Peridotite xenoliths from the Polynesian Austral and Samoa hotspots: Implications for the destruction of ancient ^{187}Os and ^{142}Nd isotopic domains and the preservation of Hadean ^{129}Xe in the modern convecting mantle.

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Abstract

The Re-Os systematics in 13 peridotite xenoliths hosted in young (<0.39 myr) rejuvenated lavas from the Samoan island of Savai'i and 8 peridotite xenoliths from 6-10 myr old lavas from the Austral island of Tubuai have been examined to evaluate the history of the oceanic mantle in this region. Modal mineralogy, trace element compositions and $^{187}\text{Os}/^{188}\text{Os}$ ratios suggest that these peridotites are not cognate or residual to mantle plumes but rather samples of Pacific oceanic lithosphere created at the ridge. Savai'i and Tubuai islands lie along a flow line in the Pacific plate, and provide two snapshots (separated by over 40 Ma in time) of Pacific mantle that originated in the same region of the East Pacific rise. Tubuai xenoliths exhibit $^{187}\text{Os}/^{188}\text{Os}$ from 0.1163 to 0.1304, and Savai'i (Samoa) xenoliths span a smaller range from 0.1173 to 0.1284. The $^{187}\text{Os}/^{188}\text{Os}$ ratios measured in Tubuai xenoliths are lower than (and show no overlap with) basalts from Tubuai. The $^{187}\text{Os}/^{188}\text{Os}$ of the Savai'i xenoliths overlap the isotopic compositions of lavas from the island of Savai'i, but also extend to lower $^{187}\text{Os}/^{188}\text{Os}$ than the lavas. $^3\text{He}/^4\text{He}$ measurements of a subset of the xenoliths range from 2.5-6.4 Ra for Tubuai and 10.8-12.4 Ra for Savai'i.

Like abyssal peridotites and xenoliths from oceanic hotspots that sample the convecting mantle, Os isotopes from the Savai'i and Tubaui xenolith suite are relatively unradiogenic, but do not preserve a record of depleted early-formed (Hadean and Archean) mantle domains expected from earlier cycles of ridge-related depletion, continent extraction, or subcontinental lithospheric mantle erosion. The lack of preservation of early-formed, depleted Os-isotopic and $^{142}\text{Nd}/^{144}\text{Nd}$ domains in the modern convecting mantle contrasts with the preservation of early-formed (early-Hadean) $^{129}\text{Xe}/^{130}\text{Xe}$ isotopic anomalies in the convecting mantle. This can be explained if the initial isotopic signatures in Re-Os and Sm-Nd systems are erased by recycling because the parent and daughter elements are retained in subducting slabs and more efficiently returned to the mantle during subduction than Xe. In this way, early-formed Os and Nd-isotopic heterogeneities could have been overprinted with, and diluted by, younger isotopic signatures. In contrast, preservation of early-formed heterogeneities in the modern convecting mantle is possible for other elements, such as Xe, that are not as efficiently recycled back into the mantle, owing to greater fluid mobility that concentrates such elements in the near-surface. Differing recycling efficiencies for Os, Nd and Xe lead to wide differences in the preservation of Hadean isotopic signatures of these elements in the modern convecting mantle. In general, incompatible elements that are fluid mobile concentrate in surface reservoirs and are more likely to preserve Hadean geochemical signatures in the convecting mantle than compatible elements and fluid immobile incompatible elements.

Keywords: Xenoliths, peridotite, $^{187}\text{Os}/^{188}\text{Os}$, hotspot, recycling, osmium, xenon,
neodymium, helium

1. Introduction

Lavas erupted at some oceanic hotspots can host peridotite xenoliths that originated in the upper mantle and were transported to the surface in upwelling magmas. Such peridotite mantle xenoliths thus offer a means to investigate the composition of the oceanic mantle beneath hotspot volcanoes—an approach pioneered in the 1970s and 1980s in classic papers on the rare earth elements (REE) and other trace elements by Fred Frey and his coworkers (Reid and Frey, 1971; Frey and Prinz, 1978; Frey, 1980). Whereas the REE in mantle peridotites are trace element tools that are sensitive to melt depletion and melt/metasomatic enrichment, the advent of the Re-Os system has provided another tool that is especially sensitive to prior melt depletion of the mantle. The Re-Os system can be used to identify and place time constraints on a wide spectrum of ancient mantle domains in the oceanic mantle that may be related to melt extraction at the ridge, the aggregated effects of continent extraction, or subcontinental lithospheric mantle erosion that may be preserved in the mantle.

^{187}Re undergoes β^- decay to ^{187}Os ($t_{1/2}=41.6$ Ga; Shen et al., 1996; Smoliar et al., 1996), and fractionation of Re from Os during mantle melting is recorded by time-integrated variability in $^{187}\text{Os}/^{188}\text{Os}$ ratios. The Re-Os system is unlike other isotopic systems in that Re is incompatible in silicate minerals and in the absence of residual sulfide is extracted from the mantle during melting, while Os is a chalcophile element and is highly compatible and retained in the mantle residue. The Re-Os isotopic system is

characterized by extreme parent daughter fractionations: mantle melts have Re/Os ratios that are orders of magnitude higher than their mantle sources, which evolve highly radiogenic $^{187}\text{Os}/^{188}\text{Os}$ over time, while depleted mantle residues remaining after melt extraction have very low Re/Os that preserve low $^{187}\text{Os}/^{188}\text{Os}$ over time. Therefore, melt-depleted, refractory peridotites are ideally suited for preserving the time-integrated history of mantle depletion by melt extraction, and low $^{187}\text{Os}/^{188}\text{Os}$ in peridotites can be used to infer a minimum age for the mantle depletion events (e.g., Walker et al., 1989). Depleted mantle domains in oceanic environments that host unusually low $^{187}\text{Os}/^{188}\text{Os}$ have been identified in xenoliths associated with oceanic hotspots (e.g., Bizimis et al., 2007), large igneous provinces (Hassler and Shimizu, 1998; Ishikawa et al., 2011) and abyssal peridotites (e.g., Harvey et al., 2006; Liu et al., 2008; Lassiter et al., 2014). The depleted mantle domains identified in the oceanic mantle have controversial origins, and are suggested to be pieces of subcontinental lithospheric mantle (e.g. Hassler and Shimizu, 1998), ancient subducted oceanic mantle lithosphere that is embedded in upwelling mantle plumes (e.g. Bizimis et al., 2007), or ancient domains of depleted, refractory regions preserved in the convecting mantle (e.g., Liu et al., 2008; Lassiter et al., 2014).

Lavas erupted at islands related to two Polynesian hotspot localities in the south Pacific—Savai'i island at the Samoan hotspot and Tubuai island at the Macdonald (or Austral) hotspot—host peridotite xenoliths that have been the focus of numerous geochemical studies (Wright, E., 1987; Hauri, 1992; Hauri et al., 1993; Hauri and Hart, 1994; Poreda and Farley, 1992; Farley et al., 1994; Farley, 1995; Burnard et al., 1998). Hauri et al. (1993) characterized two metasomatically-enriched peridotite xenoliths each

99 from the islands of Savai'i and Tubuai (Fig. 1). Clinopyroxenes separated from these four
100 xenoliths exhibit extreme REE (particularly light REE) enrichment, but concomitant
101 enrichment in the high field strength elements (HFSE; Ti, Zr, Hf, Nb) was not observed
102 (Hauri et al., 1993). Therefore, the REE enrichment of the clinopyroxenes was interpreted
103 to result from metasomatism of the xenoliths by carbonatite melts, which impart
104 characteristic REE enrichment and HFSE depletion (e.g., Hauri et al., 1993). The origin
105 of the carbonatitic melt is thought to be related to highly-enriched, small-degree melts
106 of carbonated peridotite in the mantle plumes that generated both hotspots (Hauri et al.,
107 1993; Hauri, 1992), although carbonatitic melts may be widespread in deep mantle source
108 regions for diamonds (e.g., Walter et al., 2008). Supporting this hypothesis at the Samoa
109 and Macdonald hotspots, the Sr, Nd and Pb isotopic compositions of clinopyroxenes
110 isolated from the metasomatized xenoliths were found to reflect EM2 and HIMU
111 endmember isotopic compositions associated with lavas erupted at the two hotspots that
112 generated the islands of Savai'i (Samoa hotspot) and Tubuai (Austral hotspot),
113 respectively. However, the ultra-enriched xenoliths analyzed by Hauri et al. (1993) are
114 rare among the xenolith suites from Savai'i and Tubuai, and most xenoliths from these
115 localities do not exhibit evidence for significant REE enrichment (Hauri and Hart, 1994;
116 Hauri, 1992) in keeping with their origin as oceanic lithosphere created at the ridge. To
117 better constrain the geochemical variability of the Savai'i and Tubuai xenolith suite and
118 examine the isotopic heterogeneity of the oceanic lithospheric mantle, we have
119 characterized the Os and He isotopic compositions of highly-enriched xenoliths and a
120 subset of the xenoliths from the same localities that exhibit a range of enriched and
121 depleted REE patterns (Hauri et al., 1993; Hauri and Hart, 1994; Hauri, 1992).

Re-Os isotopic systematics reveal depleted mantle domains preserved in Savai'i and Tubuai xenoliths, which have relatively low $^{187}\text{Os}/^{188}\text{Os}$ (down to 0.1163 and 0.1173, respectively). Neither the new Os-isotopic dataset presented here from Savai'i and Tubuai xenoliths, nor the global oceanic hotspot xenolith and abyssal peridotite dataset, appear to preserve a record for mantle depletion in the Hadean and Archean (e.g., Harvey et al., 2006; Lassiter et al., 2014) because they do not show the low Os isotopic compositions ($^{187}\text{Os}/^{188}\text{Os} < 0.11$) that would represent the Earth's mantle in these times (Fig. 2). This is unexpected because of evidence for the existence of early-generated depleted mantle domains from ancient continental crust (e.g., Harrison et al., 2005, 2008; Blichert-Toft and Albarède, 2008; Kemp et al., 2010; Guitreau et al., 2012; Shirey and Hanson, 1986; Stein and Hofmann, 1994; Blichert-Toft et al. 1999; Moorbath et al. 1997; Vervoort and Blichert-Toft, 1999; Caro 2011; O'Neil et al., 2011; Khanna et al., 2014).

In stark contrast to the Os-isotopic record in oceanic peridotites, $^{129}\text{Xe}/^{130}\text{Xe}$ isotopic anomalies (generated by the decay of the short-lived nuclide ^{129}I to ^{129}Xe [$t_{1/2}=15.7$ Ma]) identified in mid-ocean ridge basalts and oceanic hotspots indicate that the convecting mantle preserves heterogeneities that were generated in the mantle prior to 4.45 Ga (Staudacher and Allegre, 1982; Allegre et al., 1983; Mukhopadhyay et al., 2012; Tucker et al., 2012; Parai et al., 2012; Petö et al., 2013). In light of the preservation of early-formed Xe-isotopic anomalies in the convecting mantle, the lack of preservation of early differentiation in the Os-isotopic record, and other isotopic systems that record early differentiation (like $^{146}\text{Sm}-^{142}\text{Nd}$, $t_{1/2}=106$ Ma), is perplexing and has implications for the mechanisms of preferential preservation of isotopic heterogeneities in the mantle. We present a conceptual model that shows how different recycling efficiencies of Xe, Os

and Nd may explain the preservation of Hadean $^{129}\text{Xe}/^{130}\text{Xe}$, but not $^{187}\text{Os}/^{188}\text{Os}$ or $^{142}\text{Nd}/^{144}\text{Nd}$, in the modern convecting mantle.

2. Methods

Twenty-one xenoliths are geochemically-characterized in this study, 13 from Savai'i (Samoan island chain) and 8 from Tubuai (Austral Island chain). The Savai'i samples were erupted in a tuffaceous volcanic cone, and the Tubuai samples were recovered from large basaltic boulders (Hauri, 1992). The xenoliths from both localities are hosted in relatively young lavas: Savai'i subaerial lavas are young (<0.42 Ma; Workman et al., 2004; McDougall, 2010; Konter and Jackson, 2012) and Tubuai subaerial lavas erupted at 5.7-10 Ma (Bellon et al., 1980). With the exception of one xenolith (a pyroxenite from Savai'i), the xenoliths examined in this study are peridotites collected by E. Hauri during the 1990 field season. The peridotites generally have low modal clinopyroxene ($<5\%$ even in REE-enriched examples; Hauri and Hart, 1994; Hauri, 1992). However, several of the Tubuai xenoliths have higher modal clinopyroxene (Hauri et al., 1993; Hauri, 1992). The single pyroxenite xenolith was collected on Savai'i by M. Jackson during the 2005 field season at the same locality as the Savai'i peridotite xenoliths. Savai'i and Tubuai peridotite xenolith sample descriptions and sample locations are described in Hauri and Hart (1993), Hauri et al. (1994) and Hauri (1993).

The xenoliths were crushed in plastic bags and the freshest representative rock chips were selected under a binocular microscope. The rock chips were sonicated in milli-Q water and then powdered by hand in an agate mortar that was cleaned by grinding with silica sand followed by rinsing in Milli-Q water. Owing to the small size of the xenoliths,

only relatively small quantities of material were available for analysis and chips were selected sparingly for making powder. Up to 5 g of powder was generated from each xenolith in the manner described, but additional xenolith powder was required for duplicate and triplicate Os-isotopic analyses. Therefore, in eight cases, additional aliquots of powder (labeled “batch 2” in Table 1) were prepared in the same manner from new chips removed from different portions of the same xenoliths from which the first batch of chips (“batch 1” in Table 1) was isolated.

Os-isotopic data for 21 different xenoliths from Savai’i and Tubuai are reported in Table 1 together with previously published data on four xenoliths from these two islands. Re and Os methods are described in Jackson and Shirey (2011). Os extraction followed a Carius tube method (Shirey and Walker, 1995) where a mixed ^{185}Re - ^{190}Os spike was added to the sample with inverse aqua regia. This was followed by solvent extraction of Os (Cohen and Waters, 1996) with carbon tetrachloride. A microdistillation method was used to purify Os (Roy-Barman and Allègre, 1995). Column exchange chromatography was used to purify Re (Morgan and Walker, 1989; Walker, 1988). Os-isotopic compositions were measured by negative thermal ionization mass spectrometry (N-TIMS) on the Thermofinnigan Triton at the Department of Terrestrial Magnetism. Osmium was loaded on Pt ribbon filaments and measured as OsO_3^- . Samples with sufficient Os were measured by static faraday multicollecion, and samples with low beam intensity were analyzed by peak hopping using a single collector secondary electron multiplier (SEM). Following oxide correction to the metal composition, all runs were corrected for mass bias using the exponential law to a $^{192}\text{Os}/^{188}\text{Os}$ ratio of 3.082614 (Creaser et al., 1991). Rhenium concentrations were measured by isotope dilution and

analyzed on either the P54 or the NU plasma MC-ICP-MS at DTM. Full chemistry procedural blanks for Re and Os were <6 pg/g and <2.5 pg/g, respectively. The fraction of the total Os and Re measured that is blank is also reported in Table 1. Repeated measurements of the Johnson Matthey (J-M) Os standard during the course of this study gave a $^{187}\text{Os}/^{188}\text{Os}$ value of 0.17396 ± 0.00040 (2σ , $n=9$) for all SEM runs and 0.174012 ± 0.000039 ($n=6$) for Faraday runs. During each analytical session on the Triton, runs of xenolith $^{187}\text{Os}/^{188}\text{Os}$ ratios were adjusted for the offset between the measured and accepted (0.17399) J-M Os values, and these ratios are reported in Table 1.

Following light crushing in an agate mortar, the freshest olivine and orthopyroxene grains were removed from a subset of the xenolith samples for helium isotopic analysis. For one sample (SA-3-3), both olivine and orthopyroxene were separated for isotopic analyses. For SAV-1-41, three different olivine separates were made: olivine was separated by color (light green versus dark green) and by inclusion content. Helium isotopic measurements were carried out at Woods Hole Oceanographic Institution following methods described in Kurz et al. (2004). Helium data were obtained by crushing olivine and orthopyroxene *in vacuo*, and the new data are presented in Table 2.

The bulk of the incompatible trace element budget of the peridotite xenoliths is hosted in clinopyroxene. Clinopyroxene trace element concentrations for all of the peridotite samples reported here were measured by ion probe at the Woods Hole Oceanographic Institution and the data are reported elsewhere (Hauri and Hart, 1994; Hauri et al., 1993; Hauri, 1992). The data for the Savai'i orthopyroxenite are reported here for the first time, and were obtained in the same laboratory using the same methods as outlined in Hauri et al. (1993). The relevant trace element data are summarized in

214 Table 1.

215

216 **3. Data and observations**

217 *3.1. Re and Os concentrations and $^{187}\text{Os}/^{188}\text{Os}$ compositions*

218 Replicate Os-isotopic analyses were made on 11 whole rock peridotites (six triplicate
219 and five duplicate analyses) (Table 1). Each replicate was performed on a new aliquot of
220 powder (either from the same batch of powder, or on a different batch of powder
221 prepared from different portions of the xenolith) that was processed independently
222 through the various steps of wet chemistry and mass spectrometry (and analyzed on a
223 different filament, in a different barrel, and on a different day on the mass spectrometer).
224 The reproducibility of the $^{187}\text{Os}/^{188}\text{Os}$ replicate measurements is better than 0.8% for
225 eight samples, but is 1.5 to 2.3% for three samples (SAV-1-41, SA-3-9, TBA-4-5). This
226 far exceeds the internal precision for all the measurements, which is better than 0.1% (2
227 σ -mean), and suggests that there is isotopic heterogeneity within these peridotites. Two of
228 the three replicate analyses that have Os-isotopic reproducibility worse than 1.5%
229 represent analyses of different aliquots of the same batch of powder, which indicates that
230 there can be significant isotopic heterogeneity within a batch of powder. The third
231 example of reproducibility worse than 1.5% is for a case where different batches of
232 powder (sampling different portions of the same xenolith) have different Os-isotopic
233 compositions. The isotopic heterogeneity within a single batch of powder, or between
234 two batches of powder sampling different parts of the same xenolith, may be due to the
235 presence of trace phases, or “nuggets”, that have distinct $^{187}\text{Os}/^{188}\text{Os}$. The different
236 nuggets might represent different generations of Os in a xenolith that contains

appreciable Os. We note that, while intra-xenolith $^{187}\text{Os}/^{188}\text{Os}$ variability is up to 2% (sample TBA-4-5) in the Tubuai xenoliths and up to 2.3% (sample SAV-1-41) in the Savai'i xenoliths, the total inter-xenolith $^{187}\text{Os}/^{188}\text{Os}$ variability among different Tubuai (10.8%) and Savai'i (8.6%) xenoliths exceeds the variability observed within any single xenolith by a factor of 3.7 to 5.4, at Tubuai and Savai'i, respectively.

The concentration of Os also varies within a single powder or between different batches of powder prepared from different portions of the xenolith. Five of the eleven xenoliths with replicate measurements have Os concentrations that are reproducible within 6 to 29%. The remaining six xenoliths have Os concentrations that vary by a factor of 2.3 to 7.1, which includes comparison of analyses for different aliquots of the same powder and different powders prepared from the same xenolith. This variability can be explained by the heterogeneous distribution of Os-rich, trace phases (nuggets), both within a powder and within different regions of the same xenolith. The observation that peridotite xenoliths host isotopically-heterogeneous nuggets is well established in the literature (Shirey and Walker, 1995; Shirey and Walker, 1998; Reisberg and Meisel, 2002; Lorand et al., 2008). We note that there is no relationship between variability in Os concentrations and heterogeneity in $^{187}\text{Os}/^{188}\text{Os}$ within a xenolith: xenoliths with highly variable Os concentrations can have heterogeneous $^{187}\text{Os}/^{188}\text{Os}$ (TBA-4-5, SAV-1-42, SAV-1-30) or relatively homogenous $^{187}\text{Os}/^{188}\text{Os}$ (SAV-1-28, SA-3-3, SAV-1-46). Similarly, xenoliths with relatively homogeneous Os concentrations can have heterogeneous $^{187}\text{Os}/^{188}\text{Os}$ (SAV-1-41) or relatively homogenous $^{187}\text{Os}/^{188}\text{Os}$ (SAV-1-13).

The Savai'i xenoliths generally have lower Re than the Tubuai xenoliths: Re

concentrations vary from just below 0.001 ppb to 0.035 ppb in the Savai'i xenoliths, and from 0.030 ppb to 0.354 ppb in the Tubuai xenoliths. Replicate measurements for Re were made on four whole rock peridotites; like Os, each replicate was performed on a new aliquot of powder processed independently through the various steps of wet chemistry and mass spectrometry. Of these four xenoliths, the sample with the highest Re concentration (TBA-4-5) is also the most reproducible (the highest and lowest Re concentrations differ by only 1.1%), as is typical with the nugget effect. The xenolith with the lowest Re concentration (SAV-1-41) has the least reproducible result (the highest and lowest Re concentrations differ by over a factor of three).

There is no relationship between $^{187}\text{Os}/^{188}\text{Os}$ and Os concentrations (Fig. 2) and Re concentrations (not shown) in the new xenolith dataset. The average $^{187}\text{Os}/^{188}\text{Os}$ of the Tubuai xenolith suite (0.1262 ± 0.0035 , 1σ ; calculated by averaging the weighted average of replicate $^{187}\text{Os}/^{188}\text{Os}$ analyses for each xenolith) is higher than the average $^{187}\text{Os}/^{188}\text{Os}$ of the Savai'i xenoliths (0.1237 ± 0.0032), but the two xenolith populations overlap at the 1σ level. While the Os concentrations of the Tubuai and Savai'i xenoliths overlap, the Tubuai xenoliths extend to higher Os concentrations than the Savai'i xenoliths (Fig. 2). Two peridotite xenoliths from Samoa have remarkably low Os concentrations: SAV-1-47 has reproducibly low Os (0.043 and 0.053 ppb) and SAV-3-2 has the lowest Os concentration in the xenolith suite (0.025 ppb). Such low Os concentrations are more typically found in pyroxenites, including the Samoan pyroxenite xenolith sample SAV-05-29, which has 0.036 ppb Os.

The variability in $^{187}\text{Os}/^{188}\text{Os}$ and Os in the Tubuai and Samoa xenoliths lies approximately within the range defined by previously published data from abyssal

peridotites and peridotite xenoliths from oceanic hotspots. Excluding a single Tubuai xenolith with $^{187}\text{Os}/^{188}\text{Os}$ of 0.1304 (TBA-1-7) (not shown in Fig. 2 because Os concentrations are not available for this xenolith; see Hauri, 1992), xenoliths from Savai'i and Tubuai have $^{187}\text{Os}/^{188}\text{Os}$ lower than the primitive upper mantle (PUM). This contrasts with a subset of abyssal peridotites and peridotite xenoliths from Kerguelen ($^{187}\text{Os}/^{188}\text{Os}$ up to 0.1383) and Ontong Java Plateau (OJP) ($^{187}\text{Os}/^{188}\text{Os}$ up to 0.1404) with relatively radiogenic $^{187}\text{Os}/^{188}\text{Os}$ ratios that exceed the PUM value. The extent to which these enriched Os isotopic compositions are due to melt re-enrichment is not clear, but such a process apparently does not affect the Savai'i and Tubuai xenoliths studied here.

Peridotites with low Os concentrations (<0.1 ppb) are relatively uncommon. Xenoliths from Kerguelen (down to 0.044 ppb) and OJP (down to 0.068 ppb) can have low Os concentrations and intermediate $^{187}\text{Os}/^{188}\text{Os}$ that approach the compositions observed in the peridotite xenoliths with the lowest Os from Savai'i (down to 0.025 ppb). Oceanic peridotites with $^{187}\text{Os}/^{188}\text{Os}$ as low as (or lower than) the lowest values found in Samoa and Tubuai are also relatively uncommon, but are found at Hawaii ($^{187}\text{Os}/^{188}\text{Os}$ down to 0.1138; Bizimis et al., 2007), OJP (down to 0.1152; Ishikawa et al., 2011), and a subset of abyssal peridotites (down to 0.1139; Liu et al., 2008).

Fig. 2 shows that the $^{187}\text{Os}/^{188}\text{Os}$ of the Savai'i xenoliths overlap the isotopic compositions of lavas from the island of Savai'i, but also extend to lower $^{187}\text{Os}/^{188}\text{Os}$ than the lavas. The $^{187}\text{Os}/^{188}\text{Os}$ ratios measured in Tubuai xenoliths are lower than (and show no overlap with) basalts from Tubuai.

A subset of the Savai'i and Tubuai xenoliths exhibit different degrees of enrichment by a carbonatitic melt. Hauri et al. (1993) argued that clinopyroxene from a subset of

Savai'i and Tubuai peridotite xenoliths grew in the presence of a carbonatitic melt. The metasomatic clinopyroxenes have anomalously high REE concentrations, particularly the light REE's (LREE, with La/Yb ratios of up to 185). In contrast, Ti and Zr show strong depletions relative to the REEs. Like the LREE enrichment in clinopyroxene, Zr depletion relative to Nd (low Zr/Nd ratio; Fig. 3) in clinopyroxene suggests carbonatite metasomatism. Neither indicator of carbonatite metasomatism, low Zr/Nd or high La/Yb in clinopyroxene, correlates with $^{187}\text{Os}/^{188}\text{Os}$. For example, the Savai'i xenolith SAV-1-1 with the strongest carbonatite metasomatic signature (highest La/Yb and second lowest Zr/Nd) does have the most radiogenic $^{187}\text{Os}/^{188}\text{Os}$, but the Savai'i xenoliths with among the (SAV-1-42 and SAV-1-46) strongest carbonatite signatures have among the lowest $^{187}\text{Os}/^{188}\text{Os}$ in the Savai'i suite. The Tubuai xenolith with the highest $^{187}\text{Os}/^{188}\text{Os}$ (TBA-1-7) appears to be one of the least affected by carbonatite (it has the second lowest La/Yb and the second highest Zr/Nd), and Tubuai xenoliths that have with the strongest trace element signatures for carbonatite metasomatism have both low (TBA-4-5) and relatively high (TBA-1-9, TBA-4-11 and TBA-5-4) $^{187}\text{Os}/^{188}\text{Os}$. In summary, there is no simple relationship between trace element indicators of metasomatic enrichment and $^{187}\text{Os}/^{188}\text{Os}$ in the xenoliths, suggesting that the Os isotopic compositions measured are reliable long-term indicators of their mantle source characteristics.

3.2. Helium concentrations and isotopic data

Helium isotopes were measured in a subset of the Savai'i and Tubuai xenoliths. $^3\text{He}/^4\text{He}$ in the Savai'i xenoliths varies from 10.8 to 12.4 times atmosphere (Ra) (Figs. 4 and 5), lower than values of 12.7 and 13.7 Ra reported for Savai'i rejuvenated lavas

(Workman et al., 2004; Jackson et al., 2014). The rejuvenated lava measurements are relevant here because only rejuvenated lavas host peridotite mantle xenoliths in Samoa. These values are higher than the global upper mantle $^3\text{He}/^4\text{He}$ ratio as measured in MORB (approximately 7 to 8 Ra; Graham, 2002; Georgen et al., 2003). The new peridotite xenolith $^3\text{He}/^4\text{He}$ measurements from Savai'i are similar those of Savai'i peridotite xenoliths by crushing, which varied from 11.3 to 12.0 Ra (Farley, 1995) and from 10.2 to 11.9 Ra (Poreda and Farley, 1992). Poreda and Farley (1992) did report lower $^3\text{He}/^4\text{He}$ values for individual high temperature heating steps, but if the $^3\text{He}/^4\text{He}$ values for heating and fusion of the same sample are combined, the $^3\text{He}/^4\text{He}$ falls within the range of $^3\text{He}/^4\text{He}$ reported here for the crushing experiments. We find a weak positive relationship between $^3\text{He}/^4\text{He}$ and helium concentrations in the Savai'i xenoliths, where ^4He concentrations vary from 3.6×10^{-8} to 3.1×10^{-6} cc STP/g (and the highest concentrations are found in the Savai'i pyroxenite); the range in concentrations spans nearly two orders of magnitude, and likely reflects the variable (but generally high) density of fluid inclusions present in the xenoliths. Three visually-distinct (cloudy, clear and dark green olivine) populations of olivine isolated from SAV-1-41 exhibit factor of three variability in helium concentrations (Table 2). In Savai'i xenoliths, helium isotopic ratios exhibit no relationship with $^{187}\text{Os}/^{188}\text{Os}$ or with trace element indicators of carbonatite metasomatism (Fig. 5).

Helium isotopic values in the Tubuai xenoliths are consistently lower, and show a wider range, than the Samoan xenoliths (Figs. 4 and 5), and are lower than the upper mantle average $^3\text{He}/^4\text{He}$ ratio measured in MORB. $^3\text{He}/^4\text{He}$ in the Tubuai xenoliths (2.5 to 6.4 Ra) are also lower values than observed in Tubuai lavas: $^3\text{He}/^4\text{He}$ ratios measured

in Tubuai lavas range from 6.4 to 7.2 Ra (Graham et al., 1992; Parai et al., 2009; Hanyu et al., 2011b). However, the ^4He concentrations in the Tubuai xenoliths are all lower than the Savai'i xenoliths, and range from 1.4×10^{-9} to 2.4×10^{-8} cc STP/gram. While there is no relationship between $^3\text{He}/^4\text{He}$ and ^4He concentrations in the Tubuai xenoliths, the xenolith with the lowest ^4He concentration has the lowest $^3\text{He}/^4\text{He}$, which might be explained by radiogenic ingrowth of ^4He . We do not have Th and U abundances in these samples to test this hypothesis, but peridotites typically have very low Th and U concentrations (e.g., Recanati et al., 2012).

Figure 5 shows that there is no relationship between $^3\text{He}/^4\text{He}$ and $^{187}\text{Os}/^{188}\text{Os}$ in the dataset. The samples with paired $^3\text{He}/^4\text{He}$ and $^{187}\text{Os}/^{188}\text{Os}$ from Savai'i and Tubuai span a similar range of $^{187}\text{Os}/^{188}\text{Os}$, but the Tubuai xenoliths all have lower $^3\text{He}/^4\text{He}$.

Additionally, there is no relationship between $^3\text{He}/^4\text{He}$ and trace element indicators of carbonatite metasomatism in the Tubuai xenolith suite (Fig. 5). Furthermore, the modal abundances of the phases in the Savai'i and Tubuai xenoliths (Table 3) do not exhibit correlations with $^{187}\text{Os}/^{188}\text{Os}$ or $^3\text{He}/^4\text{He}$ (not shown).

4. Discussion

4.1. Variable Os concentrations in the Tubuai and Savai'i xenoliths:

A puzzling feature of the Savai'i xenolith suite is the observation that two peridotite xenoliths have unusually low Os concentrations ($\text{Os} \leq 0.053$ ppt). Peridotite xenoliths hosted in oceanic hotspot lavas tend to have lower Os concentrations than abyssal peridotites (Ishikawa et al., 2011), but the very low Os concentrations in two

375 Savai'i xenoliths are unusual.

376 Relatively low Os concentrations (down to 0.068 ppt) in OJP peridotite xenoliths
377 were identified by Ishikawa et al. (2011). Ishikawa et al. (2011) suggested the low Os
378 concentrations may result from the breakdown of sulfide, which is the primary host of Os
379 in peridotites (Alard et al., 2000; Burton et al., 1999; Hart and Ravizza, 1996). The
380 Savai'i xenoliths are relatively fresh, so breakdown of sulfide by surface weathering
381 processes is unlikely. The breakdown of sulfide by devolatilization during eruption and
382 exposure to oxygenated conditions (Handler et al., 1999) was also considered by
383 Ishikawa et al. (2011). We consider such an explanation unlikely in the Samoan suite
384 because the mechanism of sulfide breakdown should affect all the xenoliths, but only two
385 Savai'i xenoliths exhibit unusually low Os concentrations. One possible hypothesis is that
386 sulfides were exhausted by extensive melt extraction from the peridotites, leaving the Os
387 highly depleted in these samples. However, Os concentrations do not exhibit clear
388 correlations with indicators of melt extraction (e.g., modal abundance of spinel), and the
389 xenoliths with unusually low Os concentrations do not appear to be more refractory than
390 the other Savai'i xenoliths. Another possible explanation for the low Os concentrations is
391 percolation of sulfur-undersaturated melts through the peridotites, which would mobilize
392 sulfide and reduce the Os concentrations (Reisberg et al., 2005). However, there is no
393 relationship between Os concentrations and indicators of metasomatic enrichment (e.g.,
394 La/Yb in xenolith clinopyroxene), which argues somewhat against this hypothesis.

395 Another possibility is that the xenoliths' low Os content is an expression of the nugget
396 effect on a large scale. In other words, the heterogeneous distribution of Os carrying
397 phases is on a scale much larger than the small portion of xenolith provided by the alkali

basalt host; the Os-rich nuggets were not sampled in xenoliths with low Os concentrations, even though the nuggets were possibly present in different portions of the xenoliths analyzed. In summary, while there are several possible models for the origin of the low Os concentration in the xenoliths, there is no single mechanism that clearly explains the low Os concentrations identified in a subset of the Savai'i xenoliths.

4.2. Relationship between $^{187}\text{Os}/^{188}\text{Os}$ in xenoliths and whole rock lavas from Samoa and Tubuai.

Hauri et al. (1993) and Hauri (1992) report Sr, Nd and Pb isotopic compositions on clinopyroxene separates from the most incompatible trace element enriched xenoliths from Savai'i and Tubuai: SAV-1-1, SAV-1-28, TBA-4-11, and TBA-1-9 (the Os-isotopic data for these xenoliths are shown in Table 1). The Sr, Nd and Pb isotopic compositions of the two Savai'i xenoliths ($^{87}\text{Sr}/^{86}\text{Sr}$ is 0.710969 to 0.712838; $^{143}\text{Nd}/^{144}\text{Nd}$ is 0.512467 to 0.512516; $^{206}\text{Pb}/^{204}\text{Pb}$ is 18.845 to 18.877) are more geochemically enriched than rejuvenated lavas from Savai'i (Wright and White, 1986; Hauri and Hart, 1993; Workman et al., 2004), but they overlap with shield stage lavas from the island (Jackson et al., 2007). The Sr, Nd and Pb isotopic compositions of the Tubuai xenoliths ($^{87}\text{Sr}/^{86}\text{Sr}$ is 0.702796 to 0.702755; $^{143}\text{Nd}/^{144}\text{Nd}$ is 0.512902 to 0.512942; $^{206}\text{Pb}/^{204}\text{Pb}$ is 20.971 to 21.313) overlap with basalts from the volcano (Vidal et al., 1984; Chauvel et al., 1992; Kogiso et al. 1997; Schiano et al., 2001; Hanyu et al. 2011a, Hanyu et al., 2013). Unfortunately, Sr, Nd and Pb isotopic data were not reported for the xenoliths from Savai'i and Tubuai that exhibit less incompatible trace element enrichment, and therefore it is unknown whether these xenoliths have isotopic compositions similar to

basalts from the respective islands.

In this study, we report $^{187}\text{Os}/^{188}\text{Os}$ in xenoliths that exhibit incompatible trace element enrichment and xenoliths that are incompatible trace element depleted. The $^{187}\text{Os}/^{188}\text{Os}$ of the Tubuai xenoliths (<0.1304) is much lower than the $^{187}\text{Os}/^{188}\text{Os}$ measured in Tubuai basalts (>0.1433) (Fig. 2). Hauri and Hart (1993) noted that the $^{187}\text{Os}/^{188}\text{Os}$ of Tubuai xenoliths is similar to the range expected for oceanic lithosphere, while the $^{187}\text{Os}/^{188}\text{Os}$ of Tubuai basalts is elevated and similar to basalts from other HIMU localities (e.g., Hauri et al., 1993; Day et al., 2009; Hanyu et al., 2011a). Hauri et al. (1993) argued that, while the isotopic composition of the incompatible elements of the two most incompatible element (Sr, Nd, Pb) enriched Tubuai xenoliths reflects the composition of the metasomatic fluid, the compatible element budget (Os) of these peridotite were not significantly changed by the metasomatic fluid and preserves a lithospheric origin. The unradiogenic Os in the Tubuai xenoliths indicates that the xenoliths are not the source of the basaltic volcanism at Tubuai. Instead, while CO_2 -rich melts associated with the plume did overprint the Sr, Nd and Pb isotopic composition of the peridotite xenoliths that were entrained in the upwelling melts of from the Tubuai mantle source (Hauri et al., 1993), they did not affect the Os isotopic composition of the xenoliths.

Unlike the case of the Tubuai xenoliths and basalts, the range of $^{187}\text{Os}/^{188}\text{Os}$ in lavas from Savai'i (>0.1233 ; Hauri and Hart, 1993; Workman et al., 2004) overlaps with the $^{187}\text{Os}/^{188}\text{Os}$ xenoliths from this locality (0.1173 to 0.1284) (Fig. 2). The Savai'i xenoliths are hosted in rejuvenated lavas, and rejuvenated lavas from Savai'i have the lowest $^{187}\text{Os}/^{188}\text{Os}$ (down to 0.1233) among Samoan hotspot lavas (for comparison, the lowest $^{187}\text{Os}/^{188}\text{Os}$ in Samoan shield lavas is 0.1268; Jackson and Shirey, 2011). Indeed, Savai'i

rejuvenated lavas exhibit some of the lowest $^{187}\text{Os}/^{188}\text{Os}$ in the global OIB (ocean island basalt) database. However, xenoliths are abundant in rejuvenated lavas from Savai'i, and Jackson and Shirey (2011) suggested that xenocrystic olivines from disaggregated xenoliths (with low $^{187}\text{Os}/^{188}\text{Os}$) might contribute to the low $^{187}\text{Os}/^{188}\text{Os}$ in rejuvenated Savai'i lavas. While unradiogenic Os in oceanic hotspots is not always a result of the incorporation of xenolith material with low $^{187}\text{Os}/^{188}\text{Os}$ (Schaefer et al., 2002), this mechanism has been suggested to reduce the $^{187}\text{Os}/^{188}\text{Os}$ of the host lava at the Azores, Comoros and Canary hotspots (Widom and Shirey, 1996; Class et al., 2009; Day et al., 2010; Widom et al., 1999). We note that half of the 14 peridotite xenoliths in the Savai'i dataset have $^{187}\text{Os}/^{188}\text{Os}$ ratios lower than the lowest ratio identified in lavas from the Samoan hotspot. Therefore, even if the low $^{187}\text{Os}/^{188}\text{Os}$ of the Samoan rejuvenated lavas reflects the mantle source of rejuvenated lavas, and is not just the result of incorporation of disaggregated xenoliths, there are still 7 peridotite xenoliths with $^{187}\text{Os}/^{188}\text{Os}$ lower than Samoan lavas and are derived from a mantle domain that is unlike that sampled by any Samoan lavas. We note that the other seven peridotite xenoliths with $^{187}\text{Os}/^{188}\text{Os}$ overlapping with the range observed in Samoan lavas may also sample an Os-isotopic domain distinct from that sampled by the lavas, because the $^{187}\text{Os}/^{188}\text{Os}$ of the Savai'i rejuvenated lavas may have been reduced by incorporation of xenolith material with highly unradiogenic $^{187}\text{Os}/^{188}\text{Os}$.

4.3. Melt-rock reaction in the Tubuai and Savai'i xenoliths

Extensive carbonatite metasomatism related to the Samoa hotspot (which sources Savai'i volcanism) and Macdonald hotspot (thought to source Tubuai volcanism; Chauvel

et al., 1997) is clear in a subset of the Savai'i and Tubuai xenoliths, respectively (Hauri et al., 1993). However there is no simple relationship between indicators of metasomatism (trace element ratios measured in xenolith clinopyroxenes) and $^{187}\text{Os}/^{188}\text{Os}$ of the bulk xenolith. If the Os-isotopic compositions of the xenoliths were modified by the metasomatic fluid, we would expect the most metasomatically-enriched xenoliths (with the highest La/Yb and lowest Zr/Nd) to exhibit a relationship with whole rock $^{187}\text{Os}/^{188}\text{Os}$, but none is observed. Instead, xenoliths with the weakest metasomatic signatures exhibit both high and low $^{187}\text{Os}/^{188}\text{Os}$ (Fig. 3), suggesting that preexisting Os-isotopic heterogeneity dominates the Os-isotopic variability of the Savai'i and Tubuai xenolith suites. It is likely that the metasomatic fluid did not have sufficiently high Os concentrations to strongly modify the Os-isotopic compositions of the peridotite xenoliths, a hypothesis that is supported by measurements of low Os concentrations in oceanic carbonatites from the Canary islands (0.0047 to 0.0145 ppb; Widom et al., 1999). In summary, while the incompatible trace elements and the radiogenic isotopes of incompatible elements (He, Nd, Sr and Pb) show extensive evidence that these xenoliths interacted with their respective hotspot magmas, the data do not support recent metasomatic modification of the $^{187}\text{Os}/^{188}\text{Os}$ of the Savai'i and Tubuai xenoliths. Instead, we argue that the Os-isotopic compositions of the peridotite xenoliths reflect long-term mantle heterogeneity generated by ancient melt extraction and are not significantly modified by host melts (e.g., Walker et al., 1989; Shirey and Walker, 1998).

Helium isotopic compositions do not correlate with Os-isotopic compositions. Helium is a strongly incompatible element in the mantle (Heber et al., 2007; Jackson et al., 2013). Therefore, the lack of correlation between $^3\text{He}/^4\text{He}$ and $^{187}\text{Os}/^{188}\text{Os}$ is not

surprising, as other incompatible trace elements fail to show correlations with $^{187}\text{Os}/^{188}\text{Os}$ (Fig. 3). It is noteworthy that helium isotopic compositions do not correlate with incompatible trace element indicators of metasomatism (Fig. 5). Oceanic carbonatites can be enriched in helium (e.g., Mata et al., 2010), and helium might be expected to impart a clear metasomatic signature on the metasomatized peridotite in the same manner as the other incompatible trace elements. Indeed, the distinct He isotope ratios (Table 2) and Sr-Nd-Pb isotopic ratios (Hauri et al., 1993) of the Savai'i and Tubuai xenoliths correspond closely with those of their corresponding hotspot lavas. However, within the population of xenoliths from each island, there is no simple relationship between $^3\text{He}/^4\text{He}$ and trace element signatures, and the Os isotope signatures of the xenoliths do not generally match those of the corresponding Samoan and Tubuai lavas.

In the xenoliths from Savai'i and Tubuai, magma-mantle reaction is reflected in the precipitation of metasomatic clinopyroxene. The extensive overprinting of the incompatible elements and their isotopes (He, Sr, Nd, Pb) in a peridotite, without correlated shifts in compatible elements and their isotopes (Os), can be readily understood by considering how the budgets of incompatible and compatible elements change in a magma that undergoes melt-rock reaction with peridotite. This process was modeled by Hauri (1997), who parameterized magma-mantle interaction in terms of the solid/melt partition coefficient (K) and the Damköhler number (Da), which is a dimensionless number that is a measure of the extent of magma-mantle exchange. The Damköhler number represents the ratio of two timescales, the timescale for melt-solid exchange (t_E) versus the timescale of magma transport (t_c).

From a simple mixing standpoint, it is relatively easy to overwhelm the

compatible element budget of a magma because of the much larger concentrations of compatible elements in peridotite; the converse is true for incompatible elements, the magma overwhelms the peridotite. This simple situation becomes complicated when the “mixing” occurs during magma transport through peridotite. When the Damköhler number is low (analogous to a high melt/rock ratio), melt-rock reaction cannot keep up with the high rate of magma ascent, and magmas can be transported through the mantle at full disequilibrium while retaining even the compatible element (e.g., Os) isotopic compositions of their sources (Hauri et al., 1997). A low Damköhler number is analogous to a high melt/rock ratio. When the Damköhler number is high, magma ascent is slow enough to allow extensive equilibration between the ascending magma and surrounding peridotite, to the point of complete buffering of the melt in equilibrium with the mantle. A high Damköhler number is analogous to a low melt/rock ratio. Whether the elements in the magma approach equilibrium with the surrounding mantle (or not) depend on the partition coefficient and the Damköhler number.

Hauri (1997) demonstrated that the wide variety of highly-fractionated REE patterns and widely-varying REE/HFSE ratios in Samoan xenoliths were consistent with migration of Samoan hotspot melts into the oceanic lithosphere, with the initial melt flux dominated by carbonatite-like melts that gave way over time to a mixture of carbonatitic and basaltic trace element compositions. The best-fit Damköhler numbers estimated from the trace elements in xenolith clinopyroxenes ($Da = 1412$ decreasing to 158 with time) represent values that are high enough to fractionate the REE from each other during magma-mantle reaction. However, these Damköhler numbers are all high enough (by several orders of magnitude) that the compatible element budget of the magma is entirely

buffered by the surrounding mantle. As a result, such a magma has essentially no capacity to alter significantly the isotopic composition of compatible elements like osmium, while still retaining much of its original budget of highly incompatible elements derived from its mantle source. This kind of melt-rock interaction explains the overprinting of Sr, Nd, Pb and He isotopes in the Savai'i and Tubuai peridotite xenoliths by their respective hotspot magmas, while largely retaining the Os isotope composition of the oceanic lithosphere that was formed at a mid-ocean ridge.

Another possible explanation for the lack of relationship between $^3\text{He}/^4\text{He}$ and incompatible element indicators of metasomatism is that several different pulses of metasomatism may have overprinted the Savai'i xenolith suite (Burnard et al., 1998), possibly as the Pacific lithosphere currently under Savai'i passed over several different hotspots with different compositions (Konter and Jackson, 2012; Jackson et al., 2010). Multicomponent mixing caused by multiple pulses of metasomatic fluids with different $^3\text{He}/^4\text{He}$ ratios and variable trace element compositions may generate complicated geochemical patterns that obscure relationships among incompatible elements. This simple model may also help explain the lack of relationships between $^3\text{He}/^4\text{He}$ and $^{187}\text{Os}/^{188}\text{Os}$ and trace elements in the Tubuai xenoliths. It is not clear whether the non-atmospheric $^{129}\text{Xe}/^{130}\text{Xe}$ reported in a subset of Savai'i xenoliths analyzed by Poreda and Farley (1992) relates to metasomatic enrichment, as trace element and isotopic (i.e., Sr, Nd, Pb) indicators of carbonatite metasomatism were not analyzed in this suite of xenoliths.

In summary, the unradiogenic Os isotopic signatures of all of the Tubuai and half of the Savai'i xenoliths are lower than the lowest values identified in lavas with mantle-

derived Os-isotopic signatures at their respective islands, and this finding echoes similar observations made at the Hawaiian hotspot and the OJP. Bizimis et al. (2007) identified highly unradiogenic $^{187}\text{Os}/^{188}\text{Os}$ in peridotite xenoliths from Hawaii (down to 0.1138) that are lower than the lowest $^{187}\text{Os}/^{188}\text{Os}$ ratios identified in lavas from the hotspot. Similarly, the low $^{187}\text{Os}/^{188}\text{Os}$ found in peridotite xenoliths from the OJP (down to 0.1152; Ishikawa et al., 2011) extend to lower values than identified in lavas from the Plateau (Tejada et al., 2013). The unradiogenic $^{187}\text{Os}/^{188}\text{Os}$ identified in xenoliths, but not in lavas, from these Pacific hotspots and the OJP shows that they are not mantle residua to the lavas and thus raises a fundamental question: What is the origin of the low $^{187}\text{Os}/^{188}\text{Os}$ mantle domain sampled by the xenoliths?

4.4. Origin of the low $^{187}\text{Os}/^{188}\text{Os}$ signature in Tubuai and Savai'i xenoliths

4.4.1. Subcontinental lithospheric mantle as the origin of the low $^{187}\text{Os}/^{188}\text{Os}$ in Samoa and Tubuai xenoliths?

The relatively unradiogenic $^{187}\text{Os}/^{188}\text{Os}$ ratios reported in Savai'i (down to 0.1173) and Tubuai (down to 0.1163) xenoliths requires formation by ancient melt extraction followed by long-term preservation, but the provenance of the ancient depleted reservoir contributing to these hotspot xenoliths—SCLM or ancient depleted domains in the oceanic mantle—is uncertain. The subcontinental lithospheric mantle (SCLM) is known to host unradiogenic $^{187}\text{Os}/^{188}\text{Os}$, which is the result of ancient melt extraction from the mantle followed by long-term preservation of the melt-depleted reservoir

beneath the continents (e.g., Walker et al., 1989; Carlson and Irving, 1994; Carlson et al., 2005; Pearson and Wittig, 2008). Peridotite xenoliths from Kerguelen with relatively low $^{187}\text{Os}/^{188}\text{Os}$ (down to 0.1182) were suggested to sample slivers of SCLM incorporated into the Indian Ocean lithosphere following rifting of Gondwanaland (Hassler and Shimizu, 1998). Indeed, continental lithosphere is known to underlie portions of the Kerguelen plateau (Frey et al., 2002; Ingle et al., 2002), which makes the interpretation of the low $^{187}\text{Os}/^{188}\text{Os}$ Os compelling at Kerguelen and perhaps suggestive at other oceanic localities (e.g., Coltorti et al., 2010; Debaille et al., 2009). However, unlike Kerguelen, the Savai'i and Tubuai Islands formed in an oceanic setting far from continental margins and are not located at the locus of past continental rifting (e.g., Taylor, 2006). There is no obvious tectonic mechanism that can raft fragments of SCLM to regions of the Pacific plate now occupied by Savai'i and Tubuai, and it is unlikely that fragments of SCLM reside in this region of the south Pacific.

In addition to lacking a clear tectonic mechanism for transporting SCLM fragments to hotspot localities, the SCLM mantle is relatively rare. Bizimis et al. (2007) argued that the relative scarcity of SCLM mantle—only 2.5% of the mantle's mass—makes it far less likely to contribute to the unradiogenic $^{187}\text{Os}/^{188}\text{Os}$ observed in peridotite xenoliths from the Hawaiian hotspot than other depleted mantle reservoirs. Depleted mantle domains in the convecting mantle that may host low $^{187}\text{Os}/^{188}\text{Os}$ constitute a much larger fraction of the mantle and are therefore more likely to contribute to the peridotite xenoliths at Savai'i and Tubuai (see sections 4.4.2 and 4.4.3 below). These depleted reservoirs include the ambient upper mantle (as sampled by abyssal peridotites) and ancient subducted oceanic mantle lithosphere, where the latter may

comprise 80% (or more) of the mass of the mantle (Salters and Stracke, 2004). How these reservoirs might contribute to the unradiogenic $^{187}\text{Os}/^{188}\text{Os}$ observed in Savai'i and Tubuai xenoliths becomes the essential consideration.

4.4.2. Low $^{187}\text{Os}/^{188}\text{Os}$ in hotspot xenoliths: Ancient subducted oceanic lithosphere entrained in mantle plumes?

Peridotite xenoliths with exceptionally low $^{187}\text{Os}/^{188}\text{Os}$ (down to 0.1138) hosted in Hawaiian lavas were suggested to be derived directly from ancient subducted, depleted mantle lithosphere that is embedded in the upwelling Hawaiian mantle plume (Bizimis et al., 2007). The Hawaiian xenoliths were not considered to be peridotite fragments from the modern Pacific oceanic lithosphere because, at the time of the study, abyssal peridotites were not known to host $^{187}\text{Os}/^{188}\text{Os}$ as low as identified in the Hawaiian xenolith suite (down to 0.1138). Therefore, a different (deeper?) reservoir composed of ancient subducted oceanic lithosphere was suggested as the reservoir supplying material with lower $^{187}\text{Os}/^{188}\text{Os}$ to the Hawaiian xenolith suite. In this model, subducted oceanic mantle lithosphere, now residing in the mantle, experienced melt extraction and Re-depletion at an ancient mid-ocean ridge, and domains of mantle comprised of this material may preserve low $^{187}\text{Os}/^{188}\text{Os}$. Upwelling mantle plumes sourcing oceanic hotspot volcanism may entrain a subducted reservoir comprised of ancient subducted oceanic mantle lithosphere which itself becomes embedded in the plume (Bizimis et al., 2007) and attaches itself to the base of the preexisting oceanic lithosphere (Ishikawa et al., 2011). In this model, the subset of peridotite mantle xenoliths at hotspots that have very low $^{187}\text{Os}/^{188}\text{Os}$ are simply portions of the refractory substrate entrained in

upwelling hotspot lavas at attached to the base of the oceanic lithosphere (Bizimis et al, 2007; Ishikawa et al., 2011).

However, the discovery of low $^{187}\text{Os}/^{188}\text{Os}$ in abyssal peridotites (down to 0.1139; Liu et al. 2008), within error of the lowest values from the Hawaiian xenolith suite, makes a xenolith origin in the modern Pacific mantle lithosphere feasible. Indeed, like Hawaiian xenoliths, the $^{187}\text{Os}/^{188}\text{Os}$ ratios measured in Savai'i and Tubuai xenoliths are also well-bracketed by the range of $^{187}\text{Os}/^{188}\text{Os}$ found in the oceanic mantle lithosphere (Fig. 2), and the peridotite xenoliths from all three hotspots may originate in the oceanic lithosphere reservoir as sampled by abyssal peridotites.

Ishikawa et al. (2011) also argued that a subset of peridotite xenoliths recovered from the island of Malaita are residues from the OJP mantle plume. This subset of peridotite xenoliths with the highest mantle temperatures (and inferred greatest depths) sample fragments of the previously recycled ancient oceanic mantle lithosphere that makes up part of the OJP mantle plume. These deeper xenoliths are fragments of the plume that underplated and formed a refractory substrate attached to the base of the Pacific mantle lithosphere at ~120 Ma when the OJP erupted. Ishikawa et al. (2007) maintained that a second population of lower temperature (shallower) peridotite mantle xenoliths from the same locality are fragments of the 160 Ma Pacific lithosphere upon which the OJP was constructed, and these lower temperature xenoliths are not fragments of the underplated OJP mantle plume. The shallow and deep xenoliths were argued to exhibit different $^{187}\text{Os}/^{188}\text{Os}$ distributions, thereby supporting a distinct plume origin for the deeper xenoliths. However, there is substantial overlap in the Os-isotopic composition of the shallow and deep OJP xenolith populations: the deep xenolith population (n=12

xenoliths) spans a range of $^{187}\text{Os}/^{188}\text{Os}$ (0.1152 to 0.1270) that overlaps considerably with the shallow xenolith population (0.1161 to 0.1388; $n=60$). Additionally, the different Os-isotopic distributions between the deep and shallow xenoliths may be related to undersampling the deeper population, which consists of just 12 samples (which may not provide a representative sampling of the Os-isotopic distribution of the deeper xenoliths). It is thus not clear whether the Os-isotopic dataset of Ishikawa et al. (2007) fully supports a model where the deep OJP lithosphere has distinct $^{187}\text{Os}/^{188}\text{Os}$ from the shallow lithosphere, and if this is true, then the deep and shallow OJP xenoliths may simply sample typical oceanic mantle lithosphere. Indeed, abyssal peridotite $^{187}\text{Os}/^{188}\text{Os}$ can extend to values lower than the lowest identified in the OJP (Liu et al., 2008).

In summary, we cannot exclude the possibility that peridotite xenoliths from Hawaii and the OJP represent pieces of ancient subducted, depleted oceanic mantle lithosphere embedded in the respective mantle plumes that underplated the Pacific lithosphere. However, echoing Lassiter et al. (2014), we note that xenoliths from the hotspots with the lowest $^{187}\text{Os}/^{188}\text{Os}$ --Hawaii, OJP, Tubuai and Savai'i--all exhibit $^{187}\text{Os}/^{188}\text{Os}$ that falls within the range (or within error) of $^{187}\text{Os}/^{188}\text{Os}$ ratios measured in abyssal peridotites (Fig. 2).

4.4.3. Peridotite xenoliths at hotspots from the same heterogeneous oceanic mantle lithosphere as abyssal peridotites?

The depleted upper mantle sampled by abyssal peridotites and fracture zones has been found to host domains with low $^{187}\text{Os}/^{188}\text{Os}$. The low $^{187}\text{Os}/^{188}\text{Os}$ in many abyssal peridotite samples is too low to result from melt extraction at modern ridges, but instead

requires depletion to have occurred at up to 2 Ga (Harvey et al., 2006; Liu et al., 2008). This observation suggests that ancient depleted domains that have survived in the convecting mantle are incorporated into the oceanic mantle lithosphere at ridges and appear in the abyssal peridotite data set. Portions of the oceanic mantle lithosphere (including portions with low $^{187}\text{Os}/^{188}\text{Os}$) beneath hotspots may be dislodged by (and entrained in) upwelling hotspot lavas. This model argues that the peridotite xenoliths hosted in hotspots lavas sample the same mantle reservoir (and has the same heterogeneous Os-isotopic composition) as abyssal peridotites (e.g., Lassiter et al., 2014).

The depleted mantle sampled by abyssal peridotites is a near-perfect match for the Os-isotopic composition of the xenoliths sampled by Tubuai and Savai'i. Lassiter et al. (2014) noted that, globally, the peridotite xenolith population hosted in oceanic hotspot lavas (average of 0.1244 ± 0.0043 , 1σ standard deviation) and abyssal peridotites (average of 0.1245 ± 0.0043) have indistinguishable average $^{187}\text{Os}/^{188}\text{Os}$ and nearly identical distributions. Like the global OIB peridotite xenolith average, the average $^{187}\text{Os}/^{188}\text{Os}$ of Savai'i (0.1237 ± 0.0032 , 1σ standard deviation) and Tubuai xenoliths (0.1262 ± 0.0035) are indistinguishable from the global abyssal peridotite average compiled by Lassiter et al. (2014). Additionally, the most unradiogenic $^{187}\text{Os}/^{188}\text{Os}$ in Savai'i (down to 0.1173) and Tubuai (0.1163) are bracketed by the lowest values found in abyssal peridotites, as the depleted oceanic mantle sampled by abyssal peridotites also hosts unradiogenic $^{187}\text{Os}/^{188}\text{Os}$ (down to 0.1139; Harvey et al., 2006; Liu et al., 2008). Studies of abyssal peridotite $^{187}\text{Os}/^{188}\text{Os}$ (Snow and Reisberg, 1995; Standish et al., 2002; Harvey et al., 2006; Liu et al., 2008; Lassiter et al., 2014) demonstrate that the shallow oceanic mantle encompasses the range of Os-isotopic compositions observed in oceanic hotspot

xenoliths. The similarity in Os isotopic composition between the xenoliths in this study, other global hotspot xenoliths, and abyssal peridotites permits an interpretation whereby the xenoliths sample an ancient depleted component in the convecting mantle that has been incorporated into the oceanic lithosphere at a mid-ocean ridge. As did Lassiter et al. (2014), we conclude that the similarity of $^{187}\text{Os}/^{188}\text{Os}$ of ocean hotspot mantle xenoliths and abyssal peridotites suggests that they sample the same heterogeneous mantle reservoir.

For Tubuai and Savai'i peridotite xenoliths, it is reasonable to adopt a model in which they also sample the same mantle reservoir as global average abyssal peridotites. We favor a model where the unradiogenic Os in xenoliths from Samoa and Tubuai are simply portions of depleted domains of ancient, heterogeneous oceanic mantle that have survived intact in a convecting mantle for up to 1.8 Ga (see Table 1). These depleted domains could be regions of upper mantle that experienced melt extraction in the past, and they are incorporated into the oceanic mantle lithosphere during mantle upwelling beneath ridges. Once captured and incorporated into the oceanic lithosphere, these domains are sampled as abyssal peridotites or they are transported from mantle lithospheric depths to the surface in erupting ocean island basalts.

4.5. The paradox of the survival of early-Hadean $^{129}\text{Xe}/^{130}\text{Xe}$ signatures, but not Hadean $^{187}\text{Os}/^{188}\text{Os}$ or $^{142}\text{Nd}/^{144}\text{Nd}$, in the convecting mantle.

The low $^{187}\text{Os}/^{188}\text{Os}$ observed in Savai'i and Tubuai peridotite xenoliths, and in abyssal peridotites and peridotite xenoliths from other hotspot localities, indicates that

ancient, depleted mantle domains are preserved at multiple locations in the ocean basins. While depletion ages of up to 2 Ga are recorded at several localities, an important question is why even older depletion events, dating from the time of the Archean and even Hadean (which are clearly part of the subcontinental lithospheric mantle), are not recorded in the Os-isotopic compositions measured in peridotites from the oceanic mantle?

The geologic record provides evidence that mantle melting that led to extraction of continental and oceanic crust has occurred throughout much of the history of the Earth. $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{176}\text{Hf}/^{177}\text{Hf}$ isotopes in Archean crustal rocks (e.g., Shirey and Hanson, 1986; Stein and Hofmann, 1994; Moorbath et al. 1997; Blichert-Toft et al. 1999, Vervoort and Blichert-Toft, 1999; Caro 2011; Khanna et al., 2014) and Hf isotopes in Hadean zircons (e.g., Harrison et al., 2005, 2008; Blichert-Toft and Albarède, 2008; Kemp et al., 2010; Guitreau et al., 2012) preserve evidence for early mantle depletion by melt extraction. Concomitant melting and Re-depletion will generate reservoirs that preserve low $^{187}\text{Os}/^{188}\text{Os}$ over time. If all Re is assumed to be extracted from the mantle during melting, a model age—a Re-depletion age, called T_{RD} —can be calculated for the depletion event by calculating when the measured Os-isotopic composition of the depleted peridotite intersects the time evolution of mantle $^{187}\text{Os}/^{188}\text{Os}$ over time (Walker et al., 1989). Continental xenoliths from the SCLM preserve the lowest known terrestrial $^{187}\text{Os}/^{188}\text{Os}$, and have Re-depletion model ages that indicate melt extraction as early as the Mesoarchean (e.g., Walker et al., 1989; Shirey and Walker, 1998; Pearson et al., 2003; Carlson et al., 2005). Abyssal peridotites exhibit lower $^{187}\text{Os}/^{188}\text{Os}$ than chondritic, thus preserving evidence of past melt depletion (e.g., Snow and Reisberg, 1995; Brandon

et al., 2000; Standish et al., 2002; Harvey et al., 2006; Liu et al., 2008; Lassiter et al., 2014), but abyssal peridotites and mantle xenoliths sampling the oceanic mantle do not preserve $^{187}\text{Os}/^{188}\text{Os}$ as low as identified in the SCLM. The oldest T_{RD} model ages in the oceanic mantle are only slightly older than 2 Ga (Fig. 2) (Harvey et al., 2006; Bizimis et al., 2007; Liu et al., 2008).

It is paradoxical that the oceanic mantle does not preserve the Os-isotopic record of earlier depletion events, as the mechanisms responsible for mantle depletion are directly tied to plate tectonics, which, by some estimates, has operated for at least the past ~ 3 Ga (Shirey and Richardson, 2011; Van Kranendonk, 2010). During the time period over which modern plate tectonics is thought to have operated, the upper mantle has been continuously depleted by partial melting and melt extraction at ridges, and it is inferred that hotter mantle temperatures at 3 Ga would have resulted in higher degrees of mantle melting (and the generation of larger volumes of depleted mantle domains) compared to the more recent geological past. While larger volumes of depleted mantle production prior to 2 Ga might be expected, T_{RD} model ages in abyssal peridotites and ocean island xenoliths provide little record of mantle depletion before ~ 2 Ga. Employing a global compilation of Os-isotopic data from abyssal peridotites and ocean island peridotite xenoliths, Lassiter et al. (2014) found that various model ages based on the Re-Os system, including T_{MA} (mantle extraction model age; Shirey and Walker, 1998) and $T_{\text{Al}_2\text{O}_3}$ (“aluminachron” model age; Reisberg and Lorand, 1996), do yield older average ages for hotspot mantle xenoliths (1.3 Ga and 1.6 Ga, respectively) and abyssal peridotites (1.7 Ga and 1.5 Ga, respectively). These older ages are not unexpected, as the assumptions upon which the model ages are based tend to yield older model depletion events than T_{RD} ages

(which yield minimum ages for the depletion event, as Re is unlikely to be completely extracted from the peridotite and Re may be reintroduced into the peridotite by melt infiltration, or “metasomatism”, following the depletion event). Nonetheless, the various Os model ages still fail to capture the large volumes of depleted mantle that must have formed prior to 2 Ga (Fig. 2), and depleted reservoirs with older ages are underrepresented or absent in the oceanic hotspot mantle xenolith and abyssal peridotite datasets. If the rate of production of oceanic lithosphere and the generation of depleted mantle reservoirs was higher when the mantle was hotter, then ancient Os depletion ages might be expected to be prevalent in the geologic record. Why are old depletion ages rare or absent in the abyssal peridotite and oceanic hotspot xenolith Os-isotopic record?

The lack of very old Os-isotopic depletion ages in the oceanic mantle is a long-standing problem. A common solution is to suggest that isotopic homogenization during convective mixing has erased old Os-isotopic ages (Hauri, 2002; Harvey et al., 2006), perhaps as Re-rich crust is recycled into the mantle and rehomogenized with Re-depleted harzburgitic restite that comprises recycled oceanic mantle lithosphere (Lassiter et al., 2014) or perhaps simply because most of the Os is returned back into the mantle with the peridotitic portion of the oceanic lithosphere. If Os-isotopic evidence for early depletion events was erased by such a mixing process, the relatively young average Os-isotopic model ages available in peridotites sampling the convecting mantle may reflect mixing timescales as the oceanic mantle is homogenized by mixing.

Convective mixing also likely explains the lack of $^{142}\text{Nd}/^{144}\text{Nd}$ variability in the modern mantle. Early-formed silicate-Earth reservoirs exhibited $^{142}\text{Nd}/^{144}\text{Nd}$ variability that was generated by decay of short-lived ^{146}Sm to ^{142}Nd (^{146}Sm half life is 68-103 Ma;

789 Kinoshita et al., 2012) following early silicate Earth differentiation, and all terrestrial
 790 $^{142}\text{Nd}/^{144}\text{Nd}$ variability was generated in the lifetime of ^{146}Sm (the first 300 to 500 Ma of
 791 Earth's history) (e.g., Harper and Jacobsen, 1992; Caro et al., 2006; Boyet and Carlson,
 792 2006; Bennett et al. 2007; O'Neil et al., 2008; Rizo et al., 2013). Variability in
 793 $^{142}\text{Nd}/^{144}\text{Nd}$ diminished since the Hadean but persisted in the mantle until 2.7 Ga (e.g.,
 794 Bennett et al. 2007; Debaille et al., 2013), and $^{142}\text{Nd}/^{144}\text{Nd}$ anomalies have not been
 795 observed in younger rocks. Like early (Hadean and Archean) Re-depletion events that are
 796 no longer recorded in the osmium-isotopic composition of the oceanic mantle, early-
 797 formed $^{142}\text{Nd}/^{144}\text{Nd}$ heterogeneity has not been detected in the modern convecting
 798 oceanic mantle (Caro et al., 2006; Boyet and Carlson, 2006; Andreasen et al., 2008;
 799 Murphy et al., 2010; Cipriani et al., 2011; Caro, 2011; Jackson and Carlson, 2012). A
 800 simple model for the erasure of early-formed $^{142}\text{Nd}/^{144}\text{Nd}$ and $^{187}\text{Os}/^{188}\text{Os}$ isotopic
 801 variability is that vigorous convection in the hotter Hadean and Archean mantle was less
 802 conducive to preserving early-formed $^{142}\text{Nd}/^{144}\text{Nd}$ and Os-isotopic heterogeneities.

803 In contrast to the $^{187}\text{Os}/^{188}\text{Os}$ and $^{142}\text{Nd}/^{144}\text{Nd}$ isotopic records, which do not
 804 preserve early-formed isotopic signatures in the modern oceanic mantle, xenon isotopes
 805 record heterogeneities in the modern oceanic mantle that formed prior to 4.45 Ga,
 806 indicating that Xe is more resistant to convective mixing than Os and Nd. ^{129}Xe is the
 807 decay product of the short lived isotope ^{129}I ($T_{1/2}=15.7$ Ma), and all $^{129}\text{Xe}/^{130}\text{Xe}$
 808 variability in the Earth was generated by fractionation of I from Xe during the first ~100
 809 million years of Earth's history. Early work on mantle-derived samples showed that the
 810 $^{129}\text{Xe}/^{130}\text{Xe}$ composition of the mantle is different from the atmospheric composition
 811 (Staudacher and Allegre, 1982; Allegre et al., 1983; Poreda and Farley, 1992). A recent

812 series of papers have demonstrated that heterogeneous $^{129}\text{Xe}/^{130}\text{Xe}$ signatures are
 813 preserved in the modern oceanic mantle sampled at mid-ocean ridges and plume
 814 influenced lavas from Iceland and Rochambeau Rift (Tucker et al., 2012; Parai et al.,
 815 2012; Mukhopadhyay et al., 2012; Petö et al., 2013; Parai and Mukhopadhyay, 2015).
 816 This $^{129}\text{Xe}/^{130}\text{Xe}$ heterogeneity in the oceanic mantle was generated in the early Hadean,
 817 and has survived for 4.5 Ga in the convecting mantle to be sampled by recent MORB and
 818 OIB volcanism. Therefore, it is critical to understand how the early-formed $^{187}\text{Os}/^{188}\text{Os}$
 819 and $^{142}\text{Nd}/^{144}\text{Nd}$ isotopic record of mantle depletion was erased while $^{129}\text{Xe}/^{130}\text{Xe}$
 820 heterogeneities from the earliest Hadean persist in the modern oceanic mantle.
 821 Mukhopadhyay (2012) suggested that the erasure of Hadean $^{142}\text{Nd}/^{144}\text{Nd}$ signatures, but
 822 not Hadean $^{129}\text{Xe}/^{130}\text{Xe}$, from the convecting mantle is due to the small magnitude of the
 823 $^{142}\text{Nd}/^{144}\text{Nd}$ anomalies ($< 0.004\%$ variability in $^{142}\text{Nd}/^{144}\text{Nd}$ has been observed in
 824 terrestrial Archean terranes; e.g., Caro et al., 2006; O’Neil et al., 2008; Rizo et al., 2013)
 825 compared to the large range of $^{129}\text{Xe}/^{130}\text{Xe}$ signatures ($>10\%$ variability is observed in
 826 the modern mantle; Mukhopadhyay et al., 2012), and the smaller $^{142}\text{Nd}/^{144}\text{Nd}$ anomalies
 827 are more easily erased than the larger magnitude $^{129}\text{Xe}/^{130}\text{Xe}$ anomalies. One problem
 828 with using a “small magnitude anomaly” explanation for the Os data is that Os-isotopic
 829 heterogeneities in the mantle are large, yet early-formed Os-isotopic signatures still have
 830 been erased. For example, the $^{187}\text{Os}/^{188}\text{Os}$ of early-formed depleted mantle at 4.55 Ga
 831 was ~ 0.095 , which is $>25\%$ lower than the modern mantle (0.1296; Meisel et al., 2001).
 832 Therefore, the absence of early-formed, large magnitude $^{187}\text{Os}/^{188}\text{Os}$ excursions in the
 833 modern convecting mantle is evidence that the preservation of early geochemical
 834 heterogeneities is not necessarily a function of the magnitude of the original geochemical

835 anomaly. However, we cannot exclude the possibility that $^{142}\text{Nd}/^{144}\text{Nd}$ heterogeneity
836 exists in the modern mantle, but it is not sampled because either 1.) existing
837 instrumentation to not yet provide the necessary precision to resolve such anomalies or
838 2.) lavas sampling the modern mantle that have resolvable $^{142}\text{Nd}/^{144}\text{Nd}$ anomalies have
839 not been analyzed. Nonetheless, given the existing dataset, the $^{142}\text{Nd}/^{144}\text{Nd}$ in the modern
840 mantle is homogeneous.

841 The variable fluxes of trace elements between different terrestrial geochemical
842 reservoirs—including the crust, mantle and atmosphere—must also be an important
843 consideration when evaluating the survival of Hadean isotopic signatures in the modern
844 mantle (e.g., Galer and O’Nions, 1985; Albarède, 2001; Kellogg et al., 2002).
845 Mukhopadhyay (2012) proposed an alternative explanation for the preservation of
846 Hadean $^{129}\text{Xe}/^{130}\text{Xe}$, but not $^{142}\text{Nd}/^{144}\text{Nd}$, isotopic signatures in the modern mantle, and
847 this model may explain the lack of abundant, early-formed Os-isotopic reservoirs in the
848 mantle as well. Mukhopadhyay (2012) argued that the lower recycling efficiency of
849 xenon into the mantle compared to Nd helps preserve early-formed Xe-isotopic
850 signatures. The mantle degasses xenon to the Earth’s atmosphere during mantle melting,
851 and noble gases are not efficiently returned to the mantle during recycling (e.g.,
852 Staudacher and Allegre, 1988). As a result, during partial melting at ridges followed by
853 lithospheric return to the mantle, most of the mantle xenon enters the atmosphere. A
854 small portion of the xenon is thought to be returned to the mantle during recycling
855 (Porcelli and Wasserburg, 1995; Holland and Ballentine, 2006; Holland et al., 2009;
856 Kendrick et al. 2011; Moreira, 2013). However, xenon is not as efficiently recycled back
857 into the deep Earth as Nd and Os (Mukhopadhyay et al., 2012) because xenon is

incompatible, fluid mobile and volatile, and largely degassed to the atmosphere (Fig. 6). Therefore, early-formed Hadean ^{129}Xe signatures in the mantle are not completely overprinted by young, recycled atmospheric Xe.

Using Xe-isotopic constraints, Parai and Mukhopadhyay (2015) argue that 85 to 92% of the Xe in the convecting mantle sampled by MORB and OIB lavas is recycled atmospheric Xe that was transported into the mantle by subduction over geologic time. However, while the Xe budget of the mantle is actually dominated by recycled atmospheric Xe, the fraction of recycled Xe is not sufficient to erase Hadean $^{129}\text{Xe}/^{130}\text{Xe}$ signatures: much of the remaining Xe (8-15%) is primitive mantle Xe (after accounting for Xe produced by U and Pu fission). Unlike the case for Xe, the fractions of the Nd and Os budgets in the mantle that have been recycled are poorly constrained, but this fraction must exceed the fraction of Xe in the mantle that is recycled (i.e., >85 to 92%) such that the Hadean and Archean Nd and Os isotopic signatures have been largely overprinted in the convecting mantle.

For example, Nd is refractory and fluid-immobile and retained in the crustal portion of the oceanic lithosphere. Thus, Nd is more efficiently returned to the mantle during lithospheric recycling, and recycled Nd (meaning Nd that has gone through the plate tectonic cycle at least once) constitutes a large fraction of the mantle's Nd inventory. Assuming modern subduction rates, ancient oceanic lithosphere (including crust and mantle) recycled into the Earth's mantle since the inception of plate tectonics may reach ~80% of the mantle's mass (Salters and Stracke, 2004; Brandenburg et al., 2008). If Nd in oceanic lithosphere is immobile during subduction and quantitatively returned to the mantle in subducting slabs, then ~80% of the Nd in the mantle is recycled.

881 This is likely a minimum value: if oceanic lithosphere formation and subduction operated
882 at higher rates in the geologic past, the fraction of the Nd in the mantle that is recycled
883 exceeds 80% (and it is possible that nearly the entire Nd budget of the mantle may be
884 recycled). Thus, Hadean ^{142}Nd signatures in the mantle are more efficiently overprinted
885 by recycled material than Hadean ^{129}Xe signatures.

886 Like Nd, Os is also efficiently returned to the mantle during recycling. While Os
887 is almost certainly mobile in fluids and some can be lost from the oceanic crustal portion
888 of a downgoing slab, little Os resides in the oceanic crust. Instead, Os is compatible and
889 chalcophile and >99% of the Os in the oceanic lithosphere (assuming <100 ppt Os in the
890 oceanic crustal lithosphere and 3,000 ppt Os in the depleted mantle lithosphere) is
891 retained in the depleted oceanic mantle lithosphere portion of the slab, and is thus
892 returned to the mantle during recycling. If the concentration of osmium in the convecting
893 mantle and oceanic mantle lithosphere are similar, and if the mantle is comprised of 80%
894 recycled mantle lithosphere, then ~80% of the osmium in the convecting mantle has been
895 recycled and thus has passed through the plate tectonic cycle as least once since the onset
896 of plate tectonics. Again, this is a minimum value, and the fraction of Os in the
897 convecting mantle that is recycled may approach 100%. Thus, early-formed, Hadean
898 $^{187}\text{Os}/^{188}\text{Os}$ signatures in the mantle have been more efficiently overprinted by recycled
899 material than Hadean ^{129}Xe signatures.

900 Ultimately, the isotopic evolution of different radiogenic isotopic systems in the
901 mantle depends on the relative recycling efficiencies of parent and daughter elements.
902 However, unlike the ^{146}Sm - ^{142}Nd , and the ^{129}I - ^{129}Xe systems, where the parent isotopes
903 have been effectively extinct since the Hadean, the parent in the ^{187}Re - ^{187}Os system,

904 ^{187}Re , is still decaying and thus can modify the $^{187}\text{Os}/^{188}\text{Os}$ of the modern mantle. Like
905 Os, Re is also efficiently returned to the mantle during recycling of oceanic crust, and any
906 Re returned to the mantle will serve to overprint ancient (low) $^{187}\text{Os}/^{188}\text{Os}$ reservoirs via
907 generation of radiogenic ^{187}Os in the mantle over time. Thus, long-term recycling of
908 ^{187}Re also has played a supporting role with Os recycling in overprinting ancient (low)
909 $^{187}\text{Os}/^{188}\text{Os}$ domains in the mantle.

910 The long-term preservation of early-formed isotopic signatures of a particular
911 element is a function of not just mantle convective stirring, but also the recycling
912 efficiency of the element back into the mantle at subduction zones (Gonnermann and
913 Mukhopadhyay, 2009; Mukhopadhyay, 2012). This conceptual model permits a more
914 general description of which elements are (e.g., Xe), and are not (e.g., Nd and Os), likely
915 to preserve Hadean geochemical signatures in the Earth's mantle: Incompatible elements
916 that are fluid mobile are more likely to preserve Hadean geochemical signatures in the
917 mantle than compatible elements and fluid immobile incompatible elements.
918 Incompatible, fluid-mobile elements (like Xe) are efficiently extracted from the
919 downgoing slab and accumulate in shallow reservoirs (e.g., the crust and atmosphere)
920 instead of being returned to the mantle. Such elements are less efficiently returned to the
921 mantle than Os and Nd, and are therefore more likely to preserve early-formed,
922 primordial signatures in the mantle. The incompatible, fluid-mobile element need not be
923 volatile to preserve Hadean geochemical signatures in the mantle: incompatible, fluid-
924 mobile elements that are refractory also will be extracted from the downgoing slab,
925 concentrated in the continental crust, and not efficiently returned to the mantle. For
926 example, W is both fluid-mobile and incompatible (Arevalo and McDonough, 2008;

König et al., 2008; 2011) and the long-term preservation of early (<30 Ma) formed $^{182}\text{W}/^{184}\text{W}$ variability in 2.8 Ga komatiites (Touboul et al., 2012) and relatively modern flood basalts (Rizo et al., 2015) would be favored in the model presented here. However, non-fluid mobile incompatible elements (like Nd) and compatible elements (which, like Os, are preserved in the mantle lithosphere during recycling) are efficiently returned to the mantle during recycling and mix with, dilute, and ultimately overprint primordial signatures associated with the isotopes of such elements (Fig. 7). While high recycling efficiencies (for Nd and Os) may not be conducive to the preservation of early-formed heterogeneities, somewhat lower recycling efficiencies for incompatible fluid-mobile elements (e.g., Xe) may allow long-term preservation of early-formed Hadean signatures in the convecting mantle.

This conceptual model may be extended to many isotopic systems, including the preservation of primordial $^3\text{He}/^4\text{He}$ isotopic signatures in the deep mantle resulting from the return of noble gas depleted slabs (Gonnermann and Mukhopadhyay, 2009), to build an understanding for the role of plate tectonics and subduction in determining the preservation of early-formed, primordial reservoirs in the Earth's mantle (Figs. 6 and 7). Several radiogenic isotopic systems based on incompatible lithophile elements, including ^{87}Rb - ^{87}Sr , ^{147}Sm - ^{143}Nd , ^{176}Lu - ^{176}Hf and $^{238,235}\text{U}$ - ^{232}Th - $^{208,207,206}\text{Pb}$, are unlikely to preserve Hadean isotopic signatures in the convecting mantle: owing to variable fluid mobility during subduction, either the parent (e.g., U and Th), the daughter (e.g., Sr), or both (e.g., Sm-Nd and Lu-Hf) are likely to be efficiently returned to the mantle. Thus, these radiogenic isotopic systems are less likely to preserve memories of the Hadean in the convecting mantle. However, some deep mantle reservoirs, possibly located in the

large low shear wave velocity provinces (LLSVPs), may remain relatively unmodified by the addition of subducted material over geologic time (White, 2015). These relatively unmodified reservoirs, characterized by hosting primitive noble gas signatures (White, 2015), provide an exception where Hadean isotopic signatures can persist for elements (e.g., Pb) that may be efficiently recycled into the mantle (Jackson et al., 2010; Jackson and Carlson, 2011).

Acknowledgements. We acknowledge discussion with Roberta Rudnick, Sujoy Mukhopadhyay, Bernhard Peucker-Ehrenbrink and Jasper Konter. We thank Alison Price for help improving a figure. MGJ thanks Vincent Salters for asking why xenon-isotopic anomalies are preserved in the mantle, but early-formed Os-isotopic signatures are not. We thank Jessica Warren and Graham Pearson for sharing Os-isotopic databases for peridotites. Constructive and helpful reviews from Dave Graham, Bill White and Barry Hanan are acknowledged. MGJ thanks Fred Frey for the introduction to trace element geochemistry during coursework at MIT. Jackson acknowledges support from NSF grants EAR-1348082, EAR-1347377, EAR-1145202 and OCE-1153894. Kurz acknowledges NSF grants OCE-1259218 and OCE-1232985.

973

974 **Figure Captions:**

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976 **Figure 1. Map of the south Pacific, showing the locations of Tubuai (Austral**
977 **hotspot) and Savai'i (Samoan hotspot) islands.** Tubuai lies along the Macdonald
978 hotspot trace (Chauvel et al., 1997), and Samoa lies along the Samoan hotspot trace.
979 Pacific plate reconstructions (reproduced from Jackson et al., 2010) are based on the
980 model reported in Wessel and Kroenke (2008); the Macdonald hotspot trend is anchored
981 to Macdonald seamount, and numbers on the Macdonald hotspot reconstruction are ages
982 in Ma. The hotspot track reconstruction shows that the portion of the Pacific lithosphere
983 now occupied by the Samoan islands may have been located over the Macdonald hotspot
984 at ~50 Ma.

985

986 **Figure 2. Relationship between $^{187}\text{Os}/^{188}\text{Os}$ and Os concentrations for Savai'i and**
987 **Tubuai xenoliths.** Basalts from Tubuai (Hauri and Hart, 1993) and Savai'i (Hauri and
988 Hart, 1993; Workman et al., 2004) are plotted for comparison. Additionally, whole rock
989 abyssal peridotites (Snow and Reisberg, 1995; Brandon et al., 2000; Standish et al., 2002;
990 Alard et al., 2005; Becker et al., 2006; Harvey et al., 2006; Liu et al., 2008; Sichel et al.,
991 2008; Lassiter et al., 2014) and whole rock pyroxenite (from Hawaii; Sen et al., 2011)
992 and peridotite (from OJP, Hawaii, Kerguelen, Cape Verde, Canary Islands, and Fernando
993 de Noronha; Hassler and Shimizu, 1998; Widom et al., 1999; Meisel et al., 2001;
994 Brügman et al., 2008; Becker et al., 2006; Bizimis et al., 2007; Sen et al., 2003; Ishikawa
995 et al., 2011) xenoliths from oceanic hotspot localities are also shown. PUM is from

Meisel et al. (2001). Each xenolith from Savai'i and Samoa is represented by a single datapoint; in cases where replicate measurements have been made, the average Os concentration and the weighted (by concentration) average Os-isotopic compositions are shown. The Savai'i pyroxenite (SAV-05-25), and the highly metasomatized, enriched xenoliths examined by Hauri et al. (1993)--Savai'i (SAV-1-28) and Tubuai (TBA-1-9)--are highlighted separately. TBA-5-4 is a relatively altered xenolith, but plots together with the other Tubuai xenoliths. Rhenium depletion (T_{RD}) model ages are provided and assume a present-day PUM (primitive upper mantle) value of 0.1296 (Meisel et al., 2001) and assume that the Re/Os of the samples is 0 following melt depletion of PUM at the age indicated.

Figure 3. Trace elements ratios (Zr/Nd and La/Yb) and concentrations (La and Yb) measured in clinopyroxene versus $^{187}\text{Os}/^{188}\text{Os}$. Trace element data were measured in clinopyroxene and were obtained by ion probe. Trace element data are reported elsewhere (Hauri, 1992; Hauri et al., 1993; Hauri and Hart, 1994). Replicate Os measurements are reported in the same manner as in Fig. 2. The Savai'i pyroxenite (SAV-05-25), the altered Tubuai xenolith (TBA-5-4), and the four highly metasomatized, enriched xenoliths reported in Hauri et al. (1993)--two from Savai'i (SAV-1-28 and SAV-1-1) and two from Tubuai (TBA-1-9 and TBA-4-11)--are highlighted separately. Rhenium depletion (T_{RD}) model ages are provided as dashed lines and assume a present-day PUM (primitive upper mantle) value of 0.1296 (Meisel et al., 2001) and assumes that the Re/Os of the samples is 0 following melt depletion of PUM at the age indicated.

Figure 4. Helium isotopic ratios compared to ^4He concentrations. Helium isotopic compositions are reported for a subset of Samoan and Tubuai-xenoliths. Additionally, for comparative purposes, helium isotopic measurements for olivine and cpx phenocrysts from all Samoan lavas (Farley et al., 1992; Workman et al., 2004; Jackson et al., 2007a, 2007b; Jackson et al., 2010; Jackson et al., 2014) and two lavas from the Samoan island of Savai'i (Workman et al., 2004; Jackson et al., 2014) are shown. Helium isotopic measurements for olivine and cpx phenocrysts Tubuai basalts (Graham et al., 1992; Parai et al., 2009; Hanyu et al., 2011b) are also shown. Replicate Os measurements are reported in the same manner as in Fig. 2. Replicate measurements of $^3\text{He}/^4\text{He}$ are averaged. The Savai'i pyroxenite (SAV-05-25), and the highly metasomatized xenoliths--two from Savai'i (SAV-1-28 and SAV-1-1) and one from Tubuai (TBA-1-9)--are highlighted separately. Cpx is clinopyroxene.

Figure 5. Helium isotopic ratios compared to $^{187}\text{Os}/^{188}\text{Os}$, $(\text{La}/\text{Yb})_{\text{cpx}}$ and $(\text{Zr}/\text{Nd})_{\text{cpx}}$

Replicate Os and He measurements are treated as in Figure 4. The Savai'i pyroxenite (SAV-05-25), and the highly metasomatized xenoliths--two from Savai'i (SAV-1-28 and SAV-1-1) and one from Tubuai (TBA-1-9)--are highlighted separately. Cpx is clinopyroxene.

Figure 6. The differing recycling efficiencies of the elements determine whether Hadean isotopic signatures are preserved in the modern convecting mantle.

Compatible elements (like Os, which are retained in the mantle lithosphere) and incompatible elements that are not fluid mobile (like Nd) are efficiently returned to the

mantle during subduction, and Hadean isotopic signatures for these elements are efficiently overprinted in the mantle. However, incompatible fluid-mobile elements, both volatile (e.g., Xe) and refractory, are more efficiently extracted from the downgoing slab and accumulate in shallow geochemical reservoirs (e.g., the continental crust and atmosphere). Thus, incompatible fluid-mobile elements are not efficiently subducted into the mantle, and Hadean isotopic signatures of these are more likely to be preserved in the modern convecting mantle.

Figure 7. The conceptual model that relates survival of early-formed (Hadean) isotopic heterogeneities in the modern convecting mantle to the recycling efficiency of an element. Hadean $^{129}\text{Xe}/^{130}\text{Xe}$ anomalies are preserved in the modern convecting mantle (upper left portion of the figure) because xenon is not efficiently recycled back into the mantle (and a significant fraction of the xenon in the modern mantle reflects its primordial budget). Therefore, early-formed Hadean $^{129}\text{Xe}/^{130}\text{Xe}$ anomalies survive in the modern mantle because they have not been diluted with recycled xenon. In contrast, Nd and Os are efficiently recycled back into the mantle at subduction zones. Therefore, $^{187}\text{Os}/^{188}\text{Os}$ and $^{142}\text{Nd}/^{144}\text{Nd}$ signatures formed in the Hadean have been diluted and ultimately overprinted by continuous recycling of young Os and Nd into the mantle over geologic time.

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Table 1. Os-isotopic compositions and Re-Os concentrations of xenoliths from Savai'i (Sam

Sample ID	Powder Batch	Sample mass (g)	$^{187}\text{Os}/^{188}\text{Os}$	in-run precision 2σ (std. err., abs.)
<i>Sava'i Island (Samoa)</i>				
SA-3-9	batch 1	2.158	0.12295	0.00003
SA-3-9 rep1	batch 1	0.782	0.12118	0.00004
SA-3-9 rep2	batch 1	0.658	0.12196	0.00004
SAV-1-30	batch 1	1.672	0.11928	0.00001
SAV-1-30 rep1	batch 2	0.980	0.12023	0.00006
SAV-1-30 rep2	batch 2	0.876	0.11999	0.00005
SA-3-3	batch 1	1.906	0.12709	0.00002
SA-3-3 rep1	batch 1	0.710	0.12724	0.00010
SAV-1-41	batch 1	1.690	0.11876	0.00002
SAV-1-41 rep1	batch 2	0.746	0.11997	0.00009
SAV-1-41 rep2	batch 2	0.405	0.11732	0.00002
SAV-1-28	batch 1	1.775	0.12537	0.00005
SAV-1-28 rep1	batch 2	0.667	0.12539	0.00003
SA-3-11	batch 1	0.740	0.12663	0.00004
SAV-1-18	batch 1	1.851	0.12267	0.00001
SAV-1-18 rep1	batch 1	0.462	0.12280	0.00004
SAV-1-18 rep2	batch 2	0.911	0.12336	0.00010
SAV-1-13	batch 1	2.285	0.12542	0.00002
SAV-1-13 rep	batch 1	0.541	0.12557	0.00003
SAV-1-46	batch 1	1.894	0.11935	0.00005
SAV-1-46 rep	batch 2	0.772	0.11936	0.00006
SAV-1-47	batch 1	1.674	0.12649	0.00008
SAV-1-47 rep	batch 2	0.542	0.12733	0.00005
SAV-05-29	batch 1	0.349	0.12628	0.00008
SAV-1-42	batch 1	0.460	0.11976	0.00002
SAV-1-42 rep1	batch 1	2.386	0.11996	0.00002
SAV-1-42 rep2	batch 2	0.953	0.12055	0.00008
SAV-3-2	batch 1	2.057	0.12479	0.00009
SAV-1-1 ²		-	0.1284	-

Table 2. Helium isotopic compositions of Samoan and Tubuai xenoliths obtained by

Sample ID		Sample mass (g)	Phase analyzed	$^3\text{He}/^4\text{He}$ (R/Ra)
SA-3-9	Savai'i xenolith	0.2876	olivine	11.16
SA-3-11	Savai'i xenolith	0.2054	olivine	11.63
SA-3-3	Savai'i xenolith	0.2982	olivine	11.85
SA-3-3	Savai'i xenolith	0.2750	orthopyroxene	11.24
SA-1-47	Savai'i xenolith	0.1700	olivine	12.41
SAV-1-41	Savai'i xenolith	0.2205	cloudy olivine	11.71
SAV-1-41	Savai'i xenolith	0.1573	clear olivine	11.78
SAV-1-41	Savai'i xenolith	0.1600	dark green olivine	11.71
SAV-1-1	Savai'i xenolith	0.1122	olivine	11.16
SAV-1-28	Savai'i xenolith	0.1504	olivine	11.39
SAV-05-29	Savai'i xenolith	0.2017	orthopyroxene	11.98
SAV-1-18	Savai'i xenolith	0.2988	olivine	11.85
SAV-1-30	Savai'i xenolith	0.2986	olivine	11.58
SAV-1-46	Savai'i xenolith	0.3035	olivine	11.70
SAV-1-27	Savai'i xenolith	0.3012	olivine	11.43
SAV-1-13	Savai'i xenolith	0.3014	olivine	10.81
SAV-1-42	Savai'i xenolith	0.2972	olivine	11.53
TBA-4-5	Tubuai xenolith	0.3007	olivine	4.87
TBA-4-8	Tubuai xenolith	0.1868	olivine	6.29
TBA-2-1	Tubuai xenolith	0.2036	olivine	6.43
TBA-5-1	Tubuai xenolith	0.2677	olivine	5.20
TBA-1-9	Tubuai xenolith	0.1372	olivine	2.45

Table 3. Modal abundances of Savai'i and Tubuai xenoliths

Sample ID	olivine	orthopyroxene	clinopyroxene	spinel	Apatite	Glass
SA-3-9 ¹	76.9	22.1	0.5	0.5		
SAV-1-30 ¹	62.1	29.6	6.1	2.2		
SA-3-3 ¹	67.6	31.7	0.3	0.4		
SAV-1-41 ¹	76.1	18.9	2.7	2.3		
SAV-1-28 ¹	67.8	27.9	3.1	1.2		
SA-3-11 ¹	80.8	18.5	0.4	0.3		
SAV-1-18 ¹	56.7	38.9	2.6	1.8	trace	
SAV-1-13 ¹	70.9	24.1	3.9	1.1	trace	
SAV-1-46 ¹	76.1	18.6	3.3	2		
SAV-1-47 ¹	75.4	20.6	2.6	1.4		
SAV-05-29		99	1	—	—	<1
SAV-1-42 ¹	82.9	14.3	2.2	0.6		
SAV-3-2 ¹	83.8	15.4	0.4	0.4		
SAV-1-1 ¹	81.4	15	2.1	1	0.3	0.2
SAV-1-27 ¹	83.6	15.9	0.3	0.2		
TBA-4-4	65	15	7	10	3	—
TBA-5-1	66	24	4	5	1	<1
TBA-5-2	58	17	7	3	—	15
TBA-5-4	55	—	30	—	—	15
TBA-1-9	70	—	20	—	—	10
TBA-4-5	70	24	4	2	—	<1
TBA-4-8 ²	NA	NA	NA	NA	NA	NA
TBA-2-1	65	28	5	2	—	—
TBA-4-11	78		10	<1	8	4
TBA-1-7	70	24	4	2	—	<1

¹ Modes reported previously (Hauri and Hart, 1994).

² Remaining portion of xenolith is too small to accurately estimate modal abundances

Fig. 2

Fig. 3

Figure 4

Figure 5

Fig. 6

Fig. 7