Peridotite xenoliths from the Polynesian Austral and Samoa hotspots: Implications for the destruction of ancient $^{187}\text{Os}$ and $^{142}\text{Nd}$ isotopic domains and the preservation of Hadean $^{129}\text{Xe}$ 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 to 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 to 6.4 Ra for Tubuai and 10.8 to 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 Tubuai xenolith suites 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, geochemically-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 heterogeneities 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 (e.g., Xe) concentrate in surface
Lavas erupted at 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 elements 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 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. $^{187}\text{Re}$ undergoes $\beta$-decay to $^{187}\text{Os}$ ($t_{1/2} = 41.6$ 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 radiogenic 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 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 in oceanic environments that host unusually low $^{186}\text{Os}/^{188}\text{Os}$ 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, 1987; Hauri, 1992; Poreda and Farley, 1992; Hauri et al., 1993; Hauri and Hart, 1994; Farley et al., 1994; Farley, 1995; Burnard et al., 1998). Hauri et al. (1993) characterized two metasomatically-enriched peridotite xenoliths each from the islands of Savai’i and Tubuai (Fig. 1). Clinopyroxenes separated from these four xenoliths exhibit extreme REE (particularly light REE) enrichment, but concomitant enrichment in the high field strength elements (HFSE; Ti, Zr, Hf, Nb) was not observed (Hauri et al., 1993). Therefore, the REE enrichment of the clinopyroxenes was interpreted to result from metasomatism of the xenoliths by carbonatic melts, which impart characteristic REE enrichment and HFSE depletion (e.g., Hauri et al., 1993). The origin of the carbonatic melt is thought to be related to highly-enriched, small-degree melts of carbonated peridotite in the mantle plumes that generated both hotspots (Hauri, 1992; Hauri et al., 1993). Supporting this hypothesis, the Sr, Nd and Pb isotopic compositions of clinopyroxenes isolated from the metasomatized xenoliths were found to reflect EM2 and HIMU endmember isotopic compositions associated with lavas erupted at the two hotspots that generated the islands of Savai’i (Samoan hotspot) and Tubuai (Austral hotspot), respectively. However, the ultra-enriched xenoliths analyzed by Hauri et al. (1993) are rare among the xenolith suites from Savai’i and Tubuai, and most xenoliths from these localities do not exhibit evidence for significant REE enrichment (Hauri, 1992; Hauri and Hart, 1994) in keeping with their origin as oceanic lithosphere created at the ridge. To better constrain the geochemical variability of the Savai’i and Tubuai xenolith suite and examine the isotopic heterogeneity of the oceanic lithospheric mantle, we have characterized the Os and Hf isotopic compositions of xenoliths from the same localities that exhibit a range of enriched and depleted REE patterns (Hauri, 1992; Hauri et al., 1993; Hauri and Hart, 1994).

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 ($\text{^{187}Os}/\text{^{188}Os} < 0.11$) that would represent the Earth’s primitive and depleted mantle domains in these times (Fig. 2). This is unexpected because there is 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., 1996; 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, $\text{^{129}Xe}/\text{^{130}Xe}$ isotopic heterogeneities (generated by the decay of the short-lived nuclide $^{129}$I to $\text{^{129}Xe}$ [$t_{1/2} = 15.7$ 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, 2012; Tucker et al., 2012; Parai et al., 2012; Pețő et al., 2013). In light of the preservation of early-formed Xe-isotopic heterogeneities in the convecting mantle, the lack of preservation of early differentiation in Os-isotopes, and other isotopic systems that record early differentiation (like $\text{^{146}Sm}/\text{^{144}Nd}$; $t_{1/2} = 106$ Ma), is perplexing. We present a conceptual model that shows how different recycling efficiencies of Xe, Os and Nd may explain the preservation of Hadean $\text{^{129}Xe}/\text{^{130}Xe}$, but not $\text{^{187}Os}/\text{^{188}Or}$ or $\text{^{142}Nd}/\text{^{144}Nd}$, 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 Hauri during the 1990 field season. The peridotites generally have low modal clinopyroxene (<5% even in REE-enriched examples; Hauri, 1992; Hauri and Hart, 1994). However, several of the Tubuai xenoliths have higher enriched examples; Hauri, 1992; Hauri and Hart, 1994). The single pyroxenite xenolith was collected on Savai‘i by 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, 1994) and Hauri et al. (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 $\text{^{185}Re}/\text{^{190}Os}$ spike was added to the sample with inverse aqua regia. This was followed by
to plot together with the other Tubuai xenoliths. Rhenium are highlighted separately. TBA-5-4 is a relatively altered xenolith, Hauri et al. (1993) —Savai‘i (SAV-1-28) and Tubuai (TBA-1-9)—

concentrations and the weighted (by concentration) average Os-

isotopic compositions are shown. The Savai‘i pyroxenite (SAV-05-

1998; Sen et al., 2003; Becker et al., 2006; Bizimis et al., 2007;

Canary Islands, and Fernando de Noronha; Hassler and Shimizu,

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

concentrations 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 depletions (T_{Re}) 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.

Fig. 2. Relationship between $^{187}$Os/$^{188}$Os and Os concentrations for Savai‘i and Tubuai xenoliths. Basalts from Tubuai (Hauri and Hart, 1993) and Savai‘i (Hauri and Hart, 1993; Workman et al., 2004) are plotted for comparison. Additionally, whole rock abyssal peridotites (Snow and Reisberg, 1995; Brandon et al., 2000; Standish et al., 2002; Alard et al., 2005; Becker et al., 2006; Harvey et al., 2006; Liu et al., 2008; Sichel et al., 2008; Lassiter et al., 2014) and whole rock pyroxenite (from Hawaii: Sen et al., 2011) and peridotite (from OJP, Hawaii, Kerguelen, Cape Verde, Canary Islands, and Fernando de Noronha; Hassler and Shimizu, 1998; Widom et al., 1999; Meisel et al., 2001; Brügmann et al., 2008; Sen et al., 2003; Becker et al., 2006; Bizimis et al., 2007; Ishikawa 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

concentrations 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 depletions (T_{Re}) 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.

3. DATA AND OBSERVATIONS

3.1. Re and Os concentrations and $^{187}$Os/$^{188}$Os compositions

Replicate Os-isotopic analyses were made on 11 whole rock peridotites (six triplicate and five duplicate analyses) (Table 1). Each replicate was performed on a new aliquot of powder (either from the same batch of powder, or on a

| Table 1. Repeated measurements of the Johnson Matthey (J–M) Os standard during the course of this study gave a $^{187}$Os/$^{188}$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 $^{187}$Os/$^{188}$Os ratios were adjusted for the offset between the measured and preferred (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 analysis. 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, 1992; Hauri et al., 1993; Hauri and Hart, 1994). 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 Table 1.
Table 1
Os-isotopic compositions and Re–Os concentrations of xenoliths from Savai’i (Samoa) and Tubuai (Australs) Islands.

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Powder batch</th>
<th>Sample mass (g)</th>
<th>( { }^{187}\text{Os}/^{188}\text{Os} ) In-run precision at 2( \sigma )</th>
<th>Os conc. (ppb)</th>
<th>Os blank (% of total Os analyzed)</th>
<th>Re conc. (ppb)</th>
<th>Re blank (% of total Re analyzed)</th>
<th>( T_{\text{RD}} ) (Ga)</th>
<th>( (\text{Zr}/\text{Nd})_{\text{cpx}} )</th>
<th>( (\text{La}/\text{Yb})_{\text{cpx}} )</th>
<th>( \text{La}_{\text{cpx}} )</th>
<th>( \text{Yb}_{\text{cpx}} )</th>
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</thead>
<tbody>
<tr>
<td><strong>Savai’i Island (Samoa)</strong></td>
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<tr>
<td>SA-3-9</td>
<td>batch 1</td>
<td>2.158</td>
<td>0.12295</td>
<td>0.00003</td>
<td>0.69</td>
<td>0.17</td>
<td>0.76</td>
<td>80.0</td>
<td>1.97</td>
<td>0.26</td>
<td>0.13</td>
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<td>SA-3-9 rep1</td>
<td>batch 1</td>
<td>0.782</td>
<td>0.12118</td>
<td>0.00004</td>
<td>0.62</td>
<td>0.10</td>
<td>1.00</td>
<td></td>
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<tr>
<td>SA-3-9 rep2</td>
<td>batch 1</td>
<td>0.658</td>
<td>0.12196</td>
<td>0.00004</td>
<td>0.77</td>
<td>0.32</td>
<td>0.89</td>
<td></td>
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<td>batch 1</td>
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<td>0.11928</td>
<td>0.0001</td>
<td>0.86</td>
<td>0.14</td>
<td>1.41</td>
<td>3.3</td>
<td>34.4</td>
<td>7.2</td>
<td>0.21</td>
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<td>SAV-1-30 rep1</td>
<td>batch 2</td>
<td>0.980</td>
<td>0.12023</td>
<td>0.00006</td>
<td>0.37</td>
<td>0.18</td>
<td>0.018</td>
<td>13.2</td>
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<td>batch 2</td>
<td>0.876</td>
<td>0.11999</td>
<td>0.00005</td>
<td>0.42</td>
<td>0.59</td>
<td>0.020</td>
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<td>batch 1</td>
<td>1.906</td>
<td>0.12709</td>
<td>0.00002</td>
<td>1.31</td>
<td>0.09</td>
<td>0.35</td>
<td>0.01</td>
<td>7.5</td>
<td>9.0</td>
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<td>batch 1</td>
<td>0.710</td>
<td>0.12724</td>
<td>0.00010</td>
<td>0.43</td>
<td>0.15</td>
<td>0.005</td>
<td>36.7</td>
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<td>batch 1</td>
<td>1.690</td>
<td>0.11876</td>
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<td>0.07</td>
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<td>16.5</td>
<td>7.2</td>
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<td>0.746</td>
<td>0.11997</td>
<td>0.00009</td>
<td>1.42</td>
<td>0.05</td>
<td>0.009</td>
<td>74.7</td>
<td></td>
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<td>batch 2</td>
<td>0.405</td>
<td>0.11732</td>
<td>0.00002</td>
<td>1.58</td>
<td>0.16</td>
<td>0.0003</td>
<td>85.1</td>
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<tr>
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<td>batch 1</td>
<td>1.775</td>
<td>0.12537</td>
<td>0.00005</td>
<td>0.25</td>
<td>0.48</td>
<td>0.58</td>
<td>0.62</td>
<td>26.0</td>
<td>39.3</td>
<td>1.51</td>
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<td>batch 2</td>
<td>0.667</td>
<td>0.12539</td>
<td>0.00003</td>
<td>0.76</td>
<td>0.05</td>
<td>0.58</td>
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<td>SA-3-11</td>
<td>batch 1</td>
<td>0.740</td>
<td>0.12663</td>
<td>0.00004</td>
<td>0.98</td>
<td>0.10</td>
<td>0.002</td>
<td>77.5</td>
<td>0.41</td>
<td>62.8</td>
<td>0.72</td>
<td>0.05</td>
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<td>1.851</td>
<td>0.12267</td>
<td>0.00001</td>
<td>0.85</td>
<td>0.14</td>
<td>0.95</td>
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<td>19.2</td>
<td>0.96</td>
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<td>0.12280</td>
<td>0.00004</td>
<td>6.04</td>
<td>0.02</td>
<td>0.011</td>
<td>35.4</td>
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<td>0.12336</td>
<td>0.00010</td>
<td>2.09</td>
<td>0.03</td>
<td>0.008</td>
<td>24.9</td>
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<td>0.12542</td>
<td>0.00002</td>
<td>1.27</td>
<td>0.09</td>
<td>0.57</td>
<td>3.0</td>
<td>11.7</td>
<td>7.0</td>
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<td>0.12557</td>
<td>0.00003</td>
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(continued on next page)
### Tubuai Island (Cook-Australs)

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<th>Powder batch</th>
<th>Sample mass (g)</th>
<th>$^{187}$Os/$^{188}$Os In-run precision at 2σ</th>
<th>Os conc. (ppb)</th>
<th>Os blank (% of total Os analyzed)</th>
<th>Re conc. (ppb)</th>
<th>Re blank (% of total Re analyzed)</th>
<th>( T_{RD} ) (Ga)(^c)</th>
<th>(Zr/Nd)(_{cpx})(^d)</th>
<th>(La/Yb)(_{cpx})</th>
<th>La(_{cpx})</th>
<th>Yb(_{cpx})</th>
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\(^a\) Replicates analyses were made on either the original batch of powder (called “batch 1”), or were made on a different batch of powder (called “batch 2”) prepared from a different part of the same xenolith. All osmium standard and unknown analyses with elevated $^{185}$Re/$^{188}$Os (i.e., >0.0005) were discarded. Re and Os were measured on the same aliquot of powder.

\(^b\) Os-isotopic analyses for two Savai’i xenoliths and two Tubuai xenoliths shown in this table are reported in Hauri et al. (1993) and Hauri (1992).

\(^c\) \( T_{RD} \) ages were calculated using the PUM $^{186}$Os/$^{188}$Os value (0.1296) from Becker et al. (2006).

\(^d\) Trace element abundances in cpx are from Hauri (1992), Hauri et al. (1993) and Hauri and Hart (1994). Trace element data on SAV-05-29 are reported here for the first time. Trace element concentrations that are below the detection limit are not reported. Trace element concentrations measured on more than one (clinopyroxene) cpx grain from a single xenolith are averaged.
different batch of powder prepared from different portions of the xenolith) that was processed independently through the various steps of wet chemistry and mass spectrometry (and analyzed on a different filament, in a different barrel, and on a different day on the mass spectrometer). The reproducibility of the $^{187}\text{Os}/^{188}\text{Os}$ replicate measurements is better than 0.8% for eight samples, but is 1.5% to 2.3% for three samples (SAV-1-41, SA-3-9, TBA-4-5). This far exceeds the internal precision for all the measurements, which is better than 0.1% (2σ), and suggests that there is isotopic heterogeneity within these peridotites. Two of the three replicate analyses that have Os-isotopic reproducibility worse than 1.5% represent analyses of different aliquots of the same batch of powder, which indicates that there can be significant isotopic heterogeneity within a batch of powder. The third example of reproducibility worse than 1.5% occurred in a replicate for which different batches of powder (sampling different portions of the same xenolith) have different Os-isotopic compositions. The isotopic heterogeneity within a single batch of powder, or between two batches of powder sampling different parts of the same xenolith, may be due to the presence of trace phases, or "nuggets", that have distinct $^{187}\text{Os}/^{188}\text{Os}$. The different nuggets might represent different generations of Os in a xenolith that contains appreciable Os. The observation that peridotite xenoliths host isotopically-heterogeneous nuggets is well established in the literature (Shirey and Walker, 1995, 1998; Reisberg and Meisel, 2002; Lorand et al., 2008). We note that while intra-xenolith 187Os/188Os variability is up to 2% (sample TBA-4-5) in the Tubuai xenoliths and up to 2.3% (sample SAV-1-41) in the Savaiʻi xenoliths, the total inter-xenolith $^{187}\text{Os}/^{188}\text{Os}$ variability among different Tubuai (10.8%) and Savaiʻi (8.6%) xenoliths exceeds the variability observed within any single xenolith.

The concentration of Os also varies within a single powder and 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 6–29%. The remaining six xenoliths have Os concentrations that vary by a factor of 2.3–7.1, 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. 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, SAV-1-41, 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 Savaiʻi xenoliths generally have lower Re than the Tubuai xenoliths: Re concentrations vary from just below 0.001 ppb to 0.035 ppb in the Savaiʻi 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 the various steps of wet chemistry and mass spectrometry. Of the various steps of wet chemistry and mass spectrometry. Of the various steps of wet chemistry and mass spectrometry.

<table>
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<th>Sample ID</th>
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<th>Phase analyzed</th>
<th>$^{3}{\text{He}}/^{4}{\text{He}}$ (R/Ra)</th>
<th>In-run precision at 1σ</th>
<th>$^{4}{\text{He}}$ cc STP/g</th>
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<td>0.2876</td>
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Table 2
Helium isotopic compositions of Samoan and Tubuai xenoliths obtained by crushing in vacuo.
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; replicate $^{187}\text{Os}/^{188}\text{Os}$ analyses for each xenolith were averaged and included as a single value in the final average) 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 $^{187}\text{Os}/^{188}\text{Os}$ ratios 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. 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. Replicate Os measurements are reported in the same manner as in Fig. 2. The Savai‘i pyroxenite (SAV-05-29), 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.](http://dx.doi.org/10.1016/j.gca.2016.02.011)
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 strongest carbonatite signatures (SAV-1-42 and SAV-1-46) 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-1-9).

Fig. 4. Helium isotopic ratios compared to $^4\text{He}$ 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,b; Jackson et al., 2010a,b; 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 from 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-05-29), 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.

Fig. 5. Helium isotopic ratios compared to $^{187}\text{Os}/^{188}\text{Os}$, (La/Yb)$_{\text{cpx}}$ and (Zr/Nd)$_{\text{cpx}}$. Replicate Os and He measurements are treated as in Fig. 4. The Savai’i pyroxenite (SAV-05-29), 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.
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 have not been modified by metasomatism.

### 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 7–8 Ra; Graham, 2002; Georgen et al., 2003). The new peridotite xenolith $^{3}\text{He}/^{4}\text{He}$ measurements from Savai’i are similar to previous helium isotopic measurements 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 $^{4}\text{He}$ concentrations vary from $3.6 \times 10^{-8}$ to $3.1 \times 10^{-6}$ 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 a 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 (2.5–6.4 Ra) are also lower than values 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 $^{4}\text{He}$ concentrations in the Tubuai xenoliths are all lower than the Savai’i xenoliths, and range from $1.4 \times 10^{-9}$ to $2.4 \times 10^{-8}$ cc STP/g. While there is no relationship between $^{3}\text{He}/^{4}\text{He}$ and $^{4}\text{He}$ concentrations in the Tubuai xenoliths, the xenolith with the lowest $^{4}\text{He}$ concentration has the lowest $^{3}\text{He}/^{4}\text{He}$, which might be explained by radiogenic ingrowth of $^{3}\text{He}$. 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).
..., Fig. 5 shows there is no relationship between \(^3\text{He}/\text{He}^4\text{He}\) and \(^{187}\text{Os}/^{188}\text{Os}\) in the dataset. The samples with paired \(^3\text{He}/\text{He}^4\text{He}\) 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 \(^3\text{He}/\text{He}^4\text{He}\). Additionally, there is no relationship between \(^3\text{He}/\text{He}^4\text{He}\) 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 \(^3\text{He}/\text{He}^4\text{He}\) (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 (\(\text{Os} \leq 0.053\) ppm). 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 Savai’i xenoliths are unusual.

Relatively low Os concentrations (down to 0.068 ppm) in OJP peridotite xenoliths were identified by Ishikawa et al. (2011). Ishikawa et al. (2011) suggested the low Os concentrations may result from the breakdown of sulfide, which is the primary host of Os in peridotites (Hart and Ravizza, 1996; Burton et al., 1999; Alard et al., 2000). The Savai’i xenoliths are relatively fresh, so breakdown by surface weathering processes is unlikely. The breakdown of sulfide by devolatilization during eruption and exposure to oxygenated conditions (Handler et al., 1999) was also considered by Ishikawa et al. (2011). We consider such an explanation unlikely in the Samoa suite because the mechanism of sulfide breakdown should affect all the xenoliths, but only two Savai’i xenoliths exhibit unusually low Os concentrations. One possible hypothesis is that sulfides were exhausted by extensive melt extraction from the peridotites, leaving the Os highly depleted in these samples. However, Os concentrations do not exhibit clear correlations with indicators of melt extraction (e.g., modal abundance of spinel), and the xenoliths with unusually low Os concentrations do not appear to be more refractory than the other Savai’i xenoliths. Another possible explanation for the low Os concentrations is percolation of sulfur-undersaturated melts through the peridotites, which would mobilize sulfide and reduce the Os concentrations (Reisberg et al., 2005). However, there is no relationship between Os concentrations and indicators of metasomatic enrichment (e.g., La/Yb in xenolith clinopyroxene), which argues against this hypothesis. Another possibility is that the xenoliths’ low Os content is an expression of the nugget effect on a large scale. In other words, the heterogeneous distribution of Os-carrying phases is on a scale much larger than the small xenolith carried by the alkali basalt host. The variability of the heterogeneity increases with lower Os concentrations because the probability of sampling representative abundances of the Os-carrying phases decreases. 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-I-1, Sav-I-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 \(^{87}\text{Sr}/^{86}\text{Sr} = 0.710969–0.712838; \ ^{143}\text{Nd}/^{144}\text{Nd} = 0.512467–0.512516; \ ^{206}\text{Pb}/^{204}\text{Pb} = 18.845–18.877)\) are more geochemically enriched than the rejuvenated lavas from Savai’i that host the xenoliths (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., 2007a,b). The Sr, Nd and Pb isotopic compositions of the Tubuai xenoliths \(^{87}\text{Sr}/^{86}\text{Sr} = 0.702796–0.702755; \ ^{143}\text{Nd}/^{144}\text{Nd} = 0.512902–0.512942; \ ^{208}\text{Pb}/^{204}\text{Pb} = 20.971–21.313)\) overlap with basaltic lavas from the volcano (Vidal et al., 1984; Chauvel et al., 1992; Kogiso et al., 1997; Schiano et al., 2001; Hanyu et al., 2011a, 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 basaltic lavas 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 Sr, Nd and Pb isotopic composition of the two most incompatible element 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 the Os-isotopic composition of the xenoliths 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 CO2-rich melts associated with the plume overprinted 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 \((0.1233;\text{Hauri and Hart, 1993; Workman et al., 2004})\) overlaps with the \(^{187}\text{Os}/^{188}\text{Os}\) xenoliths from this locality \((0.1173–0.1284)\).
(Fig. 2). The Savai’i xenoliths are hosted in rejuvenated lavas, and rejuvenated lavas from Savai’i have the lowest 187Os/188Os (down to 0.1233) among Samoan hotspot lavas (for comparison, the lowest 187Os/188Os in Samoan shield lavas is 0.1268; Jackson and Shirey, 2011). Indeed, Savai’i rejuvenated lavas exhibit some of the lowest 187Os/188Os in the global OIB (ocean island basalts) 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 187Os/188Os) might contribute to the low 187Os/188Os in rejuvenated Savai’i lavas. While unradiogenic Os in oceanic hotspots is not always a result of the incorporation of xenolith material with low 187Os/188Os (Schaefer et al., 2002), this mechanism has been suggested to reduce the 187Os/188Os of the host lava at the Azores, Comoros and Canary hotspots (Widom and Shirey, 1996; Widom et al., 1999; Class et al., 2009; Day et al., 2010). We note that half of the 14 peridotite xenoliths in the Savai’i dataset have 187Os/188Os ratios lower than the lowest ratio identified in lavas from the Samoan hotspot. Therefore, even if the low 187Os/188Os 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 187Os/188Os 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 187Os/188Os overlapping with the range observed in Samoan lavas may also sample an Os-isotopic domain distinct from that sampled by the lavas, because the 187Os/188Os of the Savai’i rejuvenated lavas may have been reduced by incorporation of xenolith material with highly unradiogenic 187Os/188Os.

4.3. Melt–rock reaction in the Tubuai and Savai’i xenoliths

Extensive carbonate metasomatism related to the Samoa hotspot (which sources Savai’i’s volcanism) and the 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 187Os/188Os 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 187Os/188Os, but none is observed. Instead, xenoliths with the weakest metasomatic signatures exhibit both high and low 187Os/188Os (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 (0.0047–0.0145 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 metasomatic melts, the data do not support recent metasomatic modification of the 187Os/188Os 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 metasomatism (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 (Parman et al., 2005; Heber et al., 2007; Jackson et al., 2013). Therefore, the lack of correlation between 3He/4He and 187Os/188Os is not surprising, as other incompatible trace elements fail to show correlations with 187Os/188Os (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. Nonetheless, within the population of xenoliths from each island, there is no simple relationship between 3He/4He and trace element signatures.

In the xenoliths from Savai’i and Tubuai, magma–melt 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–melt 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–melt 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_M).

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, where the magma overwheels 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 and Kurz, 1997). 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. 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 REE 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 in the xenolith, 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 isotopic composition of the oceanic lithosphere that was formed at a mid-ocean ridge.

Another possible explanation for the lack of relationship between 3He/4He 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 (Jackson et al., 2010a, b; Konter and Jackson, 2012). Multicomponent mixing caused by multiple pulses of metasomatic fluids with different 3He/4He 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 3He/4He and 187Os/188Os and trace elements in the Tubuai xenoliths. It is not clear whether the non-atmospheric 129Xe/130Xe reported in a subset of Savai’i and Tubuai peridotite xenoliths by their respective hotspot magmas, while largely retaining the Os isotopic composition of the oceanic lithosphere that was formed at a mid-ocean ridge.

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 187Os/188Os in peridotite xenoliths from Hawaii (down to 0.1138) that are lower than the lowest 187Os/188Os ratios identified in lavas from the hotspot. Similarly, the low 187Os/188Os 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 187Os/188Os 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 187Os/188Os mantle domain sampled by the xenoliths?

4.4. Origin of the low 187Os/188Os signature in Tubuai and Savai’i xenoliths

4.4.1. Subcontinental lithospheric mantle as the origin of the low 187Os/188Os in Samoa and Tubuai xenoliths?

The relatively unradiogenic 187Os/188Os 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 187Os/188Os, 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 187Os/188Os (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 187Os/188Os Os compelling at Kerguelen and perhaps suggestive at other oceanic localities (e.g., Coltorti et al., 2010; Debaillie 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 Savai’i and Tubuai, 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 187Os/188Os observed in peridotite xenoliths from the Hawaiian hotspot than other depleted mantle reservoirs. Depleted mantle domains in the convecting mantle that may host low 187Os/188Os 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 187Os/188Os observed in Savai’i and Tubuai xenoliths becomes the essential consideration.
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 (Bizimis et al., 2007), 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 attached to the base of the oceanic lithosphere that were entrained in upwelling hotspot lavas (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 $^{187}\text{Os}/^{188}\text{Os}$ (0.1152–0.1270) that overlaps considerably with the shallow xenolith population (0.1161–0.1388; $n = 60$). 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 $^{187}\text{Os}/^{188}\text{Os}$ ratios 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 Sava’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 have 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 peridotite xenoliths sampled by Tubuai and Sava’i. Lassiter et al. (2014) noted that, globally, the peridotite xenolith population hosted in oceanic hotspot lavas (average of 0.1244 ± 0.0043, 1SD) 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 Sava’i (0.1237 ± 0.0032, 1SD) 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 $^{186}\text{Os}/^{188}\text{Os}$ (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 peridotite xenoliths. The similarity in Os isotopic composition between the peridotite xenoliths in this study, other peridotite xenoliths at hotspots globally, 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 $^{186}\text{Os}/^{188}\text{Os}$ 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, 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 Earth’s history. $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{176}\text{Hf}/^{177}\text{Hf}$ isotopes in Archean crustal rocks (e.g., Shirey and Hanson, 1986; Stein and Hofmann, 1994; Moorboth 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 Albare`de, 2008; Kemp et al., 2010; Guiterau 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 $T_{\text{RD}}$—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 $^{187}\text{Os}/^{188}\text{Os}$ (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 can exhibit lower $^{187}\text{Os}/^{188}\text{Os}$ than chondritic compositions, 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 $T_{\text{RD}}$ 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 (Van Kranendonk, 2010; Shirey and Richardson, 2011). 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, $T_{\text{RD}}$ model ages in ocean island xenoliths and abyssal peridotites provide little record of mantle depletion before ~2 Ga, and give average model ages of 0.8 Ga and 0.7 Ga, respectively. 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 $T_{\text{MA}}$ (mantle extraction model age; Shirey and Walker, 1998) and $T_{\text{Azo}}$ (“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 $T_{\text{RD}}$ 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 hartzburgitic restite that comprises recycled oceanic mantle lithosphere (Lasslier 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 

Different isotopes have different behaviors in the Earth's mantle. For example, xenon is more resistant to convective mixing than Os and Nd. 129Xe is the decay product of the short-lived isotope 129I (T_{1/2} = 15.7 Ma), and all 129Xe/130Xe variability in the Earth was generated by fractionation of I from Xe during the first 100 million years of Earth's history. Early work on mantle-derived samples showed that the 128Xe/130Xe composition of the mantle is different from the atmospheric composition (Staudacher and Allegre, 1982; Allegre et al., 1983; Poreda and Farley, 1992). A recent series of papers have demonstrated that heterogeneous 129Xe/130Xe signatures are preserved in the modern oceanic mantle and sampled at mid-ocean ridges and plume-influenced lavas from Iceland and Rocheambeau Rift (Tucker et al., 2012; Parai et al., 2012; Mukhopadhyay, 2012; Petö et al., 2013; Parai and Mukhopadhyay, 2015). This 129Xe/130Xe heterogeneity in the oceanic mantle was generated in the early Hadean, and has survived for 4.5 Ga in the convecting mantle to be sampled by recent MORB and OIB volcanism. Therefore, it is critical to understand how the early-formed 187Os/188Os and 142Nd/144Nd isotopic record of mantle depletion was erased while 129Xe/130Xe heterogeneities from the earliest Hadean persist in the modern oceanic mantle. Mukhopadhyay (2012) suggested that the erasure of Hadean 142Nd/144Nd signatures, but not Hadean 129Xe/130Xe, from the convecting mantle is due to the small magnitude of the 142Nd/144Nd anomalies (<0.004% variability in 142Nd/144Nd) has been observed in terrestrial Archean terraines; e.g., Caro et al., 2006; O’Neil et al., 2008; Rizo et al., 2013) compared to the large range of 129Xe/130Xe signatures (10% variability is observed in the modern mantle; Mukhopadhyay, 2012), and the smaller 142Nd/144Nd anomalies are more easily erased than the larger magnitude 129Xe/130Xe heterogeneities. One problem with using a “small magnitude anomaly” explanation for the Os data is that Os-isotopic heterogeneities in the mantle are large, yet early-formed Os-isotopic signatures still have been erased. For example, the 187Os/188Os of early-formed depleted mantle at 4.55 Ga was ~0.095, which is >25% lower than the modern mantle (0.1296; Meisel et al., 2001). Therefore, the absence of early-formed, large magnitude 187Os/188Os excursions in the modern convecting mantle is evidence that the preservation of early geochemical heterogeneities is not necessarily a function of the magnitude of the original geochemical anomaly. However, we cannot exclude the possibility that 142Nd/144Nd heterogeneity exists in the modern mantle, but it is not detected because either (1) existing instrumentation does not yet provide the necessary precision to resolve such anomalies or (2) lavas sampling the modern mantle that have resolvable 142Nd/144Nd anomalies have not been analyzed. Nonetheless, given the existing dataset, the 142Nd/144Nd in the modern mantle is homogeneous.

The variable fluxes of trace elements between different terrestrial geochemical reservoirs—including the crust, mantle and atmosphere—must also be an important consideration when evaluating the survival of Hadean isotopic signatures in the modern mantle (e.g., Galer and O’Nions, 1985; Albarède, 2001; Kellogg et al., 2002). Mukhopadhyay (2012) proposed an alternative explanation for the preservation of Hadean 129Xe/130Xe, but not 142Nd/144Nd, isotopic signatures in the modern mantle, and this model may explain the lack of abundant, early-formed Os-isotopic reservoirs in the mantle as well. Mukhopadhyay (2012) argued that the lower recycling efficiency of xenon into the mantle compared to Nd helps preserve early-formed Xe-isotopic signatures. The mantle degasses xenon to the Earth’s atmosphere during mantle
melting, and noble gases are not efficiently returned to the mantle during recycling (e.g., Staudacher and Allegre, 1988). As a result, during partial melting at ridges followed by lithospheric return to the mantle, most of the mantle xenon enters the atmosphere. A small portion of the xenon is thought to be returned to the mantle during recycling (Porcelli and Wasserburg, 1995; Holland and Ballentine, 2006; Holland et al., 2009; Kendrick et al., 2011; Moreira, 2013). However, xenon is not as efficiently recycled back into the deep Earth as Nd and Os (Mukhopadhyay, 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 85–92% 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 $^{129}$Xe/$^{130}$Xe signatures: much of the remaining Xe is primitive mantle Xe (even 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 is recycled are poorly constrained, but this fraction must exceed the fraction of Xe in the mantle that is recycled (i.e., >85–92%) 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). Therefore, 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. This is likely a minimum value: if oceanic lithosphere formation and subduction operated at higher rates in the geologic past, the fraction of the Nd in the mantle that is recycled exceeds 80% (and it is possible that nearly the entire Nd budget of the mantle may be recycled). Thus, Hadean $^{142}$Nd signatures in the mantle are more efficiently overprinted by recycled material than Hadean $^{129}$Xe signatures.

Like Nd, Os is also efficiently returned to the mantle during recycling. While Os is almost certainly mobile in fluids and some can be lost from the oceanic crustal portion of a downgoing slab, little Os resides in the oceanic crust. Instead, Os is compatible and chalcophile and >99% of the Os in the oceanic lithosphere (assuming <100 ppt Os in the oceanic crustal lithosphere and 3000 ppt Os in the depleted mantle lithosphere) is retained in the depleted oceanic mantle lithosphere portion of the slab, and is thus returned to the mantle during recycling. If the concentrations of osmium in the convecting mantle and oceanic mantle lithosphere are similar, and if the mantle is comprised of 80% recycled mantle lithosphere, then ~80% of the osmium in the convecting mantle has been recycled and thus has passed through the plate tectonic cycle as least once since the onset of plate tectonics. Again, this is a minimum value, and the fraction of Os in the convecting mantle that is recycled may approach 100%. Thus, early-formed, Hadean $^{187}$Os/$^{188}$Os signatures in the mantle have been more efficiently overprinted by recycled material than Hadean $^{129}$Xe signatures.

Ultimately, the isotopic evolution of different radiogenic isotopic systems in the mantle depends on the relative recycling efficiencies of parent and daughter elements. However, unlike the $^{146}$Sm–$^{142}$Nd and the $^{187}$Re–$^{129}$Xe systems, where the parent isotopes have been effectively extinct since the Hadean, the parent in the $^{187}$Re–$^{187}$Os system, $^{187}$Re, is still decaying and thus can modify the $^{187}$Os/$^{188}$Os signature of the modern mantle. Like Os, Re is also efficiently returned to the mantle during recycling of oceanic crust, and any Re returned to the mantle will serve to overprint ancient (low) $^{187}$Os/$^{188}$Os reservoirs via generation of radiogenic

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Fig. 6. The differing recycling efficiencies of the elements may 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 elements are more likely to be preserved in the modern convecting mantle.
$^{187}$Os in the mantle over time. Thus, long-term recycling of $^{188}$Re also plays a supporting role with Os recycling in overprinting ancient (low) $^{187}$Os/$^{188}$Os domains in the mantle.

The long-term preservation of early-formed isotopic signatures of a particular element is a function of not just mantle convective stirring, but also the recycling efficiency of the element back into the mantle at subduction zones (Gonnermann and Mukhopadhyay, 2009; Mukhopadhyay, 2012). This conceptual model permits a more general description of which elements are (e.g., Xe), and are not (e.g., Nd and Os), likely to preserve Hadean geochemical signatures in the Earth’s mantle: incompatible elements that are fluid mobile are more likely to preserve Hadean geochemical signatures in the mantle than compatible elements and fluid immobile incompatible elements. Incompatible, fluid-mobile elements (like Xe) are efficiently extracted from the downgoing slab and accumulate in shallow reservoirs (e.g., the crust and atmosphere) instead of being returned to the mantle. Such elements are less efficiently returned to the mantle than Os and Nd, and are therefore more likely to preserve early-formed, primordial signatures in the mantle. The incompatible, fluid-mobile element need not be volatile to preserve Hadean geochemical signatures in the mantle: incompatible, fluid-mobile elements that are refractory also will be extracted from the downgoing slab, concentrated in the continental crust, and not efficiently returned to the mantle. For 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 $^{182}$W/$^{184}$W variability in 2.8 Ga komatites (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 $^3$He/$^4$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, $^{143}$Sm/$^{144}$Nd, $^{176}$Lu/$^{176}$Hf and $^{208}$Pb/$^{206}$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 even for elements (e.g., Pb) that may be efficiently recycled into the mantle (Jackson et al., 2010a,b; Jackson and Carlson, 2011).

ACKNOWLEDGEMENTS

We acknowledge discussion with Roberta Rudnick, Sujay Mukhopadhyay, Bernhard Peucker-Ehrenbrink, Andrew Reinhard and Jasper Konter. We thank Alison Price for help improving a figure. MGJ thanks Vincent Salters for asking why xenon-isotopic heterogeneities 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-1232983.
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Please cite this article in press as: Jackson M. G. et al. 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. Geochim. Cosmochim. Acta (2016), http://dx.doi.org/10.1016/j.gca.2016.02.011


Associate editor: Shichun Huang