67	42.25	38.55	41.51	41.57	38.56
TiO_2	3.32	3.58	3.73	3.78	3.55	4.22	3.18	3.87	3.34	3.86	4.91
Al_2O_3	7.71	10.61	11.06	7.75	8.08	11.63	10.21	7.48	7.34	10.57	11.99
Fe_2O_3	4.09	5.13	6.59	6.53	7.15	6.70	6.37	13.96	11.28	13.41	13.99
FeO	6.27	6.25	5.60	5.96	5.30	6.69	5.43				
MnO	0.17	0.20	0.22	0.20	0.20	0.23	0.20	0.19	0.17	0.22	0.22
MgO	13.68	8.47	7.59	15.24	13.62	5.43	8.17	14.88	13.58	7.39	8.26
CaO	14.82	13.10	13.27	13.45	15.53	13.02	14.05	13.73	15.32	13.86	12.44
Na_2O	1.22	3.68	4.19	2.27	1.53	2.36	2.76	1.81	1.31	3.89	2.56
K_2O	3.99	2.84	1.05	0.50	0.83	1.85	0.90	0.57	3.02	1.07	0.67
P_2O_5	0.69	0.84	0.91	0.60	0.62	0.97	0.86	0.52	0.60	0.79	0.73
LOI	1.52	1.73	3.29	4.06	3.72	5.51	4.88	4.24	1.80	2.61	5.06
Total	99.16	99.10	99.29	99.23	99.81	99.28	99.26	99.79	99.26	99.40	99.38
Mg#	0.71	0.58	0.54	0.70	0.67	0.43	0.57	0.68	0.70	0.52	0.54
Rb	153	233	52.8	48.6	33.0	102	124	66.0	137	54.7	52.0
Ba	1660	1760	1510	1315	1420	1690	2970	1434	1559	1775	1253
Th	12.8	13.7	14.8	10.3	8.2	13.0	13.9	9.59	11.5	13.7	10.2
U	2.4	3.0	2.4	1.6	2.3	2.5	1.2	2.2	2.7	3.5	2.4
Ta								7.85	11.7	10.2	9.48
Nb	97	116	121	90	70	121	111	109	128	157	132
La	89	104	115	65	64	108	123	80	110	122	95
Ce	167	196	214	114	140	214	249	160	210	247	214
Sr	1008	1120	1381	887	1020	1395	2305	1090	1187	1390	1592
Nd								67.4	86.0	99.5	96.5
Hf								7.22	7.74	11.3	12.3
Zr	215	338	356	254	243	387	322	298	286	452	520
Sm								10.7	12.1	14.2	15.6
Eu								2.93	3.07	4.10	4.26
Gd								6.9	8.1	9.9	11.4
Tb								0.91	0.96	1.26	1.39
Dy								3.91	4.17	5.92	7.16
Y	16.1	22.8	25.3	17	17.2	28.5	23.5	17.3	18.6	27.8	32.4
Er								1.37	1.55	2.22	2.94
Yb								1.02	0.93	1.88	2.30
Lu								0.15	0.12	0.25	0.31
Ga	12.5	15.3	15.5	15.6	12.6	15.6	14.8	16.7	14.2	18.3	21.2
Zn	64	94	98	84	76	106	91	247	60	83	105
Co	55	44	39	68	62	44	46	94	534	116	104
Cr	802	302	185	1140	772	56	240	1236	839	208	90
Cu	158	125	84	103	85	144	55	69	54	66	51
Ni	209	98	66	361	295	56	63	328	211	47	70
Sc	41	36	37	33	43	22	27	9	15	0	5
V	338	373	425	388	355	552	359	387	351	446	427

Major-element concentrations in weight percent, trace elements in ppm. Fe concentration is expressed as both FeO and Fe_2O_3 when FeO was measured by titration or total Fe as Fe_2O_3 when it was not.

In the Alto Paranaíba alkalic province, both kimberlites and kamafugites are characterized by strong enrichment in incompatible lithophile elements with primitive-mantle-normalized U and Th abundances in the range of 200–400 and normalized Y abundances near 6 (Fig. 5). The Alto Paranaíba samples analyzed here belong to the mafurite end-member from the Mata da Corda flows (Araujo et al., 2001a). Compositionally, the mafurites are distinguished from associated ugandites primarily by higher CaO, but distinctly lower Na_2O concentrations, and commonly large enrichments in Ba (Fig. 5).

The Goiás rocks studied here span the range mafurite–leucite mafurite–ugandite, which is representative of the composition of the kamafugites of the province (except for katungite, which occurs at a single

Table 5
Major and trace elements and isotopic results for mafic–alkalic rocks from the Paraúna region and nearby intrusions, Goiás alkalic province

Sample	BIP-02	BIP-2.6	02PR 01	BIP-03	96AE 04	96AE 08	96AE 11	96AE 48F	96AE 57	96AE 58	96AE 62	MON 01	T03	02DI 02
Lat.			16.966		17.024	17.024	17.169	17.019	17.124	16.717	16.693	17.315	17.080	16.221
Long.			50.721		51.144	51.144	51.350	51.121	51.293	51.068	51.039	51.317	51.163	51.256
SiO_2	39.71	39.47	45.54	39.84	41.07	38.71	41.74	45.73	40.67	39.42	37.04	39.14	41.36	44.75
TiO_2	3.41	3.44	3.73	3.41	4.58	4.68	3.74	4.00	5.63	4.16	5.56	4.91	4.75	3.64
Al_2O_3	9.32	9.24	9.48	8.93	8.90	9.40	7.42	12.39	10.11	10.11	6.27	7.89	9.1	12.31
Fe_2O_3	8.13	7.84	12.17	7.89	12.52	16.07	13.45	13.26	13.43	16.21	15.91	13.90	15.82	13.62
FeO	5.88	6.07		6.14										
MnO	0.26	0.26	0.16	0.26	0.18	0.25	0.20	0.23	0.19	0.24	0.21	0.17	0.23	0.19
MgO	10.56	10.65	7.90	10.82	15.63	10.42	16.59	6.53	13.05	10.93	16.80	12.98	9.6	6.98
CaO	14.09	14.13	13.17	13.64	11.81	13.89	12.64	11.05	12.19	13.60	11.27	14.02	15.58	10.99
Na_2O	3.00	2.80	1.53	2.97	2.78	2.51	2.50	3.77	3.18	1.88	2.26	0.64	1.57	2.75
K_2O	1.48	1.24	2.97	1.21	1.26	3.04	0.39	2.49	0.79	1.81	3.02	0.85	1.05	2.09
P_2O_5	1.12	1.12	0.56	1.11	0.47	0.91	0.49	0.85	0.60	1.07	0.79	0.55	0.93	0.51
LOI	2.80	3.21	1.84	3.26	2.76	0.70	1.84	2.80	2.93	3.29	0.24	4.16	3.36	1.37
Total	99.76	99.47	99.15	99.48	99.20	99.88	99.16	100.30	99.85	99.43	99.13	99.22	99.99	99.21
Mg#	0.59	0.59	0.56	0.59	0.71	0.56	0.71	0.49	0.66	0.57	0.68	0.65	0.55	0.50
Rb	54	56	66	68	103	86	36	48	80	51	98	44	79	36
Ba	1295	1260	1080	1225	1654	1463	879	2574	1692	896	1232	1568	1472	945
Th	15.2	15.1	11.9	16.0	8.64	14.5	7.35	11.1	7.83	12.2	13.0	8.68	5.00	7.16
U	3.3	2.6	3.0	2.3	1.9	3.4	1.5	2.3	1.9	2.8	3.1	2.2		1.8
Ta			7.58		6.83	11.1	5.22	7.31	6.08	8.60	9.73	6.91		6.66
Nb	133	132	108	134	106	178	86	124	99	136	160	99	96	83
La	120	119	76	119	75	125	62	89	65	109	111	71	82	67
Ce	216	217	165	216	147	242	125	179	131	220	228	154	189	145
Sr	1725	1633	958	1367	1098	1689	807	1275	1228	1901	1416	714	1089	906
Nd			76.1		63.6	106	59.6	82.2	60.0	108	106	70.6	81.6	70.3
Hf			9.59		6.33	11.2	6.29	7.88	6.60	13.4	11.0	8.88		8.63
Zr	506	502	330	501	259	475	266	343	273	587	486	315	273	371
Sm			12.8		9.4	15.8	9.6	13.2	9.7	18.6	16.9	11.7	12.5	11.5
Eu			3.18		2.68	4.29	2.72	3.99	2.87	5.22	4.55	3.15	3.70	3.27
Gd			9.6		6.0	10.5	7.2	9.7	6.9	14.3	11.2	8.4	9.08	8.9
Tb			1.23		0.79	1.35	0.92	1.28	0.91	1.93	1.41	1.12		1.23
Dy			6.40		3.52	6.23	4.26	6.08	4.22	9.25	6.19	4.72	6.36	5.59
Y	34.1	34.4	30.9	35.0	15.2	27.9	19.2	28.6	18.6	43.7	25.7	20.2	21.0	29.1
Er			2.86		1.29	2.39	1.66	2.51	1.57	3.55	1.97	1.59	2.04	2.59
Yb			2.25		0.94	1.76	1.22	2.03	1.18	2.60	1.35	1.29	2.19	2.08
Lu			0.40		0.13	0.25	0.17	0.30	0.17	0.35	0.17	0.16	0.62	0.31
Ga	20.1	19.8	18.4	19.0	15.8	21.2	15.7	14.8	18.9	22.0	19.7	16.2		19.2
Zn	113	114	39	111	112	148	113	126	126	158	141	69		75
Co	49	52	114	50	73	62	79	41	65	57	94	73	58	155
Cr	345	360	375	346	1341	85	1697	132	537	152	965	641	275	444
Cu	58	57	76	55	81	92	110	30	86	45	93	96		50
Ni	183	193	98	204	453	98	495	34	301	155	491	111	65	92
Sc	37	37	0	39	37	41	34	19	29	24	25	17		11
V	334	335	346	325	413	421	357	370	474	277	421	445	481	329

Major elements in wt.%, trace elements in ppm.

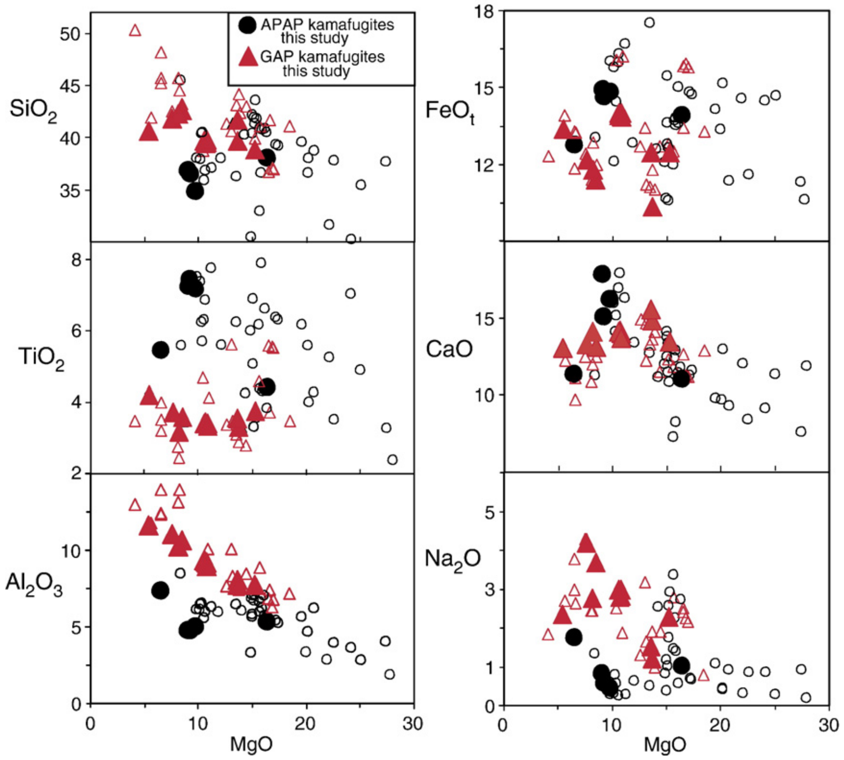

Fig. 4. Variation of MgO versus SiO_2, TiO_2, Al_2O_3, FeO, CaO and Na_2O in kamafugites from the Alto Paranaíba (APAP) and Goiás alkalic provinces (GAP). Smaller open symbols are from previous studies of the rocks of these provinces including (Gibson et al., 1995a,b; Carlson et al., 1996; Junqueira-Brod, 1998; Araujo et al., 2001a; Sgarbi and Gaspar, 2001).

locality). Compared to the Alto Paranaíba kamafugites, the Goiás kamafugites have higher Al_2O_3, and Na_2O, but lower TiO_2 and P_2O_5 (Tables 3 4 and 5; Fig. 4). The Goiás kamafugites have similar abundances of iron as do the Alto Paranaíba samples, but have much lower Fe^{3+}/Fe^{2+} ratios and loss on ignition, most likely reflecting the advanced alteration state of the Mata da Corda flows of the Alto Paranaíba alkalic province. Al_2O_3 shows a negative correlation with MgO abundances in the alkalic rocks from both provinces. The Goiás kamafugites have smooth primitive mantle-normalized incompatible-element patterns with less severe enrichment in highly incompatible elements compared to the Alto Paranaíba kamafugites (Fig. 5). They also do not show the extreme enrichments in Ba seen in some Alto Paranaíba alkalic province samples.

The Alto Paranaíba and Goiás kamafugites have overlapping in Sr and Nd isotopic composition ($^{87}Sr/^{86}Sr$ between 0.70445 and 0.70667, εNd between −1.0 to −5.3; Tables 3 and 6; Fig. 6), but other compositional types from the Alto Paranaiba extend this range to significantly lower εNd (Fig. 6). The two Goias samples with $^{87}Sr/^{86}Sr$ >0.706 plot well to the right (high $^{87}Sr/^{86}Sr$) of the Nd–Sr array (Fig. 6), most likely reflecting some degree of alteration in these samples. Surprisingly, given the fairly altered state of the Mata

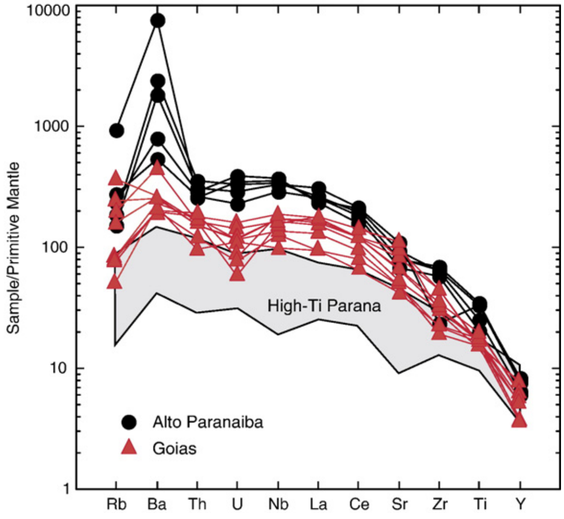

Fig. 5. Primitive mantle-normalized (McDonough and Sun, 1995) incompatible trace-element abundances in Alto Paranaíba and Goiás alkalic province rocks in comparison with high-Ti Paraná flood basalts (Mantovani et al., 1985; Piccirillo et al., 1989).

Table 6
Sr and Nd isotopic compositions of mafic–alkalic rocks from the Goiás alkalic province

Sample	Rb	Sr	^{87}Rb/^{86}Sr	^{87}Sr/^{86}Sr	^{87}Sr/^{86}Sr(i)	Sm	Nd	^{147}Sm/^{144}Nd	^{143}Nd/^{144}Nd	εNd(i)	T_{DM} (Ga)
DG-10A	159	1047	0.439	0.705575	0.705045	11.89	83.65	0.0859	0.512357	−4.2	0.99
DG-11A	239	1136	0.608	0.705526	0.704792	16.48	82.49	0.1207	0.512356	−4.6	1.36
DG-11C	55.8	1444	0.112	0.705329	0.705194	14.34	97.49	0.0889	0.512348	−4.4	1.03
DG-14A	53.8	930	0.167	0.704750	0.704548	10.51	67.73	0.0938	0.512472	−2.1	0.91
DG-16C	39.7	1199	0.096	0.704986	0.704870	10.42	68.16	0.0924	0.512443	−2.6	0.94
DG-18A	107	1429	0.216	0.705406	0.705145	15.93	100.4	0.0959	0.512481	−1.9	0.92
LM-17	128	2306	0.160	0.706673	0.706480	12.99	89.83	0.0874	0.512382	−3.8	0.98
01SB		1066		0.704925		10.38	67.03	0.0937	0.512490	−1.8	0.84
02SB		1132		0.705638		11.33	79.45	0.0862	0.512354	−4.4	0.95
BIP-02	56.2	1742	0.093	0.704761	0.704648	17.08	105.8	0.0976	0.512417	−3.2	1.01
BIP-2.6	58.7	1659	0.102	0.705048	0.704924	17.03	105.6	0.0975	0.512420	−3.1	1.01
02PR01	61.1	915	0.193	0.706397	0.706177	11.29	67.24	0.1015	0.512489	−1.9	0.90
BIP-03	68.9	1353	0.147	0.705039	0.704861	16.60	102.9	0.0975	0.512377	−4.0	1.06
96AE04	55.3	1000	0.160	0.705260	0.705080	8.93	53.94	0.0914	0.512406	−3.4	0.98
96AE08	49.6	1605	0.089	0.704980	0.704880	15.10	99.86	0.0901	0.512362	−4.3	1.02
96AE11	25.8	734	0.102	0.704570	0.704450	9.24	55.57	0.1005	0.512446	−2.8	1.00
96AE48F	34.2	1272	0.078	0.704890	0.704800	12.98	80.04	0.0980	0.512436	−2.9	0.99
96AE57	37.0	1172	0.091	0.705020	0.704920	17.12	113.4	0.0913	0.512423	−3.1	0.96
MON01	47.2	640	0.213	0.704893	0.704650	9.57	56.97	0.1016	0.512535	−1.0	0.84
T03	56.2	1223	0.133	0.704717	0.704566	15.00	93.15	0.0973	0.512527	−1.1	0.82
02DI02	35.5	906	0.113	0.704710	0.704580	11.12	66.54	0.1010	0.512456	−2.6	0.99

All trace-element concentrations reported in ppm. εNd(i) is calculated for an age of 85 Ma.

da Corda flows, the Alto Paranaiba kamafugites plot in the middle of the Nd–Sr mantle array. Nd isotopic compositions in the Goiás kamafugites, in particular, overlap those of both the high-Ti groups of Paraná flood basalts and the range observed in the Walvis Ridge oceanic basalts (Fig. 6). The Goiás kamafugites also have ^{176}Hf/^{177}Hf (Table 7) that overlap the few analyses available for Tristan da Cunha basalts (Patchett and Tatsumoto, 1980). Os isotopic compositions of both Goiás (Table 7) and Alto Paranaíba mafic–alkalic rocks (Carlson et al., 1996; Araujo et al., 2001a) range widely, from initial values as low as ^{187}Os/^{188}Os = 0.1122 in Alto Paranaíba kimberlites and 0.1142 in Goiás kamafugites to as high as 0.3045 in Alto Paranaíba and 0.6540 in Goiás kamafugites. Most samples have initial ^{187}Os/^{188}Os below 0.16. Both the Goiás and Alto Paranaíba mafic–alkalic rocks have moderately high Os concentrations and show only a very broad, if any, negative-correlation between Os concentration and isotopic composition. For those samples with ^{187}Os/^{188}Os <0.16, Os concentrations range from 0.088 to 1.25 ppb in the Goiás kamafugites with a similar range for Alto Paranaíba kamafugites extending to concentrations approaching 2 ppb in the Alto Paranaíba kimberlites (Carlson et al., 1996). Re concentrations also range widely, from 0.033 to over 6.3 ppb, but the majority of samples have surprisingly low Re concentrations of below 0.4 ppb. With the relatively high Os concentrations of these samples, the low Re concentrations result in low ^{187}Re/^{188}Os ratios with a median of 1.02 in the Alto Paranaíba mafic–alkalic rocks and 2.00 in the Goiás kamafugites. Given

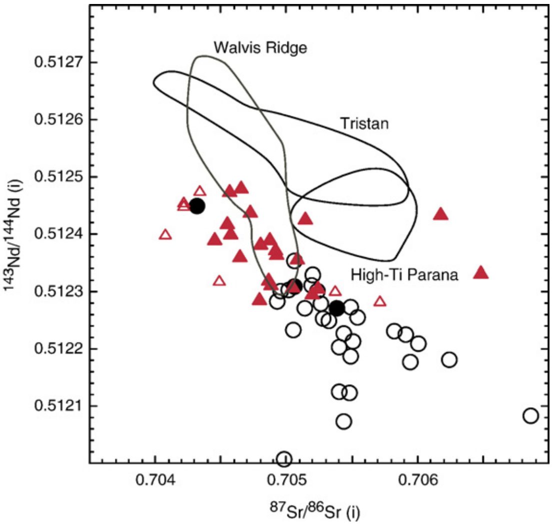

Fig. 6. Initial ^{87}Sr/^{88}Sr versus ^{143}Nd/^{144}Nd in kamafugites from the Alto Paranaíba (black circles) and Goiás (red triangles) alkalic provinces. Small open symbols are previous data from these areas (Gibson et al., 1995a,b; Carlson et al., 1996; Araujo et al., 2001a). Fields for Tristan/Walvis Ridge basalts from O'Nions et al. (1977), Richardson et al. (1982) and Gibson et al. (2005) and Paraná basalts from Mantovani et al. (1985) and Piccirillo et al. (1989). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Table 7
Lu–Hf and Re–Os isotopic results for Goiás samples

Sample	Lu	Hf	$^{176}Lu/^{177}Hf$	$^{176}Hf/^{177}Hf$	2σ	εHf(i)	Re	Os	$^{187}Re/^{188}Os$	$^{187}Os/^{188}Os$	2σ	$^{187}Os(i)/^{188}Os$
DG-16C							5	571	0.04	0.14523	0.00028	0.1452
DG-18A							23	372	0.30	0.14568	0.00028	0.1453
01SB	0.133	6.536	0.00287	0.282713	0.000094	−0.4	113	175	3.11	0.13642	0.00047	0.1323
02SB	0.135	6.469	0.00294	0.282616	0.000010	−3.9	435	544	3.88			
03SB							5927	72	460	1.352	0.026	0.7386
05SB							44	16	13.6	0.17893	0.00068	0.1608
BIP-02							31	80.8	1.98	0.6568	0.0013	0.6540
BIP-2.6							87	870	0.49	0.15875	0.00030	0.1581
02PR01	0.306	7.601	0.00568	0.282746	0.000005	0.6	430	126	16.4	0.15613	0.00022	0.1342
BIP-03							25	1246	0.10	0.14256	0.00028	0.1424
96AE04	0.118	6.066	0.00274	0.282945	0.000009	7.8	86	206	2.01	0.13268	0.00022	0.1300
96AE08	0.213	10.47	0.00287	0.282724	0.000006	0.0	1199	13	499	0.99796	0.07689	0.3328
96AE11	0.158	6.372	0.00349	0.282759	0.000007	1.2	39	151	1.26	0.14582	0.00028	0.1441
96AE48F	0.270	7.733	0.00494	0.282720	0.000007	−0.3	390	21	92.0	0.38076	0.01021	0.2580
96AE57	0.153	6.801	0.00318	0.282802	0.000007	2.7	104	88	5.64	0.13737	0.00046	0.1298
96AE58	0.317	13.18	0.00340	0.282758	0.000006	1.1	261	52	24.5	0.18506	0.00053	0.1524
96AE62							1602	741	10.6	0.12836	0.00024	0.1142
MON01	0.139	6.446	0.00305	0.282732	0.000007	0.2	462	109	20.6	0.20326	0.00038	0.1758
T03	0.243	10.30	0.00333	0.282580	0.000018	−5.2	254	81	15.2	0.17027	0.00077	0.1499
02DI02	0.300	8.137	0.00520	0.282741	0.000006	0.4	523	218	11.6	0.14843	0.00022	0.1329

Lu and Hf concentrations in ppm, Re and Os in ppb.

the wide range of Re/Os ratios in these rocks and the inclusion of kimberlitic compositions in the Alto Paranaíba median, these median Re/Os ratios likely are not significantly different.

6. Discussion

6.1. Lithospheric mantle composition and history

Most of the xenoliths from the Alto Paranaíba alkalic province show the magmaphile element (e.g. Ca, Al) depletion, high Mg# and old Re-depletion model ages typical of old lithospheric mantle (Carlson et al., 2005). Re-depletion model ages are calculated by assuming that the Re/Os ratio of the xenolith was zero before its journey to the surface. This assumption is made in order to get around the problem of Re addition through infiltration of xenoliths by their commonly Re-rich host magmas (Walker et al., 1989). Alto Paranaíba sample X219 provides a good example of this problem. This sample has quite high Re concentration and consequently an Re/Os ratio substantially higher than estimated even for the fertile mantle, let alone the Re-depleted mantle sampled by the other Alto Paranaíba xenoliths. This sample, however, has the lowest measured $^{187}Os/^{188}Os$ of the xenoliths. If all the Re measured in this sample was introduced during xenolith entrainment, then the Re-depletion model age of 3.17 Ga accurately reflects the time of Re-depletion of this sample. On the other hand, if the Re/Os ratio of this sample was non-zero prior to entrainment in the host magma, the Re-depletion model age underestimates the true time of melt-depletion. The maximum age of Re-addition can be estimated by noting that the very high Re/Os ratio, but low measured $^{187}Os/^{188}Os$, would lead the $^{187}Os/^{188}Os$ of this sample to go below that of the Earth's initial Os isotopic composition ($^{187}Os/^{188}Os = 0.09524$, Smoliar et al., 1996) in 344 Ma, or 264 Ma ago, which provides an upper bound to the timing of metasomatism. The likely lower age bound to this Re addition is through interaction with the host magma at ~80 Ma. Since Re-depletion model ages provide only lower bounds to the true time of melt depletion, the 1.92 to 3.17 Ga Re-depletion ages of the Alto Paranaíba xenoliths strongly support previous suggestions (Carlson et al., 1996; Araujo et al., 2001a) that the Alto Paranaíba alkalic province erupted through the Paleoproterozoic to Neoarchean basement of the São Francisco Craton even though the surface exposures of basement in this area are Neoproterozoic rocks associated with the Brasiliano metamorphic belts. The composition of these xenoliths further suggests that the lithospheric mantle of the São Francisco craton consists of melt-depleted peridotites, as do many other ancient continental cratons.

The chemical and isotopic characteristics of the Alto Paranaíba xenoliths are in stark contrast to the xenoliths from Goiás, the majority of which have compositions close to estimates of the fertile mantle (Fig. 3). Most of the Goiás xenoliths have $^{187}Os/^{188}Os$ overlapping the values seen in modern samples of the mantle, for example abyssal peridotites (Reisberg et al., 1991; Snow and Reisberg, 1995) and because their Os isotopic compositions and particularly Re/Os ratios are near estimates of the fertile mantle, the Re–Os results place no firm limits on the age of these mantle samples.

Two of the Goiás xenoliths, however, appear to be slightly melt-depleted with marginally lower CaO, Al_2O_3 and $^{187}Os/^{188}Os$ compared to the other samples. The Re-depletion model ages for both samples are ~1.2 Ga. Given the limited amount of melt depletion recorded by the major element composition of these samples, they likely had significant Re/Os ratios prior to entrainment in the host magma. If so, their Re-depletion model ages underestimate the true age of depletion. This contention is supported by the one Goiás kamafugite (sample 96AE58) that has an initial $^{187}Os/^{188}Os$ below that measured for any of the Goiás xenoliths. The Re-depletion model age of kamafugite 96AE58 is 2.01 Ga, suggesting that Meso- to Paleoproterozoic lithospheric mantle may underlie the Goiás alkalic province. Whether or not the more fertile Goiás xenoliths also date to the Proterozoic or represent younger mantle in this area cannot be determined with the available data. The compositional characteristics of all the Goiás xenoliths suggest that the lithospheric mantle underlying the Goiás alkalic province may be significantly more fertile than that beneath the Alto Paranaíba alkalic province.

6.2. Lithospheric mantle imprint on magma composition

The distinct isotopic and compositional characteristics of the xenoliths studied here allow us to explore the potential consequences that the lithospheric mantle could have on the chemical and isotopic composition of magmas either sourced in this lithosphere, or erupted through it, by comparison with the composition of the kamafugites from each province. Obviously, because most of the spinel peridotites studied here derive from shallow depths in the lithosphere, the xenoliths are not likely to directly represent the composition of the kamafugite sources, which, given their very steep incompatible-element patterns, most likely derive from the garnet stability field. Nevertheless, in the following we explore whether the compositional distinction between the Goiás and Alto Paranaíba peridotites reflect general properties of the lithospheric mantle beneath these areas that might be reflected in magma geochemistry.

The most obvious compositional distinctions between the Goiás and Alto Paranaíba kamafugites are the higher Na_2O and Al_2O_3 concentrations of the Goiás samples at a given MgO concentration (Fig. 4). These features are shared with the xenoliths from these respective localities. The other clear compositional distinction in the xenoliths, the higher CaO concentration of the Goiás peridotites, is not expressed in the kamafugites. These major element trends cannot be explained by fractional crystallization of any combination of the main phases in these rocks (olivine, clinopyroxene, phlogopite, perovskite) of the compositions reported by Araujo et al. (2001a). The fact that the Goiás kamafugites show lower degrees of enrichment in the highly incompatible trace elements than do the Alto Paranaíba kamafugites suggests that these compositional differences are related to different degrees of partial melting with the Goiás kamafugites being derived by slightly higher degrees of melting of more fertile sources.

The range in isotopic composition (Fig. 7) in the Brazilian mafic–alkalic rocks shows clearly that the compositional variability is not only due to differences in partial melting of similar sources. The Goiás kamafugites show little correlation between composition and radiogenic isotope composition, but when the compositional range is extended by adding the wide variety of rock types erupted in the Alto Paranaíba province, there are broad correlations in Nd and Os isotopic

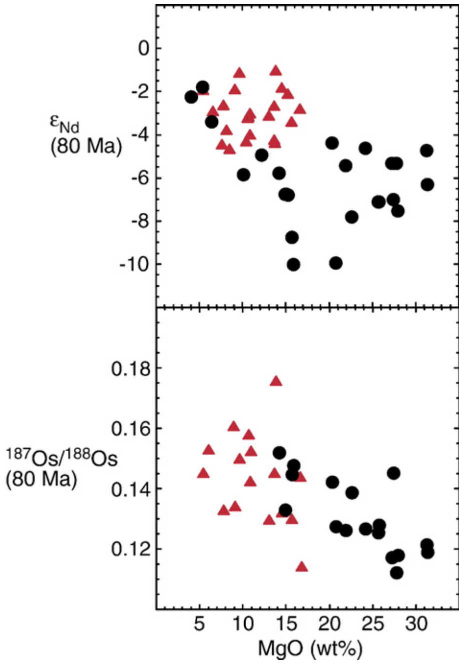

Fig. 7. Initial $^{187}Os/^{188}Os$ and εNd versus MgO concentrations for kamafugites from the Goiás (red triangles) and Alto Paranaíba (black circles) alkalic provinces. Additional Alto Paranaíba data from Carlson et al. (1996) and Araujo et al. (2001a). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

composition and MgO content (Fig. 7). The high-MgO end of the correlation has low $^{187}Os/^{188}Os$ and $^{143}Nd/^{144}Nd$ as might be expected for old, metasomatized, but Re-depleted, lithospheric mantle peridotite. The other end of the array has Sr and Nd isotopic composition approaching values observed for rocks derived from the convecting mantle, such as ocean island basalts, yet has Os isotopic composition considerably higher than observed in mantle peridotite. In these diagrams, old continental crust would plot towards the high $^{187}Os/^{188}Os$, low MgO end of the kamafugite data, but would also plot at the low $^{143}Nd/^{144}Nd$ end of the diagram and thus is an unlikely contributor to the isotopic range observed in these magmas.

6.3. The need for an olivine-poor source component

The wide range in $^{187}Os/^{188}Os$ of the mafic–alkalic rocks is difficult to explain if the sources of these magmas consisted only of peridotite, because peridotite usually has very low Re/Os ratios that do not allow it to evolve large variations in $^{187}Os/^{188}Os$ (Shirey and Walker, 1998; Carlson, 2005). In the Alto Paranaíba province, only the kimberlites have $^{187}Os/^{188}Os$ approaching the low values seen in the peridotite xenoliths. Two of the Goiás kamafugites have $^{187}Os/^{188}Os$ near the value measured in Goiás peridotite xenoliths, and one has lower $^{187}Os/^{188}Os$, but most have higher $^{187}Os/^{188}Os$. Most of the kamafugites have $^{187}Os/^{188}Os$ higher than observed for all peridotites world-wide (Carlson et al., 2005), and some have $^{187}Os/^{188}Os$ higher than all but low Os content, crustally-contaminated, ocean island basalts (Reisberg et al., 1993; Widom and Shirey, 1996).

Two materials potentially can contribute high $^{187}Os/^{188}Os$ to these melts; old crust introduced by crustal contamination, or olivine-poor components (e.g. pyroxenite, eclogite, glimmerite or various other metasomatic assemblages) in the mantle source of these magmas. Several features of these magmas argue against crustal contamination and instead for an olivine-poor component in the mantle source. Of the samples with $^{187}Os/^{188}Os$ higher than 0.16, two Alto Paranaíba kamafugites with $^{187}Os/^{188}Os$ near 0.30 have quite high Os concentrations (~0.5 to 0.6 ppb — Carlson et al., 1996). To reach this isotopic composition through crustal contamination would require the addition of over 60% of Archean crust (with an assumed $^{187}Os/^{188}Os = 1.56$, Os concentration = 0.05 ppb). The Goiás kamafugite with the highest initial $^{187}Os/^{188}Os$ (0.654) also has $^{87}Sr/^{86}Sr$ (0.7046) on the low side and $^{143}Nd/^{144}Nd$ ($\varepsilon Nd = -3.2$) on the high side of the distribution of the Goiás samples, again making it difficult to explain the elevated $^{187}Os/^{188}Os$ by crustal contamination.

Another possible contributor of high $^{187}Os/^{188}Os$ to mantle derived magmas are olivine-poor lithologies in the mantle source of the magmas. For want of a better term, we include in this category the variety of olivine-poor assemblages that can be introduced into the mantle through melt or fluid metasomatism that are rich in minerals such as pyroxene, phlogopite, amphibole, carbonate, phosphate and Ti-rich oxides. Compared to peridotite, these olivine-poor components generally have high Re/Os ratios, low Os concentrations, and thus, with time, will develop radiogenic Os isotopic compositions (Reisberg et al., 1991; Pearson et al., 1995; Shirey and Walker, 1998; Carlson, 2005). Because the high Re/Os ratios of such materials usually are determined by low Os, rather than unusually high Re, concentrations, in order for them to significantly influence the Os isotopic composition of a source composed of a mixture of peridotite plus olivine-poor components, the ratio of olivine-poor material to peridotite has to be high, and the olivine-poor component has to be old. For example, a rock with 2 ppb Re and 0.1 ppb Os, typical concentrations for a primitive ocean island basalt (Hauri and Hart, 1997), with Os isotopic composition equal to that of the fertile mantle at the time of its formation, would have a present day $^{187}Os/^{188}Os$ of 0.208, 1.714, and 4.956 for formation ages of 45, 900, and 2700 Ma, respectively. These indeed are radiogenic values, but if mixed with a peridotitic lithosphere similar to that sampled by the Alto Paranaíba xenoliths (average 4.23 ppb Os and $^{187}Os/^{188}Os = 0.1128$), would require 96%, 50%, and 24% by weight, respectively, of the different aged olivine-poor components quoted above in order for the $^{187}Os/^{188}Os$ of the mixture to exceed 0.15. These very high percentages of olivine-poor component in the mixture do not necessarily have to represent the amount of this material in the source if the olivine-poor component melts at lower temperature than the peridotite and thus contributes a larger percentage of the melt (Lloyd et al., 1985; Foley, 1992b; Hirschmann et al., 2003). This possibility is particularly likely for a highly-melt-depleted peridotite end-member like the Alto Paranaíba xenoliths that will have a high solidus temperature. Nevertheless, for the samples with the higher Os isotopic compositions ($^{187}Os/^{188}Os > 0.2$), the Os isotopic results require that the melts be dominated by the olivine-poor component in the mantle rather than peridotite if the peridotitic component is similar in composition to the Alto Paranaíba peridotites. For a peridotitic component compositionally similar to the more fertile Goiás xenoliths, $^{187}Os/^{188}Os > 0.15$ in a mixed source of peridotite plus olivine-poor components would require 90%, 26%, and 10% of the olivine-poor component of 45, 900 and 2700 Ma age, respectively. These relative abundances of the olivine-poor component are smaller, but still show the important role that must be inferred for an olivine-poor component in the genesis of these magmas. Whether or not ultramafic–potassic melts can be derived from such olivine-poor compositions in the mantle has been the subject of several experimental studies (Lloyd et al., 1985; Foley, 1992a,b).

At these high ratios of olivine-poor component to peridotite, the incompatible elements in the melt will be derived predominantly from the olivine-poor component. Given the quite distinct compositions and ages of the lithospheric peridotites sampled in the Alto Paranaíba and Goiás Provinces, if the olivine-poor source component is of a similar age and compositional variation to the peridotites, the limited spread in Sr and Nd isotopic composition of the kamafugites, particularly those from Goiás, is unexpected. The Alto Paranaíba mafic–alkalic rocks do trend towards higher ^{87}Sr/^{86}Sr and lower ^{143}Nd/^{144}Nd than the Goiás kamafugites (Fig. 6), which could indicate the involvement of an older, enriched, lithospheric component in the Alto Paranaíba samples, but also one increasingly dominated by peridotite because many of the Alto Paranaiba mafic–alkalic rocks have lower ^{187}Os/^{188}Os than do the Goias mafic–alkalic samples (Fig. 8). However, the Nd isotopic difference between the Goiás and Alto Paranaíba mafic–alkalic rocks is small, particularly at the low-MgO end of the compositional spectrum (Fig. 7), given the dramatically different compositional characteristics of the lithospheric mantle sampled by the xenoliths from these areas. One way to explain the limited spread in Nd isotopic composition is to introduce the olivine-poor component recently, and regionally, through the injection of melts into preexisting peridotitic lithospheric mantle. Because the Nd/Os concentration ratios of most mantle melts and mantle peridotite are very different from one another, Nd–Os isotope mixing curves will be strongly hyperbolic (Fig. 8). As a result, the Nd isotopic composition of the olivine-poor end-member is reasonably well constrained by the kamafugite data, but the Os isotopic composition of this component is not. Similarly, the Os isotope composition of the peridotitic end-member is constrained by the kamafugite data, but the Nd isotopic composition of this end-member is not.

Fig. 8 explores the consequences of adding various composition melt components to the lithospheric mantle at various times in the past. There is considerable flexibility in this modeling given the range of Rb, Sr, Sm, Nd, Re and Os concentrations and Sr, Nd and Os isotopic compositions possible for both the melt and peridotitic end-members (Table 8). Consequently, the model results can, at best, be considered plausibility arguments. If the melt was introduced into the lithospheric mantle at the time of Paraná flood-basalt volcanism (~130 Ma), an unrealistically high Re/Os ratio in this material is required in order to evolve Os isotopic composition as high as observed in the highest ^{187}Os/^{188}Os kamafugites. Using Re and Os concentrations for this component similar to those estimated for the average ocean island basalt (Table 8 — Hauri and Hart, 1997), a Paraná-age melt component will have ^{187}Os/^{188}Os near 0.21 at 85 Ma. If mixed with an old, depleted, peridotitic component like the Alto Paranaíba xenoliths, kamafugites with ^{187}Os/^{188}Os ~0.15 would

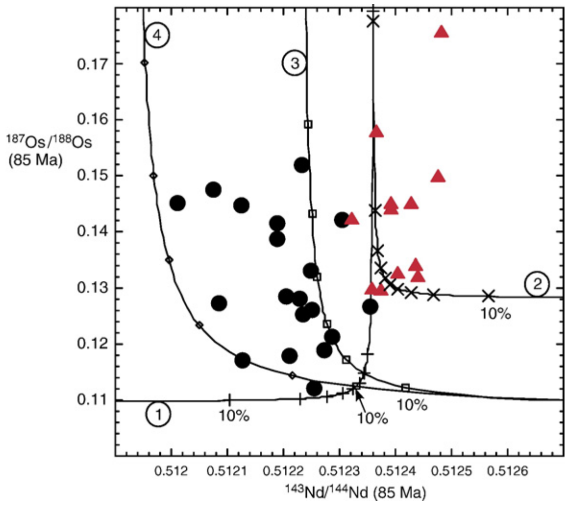

Fig. 8. Os versus Nd isotopic composition illustrating the mixing relationships between peridotite and an olivine-poor component in the source of the mafic–alkalic magmas. Data for the mafic–alkalic rocks from the Alto Paranaíba and Goiás alkalic provinces are shown by the black circles and red triangles, respectively. The four mixing curves shown correspond to the end-members listed in Table 8 and are identified on this figure as 1) Archean peridotite with low ^{187}Os/^{188}Os and ^{143}Nd/^{144}Nd intruded at 130 Ma by a melt with Sm/Nd ratio similar to high-Ti Paraná basalt and Re and Os concentrations similar to typical mantle-derived mafic melts (e.g. Hauri and Hart, 1997); 2) modern fertile peridotite intruded at 130 Ma by the same melt as used in model 1; 3) depleted (low ^{187}Os/^{188}Os, high ^{143}Nd/^{144}Nd) Archean peridotite intruded at 900 Ma by a low Sm/Nd, high LIL-content, low Os content, melt; or 4) the same components as 3, but a melt intrusion age of 1700 Ma. All mixing curves shown assume that mixing occurred at the ~85 Ma eruption age of the Brazilian mafic–alkalic rocks. Elemental concentrations and isotopic compositions for these end-members are given in Table 8. Marks along each curve give 10% increments in the amount of olivine-poor component in the mixture. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Table 8
End-member compositions for mixing model shown in Fig. 8

	Sm (ppm)	Nd (ppm)	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$ (Tintr)	$^{143}Nd/^{144}Nd$ (85 Ma)	Re (ppb)	Os (ppb)	$^{187}Re/^{188}Os$	$^{187}Os/^{188}Os$ (Tintr)	$^{187}Os/^{188}Os$ (85 Ma)
1) Moderate Sm/Nd melt intruding Archean LREE-rich (low Sm/Nd) peridotite at 130 Ma										
Peridotite	0.14	1	0.085	0.51150	0.51152	0.1	4	0.13	0.1096	0.1097
Basalt	5.3	23	0.139	0.51232	0.51236	2	0.1	106	0.1285	0.2077
2) Moderate Sm/Nd melt intruding modern fertile peridotite at 130 Ma										
Peridotite	0.44	1.4	0.197	0.51280	0.51286	0.25	3	0.44	0.1280	0.1283
Basalt	5.3	23	0.139	0.51232	0.51236	2	0.1	106	0.1285	0.2077
3) LREE enriched (low Sm/Nd) melt intruding depleted Archean peridotite at 900 Ma										
Peridotite	0.35	1	0.212	0.51156	0.51278	0.1	4	0.13	0.1079	0.1097
Basalt	5.3	30	0.107	0.51160	0.51223	2	0.1	106	0.1230	1.7143
4) LREE enriched (low Sm/Nd) melt intruding depleted Archean peridotite at 1700 Ma										
Peridotite	0.35	1	0.212	0.51053	0.51278	0.1	4	0.13	0.1062	0.1097
Basalt	5.3	30	0.107	0.51079	0.511921	2	0.1	106	0.1200	3.015

Tintr gives the isotopic composition of the basalt at the time of its intrusion into the lithospheric mantle.

require ~92% of the melt end-member in the mixed source. Making the melt component older lowers the proportion of this component needed in order to match the Os isotope range observed in the kamafugites. For example, if the melt component was added during the 900 Ma convergence-related volcanism in the Brasiliano belt (Pimentel et al., 1999), then $^{187}Os/^{188}Os$ ~0.15 is reached in a source composed of a nearly equal mix of olivine-poor end-member and Archean depleted peridotite. If the melt component is as old as the circa 1700 Ma basement of the Brasiliano belt, less than 34% of this component is needed in the mixture to reach $^{187}Os/^{188}Os$ ~0.15 (Fig. 8). In this type of model, the predominance of high $^{187}Os/^{188}Os$ values for the Goiás kamafugites could reflect either a routinely high olivine-poor component in these magmas, or the lack of a low $^{187}Os/^{188}Os$ peridotitic component as is suggested by the data for the Goiás peridotite xenoliths (Fig. 8). In any case, the high $^{187}Os/^{188}Os$ of the kamafugitic rocks provide additional support for the idea that the source regions of mantle-derived rocks are not just peridotite, but can contain substantial amounts of olivine-poor components.

7. Conclusions

Peridotite xenoliths from the Alto Paranaíba and Goiás mafic–alkalic provinces reveal the existence of distinct shallow lithospheric mantles beneath these areas. The Alto Paranaíba xenoliths are depleted in Ca, Al and Re, and have high Mg#, consistent with them being the residues of extensive partial melt removal. Old Re-depletion model ages (average 2.4 Ga) for these xenoliths indicate that the Alto Paranaíba alkalic province is underlain by Paleoproterozoic to Neoarchean lithospheric mantle of the São Francisco Craton. In contrast, spinel–peridotites from the Goiás alkalic province have compositions similar to fertile mantle, and the majority have Re–Os systematics close enough to estimates of the modern fertile mantle that they do not yield reliable model age information. Two Goiás peridotites have lower $^{187}Os/^{188}Os$ and define Re-depletion model ages of ~1.2 Ga that, along with an Re-depletion model age of 2.0 Ga for a Goiás kamafugite, indicate that at least some of the lithospheric mantle in this area may date to the Mesoproterozoic assembly of the Brasiliano mobile belt basement.

Mafic–alkalic magmatism in the Goiás and Alto Paranaíba provinces shows some compositional overlap, but also shows compositional distinctions that mirror those of the mantle xenoliths from the two provinces implying the involvement of shallow lithospheric mantle in defining the composition of these magmas. The very high $^{187}Os/^{188}Os$ in some kamafugites from both provinces suggests that the source of these magmas is not only peridotite, but includes olivine-poor components likely introduced by melt metasomatism. The $^{187}Os/^{188}Os$ of the majority of the kamafugites is significantly higher than seen in any peridotite and again points to a significant mass fraction of olivine-poor component in the mantle source of these magmas. The Alto Paranaíba mafic–alkalic rocks extend to lower $^{187}Os/^{188}Os$ than do the Goiás kamafugites, which, in part, reflects the involvement of older peridotitic mantle lithosphere similar to that sampled by the Alto Paranaíba xenoliths. The more Mg-rich, and lower $^{187}Os/^{188}Os$, of the kimberlitic end of the compositional range seen in the Alto Paranaíba province also indicates a source dominated by peridotite. Given the limited range in Sr and Nd isotopic composition of the mafic–alkalic magmas, the olivine-poor component in the source of these magmas could have been added by introduction of melts into the lithospheric mantle on a regional scale, thereby overprinting the incompatible-element characteristics, and the isotopic systems based on them, of the varying age and composition lithospheres of these provinces. Melt metasomatism of the mantle lithosphere as late as Paraná flood-basalt volcanism is possible, but would require that the kamafugites basically be remelts of the introduced olivine-poor components in the lithosphere, which is unlikely given the major element characteristics of these magmas. Pushing melt infiltration back to the 900–1700 Ma events involved in the formation of the Brasiliano belt would allow the kamafugites to be melts of a source composed of roughly equal amounts of lithospheric peridotite and metasomatic veins. If this model is correct, the implication is that both the Goiás and Alto Paranaíba mafic–alkalic magmatism is the result of remelting of Brasiliano belt lithospheric mantle composed of peridotite veined with the crystallization products of infiltrating Mesoproterozoic melts (e.g. Gibson et al., 1995a; Carlson et al., 1996). Since this mixed lithospheric mantle also would have isotopic compositions similar to the Paraná flood basalts, the results presented here suggest that the Paraná high-Ti source, and that of the isotopically similar oceanic basalts of the Walvis Ridge, formed in the Brazilian lithosphere in the Proterozoic.

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