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Osmium isotopic compositions of mantle xenoliths: A global perspective

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Abstract—The Re-Os isotopic systematics of spinel- and garnet-bearing mantle xenoliths from North and Central America, Europe, southern Africa, Asia, and the Pacific region were determined to define more precisely the isotopic composition of a hypothetical primitive upper mantle (PUM). When plotted against Al_2O_3 or other melt depletion indicators, the $^{187}\text{Os}/^{188}\text{Os}$ ratios of all xenolith suites display positive trends toward a uniform composition at a fertile mantle composition. A $^{187}\text{Os}/^{188}\text{Os}$ value for PUM of 0.1296 ± 0.0008 (level of confidence 95%) was defined on the basis of 117 spinel-bearing xenoliths from this work and data from the literature, including data for massif peridotites. The $^{187}\text{Os}/^{188}\text{Os}$ ratio of the PUM is similar to the range of compositions defined by ordinary and enstatite chondrites, not carbonaceous chondrites. Spinel-bearing mantle peridotites sampled by volcanism and peridotite massifs appear to have been extracted from a common fertile source (PUM) between 1 and 2 Ga ago. These peridotites now form part of the subcontinental lithospheric mantle underlying continental crust of similar or greater formation age. Copyright © 2001 Elsevier Science Ltd

1. INTRODUCTION

Xenoliths derived from the mantle and brought to the surface via volcanism are among the few types of materials that permit direct study of the chemical composition of the Earth's upper mantle. By examining the isotopic composition of mantle samples, a time dimension is added to the estimate of the chemical composition of domains within the mantle. This is important for understanding the chemical evolution of the terrestrial mantle from a primitive composition, created during Earth's early history, to the present differentiated states of the mantle and crust. Because of these aspects, mantle xenoliths have been vigorously studied for major and trace element compositions, mineral compositions, and lithophile isotope systems such as Rb-Sr, U-Pb, and Sm-Nd (Menzies et al., 1987; McDonough and Frey, 1989). The addition of the Re-Os isotope system ($^{187}\text{Re} \rightarrow ^{187}\text{Os} + \beta^-$; $\lambda = 1.666 \cdot 10^{-11} \text{a}^{-1}$; Smoliar et al., 1996) to the study of xenoliths has proven important because parent and daughter elements have contrasting geochemical behaviors during partial melting processes in the mantle (e.g., Morgan, 1986).

Although Re is a mildly incompatible element, Os usually remains compatible in the residue (Shirey and Walker, 1998). This characteristic of the Re-Os isotope system can provide unique information about the timing of melt depletion in the mantle. The system is normally immune to subsequent processing within the mantle because the Os model dating technique used for peridotites is based on the loss of the parent isotope and the resulting retardation of the growth of radiogenic ^{187}Os (e.g., Walker et al., 1989). For the system to be disturbed, either Re must be added during some later event, or Os with a different isotopic composition must be added to the system.

Evidence from previous studies suggests that these situations are rare (Walker et al., 1989; Pearson et al., 1995b) and restricted to specific tectonic settings such as mantle lithosphere above plumes (e.g., Chesley et al., 1999) or above subducted mantle wedges (Brandon et al., 1996). In addition to the normal closed-system behavior within the mantle, the high abundance of Os in most ultramafic xenoliths relative to silica-rich crustal sediments and S-saturated basaltic and alkaline melts makes the Os isotope system robust to late-stage contamination that can hamper the usefulness of the lithophile isotope systems for the study of mantle evolution.

Over the previous decade, a number of studies have examined in detail the Os isotopic compositions of peridotitic xenoliths from a variety of mantle domains, including domains underlying southern Africa (Pearson et al., 1995a; Olive et al., 1997; Walker et al., 1989), Tanzania (Chesley et al., 1999), the Siberian craton (Pearson et al., 1995b), the Wyoming craton of North America (Carlson and Irving, 1994), the Canadian Cordillera (Peslier et al., 2000), the Baikal rift (Pearson et al., 1998), and southern Australia (Handler et al., 1997). The Os isotopic systematics of peridotitic xenoliths associated with ocean islands (Hassler and Shimizu, 1998) and arc systems (Brandon et al., 1996; Parkinson et al., 1998) have also been examined. Here, Re-Os isotopic data for a suite of 39 previously well-characterized peridotite xenoliths from worldwide localities are reported. The purpose of this study is to expand the Re and Os concentration and the Os isotopic database for well-characterized upper mantle samples from a variety of different settings and to begin the task of synthesizing the rapidly expanding Os isotopic database for upper mantle peridotites to understand the causes of the global variations.

Twenty of the samples examined here were collected during the course of the Basaltic Volcanism Study Project (BVSP; assembled by A.J.I. in 1978; Project, 1981). The BVSP was an important advance in the understanding of the mantle because

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Table 1. Re-Os isotopic compositions of peridotite xenoliths.

Site	Sample	Location	Re (ng/g)	ISD (ng/g)	Os (ng/g)	$^{187}\text{Os}/^{188}\text{Os}$	$^{187}\text{Re}/^{188}\text{Os}$	γ_{Os}	Al_2O_3 (% w/w)	Detection mode ^a	
United States											
		Kilbourne Hole, New Mexico	0.236	0.018	2.88	0.12903	0.394	0.032	1.5	4.29	SCIC
	UM-2	Kilbourne Hole, New Mexico	0.253	0.019	3.11	0.12544	0.392	0.031	-1.3	3.36	SCIC
	UM-2	Kilbourne Hole, New Mexico	0.251	0.020	3.13	0.12563	0.386	0.031	-1.1		SCIC
	UM-3	Kilbourne Hole, New Mexico			2.48	0.12592			-0.9	2.86	SCIC
	UM-3	Kilbourne Hole, New Mexico	0.118	0.006	2.48	0.12586	0.228	0.032	-0.9		SCIC
	UM-4	Kilbourne Hole, New Mexico	0.256	0.019	1.58	0.12738	0.781	0.116	0.2	3.57	SCIC
		Kilbourne Hole, New Mexico	0.084	0.012	1.61	0.12691	0.253	0.038	-0.1		SCIC
	UM-5	Kilbourne Hole, New Mexico	0.220	0.031	2.56	0.12286	0.414	0.060	-3.3	3.07	SCIC
	UM-6	Kilbourne Hole, New Mexico	0.176	0.011	2.77	0.12675	0.306	0.021	-0.3	4.07	SCIC
		Kilbourne Hole, New Mexico	0.169	0.011	2.88	0.12463	0.282	0.024	-2.0	3.51	MCFC
	UM-7	Kilbourne Hole, New Mexico	0.329	0.022	2.98	0.12772	0.532	0.038	0.5	3.20	SCIC
		Kilbourne Hole, New Mexico	0.213	0.027	2.31	0.12977	0.445	0.058	2.1	3.47	SCIC
		Kilbourne Hole, New Mexico	0.211	0.030	2.21	0.12968	0.462	0.065	2.1		SCIC
	UM-8	Kilbourne Hole, New Mexico	0.216	0.010	3.07	0.12498	0.339	0.017	-1.7	3.37	SCFC
	UM-9	Kilbourne Hole, New Mexico	0.432	0.016	3.28	0.12711	0.634	0.027	0.0	3.19	SCIC
		Kilbourne Hole, New Mexico	0.053	0.010	0.85	0.12456	0.302	0.057	-2.0	3.36	MCFC
		Kilbourne Hole, New Mexico	0.202	0.012	4.07	0.11700	0.239	0.014	-7.9	0.92	MCFC
	UM-10	San Carlos, Arizona	0.012	0.006	1.84	0.12326	0.031	0.017	-3.0	1.37	SCIC
	UM-11	San Carlos, Arizona	0.190	0.009	2.05	0.12046	0.445	0.024	-5.1	1.54	SCIC
	UM-12	Cochise Crater, Arizona			3.84	0.12626			-0.6	3.35	SCIC
	UM-12	Cochise Crater, Arizona	0.277	0.023	3.87	0.12793	0.345	0.029	0.7		SCIC
		Dish Hill, California	0.090	0.010	1.12	0.12654	0.386	0.043	-0.4		SCIC
Mexico											
	UM-14	San Quintín			2.74	0.12253			-3.5	2.86	SCIC
	UM-14	San Quintín	0.268	0.031	2.81	0.12288	0.459	0.055	-3.3		SCIC
		San Quintín	0.449	0.026	1.05	0.12866	2.066	0.120	1.4	3.23	MCFC
		San Quintín	0.106	0.008	1.20	0.13205	0.425	0.032	4.0	3.91	MCFC
		San Quintín	0.054	0.010	1.39	0.12292	0.187	0.034	-3.1	1.81	MCFC
Pacific											
	UM-1	Nunivak Island, Alaska	0.370	0.078	3.50	0.12729	0.508	0.107	0.2	2.16	SCIC
	UM-13	Salt Lake Crater, Hawaii	0.131	0.012	1.85	0.12299	0.341	0.032	-3.2	2.23	SCIC
	UM-15	Mount Quincan, Australia	0.377	0.047	1.45	0.12718	1.257	0.158	0.1	2.89	SCIC
Africa											
	UM-16	Matsoku, Lesotho	0.372	0.014	5.37	0.10968	0.333	0.014	-13.7	1.52	MCFC
	UM-17	Matsoku, Lesotho	0.194	0.010	2.67	0.11396	0.350	0.020	-10.3	2.45	SCIC
	UM-18	Matsoku, Lesotho	0.086	0.012	4.64	0.10952	0.089	0.012	-13.8	1.25	SCIC
	UM-19	Matsoku, Lesotho	0.088	0.009	13.1	0.11134	0.032	0.003	-12.4	1.41	MCFC
	UM-20	Lashaine, Tanzania	0.028	0.009	0.046	0.11650	2.87	0.92	-7.2	1.27	SCIC
Europe											
		Massif Central, France	0.166	0.010	1.63	0.13363	0.491	0.030	5.2	3.09	MCFC
		Massif Central, France	0.042	0.010	3.19	0.11842	0.063	0.015	-6.8	0.97	MCFC
		Massif Central, France	0.156	0.010	5.21	0.12771	0.144	0.010	0.5	4.24	MCFC
		Dreiser Weiher, Germany	0.006	0.016	1.62	0.11803	0.018	0.047	-7.1	0.64	SCIC
		Dreiser Weiher, Germany	0.069	0.009	2.62	0.12242	0.127	0.017	-3.7	2.11	SCIC
		Dreiser Weiher, Germany	0.115	0.012	2.38	0.12968	0.232	0.024	2.1	4.29	SCIC
		Kapfenstein, Austria	0.020	0.010	0.908	0.12451	0.091	0.046	-2.0	2.61	MCFC
Asia											
		Tariat, Mongolia	0.119	0.013	2.74	0.12830	0.209	0.023	1.0	4.34	SCIC
		Tariat, Mongolia	0.138	0.013	2.53	0.12719	0.263	0.026	0.1	4.45	SCIC
		Hannuoba, China	0.369	0.020	3.86	0.12632	0.460	0.026	-0.6	2.68	SCIC
		Hannuoba, China	0.189	0.010	3.49	0.11826	0.261	0.015	-6.9	1.27	SCIC

SCIC, single collector ion counting (Bobcat I, UMD); SCFC, single collector faraday sup (Bobcat I, UMD); MCIC, multicollector ion counting ICPMS (Nu Instruments, Bern); MCFC, multicollector Faraday cup (VGSector54).

it led to the production of high-quality major and trace element data, as well as Sr, Nd and O isotope data, for mantle samples and for mantle-derived magmatic systems. The BVSP collection comes largely from North America, but it also includes samples from Hawaii, southern Africa, and Australia. Samples from this collection have the sample number prefix "UM" (Table 1). In addition to these samples, xenoliths from the Dreiser Weiher locality, Rhein Graben, Germany; Montbriçon in the Massif Central, France; Montferrier in Languedoc,

France; Kapfenstein, Styria, Austria; Hannuoba, China; and the Tariat depression of the Baikal Rift system, Mongolia, were also examined.

Linear correlations between indicators of melt depletion and $^{187}\text{Os}/^{188}\text{Os}$ have been observed for xenolith suites and orogenic peridotites from several locations, although mainly from Archean and Proterozoic cratons (Walker et al., 1989; Pearson et al., 1995b; Reisberg and Lorand, 1995; Meisel et al., 1996). These trends have been used to estimate the ages of, and

demonstrate the persistence of, subcontinental lithospheric mantle (SCLM) keels underlying ancient cratons. In a previous study, correlations defined by spinel-bearing xenoliths from three rift settings were used to define the Os isotopic composition of the Earth's primitive upper mantle (PUM; Meisel et al., 1996). In the past, PUM has been used to describe several types of mantle. Here, PUM is defined as the composition of a hypothetical upper mantle reservoir that has never experienced a melt depletion or enrichment event. As such, true PUM almost certainly does not exist anywhere within the Earth. For some chemical parameters, however, including Re/Os, some forms of nearly primitive mantle may approximate the characteristics of PUM. This may be especially true if a portion of the mantle melted only slightly at some stage in its past. Meisel et al. (1996) argued that the fact that xenolith suites from chemically distinct locations from three continents define a uniform $^{187}\text{Os}/^{188}\text{Os}$ composition for the modern PUM of 0.1290 ± 0.009 indicates that a profound connection must exist between these sampled reservoirs. Thus, an additional aim of this study is to constrain more precisely the Os isotopic composition of the PUM and to consider processes that may have led to its composition.

2. GEOLOGIC SETTINGS AND PETROLOGY

The xenoliths examined in this study can be divided into two major groups. One group is mainly composed of spinel-bearing peridotites and is hosted in alkalic basalts from young continental extensional (rift), oceanic regions, and continental margins. The second group consists mainly of garnet peridotites from kimberlitic intrusions in old continental (mainly cratonic) provinces (e.g., Nixon and Davies, 1987). The tectonic environments within which the two groups occur are also distinguishable in their age of formation: the spinel-bearing xenoliths are hosted by magmas intruding Proterozoic and Phanerozoic crust, whereas the kimberlitic magmas hosting the garnet peridotite xenoliths normally penetrate thick Archean lithosphere.

2.1. Spinel-Bearing Xenoliths

A major part of the xenolith suite examined in this study was collected in the southwestern United States. Sampling locations included the following: Dish Hill, California (DH132); Cochise Crater, Arizona (UM-12); San Carlos, Arizona (UM-10 and UM-11); and especially the Kilbourne Hole volcanic field, New Mexico (12 samples: UM-2 to UM-9, KH77-1, KH77-14, KH80-100, and KH80-106). These rocks were described in detail by Irving (1980), BVSP, and Irving and Carlson (personal communication). The host rocks are Quaternary alkalic basalts (basanites). Most samples have textures that are allotriomorphic- (proto-) granular to porphyroclastic, with coarse average grain sizes ranging from 1 to 3 mm. In contrast, three samples (UM-5, UM-7, and UM-8) are fine grained (grains of ~ 0.5 mm maximum dimension) and display prominent preferred orientation of tabular grains. These have been interpreted to be pieces of texturally and chemically modified wall rock from adjacent pyroxenitic dikes (Irving, 1980). Most of these samples are relatively rich in clinopyroxene and Al, and they have rather fertile major element compositions. One exception to this is sample KH80-106, a harzburgite with an Al_2O_3

concentration of $<1\%$ (w/w). Strontium and Nd isotopic compositions for some of the southwestern United States imply a complicated history of depletion and refertilization (Roden et al., 1988; Irving and Carlson, unpublished data). Long-term light rare earth element (LREE) depletion and resulting positive ϵNd values indicate derivation from MORB-like mantle (DMM) sources (Roden et al., 1988).

Four variably melt-depleted samples were examined from San Quintín, northwestern Mexico (UM-14, BCN-130, BCN-200, and BCN-203). Sample UM-14 was found in a host basanite that erupted near a continental margin rift. Its texture has been described as porphyroclastic gneissose (Basu, 1975) and is LREE depleted (Irving, 1980). The petrology of the remaining three samples is described by Luhr and Aranda-Gómez (1997).

Three anhydrous samples (fertile to depleted) from the Dreiser Weiher location, Westeifel, Germany, were previously studied by Stosch and Seck (1980), Stosch and Lugmair (1986), and Stosch et al. (1986). The Quaternary basaltic host lavas (basanites, nephelinites, and leucitites) erupted during the breakup of the Rhein Graben system.

Two samples (MBR 11 and MBR X) have been studied in the Montbriançon Quaternary scoria cone (Devès, Massif Central). MBR 11 is a LREE-depleted ($\text{La}_n/\text{Lu}_n = 0.8$) protogranular lherzolite. MBR X a secondary protogranular harzburgite with strong large ion lithophile element enrichment ($\text{La}_n/\text{Lu}_n = 10$; Lorand and Alard, personal communication). Modal metasomatism phases are lacking. MTF 37 is a LREE-depleted ($\text{La}_n/\text{Lu}_n = 0.3$), highly fertile porphyroclastic lherzolite collected in the Oligocene absarokite neck of Montferrier (Fabriès et al., 1987), a minor volcanism predating the plume-related Pliocene Quaternary volcanism of the Massif Central (see Granet et al., 1995). This sample contains several percent Ti-pargasitic amphibole.

Sample KF2 (Kapfenstein, Styria, Austria) is from the late tertiary volcanic province that extends from Slovenia, south-east Styria, south Burgenland (both Austria), into the Balaton area in Hungary. The host of this amphibole-bearing, porphyroclastic-deformed xenolith was a basanite (Kurat et al., 1980, 1991).

Two fertile samples (MHP79/2 and Mo101) are from the Tariat Depression of central Mongolia, associated with the Baikal rift system. The two samples are described as coarse-grained protogranular (Preß et al., 1986). Two samples are from Hannuoba Cenozoic basanites located in northeast China and were part of the suite studied by Song and Frey (1989). DM1-4 has olivine and pyroxene forming a very slightly elongated weak foliation. Pyroxene and spinel neoblasts surround large olivine aggregates, forming a tabular mosaic-porphyroclastic texture. DM1-7 is less fertile than DM1-4 (1.3% (w/w) vs. 2.7% (w/w) Al_2O_3) and has a coarse-tabular texture, with olivine and pyroxene slightly elongated.

Sample UM-15 is from Mount Quincan, Australia. Texturally, it is similar to the samples from Kilbourne Hole. Sample UM-1 is a hydrous sample from Nunivak Island (Alaska) that contains pargasitic amphibole, biotite, and apatite. It is medium grained with a preferred orientation of tabular grains (Francis, 1976).

All of the above-mentioned xenoliths are from regions of prehistoric continental rifting (e.g., Rio Grande rift: Kilbourne

Hole, San Carlos, Cochise Crater, Dish Hill, the Rhein Graben or the Baikal rift) or present near rifting continental margins (San Quintín, Nunivak Island). One exception to this is sample UM-13, from the nephelinite tuff at Salt Lake Crater (Hawaii). It is distinct from the other xenoliths in that it was not hosted by a magma generated in a rift setting and is found in a non cratonic, intraplate setting.

2.2. Garnet Peridotite Xenoliths from Cratonic Settings

The five depleted (Al_2O_3 poor) samples studied were all part of the BVSP. They include four garnet lherzolites from the Cretaceous Matsoku kimberlite pipe (Lesotho) and one small sample of garnet-spinel harzburgite from the Quaternary Lashaine ankaramite-carbonatite tuff ring (Tanzania).

The four coarse-grained garnet lherzolites (UM-16, UM-17, UM-18, and UM-19; equivalent to LBM 8, LBM 9, LBM 10, and LBM 11 of Cox et al. (1973) are rich in olivine and enstatite and contain very minor primary phlogopite (up to 1.3% (v/v)). Petrographic evidence for metasomatism is not apparent in these samples, and their pristine nature is noteworthy. Cryptic metasomatism, however, expressed by the LREE/heavy rare earth element enrichments up to La^n/Lu^n 10-33, is evident.

The Lashaine mantle xenoliths were carried by the Holocene ankaramite that occurs ~100 km to the east of the Archean Tanzanian craton within the south extending east African rift. Although erupted within a rift, there are indications that the source of the xenoliths is old lithospheric mantle (Burton et al., 2000). In addition, by use of Os isotopic data, Chesley et al. (1999) concluded that the lithosphere sampled by these xenoliths was at one time enriched in a plume component. Sample UM-20 from this locality has ~5% (v/v) pyrope garnet and 4% (v/v) chromite, with 6% (v/v) secondary phases (described as "clay"). Because of its small size, the whole rock composition of this sample is probably not representative of a significant mantle domain, which may explain the very low PGE abundances reported previously (Morgan et al., 1981).

3. ANALYTICAL METHODS

Samples that were not received as powder were finely ground in an agate mortar and pestle. Approximately 1 to 2 g of sample powder was weighed into Carius tubes and spiked with a mixed Re-Os spike. The digestion and separation techniques followed the methods described in Shirey and Walker (1995). More than 3 g of sample was unfavorable because a silicate gel formed during the 6- to 10-h digestion at 240°C. Formation of a gel was only observed for fresh unserpentinized peridotites. This gel prohibited the complete attack (leach) of the remaining silicates but did not affect the reproducibility of the Os concentration or $^{187}\text{Os}/^{188}\text{Os}$. Thus, most of the Os-bearing phases are presumed to be interstitial and effectively attacked by the acid mix. Treating the tubes after cooling with an ultrasonic bath reversed the gel to a solution. Osmium was separated by double distillation and further purified with microcolumns. Rhenium was collected and purified by anion exchange columns.

Most of the Re and Os isotopic compositions were measured with a single collector mass spectrometer with an electron multiplier used in a pulse-counting mode in the Isotope Geochemistry Laboratory (IGL) at the University of Maryland. Some of the higher Os abundance samples were measured via static multiple Faraday cups on a Sector 54 multicollector mass spectrometer, also at the University of Maryland. External reproducibility for the $^{187}\text{Os}/^{188}\text{Os}$ for the in house Os standard solution at the Isotope Geochemistry Laboratory by use of the single collector instrument in the pulse counting mode is about $\pm 0.25\%$

(2σ). The average standard value for $^{187}\text{Os}/^{188}\text{Os}$ obtained by the multiplier (0.1140) was ~0.2% higher than the value obtained via Faraday cup (0.1138). The reproducibility for the Faraday cups for this level of Os is approximately $\pm 0.1\%$ (2σ).

Total analytical blanks for the procedure during the course of this study were 40 ± 20 pg for Re and 7 ± 5 pg for Os. Osmium blank corrections were generally minimal ($\leq 1\%$), except for the case of UM-20. Rhenium blank corrections, however, were significant for most rocks and limit the uncertainty in $^{187}\text{Re}/^{188}\text{Os}$ from ± 5 to 50%. All data are blank corrected, and cited uncertainties are combined uncertainties calculated according to the rules of EURACHEM (1995) (Table 1).

4. RESULTS

4.1. Concentration Data

Concentrations of Os in the entire suite range from 0.05 to 13 ng/g, averaging 2.9 ng/g (Table 1). They do not show a normal distribution (Fig. 1a). The extreme high and low concentrations are only found in Archean garnet and garnet spinel peridotite samples, respectively. Excluding the old cratonic garnet peridotites, the average drops to 2.6 ng/g, with a standard deviation of 1.0 ng/g and a range of 0.85 to 5.2 ng/g. The average for this suite is significantly lower than the reported mantle average for Os of 3.3 ng/g reported by Morgan et al. (1981) and Morgan (1986). Carius tube data for individual samples that had been previously analyzed by radiochemical neutron activation analysis (RNAA) and reported by Morgan et al. (1981) and Morgan (1986), were on average ~10% lower than the RNAA results (Fig. 1b). This bias may be due to the incomplete attack of Os-bearing phases such as spinel during acid digestion, even under the elevated internal pressure of the Carius tube digestion. Incomplete digestion of different platinum group minerals can be problematic if these phases are isotopically heterogeneous, if Re is sited in other phases, or if Re is leached preferentially from the same phases. For example, recent mass balance calculations performed on mantle peridotite xenoliths indicate that whereas Os is sited almost exclusively in sulfides, Re may exhibit a greater preference for either garnet (Righter and Hauri, 1998) or olivine and pyroxene (Burton et al., 1999; Handler and Bennett, 1999). Analysis of duplicate analyses suggests that the selective attack of phases may affect the measured $^{187}\text{Re}/^{188}\text{Os}$ but probably did not affect the Os isotopic composition of whole rock samples, given the limited differences in concentration between the two techniques.

The average Re concentration of the entire sample population is 0.19 ng/g, with a standard deviation of 0.11 ng/g (Fig. 1a). As with Os, the average is lower than the reported upper mantle estimate of 0.25 ng/g (Morgan, 1986). It should be noted, however, that the estimate was based on an extrapolation of Re concentrations from a variety of depleted xenoliths to a hypothetical fertile composition for MgO. This was necessary because Re is an incompatible element under mantle melting conditions. Compared with the RNAA results, our new Re data are on average higher, but there are much greater variations than with Os, and this difference may be an artifact resulting from a nugget effect (Fig. 1c).

Some xenoliths have unsupported $^{187}\text{Os}/^{188}\text{Os}$, with ratios that are similar to or more radiogenic than the chondritic average of 0.1270 but subchondritic $^{187}\text{Re}/^{188}\text{Os}$ (average chondritic is 0.403). For other samples, the opposite is true. Possible reasons for these characteristics include the following:

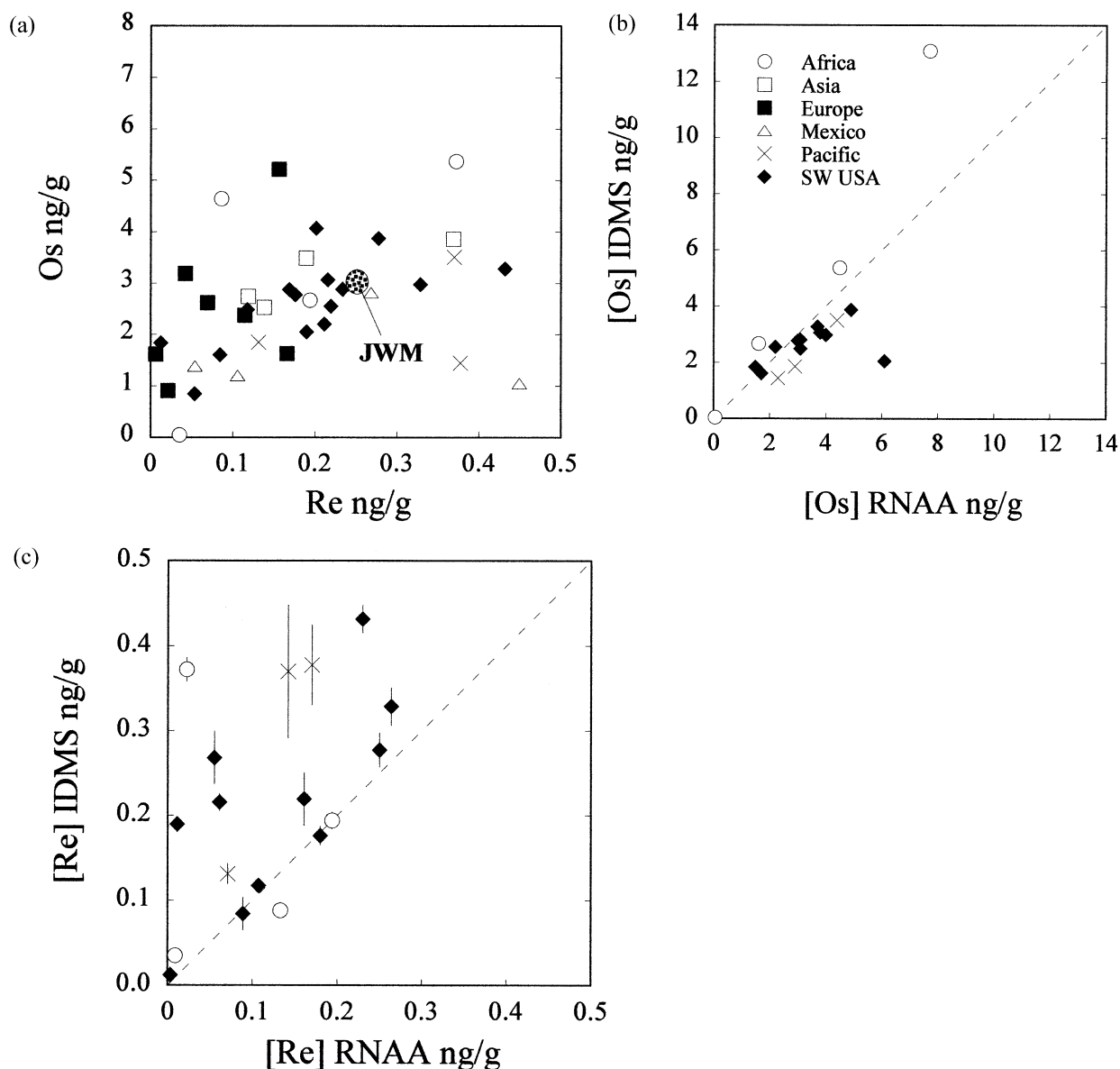


Fig. 1. (a) Osmium vs. Re concentrations of all mantle xenoliths from this work. JWM is the primitive mantle composition defined by Morgan et al. (1981). Analytical uncertainties resulting from blank corrections are shown by the error bars for Re. Uncertainties for Os are smaller than the symbol size. (b) Comparison of isotope dilution TIMS data after Carius tube digestion with Os concentration from Morgan et al. (1981) prepared by radiochemical neutron activation analysis. An offset of spinel peridotites toward higher Os concentrations as measured by radiochemical neutron activation analysis (RNAA) is possibly due to incomplete acid digestion for the isotope dilution results. (c) Rhenium isotope dilution data for spinel peridotites are higher or similar to the RNAA data from Morgan et al. (1981). The reason for the large scatter is unclear. Analytical uncertainties resulting from blank corrections are shown by the error bars for Re. Uncertainties for Os are smaller than the symbol size. SW USA, southwestern United States.

contamination of the xenoliths during uplift in a Re-rich magma; weathering of sulfides and thus loss of Re after eruption; loss of Re during or shortly before eruption; or analytical problems, such as erratic blank. In any case, losses or gains of Re in most xenoliths must have occurred recently, i.e., before, during, or after eruption, because most $^{187}\text{Os}/^{188}\text{Os}$ are lower than PUM (i.e., ≤ 0.129) and, as will be seen below, in most instances display good linear correlations with melt depletion indicators.

Duplicate digestions of six spinel-bearing peridotites gave

limited deviations between 0.2 and 4.5 relative percentages for Os concentrations, suggesting that the phase or phases containing the Os are homogeneously distributed throughout the 1.0- to 2.0-g-size aliquots of sample powder analyzed. Rhenium differences between two duplicates, however, were different by as much as 50 relative percent. These differences suggest that Re is not homogeneously distributed within the powders, perhaps reflecting a nugget effect, variable contamination, or weathering effects.

Direct comparison of some of the new data with isotope

dilution data collected by other researchers is also possible. For example, Matsoku samples UM-18 and UM-19 (LBM 10 and LBM 11, respectively) were also analyzed by Olive et al. (1997) using a low-temperature and low-pressure microwave digestion method. They obtained much lower Os concentrations of 2.63 vs. our 4.64 ng/g (UM-18) and 2.33 vs. 13.1 ng/g (UM-19). UM-19 was analyzed twice by RNAA (Morgan et al., 1981) and yielded 7.7 and 10.8 ng/g, respectively. The large scatter in Os abundance is evidently caused by a nugget effect (in contrast to Os in spinel peridotite xenoliths), but an influence caused by the different digestion techniques cannot be excluded. Rhenium concentrations, on the other hand, are ~30% lower for Carius tube digestion (this work) compared with those of Morgan et al. (1981).

Sample UM-20, from Lashaine, Tanzania, has the lowest Os concentration among the samples measured in the present study (Table 1). This is also in accord with the low overall PGE abundances in this rock (Morgan et al., 1981). It has been demonstrated that percolation of peridotites by high melt fractions (>1%) at high temperature (1200°C) and fO_2 (FMQ + 2 log units) are capable of removing PGE host minerals from mantle peridotites (Brandon et al., 1996; Brandon, personal communication). For Lashaine, such melt interaction probably did not change the Os isotopic composition, given that it would have been a net removal process.

4.2. Isotopic Data

The $^{187}\text{Os}/^{188}\text{Os}$ ratios of all samples lie within the range 0.1095 to 0.1336. Of note, all but 2 of the 39 xenoliths examined here have ratios < 0.130, the maximum estimate of the modern PUM reported by Meisel et al. (1996). The corresponding $^{187}\text{Re}/^{188}\text{Os}$ ratios encompass a large range, with 14 out of 39 samples having ratios that are suprachondritic (i.e., >0.4). For individual suites of xenoliths, the scatter in $^{187}\text{Os}/^{188}\text{Os}$ vs. $^{187}\text{Re}/^{188}\text{Os}$ (Fig. 2) is much greater than has been noted for massif peridotites (Reisberg and Lorand, 1995). This is probably an indication of late-stage open-system behavior of Re in some of the xenoliths.

A dependence of $^{187}\text{Os}/^{188}\text{Os}$ with the melt depletion indicator Mg was first observed for peridotite xenoliths from the Kaapvaal craton, southern Africa (Walker et al., 1989), and for massif peridotite samples from Ronda and the Pyrenees (Reisberg et al., 1991; Reisberg and Lorand, 1995). Other melt depletion indicators, such as Mg/Si, Al/Mg, Al/Si, olivine or clinopyroxene abundance, Al_2O_3 concentration in clinopyroxene or whole rock, V, and Yb, Cr also normally show good linear trends with the isotopic composition. Ratios are less dependent on the mineral abundance in a given aliquot of sample and thus sometimes show better trends than absolute concentration data. Plots of $^{187}\text{Os}/^{188}\text{Os}$ vs. whole rock Al_2O_3 and Al/Mg are shown in Figure 3 for the different locales of this study. Melt depletion indicators determined on single minerals, such as the Al concentration in clinopyroxene, should be even more reliable than whole rock data, but data are less abundant. For the suites where mineral data have been published, all show good linear correlations with $^{187}\text{Os}/^{188}\text{Os}$ (Fig. 4).

The Os isotopic compositions of two samples seem to be independent of secondary enrichment in highly incompatible

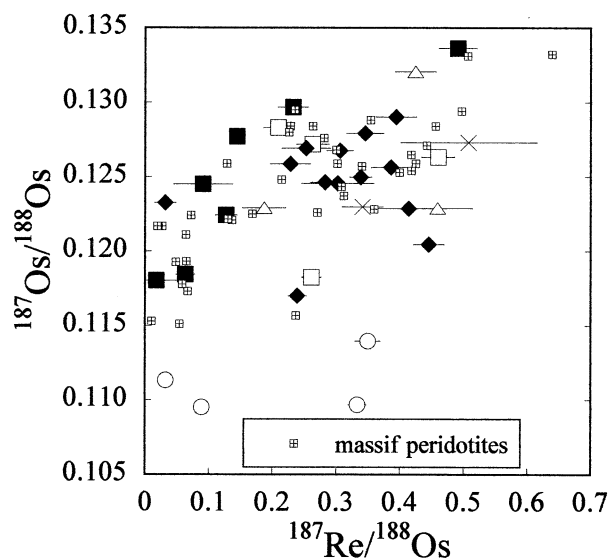


Fig. 2. $^{187}\text{Re}/^{188}\text{Os}$ vs. $^{187}\text{Os}/^{188}\text{Os}$ plot for all data from this work, plus massif peridotite data (Reisberg and Lorand, 1995; Burnham et al., 1998). Although scattered, a depletion trend toward lower isotopic composition of the spinel peridotite data can be seen.

trace elements. Cryptic or modal metasomatism, e.g., phlogopite in Nunivak island xenoliths or pargasite in Mtf37, must have occurred either shortly before eruption or must have had little influence on the Re abundance. This lack of metasomatic enrichment in ^{187}Os is unlike certain peridotite xenoliths with suprachondritic $^{187}\text{Os}/^{188}\text{Os}$ ratios reported for some locales that likely sample mantle wedge materials above a dewatering slab (e.g., Brandon et al., 1996).

Replicate measurements of digestions of six independent samples measured in the course of this study gave relative differences between duplicate measurements that are within the expected external reproducibility of a comparable quantity of an Os standard solution. The one exception is UM-12, for which duplicate analyses differed by ~1.3%. As with concentration data, the isotopic data of UM-18 and UM-19 in Table 1 (LBM 10 and LBM 11, respectively) can be compared directly with the data of Olive et al. (1997). The new results are significantly lower and do not overlap with the Olive et al. data within the cited uncertainties. Whereas their data are identical within uncertainty (0.1142 ± 0.0004 and 0.1156 ± 0.020), our data are distinctly different and lower (0.1095 ± 0.0001 and 0.1113 ± 0.0003). Good correlations with melt depletion indicators with $^{187}\text{Os}/^{188}\text{Os}$ for the four Matsoku of this work (Fig. 3) give confidence that the relative difference between these two Matsoku samples are meaningful. As noted with regard to the concentration data, the study by Olive et al. used a less vigorous digestion technique that may not have liberated all Os and may have left a nonradiogenic component.

4.2.1. Spinel-Bearing Xenoliths

Sample suites from the Massif Central, Kilbourne Hole (and other southwestern U.S. localities), Dreiser Weiher, San Quintín, Tariat, and Hannuoba span almost the whole range from very depleted (i.e., low Al_2O_3 and subchondritic $^{187}\text{Os}/^{188}\text{Os}$)

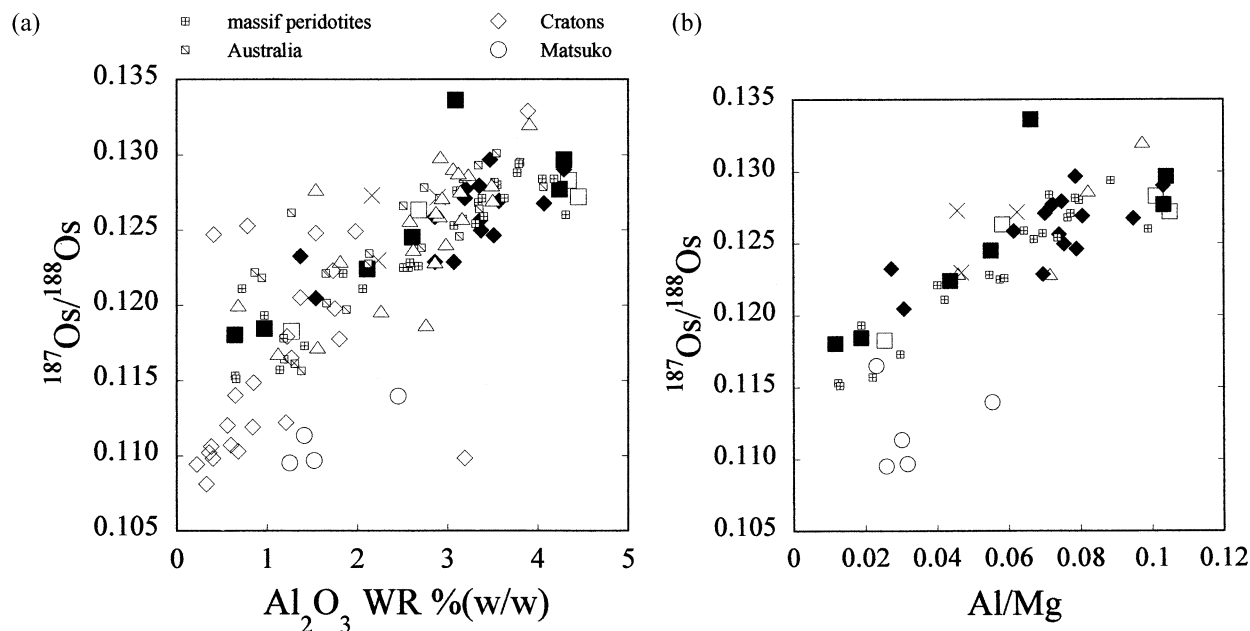


Fig. 3. Osmium isotopic composition vs. whole rock Al_2O_3 abundances and Al/Mg, respectively. Depletion trends for spinel peridotites are more apparent than in the isochron plot (Fig. 2). Data for kimberlite-hosted xenoliths, mainly garnet bearing, tend to scatter throughout the entire diagram but plot mostly at the lower end of the $^{187}\text{Os}/^{188}\text{Os}$ range. Craton data include isotope data from the literature (Carlson and Irving, 1994; Pearson et al., 1995a,b; Chesley et al., 1999) and unpublished Al_2O_3 data (Carlson and Pearson, personal communication).

to fertile ($\text{Al}_2\text{O}_3 \sim 3\text{--}4\%$ (w/w)) and $^{187}\text{Os}/^{188}\text{Os}$, similar to PUM. Data for xenoliths from each general area form linear trends with similar slopes in the $^{187}\text{Os}/^{188}\text{Os}$ vs. Al_2O_3 and Al/Mg diagrams (Fig. 3; Table 2). For all these suites, the presence of only two samples (MBR 11 and BCN-200) with Os isotopic compositions that are significantly higher (though only

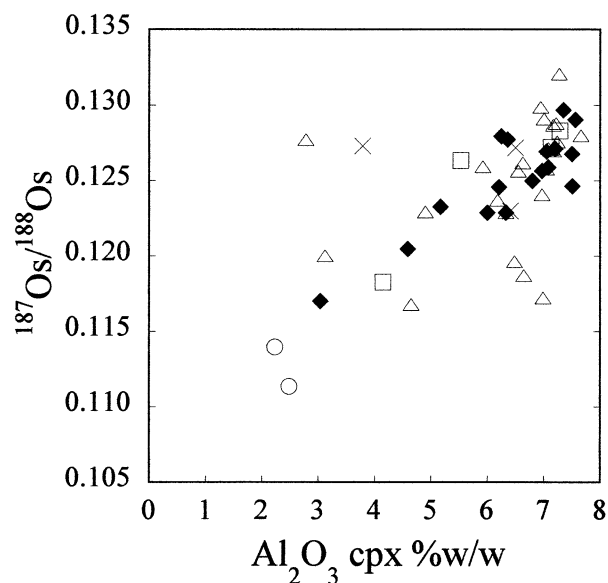


Fig. 4. Although dependent on temperature, a depletion trend of $^{187}\text{Os}/^{188}\text{Os}$ whole rock data with Al_2O_3 concentrations of clinopyroxene is visible.

marginally so) than the earlier estimate for the PUM maximum suggests that the mantle materials represented by this group of samples are rarely enriched in a Re-bearing melt or fluid component after their original depletion event. This is in contrast to some rare peridotites from subduction-related volcanic hosts that plot above the PUM estimate (Brandon et al., 1996), although normally with very different major and trace element characteristics. The ^{187}Os enrichment in these instances can be explained by addition of a crustal Os component in the mantle wedge overlying the subduction zone.

Mid-Proterozoic formation ages for the massif peridotites from Ronda and the Pyrenees were postulated by Reisberg and Lorand (1995) on the basis of the Re depletion model age of the y-axis intercept in the Al_2O_3 vs. $^{187}\text{Os}/^{188}\text{Os}$ diagram. On the basis of the same type of model age calculation, Proterozoic model depletion ages ranging from 1.8 Ga (San Quintín) to 1.3 Ga (southwestern United States) can be proposed for the SCLM sources of these xenolith suites (Table 2).

The Hawaiian xenolith from Salt Lake crater (UM-13) yields a rather low $^{187}\text{Os}/^{188}\text{Os}$ of 0.1230. The xenolith probably represents material removed from the underlying asthenospheric mantle, oceanic lithosphere, or a cumulate from the lava system. In any event, the depleted isotopic composition suggests moderate-term (several hundreds of million years) evolution with a subchondritic Re/Os ratio. The isotopic composition is within the range reported from abyssal peridotites (Snow and Reisberg, 1995).

4.2.2. Garnet Peridotite Xenoliths from Archean Cratons

The five samples from the Kaapvaal (Matsoku) and the Tanzanian cratons have highly variable Os concentrations (0.05

Table 2. Linear regressions of the xenoliths suites and estimated $^{187}\text{Os}/^{188}\text{Os}$ composition of primitive upper mantle.

Region	Linear regression ^a			$^{187}\text{Os}/^{188}\text{Os}$ at 4.2% (w/w) Al_2O_3	Confidence interval ^b		<i>n</i>	T_{RD} (Ga)	Crustal ages ^d (Ga)
	MO	MI	R		Lower	Upper			
Pacific	0	1					2		
Asia	0.1163	0.002745	0.907	0.1278			4	1.58	
Australia	0.1158	0.003481	0.765	0.1304	0.1274	0.1334	21	1.65	0.6, 1.9 ^e
Europe	0.1160	0.003019	0.990	0.1287	0.1273	0.1300	6	1.62	1.5 ^f
Southwestern United States	0.1165	0.002865	0.820	0.1285	0.1269	0.1301	17	1.55	
Mexico	0.1151	0.003735	0.717	0.1308	0.1279	0.1336	23	1.75	
Massif	0.1141	0.003685	0.930	0.1296	0.1286	0.1306	32	1.90	1.9–2.6 ^g
Canada	0.1203	0.002404	0.936	0.1304	0.1294	0.1314	12	0.99	0.5; 1.9–2.3 ^h
Average	0.1163			0.1295					
Standard deviation	0.0019			0.0011					
All	0.1160	0.003253	0.812	0.1296	0.1289	0.1304	117	1.62	
Upper	0.1148							1.80	
Lower	0.1172							1.45	

^a Linear regression, $(^{187}\text{Os}/^{188}\text{Os}) = \text{MO} + \text{MI} \cdot \text{Al}_2\text{O}_3$.

^b Confidence interval at 95% level with upper and lower limits.

^c T_{RD} , rhenium depletion model age after Shirey and Walker (1998).

^d Crustal ages, ages found in the overlying crust.

^e Handler et al. (1997).

^f Stosch and Lugmair (1986).

^g Reisberg and Lorand (1995).

^h Peslier et al. (2000).

and 13 ng/g) and low, but a limited range, in $^{187}\text{Os}/^{188}\text{Os}$ (0.1095–0.1165). The isotopic compositions observed for the Kaapvaal samples are similar to those reported previously (Walker et al., 1989; Pearson et al., 1995a; Olive et al., 1997), with T_{RD} ages, calculated after the method of Shirey and Walker (1998), as old as 2.8 Ga. The results are also similar to the range of Os isotopic compositions reported for xenoliths from the Siberian craton (Pearson et al., 1995b), the Wyoming craton (Carlson and Irving, 1994), and the Tanzanian craton samples that escaped extensive reequilibration with recent plume-derived Os (Chesley et al., 1999). The samples studied here are all characterized by low Mg/Si, and thus by a restitic–cumulitic character (Pearson et al., 1995a). As a consequence, the data plot in a field that is distinctly different from off cratonic samples in most $^{187}\text{Os}/^{188}\text{Os}$ vs. element or ratio plots (Fig. 3). This suite includes no samples with fertile or approaching fertile compositions. Consequently, unlike with the Proterozoic suites, extrapolation to a fertile end member is not a viable enterprise with these samples.

5. DISCUSSION

5.1. $^{187}\text{Os}/^{188}\text{Os}$ of Modern PUM and Persistence of Apparently Fertile Mantle

As shown above, xenolith suites from seven different suites (southwestern United States, Hannuoba, Tariat, Dreiser Weiher, San Quintín, and the Massif Central) all define generally linear trends on plots of $^{187}\text{Os}/^{188}\text{Os}$ vs. melt depletion indicators with a common intersection point (Fig. 3). Similar linear trends are also noted for xenolith suites from south Australia (Handler et al., 1997) and Canadian Cordillera (Peslier et al., 2000), as well as for orogenic lherzolites massif peridotites from the Pyrenees (Reisberg and Lorand, 1995) and Ronda (Reisberg et al., 1991). Linear trends have also been reported for systems that apparently became melt depleted after

the Proterozoic, as for orogenic peridotite massifs from the eastern Alps of Austria (Meisel et al., 1997) and several suites from southern Australia (Handler et al., 1997). The new data presented here, together with previously published data, were grouped per region. In all, seven groups are considered: Asia, Australia, Europe, southwestern United States, Mexico, massif, and Canada. The Al_2O_3 and $^{187}\text{Os}/^{188}\text{Os}$ data for each group were individually regressed (Fig. 5). Each trend was extrapolated up to a fertile composition Al_2O_3 content of 4.2% (w/w). Results show intercepts of $^{187}\text{Os}/^{188}\text{Os}$ at the fertile Al_2O_3 composition ranging from 0.1278 to 0.1308. A similar value (0.1301) is reported for the most fertile Vitim samples (Pearson et al., 1998). A compilation of regression results for the different suites is provided in Table 2.

All of the above-mentioned suites likely represent Proterozoic SCLM. Extrapolations for samples derived from Archean SCLM are more problematic. This is not only true for our limited suite of samples (see above) but for published suites from southern Africa (Pearson et al., 1995a; Carlson et al., 1999), the Wyoming Province of North America (Carlson and Irving, 1994), Tanzania (Chesley et al., 1999), and the Siberian Craton (Pearson et al., 1995b). With the exception of several samples from the Tanzanian suite (Chesley et al., 1999), samples derived from SCLM underlying Archean cratons all have undergone extensive melt depletion; thus, these suites do not include samples with nearly fertile compositions. Consequently, extrapolating these data to a fertile composition is not meaningful (Fig. 5). The absence of fertile xenoliths likely reflects more extensive melting during the Archean, compared with the Proterozoic and Phanerozoic.

As discussed by Meisel et al. (1996), the intersection foci of the linear trends at fertile compositions for the Proterozoic suites is an important constraint on the modern $^{187}\text{Os}/^{188}\text{Os}$ of the PUM. The much larger database presented here, including data from other studies, remains consistent with the earlier

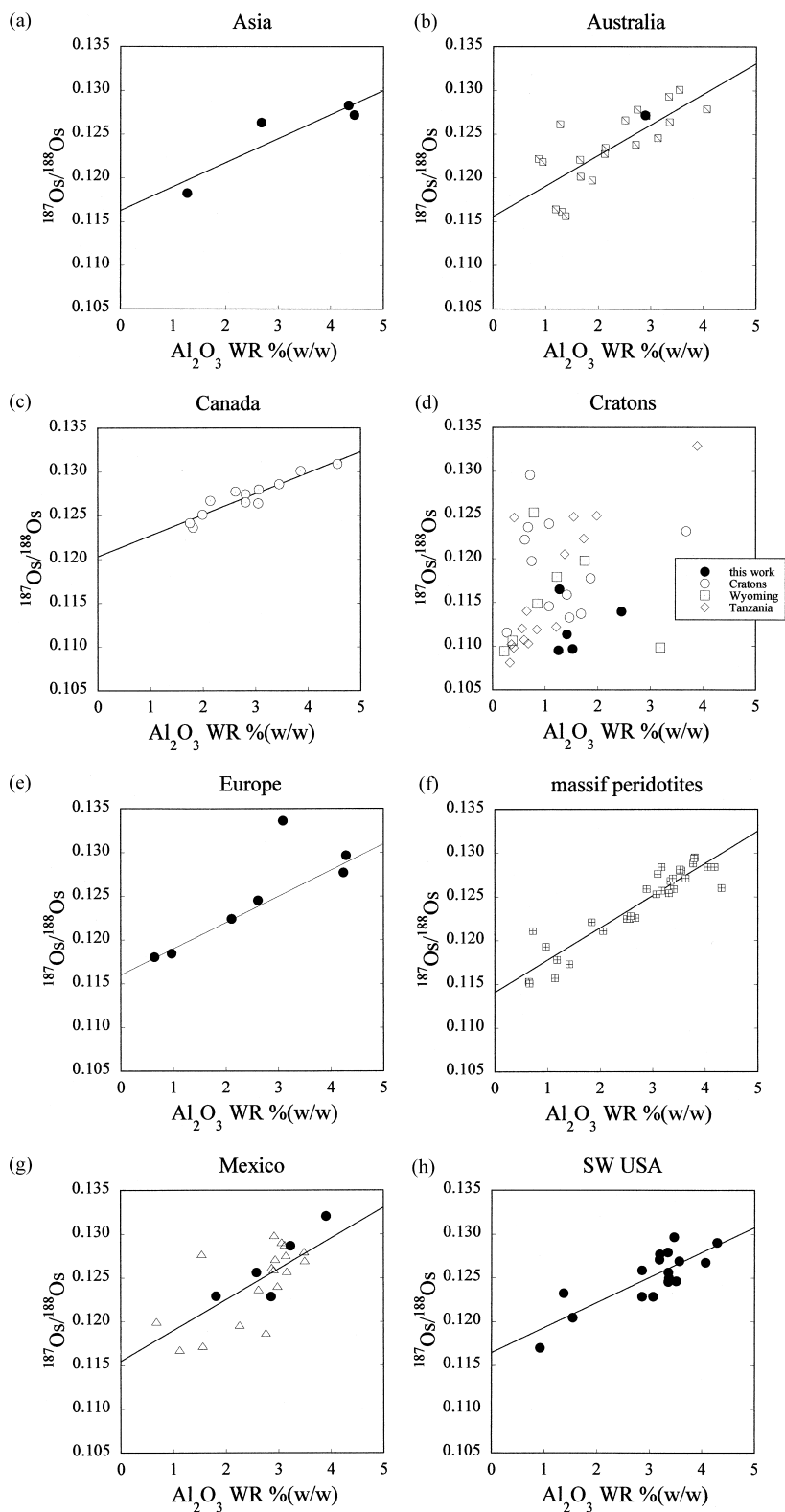


Fig. 5. Correlation graphs for all regions include all suites but exclude the Nunivak and Hawaiian samples. Samples from this work are marked with filled black circles. Also displayed are correlations of European massif peridotites (Reisberg and Lorand, 1995; Burnham et al., 1998), additional Mexican samples (unpublished data), Canadian cordillera (Peslier et al., 1998), and samples from cratonic regions including Siberia (Pearson et al., 1995b), Kaapval (Pearson et al., 1995a), Tanzania (Chesley et al., 1999), Wyoming (Carlson and Irving, 1994), and Australia (McBride et al., 1996; Handler et al., 1997). Only the data for pure lherzolites are plotted. The individual regression data for each correlation line are given in Table 2.

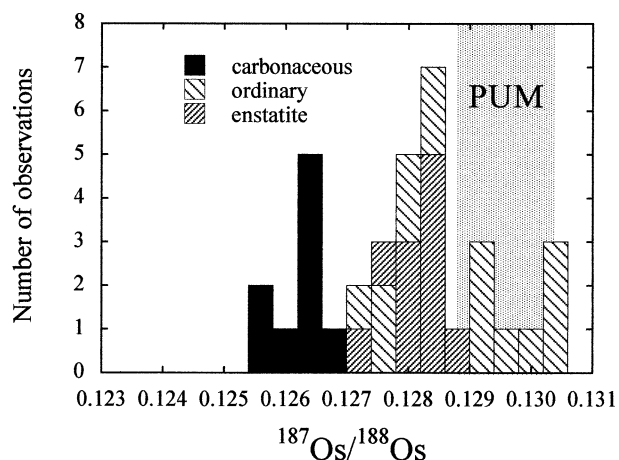


Fig. 6. A comparison of chondrite data and unpublished data with the new isotopic composition of the primitive upper mantle (PUM). PUM is significantly more radiogenic than carbonaceous chondrites but is well within the range of compositions defined by ordinary and enstatite chondrites.

value. The new, refined PUM $^{187}\text{Os}/^{188}\text{Os}$ ratio of 0.1296 ± 0.0008 (95% confidence interval and intercept of the correlation of 117 samples at 4.2% w/w Al_2O_3) is only marginally higher than the earlier reported value of 0.1290 ± 0.0009 . As has been noted previously (Walker et al., 1998, 1999), this composition is most similar to that of enstatite and ordinary chondrites, not carbonaceous chondrites (Fig. 6). Thus, if the highly siderophile element composition of the PUM was set by late accretion (Chou, 1978; Morgan, 1986), the average parental component for the highly siderophile elements in the upper mantle, including Re and Os, had a nebular history most like that of ordinary or enstatite chondrites. The greatly enlarged database presented here further substantiates the earlier estimate. Similar late veneer components have been postulated from recent platinum group element studies of spinel peridotites from orogenic massifs (Pattou et al., 1996) and basalt-hosted xenoliths (Rehkämper et al., 1997).

A surprising aspect of the xenolith database for Os is that there exists a moderate proportion of xenolith samples that have both fertile compositions and Os isotopic compositions that are very similar to the PUM estimate (UM7, UM12, and Ib8). The postulated present PUM composition of $^{187}\text{Os}/^{188}\text{Os}$ 0.1296 is within the range of chondrites sampled by the Earth in the last centuries. This implies that the Re-Os systematics of some fertile samples were minimally affected throughout the evolution of the Earth's mantle. This is a somewhat unexpected observation because all mantle samples that were studied for Os isotopes have in common that they underwent at least one melt extraction event. For example, even the most fertile end members of the studied xenolith suites are characterized by positive MORB-like ϵ_{Nd} values (Stosch and Lugmair, 1986; Roden et al., 1988; Song and Frey, 1989), and consequently, they experienced long-term LREE depletion relative to the Mg/Si ratio that is higher than chondrites. Thus, some samples either experienced little depletion of Re during events that removed a substantial proportion of LREE, or the materials were somehow refertilized at some stage with melt components, including Re, but not LREE or other highly incompatible

trace elements. The former possibility could potentially be achieved via low degrees of melting where the mild incompatibility of Re would lead to little depletion of Re. The latter possibility requires that the duration of the depleted period (i.e., when Re was depleted) was relatively short (<500 Ma), and it requires that the added material bring the bulk composition of the rock back to within the boundaries of fertile mantle. Refertilization of larger mantle domains is a very unlikely process, yet it almost certainly occurs in some smaller mantle domains. Although it is an unlikely process to erase an earlier depletion history in Re-Os systematics, this is a process that cannot, at this time, be excluded from consideration.

5.2. Comparison of SCLM with the DMM and Implications for Mass Balance of Re in the Upper Mantle

The fact that all of the groups considered suggest a similar, undepleted composition for $^{187}\text{Os}/^{188}\text{Os}$ is remarkable because the groups sample SCLM from five different continents that was isolated, in some cases, nearly 1 billion years apart. If Re, together with melt, was continually extracted and permanently removed from the convecting upper mantle, isolation of SCLM at different times in Earth's history might be expected to lead to long-term variations in the Os isotopic compositions of samples that were not depleted in Re within each reservoir. For example, a small domain that did not experience a significant change in Re/Os when isolated from the convecting upper mantle during the early Proterozoic (e.g., the Pyrenees) might be expected to evolve to a more radiogenic composition than material more recently isolated (e.g., the Canadian Cordillera), because of the potential long-term loss of Re from the upper mantle because of its mild incompatibility. In this scenario, Re is gradually removed from the upper mantle as it is selectively incorporated in continental crust or subducted in oceanic crust into the lower mantle. The lack of an observable difference between early and late Proterozoic trends suggests either that the data are too uncertain to detect the differences between older and younger reservoirs, or that the long-term Re depletion of the upper mantle is insufficient to cause a detectable difference in old vs. young reservoirs. With the current database, the data are right at the level of resolution. For example, a decrease of 20% in the Re/Os of a mantle reservoir for ~1 Ga would lead to a $^{187}\text{Os}/^{188}\text{Os}$ that would only be ~1% lower than an undepleted reservoir at the end of the period. Consequently, the potential slowing of the growth trajectory of the convecting upper mantle is not quite resolvable.

The long isolation and Re depletion history of Archean and Proterozoic SCLM is contrasted with the apparent chondritic Os isotopic composition of the DMM, as constrained via studies of modern MORB glasses and sulfide inclusions (Roy-Barman and Allègre, 1994; Schiano et al., 1997; Roy-Barman et al., 1998) and ophiolites of various age (Walker et al., 1998). These materials have $^{187}\text{Os}/^{188}\text{Os}$ ratios that range to as low as 0.127 for modern rocks, and compositions for chromitites from ophiolites of variable age do not stray more than ~2% below the average evolution of chondrites. Although many low Os abundance samples of MORB glass may have more radiogenic Os than their source because of seawater contamination (modern seawater $^{187}\text{Os}/^{188}\text{Os}$ is ~1; Sharma et al., 1997), the

constancy of ratios at the low end of the MORB field appears to define a dominant composition for the DMM of ~ 0.127 to 0.130 (Roy-Barman et al., 1998), similar to our estimate for PUM.

Despite the evidence for chondritic evolution of DMM from MORB and ophiolite chromitites, it is important to note that samples of presumably young abyssal peridotites have Os isotopic compositions that range from chondritic to as depleted as some SCLM xenoliths (0.120 – 0.129 ; Martin, 1991; Roy-Barman and Allègre, 1994; Esperança et al., 1999; Brandon et al., 2000) and also that the average $^{187}\text{Os}/^{188}\text{Os}$ of abyssal peridotite compositions (0.124 – 0.125 ; Snow and Reisberg, 1995) is significantly lower than the PUM estimate and the DMM estimate based on MORB and ophiolite chromitites. Consequently, there remains considerable debate about the true average Os isotopic composition of the convecting upper mantle. The $^{187}\text{Os}/^{188}\text{Os}$ composition range of abyssal peridotites may thus either indicate that the magnitude of Os isotopic heterogeneities within the DMM is quite large, a range evidently not sampled by the MORB or ophiolites that have been examined, or that the range reflects local heterogeneities, perhaps resulting from ancient depletion events, which is not representative of the bulk DMM. This issue remains to be resolved.

If it is assumed that the Os isotopic compositions reported for MORB and ophiolites is representative of the average DMM, then it must be concluded that the PUM and the DMM, at present, have indistinguishable Os isotopic compositions. This in turn requires that the convecting upper mantle has experienced no significant long-term history of Re removal. If this is true, why isn't Re removed from the convecting upper mantle, given its incompatible trace element behavior?

There are several possible mechanisms that permit Re removal from the mantle but that may result in no major change in the Re/Os of the convecting upper mantle. First, consider the formation of continental crust. Evidence presented here and in previous publications (e.g., Walker et al., 1989) indicates that the presence of chemically isolated keels beneath cratonic areas is probably the rule rather than the exception. Thus, Re removal into continental crust is likely dominated by removal of Re from SCLM underlying the new crust, some portion of which is evidently not mixed back into the upper convecting mantle, and hence becomes chemically isolated. Isolation of lithospheric mantle from the convecting upper mantle system is only possible in keels below the continental crust. Such SCLM material must be buoyant as a result of the removal of Al- and Fe-rich melts, leaving a Mg-rich, less dense material behind. Recent estimates of the Re concentration in the continental crust are only 0.20 ng/g for the lower continental crust and 0.40 ng/g for the upper continental crust, respectively (Saal et al., 1998). Bulk continental crust must therefore have a Re concentration that is similar to that of the convecting upper mantle. Continental crust has $<1\%$ of the mass of the entire mantle, and the Re concentration difference between crust and mantle— 0.20 to 0.40 vs. 0.25 ng/g, respectively—is small. Thus, formation of continental crust should have had little impact on the concentration of Re in the convecting upper mantle.

Another mechanism that may lead to a retention of Re in the upper convecting mantle is rapid remixing of oceanic crust back into the upper mantle. If oceanic basaltic crust is not

removed to the lower mantle and isolated for long periods of time, there should be little net removal of Re from the upper mantle, with only the Re present in the oceanic crust isolated for a short term.

Finally, a study of eclogites thought to be pieces of ancient subducted oceanic crust suggests that up to 60% of Re is removed from downgoing slabs (Becker, 2000). The removed Re may end up in the overlying mantle wedge during dehydration and ultimately become convectively remixed into the upper mantle.

5.3. Implications for the Growth of Continental Crust

The y-axis intercepts for the groups that were regressed (Table 2) suggest similar melt depletion ages ranging only from 1.9 Ga (peridotite massifs) to 1.0 Ga (Canadian Cordillera). These y-axis intercepts can most easily be explained by melt depletion events of most spinel-bearing mantle xenolith systems within a relatively limited period of time, <1 billion years. Although the xenoliths investigated in this study are not necessarily random samples of SCLM, the intense melt depletion and formation of SCLM recorded in these suites could be the consequence of an important continental crust-forming period that occurred globally while the Earth was still sufficiently hot.

This observation may contradict some estimates of continental crust growth. According to Nägler and Kramers (1998), most of the continental crust growth ($\sim 70\%$) took place before 2.0 Ga, whereas the $^{187}\text{Os}/^{188}\text{Os}$ - Al_2O_3 regression ages suggest that a significant amount of SCLM was formed after 2.0 Ga. If it is assumed that the SCLM Os model ages are accurate, then the continental crust may have served as an anchor for the later-forming SCLM. This implies that the continental crust and the underlying SCLM are complementary neither in chemical composition nor in age. Alternatively, spinel-bearing xenoliths may represent the remaining younger mantle (30%) that could have been added to continental roots during rifting. It should be noted that the Re/Os is lowered by large degrees of melting, as reflected in MORB. This process does not change the Sm/Nd, whereas low degrees of partial melting leading to the formation of continental crust fractionates Sm/Nd but not Re/Os. Thus, continent formation may not leave any traces in the Re-Os isotope systems. This problem will ultimately have to be addressed via careful studies of the relations between SCLM ages and the age of overlying crust for individual cratonic domains (e.g., Peslier et al., 2000).

6. SUMMARY

A relatively uniform PUM $^{187}\text{Os}/^{188}\text{Os}$ ratio of 0.1296 ± 0.0008 is indicated by this study. This composition most likely indicates that Re and Os were added to the Earth as a late veneer after core formation. The late veneer was composed of materials whose nebular history of highly siderophile elements was similar to that of ordinary and enstatite chondrites. It appears that some fertile portions in the upper mantle escaped depletion during the formation of Proterozoic SCLM from chondritic asthenosphere. Although slightly depleted with respect to large ion lithophile elements, this process evidently had little or no effect on the Re-Os system.

The potential similarity in Os isotopic composition between the PUM and the DMM suggests that little Re-enriched ocean crust is stored in the lower mantle and isolated for long periods of time from the convecting upper mantle. Either Re is removed from downgoing slabs during dewatering, or subducted slabs are largely recycled back into the convecting upper mantle.

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