A radiogenic isotopic (He-Sr-Nd-Pb-Os) study of lavas from the Pitcairn hotspot: Implications for the origin of EM-1 (enriched mantle 1)

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Abstract

We present new He-Sr-Nd-Pb-Os isotopic compositions and major and trace-element concentrations for ten subaerially-erupted lavas and one seamount lava associated with the Pitcairn hotspot. The most geochemically-enriched lavas at the Pitcairn hotspot have signatures that are consistent with recycled sediments derived from upper continental crust. Pitcairn lavas have elevated Ti, which also supports the presence of a mafic protolith in the Pitcairn mantle. A subset of Pitcairn seamount samples, including the seamount sample presented here, are tholeiitic. Tholeiitic lavas are uncommon at ocean hotspots located far from mid-ocean ridges. Like tholeiites erupted at Hawaii, the presence of tholeiites in the Pitcairn magmatic suite can be

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explained by melting a silica-saturated recycled mafic component in the Pitcairn mantle source. We also present the highest $^3\text{He}/^4\text{He}$ ratio (12.6 Ra, ratio to atmosphere) from the Pitcairn hotspot. This sample anchors the high $^{206}\text{Pb}/^{204}\text{Pb}$ portion of the Pitcairn array and provides evidence for a plume component in the Pitcairn mantle. In contrast, Pitcairn lavas that have the lowest $^{206}\text{Pb}/^{204}\text{Pb}$ are the most geochemically enriched, and have the highest $^{87}\text{Sr}/^{86}\text{Sr}$ and lowest $^{143}\text{Nd}/^{144}\text{Nd}$ in the Pitcairn suite; these EM-1 end-member lavas have MORB-like $^3\text{He}/^4\text{He}$ (~ 8 Ra, ratio to atmosphere). Recycled oceanic crust and sediment suggested to be in the Pitcairn EM-1 mantle are expected to have low $^3\text{He}/^4\text{He}$ (< 0.1 Ra). Therefore, the higher, MORB-like $^3\text{He}/^4\text{He}$ in Pitcairn EM-1 lavas is paradoxical, but might be explained by diffusive exchange of helium, but not the heavy radiogenic isotopes, with the ambient mantle over billion-year timescales.

Keywords: Pitcairn, EM-1, sediment recycling, pelagic clays, crustal recycling, tholeiite, FOZO, $^3\text{He}/^4\text{He}$, ocean island basalt (OIB), hotspot, Hawaii

1. Introduction

Geochemical analyses of ocean island basalts (OIB) erupted at hotspots reveal that the Earth's mantle is compositionally heterogeneous (e.g., Gast et al., 1964; Zindler and Hart, 1986; Hofmann, 1997, 2003; White, 2010; Stracke, 2012). Compilations of radiogenic isotopic measurements made on OIB show that several distinct isotopic compositions emerge from the global dataset (Zindler and Hart, 1986), including EM-1 (enriched mantle 1), which is characterized by the unradiogenic $^{206}\text{Pb}/^{204}\text{Pb}$ ratios with moderately high

²⁰⁸Pb/²⁰⁴Pb and ⁸⁷Sr/⁸⁶Sr, EM-2 (enriched mantle 2) which is character-
 ized by intermediate ²⁰⁶Pb/²⁰⁴Pb ratios and highly radiogenic ⁸⁷Sr/⁸⁶Sr, and
 HIMU (high μ , or high ²³⁸U/²⁰⁴Pb), which is characterized by the the most
 radiogenic Pb isotopic compositions in the oceanic mantle. Geochemically-
 depleted compositions are also sampled by hotspots, and the origin of the
 depleted component in plumes is not well understood and may host elevated
 (primordial) ³He/⁴He ratios (Hart et al., 1992; Farley et al., 1992; Hanan
 and Graham, 1996). Mixing of the EM-1, EM-2 and HIMU end-members
 with the depleted composition is thought to generate much of the diversity
 in OIB.

While the field of mantle geochemistry has advanced to a state of careful
 description and classification of the various mantle species, and their pos-
 sible mixing relationships, the origin of these species still remains poorly
 understood. Models explaining the sources of the mantle end-members often
 call for subduction injection of surface materials, including continental crust,
 oceanic crust, and sediments into the mantle over geologic time (Hofmann
 and White, 1982). Following mixing with ambient mantle, the subducted
 protoliths are sampled by regions of upwelling mantle, called plumes, and
 melted beneath hotspots. Alternative hypotheses for the origin of the various
 end-members include delamination of subcontinental lithosphere (McKenzie
 and O’Nions, 1983; Mahoney et al., 1991), CO₂-flux induced melting of pris-
 tine mantle domain in the lower mantle (Collerson et al., 2010), and more
 exotic metasomatic processes, where fluids in the mantle impart enriched
 geochemical signatures on specific domains of the mantle (e.g., Menzies and
 Murthy, 1980; Menzies, 1983; Niu and O’Hara, 2003; Workman et al., 2004;

34 Geldmacher et al., 2008; Salters and Sachi-Kocher, 2010; Pilet et al., 2011).
 35 However, a consensus is emerging for the origin of EM-2 as recycled marine
 36 sediment (Workman et al., 2008) with a terrigenous origin (Jackson et al.,
 37 2007a). Similarly, the long-standing hypothesis that HIMU represents re-
 38 cycled oceanic crust (Hofmann and White, 1982; Zindler et al., 1982; Eiler
 39 et al., 1997) is gaining strength (Hanyu et al., 2011; Kawabata et al., 2011;
 40 Cabral et al., 2013; Hanyu et al., 2014).

41 Among all mantle end-members, no consensus has emerged for the origin
 42 of EM-1 (e.g., Hofmann, 1997; Hart, 2011) and yet numerous mechanisms
 43 have been proposed (e.g., Woodhead and McCulloch, 1989; Weaver, 1991;
 44 Mahoney et al., 1991; Chauvel et al., 1992; Woodhead and Devey, 1993;
 45 Eiler et al., 1995; Rehkämper and Hofmann, 1997; Lassiter and Hauri, 1998;
 46 Blichert-Toft et al., 1999a; Gasperini et al., 2000; Tatsumi, 2000; Eisele et al.,
 47 2002; Escrig et al., 2004; Hanan et al., 2004; Honda and Woodhead, 2005;
 48 Brandenburg et al., 2008; Geldmacher et al., 2008; Salters and Sachi-Kocher,
 49 2010; Collerson et al., 2010; Willbold and Stracke, 2010). Here, armed with a
 50 new geochemical dataset from the Pitcairn hotspot, we argue for the presence
 51 of recycled sediment in the Pitcairn EM-1 mantle (e.g., Eisele et al., 2002;
 52 Honda and Woodhead, 2005), and we suggest that Pitcairn lavas also sample
 53 recycled oceanic crust. We also report new helium data on Pitcairn lavas,
 54 including the highest $^3\text{He}/^4\text{He}$ ratio from Pitcairn, and the new data provide
 55 insights into the evolution of the poorly-understood high $^3\text{He}/^4\text{He}$ mantle
 56 and EM-1 mantle reservoirs.

57 2. Methods

58 The ten subaerial lavas from Pitcairn island reported here were collected
59 by Ken Farley during the 1989 Helios Expedition. We do not report ages
60 for the samples in this study, but subaerial lavas from Pitcairn have ages
61 between 0.4 and 1.0 Ma (Duncan et al., 1974). Figure 1 shows locations for
62 the subaerial samples presented in this study. Based on comparison with
63 the geological formations as mapped by Carter (1967), the samples in this
64 study were collected from the Tedside (Pit-1, Pit-3, Pit-4, Pit-7, Pit-8) and
65 Pulawana volcanic formations (Pit-6, Pit-16, Pit-11, Pit-12, Pit-13). Samples
66 Pit-6, 12, 13 and 16 were collected in place from outcrops. Samples Pit-1, 3,
67 4a, 7, 8 and 11 were collected as cobbles on the wave cut terrace; it is very
68 likely that cobbles on the terrace derive from the overlying cliff as there are
69 no rivers to transport cobbles and longshore transport is not plausible.

70 We also report new geochemical data on a single submarine sample, 65DS-
71 4, dredged between 1503 and 1276 m water depth on Adams Seamount
72 ($25^{\circ}21'26''$, $125^{\circ}18'59''$), located ~ 90 km ESE of Pitcairn. This sample
73 was recovered aboard the FS Sonne on December 1989 on the 65th cruise of
74 this vessel (Stoffers et al., 1990; Woodhead and Devey, 1993). The seafloor
75 extending ~ 90 km ESE of Pitcairn Island is populated by a submarine vol-
76 canic field hosting approximately 90 volcanic cones and seamounts, includ-
77 ing Adams Seamount (Hekinian et al., 2003), the largest seamount in this
78 volcanic field (Woodhead and Devey, 1993; Hekinian et al., 2003). Several
79 of these volcanic features exhibit evidence of being active, and submarine
80 dredges returned samples with fresh volcanic rocks and glass; there was little
81 evidence for sediment cover (Stoffers et al., 1990). Seafloor backscatter im-

82 ages revealed that several of the volcanic edifices are highly reflective, further
 83 supporting a young age for these seamounts, and substantiating the hypoth-
 84 esis that that these seamounts (including Adams Seamount) mark the active
 85 portion of the Pitcairn hotspots (Woodhead and McCulloch, 1989; Stoffers
 86 et al., 1990; Hekinian et al., 2003). K/Ar age determinations of lavas recov-
 87 ered from Adams Seamount range from 2 to 7 kyr (Guillou et al., 1997) and
 88 volcanic rocks recovered from this seamount exhibit evidence of only slight
 89 alteration (Hekinian et al., 2003). Additionally, geochemical data obtained
 90 on seven dredges of Adams Seamount, including new data presented here on
 91 sample 65DS-4, indicate that this seamount has a Pitcairn hotspot pedigree
 92 (Woodhead and Devey, 1993). Given this significant body of evidence, we
 93 adopt the finding that Adams Seamount marks the active end of the Pitcairn
 94 hotspot.

95 Cut rock slabs were crushed in plastic bags to avoid contamination with
 96 metal from the hammer, and the freshest chips were separated and powdered
 97 in agate. With the exception of sample 65DS-4, major elements and a subset
 98 of trace elements were measured on powders by XRF in the laboratory of
 99 Michael Rhodes at the University of Massachusetts Amherst, and the meth-
 100 ods and associated errors are reported in Rhodes (1988). The remaining trace
 101 elements were measured by ICP-MS at the Geoanalytical Lab at Washington
 102 State University (WSU) (Knaack et al., 1994). For sample 65DS-4, major
 103 and trace elements were measured at WSU by XRF and ICP-MS (analyti-
 104 cal methods and errors are reported in Johnson et al., 1999; Knaack et al.,
 105 1994). At WSU, precision for the analyses of SiO_2 , Al_2O_3 , TiO_2 and P_2O_5
 106 in basalts is 0.11–0.33% (1σ) of the amount present, and 0.38–0.71% (1σ)

107 of the amount present for FeO, MgO, CaO, Na₂O, MnO and K₂O (Johnson
 108 et al., 1999). Trace element analyses by ICP-MS at WSU have a precision of
 109 0.77–3.2% (1 σ) for all elements except for Th (9.5%) and U (9.3%) (Knaack
 110 et al., 1994). Sample 65DS-4 is an olivine cumulate, and we present XRF
 111 major element data on the whole rock and on (olivine-free) matrix separated
 112 from the sample.

113 For the subaerial lavas, chemical separation protocols and Sr and Nd
 114 isotopic measurements were made at the Woods Hole Oceanographic Insti-
 115 tution and follow the same protocols outlined in Hauri and Hart (1993); Sr
 116 and Nd isotopic data have never been reported for these samples. Sr-isotopic
 117 compositions were fractionation corrected relative to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$, and
 118 $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were adjusted to an SRM987 value of 0.710240. The exter-
 119 nal precision for the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio is estimated to be 45 ppm, and is based
 120 on repeat analyses of the SRM987 Sr standard. Nd-isotopic compositions
 121 were fractionation corrected relative to a $^{146}\text{Nd}/^{150}\text{Nd}$ value of 0.7219. The
 122 $^{143}\text{Nd}/^{144}\text{Nd}$ ratios were adjusted to a value of 0.511847 for the La Jolla Nd
 123 standard. The external precision on the $^{143}\text{Nd}/^{144}\text{Nd}$ is estimated to be 40
 124 ppm, and is based on repeat measurements of the La Jolla Nd standard. Pb
 125 isotopic data for the subaerial samples were previously reported for all sam-
 126 ples in Eiler et al. (1995), except for sample Pit-6, which is reported here;
 127 analytical procedures and fractionation correction are the same as outlined in
 128 Hauri and Hart (1993), and Pb-isotopic ratios for sample Pit-6 were adjusted
 129 to SRM981 values from (Todt et al., 1996) to be consistent with the data
 130 reported in Eiler et al. (1995). Sr, Nd and Pb chemical separation protocols
 131 and isotopic analyses for the submarine lava, 65DS-4, follow those outlined

132 in Hart and Blusztajn (2006); Sr, Nd and Pb isotopic measurements were
133 made on the Neptune multi-collector ICP-MS at WHOI. After adjusting to
134 0.710240 and 0.511847 for SRM987 and La Jolla Nd standards, respectively,
135 the external precision is estimated to be between 15-25 ppm. Pb-isotopic
136 measurements were adjusted to the SRM 981 values reported in (Todt et al.,
137 1996), and external reproducibility is estimated to be better than 120 ppm
138 for the $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios.

139 Osmium isotopic analyses were performed on 315 to 820 mg of olivine
140 separated from six Pitcairn samples. Samples were dissolved by HF-HCl
141 dissolution, subjected to a gentle dry down at $<100^\circ\text{C}$, then loaded in Car-
142 ius Tubes as described in Shirey and Walker (1995); chemical separation of
143 Os is as described in Shirey and Walker (1995). The isotopic analyses were
144 made at DTM by negative thermal ionization mass spectrometry (N-TIMS)
145 (Hauri and Kurz, 1997). In-run precision was better than 0.1% (2σ standard
146 error of the mean) on the $^{187}\text{Os}/^{188}\text{Os}$ ratio for all runs. Total procedural
147 blanks were <5 pg Os and were within analytical errors. Analyses of the
148 DTM osmium standard yielded an average $^{187}\text{Os}/^{188}\text{Os}$ of 0.11378 ± 10 (2σ ,
149 $n=5$). Helium isotopic measurements were made on 0.9 to 3.8 g of olivine
150 phenocrysts by crushing in vacuo and follow the methods outlined in Farley
151 et al. (1993). Uncertainties on the $^3\text{He}/^4\text{He}$ ratios are estimated to be ± 0.5
152 Ra; the estimated uncertainties account for both blank correction and uncer-
153 tainty in standards of similar size to samples. The sample with the highest
154 $^3\text{He}/^4\text{He}$, Pit-16, was measured twice on more than 4 g of olivine, on coarse
155 (11.8 Ra) and fine (13.3 Ra) olivine fractions (the average value of the two
156 measurements, 12.6 Ra, is used hereafter), to verify the elevated $^3\text{He}/^4\text{He}$

ratio; the raw ratios were 10.5 and 11.5 Ra, so even without blank correction the sample is unusual. This sample was collected on a sea-cliff, where erosion enhanced by continuous wave action precludes long-term exposure at the surface, which enables us to exclude a cosmogenic helium origin for the elevated isotopic ratio in this sample.

The new isotopic data are presented in Table 1, together with previously published data on these samples. Additionally, in Table 2 we report major element abundances for olivine phenocrysts from submarine sample 65DS-4; these data complement previously-published olivine compositional data on subaerially-erupted lavas from the hotspot (Eiler et al., 1995). Microprobe analyses were made on Cameca SX-100 at UC Santa Barbara with the following operating conditions: 15 kV accelerating voltage, 15 nA beam current and 1 μm beam diameter.

3. Data

All of the samples in this study are porphyritic in texture. According to a TAS classification (Figure 2), subaerial Pitcairn lavas are predominantly alkalic and span the whole suite of compositions from alkaline basalts to trachy-andesites. Most of the submarine lavas are also alkalic, but a limited number of seamount samples cross the alkali-tholeiite division of Macdonald and Katsura (1964) and are tholeiitic, as was noted by Hekinian et al. (2003). The new seamount sample, 65DS-4, is clearly tholeiitic, based on its whole-rock and matrix major element composition.

Figure 3 shows variation diagrams of MgO versus major element oxides. Olivine fractionation dominates the major element variations for bulk rock

181 compositions $> 8\text{-}10$ wt.% MgO. At $\text{MgO} < 8\text{-}10$ wt.%, clinopyroxene frac-
 182 tionation is apparent from reduced $\text{CaO}/\text{Al}_2\text{O}_3$ ratios. Fractionation of Fe-
 183 and Ti-rich phases is evident in lavas with $\text{MgO} < 5$ wt.%.

184 The panels in Figure 4 display isotopic compositions for the new Pit-
 185 cairn hotspot dataset compared to previously published Pitcairn data and
 186 other OIB. Separate fields are designated for Samoa, which represents the
 187 extreme EM-2 end-member, and for Hawaii. A separate distinction within
 188 the Hawaiian field is made for Koolau lavas because these lavas host an
 189 enriched end-member that has an EM-1 pedigree. However, Pitcairn and
 190 Hawaiian Koolau lavas have distinct EM-1 flavors. In both, $^{208}\text{Pb}/^{204}\text{Pb}$
 191 versus $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams, Pitcairn
 192 and Hawaii form two separate fields, and highlight the difference in the EM-
 193 1 composition between the two hotspots. However, in $^{143}\text{Nd}/^{144}\text{Nd}$ versus
 194 $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic space, Pitcairn overlaps with Hawaiian Koolau lavas, high-
 195 lighting the complementary EM-1 signatures present at both hotspots, in
 196 spite of their different Pb-isotopic compositions. The plot of $^{208}\text{Pb}/^{204}\text{Pb}$ ver-
 197 sus $^{206}\text{Pb}/^{204}\text{Pb}$ reveals the presence of a third component in Pitcairn lavas
 198 (located at the lowest vertex of the triangle outlining the Pitcairn dataset
 199 in Pb-isotope space in Figure 4), and this component is characterized by
 200 lower $^{208}\text{Pb}/^{204}\text{Pb}$ at a given $^{206}\text{Pb}/^{204}\text{Pb}$. Curiously, this third component
 201 is only visible in $^{208}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ isotope space, and is not
 202 evident in other isotope projections. Sample 65DS-4 defines the end-member
 203 composition for this new component, and it plots closer to Hawaiian Koolau
 204 lavas in Pb-isotopic space than any previously examined Pitcairn lava. We
 205 emphasize that a subset of prior analyses of lavas from the Pitcairn hotspot

206 have Sr, Nd and Pb isotopic compositions that are similar to 65DS-4, but
 207 this sample slightly expands the field of the Pitcairn hotspot in $^{208}\text{Pb}/^{204}\text{Pb}$
 208 versus $^{206}\text{Pb}/^{204}\text{Pb}$ isotopic space.

209 Figure 5 shows $^3\text{He}/^4\text{He}$ ratios in olivine in subaerial samples (reported
 210 here for the first time) and seamount glasses (reported in Honda and Wood-
 211 head, 2005), plotted against $^{206}\text{Pb}/^{204}\text{Pb}$. At a given $^{206}\text{Pb}/^{204}\text{Pb}$ ratio, the
 212 $^3\text{He}/^4\text{He}$ of the olivines is bracketed by the $^3\text{He}/^4\text{He}$ of the glass dataset; only
 213 the Pit-16 olivine $^3\text{He}/^4\text{He}$ measurement is higher than any of the $^3\text{He}/^4\text{He}$
 214 measurements on glasses. The $^3\text{He}/^4\text{He}$ of the Pitcairn olivines presented here
 215 show no evidence for reduced $^3\text{He}/^4\text{He}$ by radiogenic ingrowth of ^4He . Fig-
 216 ure 5 also shows a possible relationship between $^{206}\text{Pb}/^{204}\text{Pb}$ and $^3\text{He}/^4\text{He}$,
 217 where the highest $^3\text{He}/^4\text{He}$ ratio is found in the sample with the highest
 218 $^{206}\text{Pb}/^{204}\text{Pb}$. In comparison, the two lavas with the lowest $^{206}\text{Pb}/^{204}\text{Pb}$ have
 219 the lowest $^3\text{He}/^4\text{He}$ ratios, 7.7 Ra (measured in a glass) and 8.0 Ra (measured
 220 in olivine), indicating that EM-1 lavas at Pitcairn have MORB-like $^3\text{He}/^4\text{He}$.

221 Figure 6 shows $^{187}\text{Os}/^{188}\text{Os}$ ratios and Os concentrations measured on
 222 olivines from samples from this study compared to the whole rock measure-
 223 ments of Pitcairn lavas from previous studies (Reisberg et al., 1993; Eisele
 224 et al., 2002). $^{187}\text{Os}/^{188}\text{Os}$ ratios in olivines from this study range from 0.1342
 225 to 0.1432, which falls within the range previously identified in whole-rock
 226 measurements of the most Os-rich Pitcairn samples (0.131 to 0.148 for sam-
 227 ples with Os > 50 pg/g) (see Figure 2b in Eisele et al., 2002). As observed
 228 by Eisele et al. (2002), there are no clear relationships between Os isotopic
 229 ratios and the other radiogenic isotopic compositions presented in this study.
 230 Rhenium was not measured on the olivines, so an age-correction cannot be

performed. However, recent studies on magmatic olivines from Samoa (Jackson and Shirey, 2011), Iceland (Debaille et al., 2009) and the Cook Islands (Hanyu et al., 2011) indicate that Re/Os ratios in magmatic olivines are low, and the subaerial lavas from Pitcairn are young, so any age-correction to the $^{187}\text{Os}/^{188}\text{Os}$ ratio of the olivine will be small.

Figure 7 displays SiO_2 data plotted against $^{206}\text{Pb}/^{204}\text{Pb}$ for Pitcairn, together with the global OIB dataset from Jackson and Dasgupta (2008) and a Hawaiian dataset from Jackson et al. (2012). With the exception of the matrix analysis from sample 65DS-4, only Pitcairn samples with $\text{MgO} > 9\%$ are plotted, and all lavas were corrected for olivine fractionation. The data show that, with Hawaii, Pitcairn lavas anchor the low $^{206}\text{Pb}/^{204}\text{Pb}$, SiO_2 -rich portion of the global OIB array.

4. Discussion

4.1. Recycled sediment in the source of Pitcairn EM-1 lavas

Previous work on the EM-1 signature hosted in Pitcairn lavas (Woodhead and McCulloch, 1989; Woodhead and Devey, 1993; Eisele et al., 2002; Honda and Woodhead, 2005) argued that the data are consistent with sediment addition to the mantle source: EM-1 Pitcairn lavas have high $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{187}\text{Os}/^{188}\text{Os}$, low $^{143}\text{Nd}/^{144}\text{Nd}$, and Nb-depletion ($\text{Nb}/\text{Nb}^* < 1$) (e.g. Eisele et al., 2002; Honda and Woodhead, 2005). Additionally, EM-1 lavas from Pitcairn tend to have low Ce/Pb ratios (e.g. Honda and Woodhead, 2005), which is a characteristic of marine sediments (e.g. Plank and Langmuir, 1998).

More recently, Willbold and Stracke (2010) argued for recycling of lower

continental crust to generate EM-1 signatures in oceanic lavas, which follows prior work suggesting a similar origin for the EM-1 mantle (Tatsumi, 2000; Escrig et al., 2004; Hanan et al., 2004; Willbold and Stracke, 2010). Willbold and Stracke (2010) relied on Eu/Eu^* to distinguish between upper and lower continental crust in the mantle source of Pitcairn lavas: Upper continental crust has negative Eu/Eu^* , and lower continental crust has positive Eu/Eu^* . Intriguingly, the Pitcairn lavas in Willbold and Stracke (2010) (refer to Figure 4 of said reference) show no clear Eu-anomaly, and most of the Pitcairn lavas in Willbold and Stracke (2010) actually have Eu/Eu^* values slightly less than one, which is consistent with the addition of recycled upper continental crust. While Eu/Eu^* may be useful for distinguishing a lower continental crust component in several EM-1 hotspots globally, it may not support a lower continental crust component at the Pitcairn hotspot, and permits a model whereby upper continental crust and derivative sediments exist in the Pitcairn mantle.

Sediments derived from upper continental crust have the necessary geochemical ingredients for generating an EM-1 mantle source sampled by Pitcairn lavas. However, upper continental crust and derivative sediments are highly heterogeneous (Plank and Langmuir, 1998; Rudnick and Gao, 2003). To better characterize the type of the crustal material that contributes to the EM-1 source, we point out two distinctive geochemical features of EM-1 as sampled by Pitcairn lavas: 1. Pitcairn lavas exhibit a range of highly unradiogenic $^{206}\text{Pb}/^{204}\text{Pb}$ ratios associated with relatively high $^{208}\text{Pb}/^{204}\text{Pb}$, and 2. The most geochemically-enriched Pitcairn lava with Hf-isotopic data departs from the global array formed by $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{176}\text{Hf}/^{177}\text{Hf}$ and

trends toward a component with high $^{176}\text{Hf}/^{177}\text{Hf}$ at a given $^{143}\text{Nd}/^{144}\text{Nd}$
 (Eisele et al., 2002). The former geochemical characteristic indicates that the
 protolith contributing to the enriched mantle source beneath Pitcairn must
 also have high Th/U (Eisele et al., 2002). Sediments can have high Th/U,
 and pelagic clays have among the highest Th/U, as they are the weathering
 products of mature upper continental crust (where U has been preferentially
 mobilized relative to Th over time) (Plank and Langmuir, 1998). Eisele
 et al. (2002) demonstrated that the Th/U ratio in pelagic sediments will
 generate Pb-isotopic systematics appropriate for Pitcairn hotspot lavas over
 time. Pelagic clays also satisfy the $^{176}\text{Hf}/^{177}\text{Hf}$ systematics in Pitcairn lavas
 since they do not contain detrital zircon, a Hf-rich phase. The presence of de-
 trital zircon in a sediment would yield lower Lu/Hf ratios at a given Sm/Nd
 ratio, and the departure of Pitcairn lavas from the mantle array suggests
 the presence of a zircon-poor pelagic sediment. Therefore, among possible
 subducted protoliths in the EM-1 mantle, pelagic sediments have the com-
 bination of high Th/U and high Lu/Hf required to generate the Nd, Hf and
 Pb isotopic compositions in Pitcairn EM-1 lavas. Therefore, based on the
 geochemical characteristics inferred for the Pitcairn EM-1 mantle source, in-
 corporation of a pelagic sedimentary protolith into the mantle sourced by
 Pitcairn EM-1 lavas remains a plausible model for the origin of this mantle
 end-member (Woodhead and McCulloch, 1989; Woodhead and Devey, 1993;
 Eisele et al., 2002; Honda and Woodhead, 2005).

4.2. *The presence of subducted oceanic crust in the EM-1 Pitcairn mantle*

Eisele et al. (2002) modeled the mantle source of the EM-1 Pitcairn man-
 tle and found that, in addition to supporting a role for recycled pelagic

305 sediment in the EM-1 mantle, the model results permit the presence of a
306 mafic component in the EM-1 Pitcairn mantle. High concentrations of TiO_2
307 in OIB lavas, including lavas from the Pitcairn suite, have also been used to
308 argue for a recycled mafic component in their mantle sources (Prytulak and
309 Elliot, 2007). This is because melting of a primitive or depleted mantle peri-
310 dotite cannot generate the elevated TiO_2 observed in primitive OIB lavas
311 from many hotspots, even at very low degrees of melting. We emphasize
312 that elevated Ti concentrations do not translate to particularly striking Ti,
313 Ta and Nb (TITAN) anomalies (Jackson et al., 2008) in the new dataset; this
314 is because the concentrations of these elements are generally complemented
315 by higher concentrations of elements of similar incompatibility. While the
316 presence of a mafic component in the Pitcairn mantle is indicated from el-
317 evated TiO_2 in Pitcairn lavas, it is important to evaluate the nature of this
318 mafic component.

319 With the exception of the tholeiitic seamount samples from the Pitcairn
320 hotspot (including sample 65DS-4), most Pitcairn lavas are alkalic and silica
321 undersaturated, which suggests the mafic component in the mantle source
322 is silica-undersaturated (i.e., a silica-deficient pyroxenite; Herzberg, 2011).
323 This silica-undersaturated mafic component may be responsible for elevating
324 the TiO_2 in alkalic Pitcairn lavas.

325 The less common tholeiitic samples from the Pitcairn hotspot offer im-
326 portant clues to silica-saturated recycled mafic compositions in the Pitcairn
327 mantle. Together with Koolau lavas from Hawaii (e.g., Hauri, 1996), lavas
328 from the Pitcairn hotspot, in particular the submarine Pitcairn seamount
329 lavas (e.g., Woodhead and Devey, 1993; Eisele et al., 2002; Hekinian et al.,

2003), anchor the SiO₂-rich portion of the global OIB array (Figure 7, Jackson and Dasgupta (2008). The similarity in SiO₂ contents in both the Pitcairn seamount lavas (including sample 65-DS4) and Koolau lavas may suggest that mantle sources at both EM-1 localities may generate SiO₂-rich lavas. The observation of tholeiitic samples at the Pitcairn hotspot is important and can give important clues to the nature of the EM-1 mantle, as oceanic hotspot lavas erupted far from mid-ocean ridges (i.e., excluding near-ridge hotspots like Iceland and Galapagos) are predominantly alkalic. Among hotspots erupted far from mid-ocean ridges, excluding Hawaii, tholeiitic lavas are uncommon (<10%). Of 14 such hotspots tabulated by Jackson and Dasgupta (2008), all plot, on average, in the alkali basalt field of Macdonald and Katsura (1964).

The new tholeiitic Pitcairn seamount sample, 65DS-4, not only has the highest olivine-corrected SiO₂ abundance in the Pitcairn dataset, but it also extends the Pitcairn field closer to the Koolau EM-1 field in Pb-isotopic space than any previously examined Pitcairn lava (Figure 4). The geochemical affinities suggest there may be broad similarities, including geochemically-enriched Sr and Nd isotopic compositions and elevated SiO₂, in the mantle sources sampled by the low ²⁰⁸Pb/²⁰⁴Pb component in Pitcairn (which is exhibited in purest form by 65DS-4; Figure 4) and the EM-1 component in Koolau lavas. The origin of the SiO₂-rich nature of Koolau lavas from Hawaii is argued to be the result of eclogite melting in the Hawaiian mantle plume (Sobolev et al., 2005; Hauri, 1996; Herzberg, 2011), where the eclogite is suggested to be recycled oceanic crust that preserved low U/Pb ratios through the subduction zone and generated relatively low time-integrated ²⁰⁶Pb/²⁰⁴Pb

355 (Jackson and Dasgupta, 2008; Jackson et al., 2012). Hauri (1996) showed
 356 that existing peridotite melting experiments at pressures > 2 GPa (i.e., pres-
 357 sures that are slightly lower than the pressure at the base of mature oceanic
 358 lithosphere beneath Pitcairn and Hawaii) cannot generate the major element
 359 compositions observed in Koolau lavas, and more recent compilations of ex-
 360 perimental data support this observation (see discussion in Jackson et al.,
 361 2012). Hauri (1996) Hauri (1996) also showed that elevated FeO at a given
 362 SiO_2 in Koolau lavas is a result of eclogite melting. Given existing experi-
 363 ment constraints, the high FeO and SiO_2 -rich tholeiites from Hawaii are best
 364 explained by melting a silica-saturated, mafic protolith in the mantle source
 365 (Hauri, 1996; Jackson et al., 2012). Sample 65DS-4 and a subset of Pitcairn
 366 seamount samples plot within or close to the Hawaiian Koolau field in a plot
 367 of SiO_2 vs. FeO (Figure 8); sample 65DS-4 plots in the field for Mauna
 368 Loa lavas, which also host elevated FeO at a given SiO_2 abundance, and like
 369 Koolau lavas, are argued to be melts of an eclogite-bearing mantle source
 370 (Hauri, 1996; Lassiter and Hauri, 1998; Blichert-Toft et al., 1999a; Huang
 371 and Frey, 2005; Sobolev et al., 2005; Herzberg, 2006; Jackson et al., 2012).
 372 Therefore, we suggest that a subset of EM-1 Pitcairn lavas, which have SiO_2
 373 abundances that anchor the high- SiO_2 portion of the data arrays in Figures
 374 7 and 8, sample a mantle source that is lithologically similar to the mantle
 375 source of high- SiO_2 Hawaiian lavas (Mauna Loa and Koolau).

376 *4.3. Helium isotopic constraints on the mantle source beneath Pitcairn*

377 *4.3.1. Origin of the depleted high $^3\text{He}/^4\text{He}$ component hosted in Pitcairn*
378 *lavas: Entrained depleted upper mantle or a depleted plume (FOZO?)*
379 *component?*

380 The new data provide a window into the high $^3\text{He}/^4\text{He}$ component in the
381 Pitcairn suite. The Pitcairn lava (Pit-16) with the highest $^3\text{He}/^4\text{He}$ (12.6
382 Ra, ratio to atmosphere) is also the sample with the highest $^{206}\text{Pb}/^{204}\text{Pb}$
383 that has been examined for helium isotopes (Figure 5). Additionally, with
384 increasing $^{206}\text{Pb}/^{204}\text{Pb}$, Pitcairn lavas exhibit increasingly depleted geochem-
385 ical signatures, including higher $^{143}\text{Nd}/^{144}\text{Nd}$ and lower $^{87}\text{Sr}/^{86}\text{Sr}$ (Figure 4).
386 While the high $^3\text{He}/^4\text{He}$ Pitcairn sample, Pit-16, is not itself geochemically-
387 depleted ($^{143}\text{Nd}/^{144}\text{Nd}$ is 0.512590, $^{87}\text{Sr}/^{86}\text{Sr}$ is 0.704687), it plots in a region
388 of isotopic space that is closer to the most depleted Pitcairn component (but
389 farther from the EM-1 domain) than other lavas from Pitcairn that have been
390 characterized for helium isotopes (Figure 4). Therefore, going in the direc-
391 tion of the depleted Pitcairn component in isotopic space, $^3\text{He}/^4\text{He}$ appears
392 to increase. Since MORB lavas tend to have $^3\text{He}/^4\text{He}$ ratios that cluster
393 near 8 ± 1 Ra (e.g., Graham et al., 1992), the new $^3\text{He}/^4\text{He}$ measurement
394 suggests that the depleted component in the Pitcairn mantle is unlike the
395 typical upper mantle material sampled at mid-ocean ridges. Instead, this
396 mixing component may be the high $^3\text{He}/^4\text{He}$ component that is proposed to
397 reside in the lower mantle (Hart et al., 1992).

398 Based on radiogenic Sr-Nd-Pb isotopic compositions, four distinct man-
399 tle components EM-1, EM-2, HIMU and the depleted MORB mantle are sug-
400 gested to encompass the global OIB dataset, and these end-members have

401 been plotted as the vertices of a three dimensional tetrahedron in Sr-Nd-Pb
402 isotopic space (e.g., Hart et al., 1992). Each hotspot forms arrays within
403 the tetrahedron that appear to converge on a common component with geo-
404 chemically depleted Sr and Nd isotopic compositions, and Hart et al. (1992)
405 argued that this component, called FOZO (Focus Zone), is distinct from (i.e.,
406 less geochemically-depleted than) the MORB mantle and has high $^3\text{He}/^4\text{He}$.
407 Indeed, the Pitcairn lava with the highest $^3\text{He}/^4\text{He}$ is far from being as de-
408pleted as the MORB mantle.

409 Like at other hotspot suites, geochemically-depleted lavas from the Pit-
410cairn hotspot appear to converge on the common component of Hart et al.
411(1992), but the range of $^3\text{He}/^4\text{He}$ data in the subset of depleted Pitcairn
412lavas examined by Honda and Woodhead (2005) were inconsistent with a
413FOZO component in the Pitcairn hotspot. Instead, Honda and Woodhead
414(2005) proposed that the MORB-mantle is the depleted component sampled
415by geochemically-depleted Pitcairn lavas. The new $^3\text{He}/^4\text{He}$ isotopic data
416presented here suggest that the depleted component in the Pitcairn mantle
417is not like the mantle sampled at mid-ocean ridges. Instead, the terrestrial
418mantle reservoir hosting high $^3\text{He}/^4\text{He}$, like that sampled by sample Pit-16, is
419proposed to reside in the lower mantle (Hart et al., 1992). If this is the case,
420then the depleted Pitcairn component was incorporated into the Pitcairn
421plume in the lower mantle, and is not a component entrained in the upper
422mantle. While the $^3\text{He}/^4\text{He}$ ratios in the geochemically-depleted Pitcairn
423lavas are not extraordinarily elevated, they do provide hints of a non-MORB
424component in the geochemically depleted domains of the Pitcairn plume.
425Honda and Woodhead (2005) showed from neon isotope data that Pitcairn

426 samples of clear EM-1 signature had a major component of solar neon, but
427 the data did not fall near the N-MORB correlation line.

428 4.3.2. MORB-like $^3\text{He}/^4\text{He}$ in the EM-1 mantle

429 A notable feature of the new dataset is the observation that the Pitcairn
430 lavas with the most geochemically-enriched isotopic signatures have $^3\text{He}/^4\text{He}$
431 similar to MORB. The two Pitcairn lavas that have the most extreme EM-1
432 signatures (i.e., lowest $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ and highest $^{87}\text{Sr}/^{86}\text{Sr}$),
433 and that have also been characterized for helium isotopic ratios, are found
434 to have MORB-like $^3\text{He}/^4\text{He}$ (Figure 5). However, the other mantle end-
435 members associated with recycled crustal materials, EM-2 and HIMU, have
436 $^3\text{He}/^4\text{He}$ ratios lower than MORB, as do lavas from several other hotspot
437 localities (e.g., Tristan, Gough, Comores, Canaries) that do not sample end-
438 member heavy radiogenic isotope compositions (Kurz et al., 1982; Graham
439 et al., 1992; Hanyu and Kaneoka, 1997; Barfod et al., 1999; Class et al.,
440 2005; Jackson et al., 2007a; Parai et al., 2009; Day and Hilton, 2011; Moreira
441 et al., 2011). $^3\text{He}/^4\text{He}$ ratios range down to 4.3 Ra in Samoan EM-2 lavas
442 with $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.7186 (Jackson et al., 2007a, 2009), and the low $^3\text{He}/^4\text{He}$ is
443 interpreted to be the result of the incorporation of terrigenous sediment with
444 low $^3\text{He}/^4\text{He}$ (down to 0.01 Ra; Podosek et al., 1980; Ballentine and Burnard,
445 2002) into peridotitic mantle (>8 Ra); such sediment is U and Th-rich and
446 will generate ^4He , and the mixture of sediment and mantle peridotite will
447 evolve relatively low $^3\text{He}/^4\text{He}$ in the EM-2 reservoir over time (Jackson et al.,
448 2009). Similarly, lavas sampling the HIMU mantle generally have $^3\text{He}/^4\text{He}$
449 < 7 Ra (e.g., St. Helena, Mangaia, Tubuai, older series at Rurutu; Graham
450 et al., 1992; Hanyu and Kaneoka, 1997; Parai et al., 2009; Hanyu et al., 2011)

451 or <8 in OIB with less extreme HIMU signatures (e.g., Canaries; Day and
 452 Hilton, 2011). The sub-MORB $^3\text{He}/^4\text{He}$ ratios may relate to the presence of
 453 recycled U and Th-rich oceanic crust into the HIMU mantle (e.g., Hofmann
 454 and White, 1982). Note that this U-Th enrichment is opposite to the low
 455 U/Pb assumed above for the eclogitic component of Koolau EM-1 lavas. If the
 456 high HIMU mantle is host to recycled oceanic crust, which is degassed and
 457 will have highly radiogenic $^3\text{He}/^4\text{He}$ (<0.1 Ra; Moreira and Kurz, 2001), an
 458 important question is how lavas sampling this mantle reservoir have $^3\text{He}/^4\text{He}$
 459 that is significantly higher than recycled oceanic crust. A number of solutions
 460 have been presented that attempt to explain the observation of sub-MORB
 461 $^3\text{He}/^4\text{He}$ ratios in HIMU lavas that are not as low as degassed oceanic crust
 462 (e.g., Hanyu and Kaneoka, 1997; Hanyu et al., 1999; Hilton et al., 2000;
 463 Moreira and Kurz, 2001; Class et al., 2005; Hart et al., 2008; Parai et al.,
 464 2009; Day and Hilton, 2011; Hanyu et al., 2011; Moreira et al., 2011).

465 Like the HIMU and EM-2 reservoirs, Pitcairn EM-1 lavas are also sug-
 466 gested to host recycled crustal components, including recycled sediment and
 467 oceanic crust. However, Pitcairn EM-1 lavas have MORB-like $^3\text{He}/^4\text{He}$,
 468 which is higher than end-member HIMU and EM-2 lavas and is therefore
 469 notable. Recycled crustal materials, like those suggested to exist in the EM-
 470 1 mantle beneath Pitcairn, should generate ^4He by U and Th decay over time,
 471 and this should generate mantle reservoirs with low $^3\text{He}/^4\text{He}$ (<0.1 Ra). A
 472 clue to the apparent paradox of MORB-like $^3\text{He}/^4\text{He}$ in the Pitcairn may
 473 be provided by the higher diffusivity of He relative to the other radiogenic
 474 isotopic tracers (Sr, Nd and Pb). Owing to the high diffusivity of helium,
 475 Hart et al. (2008) suggested that mantle heterogeneities with length-scales

476 smaller than 5 to 10 km will not preserve their helium isotopic signatures
 477 over time scales of 1.5 Ga, but their Sr, Nd and Pb isotopic signature will
 478 remain relatively unchanged. If the recycled sediment heterogeneities in the
 479 EM-1 mantle are distributed over relatively short length scales (5 to 10 km
 480 or less), and if the EM-1 mantle is sufficiently old (>1.5 Ga), then the he-
 481 lium isotopic composition of the EM-1 mantle may become overwhelmed by
 482 the $^3\text{He}/^4\text{He}$ signature of the mantle reservoir hosting the recycled materi-
 483 als, here inferred to be ambient upper mantle, which is 8 Ra. Of course,
 484 this requires that the ^4He production of the recycled component is low com-
 485 pared to the ambient mantle helium concentration (Hart et al., 2008). If the
 486 time-integrated helium concentrations of the upper mantle greatly exceed the
 487 concentrations of the recycled oceanic crust and sediment (Day and Hilton,
 488 2011), and if the length scales of the recycled materials are small ($\ll 5$ -10
 489 km; Hart et al., 2008), the $^3\text{He}/^4\text{He}$ of the resulting mantle reservoir is more
 490 likely to have a $^3\text{He}/^4\text{He}$ similar to the ambient upper mantle. A diffusive
 491 decoupling mechanism like that proposed by Hart et al. (2008) may explain
 492 the presence of MORB-like $^3\text{He}/^4\text{He}$ in a mantle reservoir that hosts recycled
 493 oceanic crust and sediments. Because neon diffuses many orders of magni-
 494 tude slower than helium (in olivine, Cherniak et al., 2014), the solar neon
 495 signature of Honda and Woodhead (2005) could be retained in the recycled
 496 material while a MORB-like helium is imprinted by diffusive exchange. Of
 497 course, a more sinister explanation would simply postulate a helium (but not
 498 Sr, Nd or Pb) isotope exchange in the uppermost mantle between migrating
 499 EM-1 melts (or magma chambers) and the ambient MORB mantle country
 500 rocks. It is of course possible that EM-1 has the same helium value as MORB

501 by coincidence.

502 5. Conclusions

503 The extreme enriched mantle EM-1 signature in Pitcairn lavas likely
504 hosts an enriched protolith derived from upper continental crust, and this
505 protolith exhibits many geochemical similarities with pelagic sediment. El-
506 evated primary-melt Ti in the EM-1 lavas suggests the presence of recycled
507 pyroxenite in the mantle source. We identify a tholeiitic sample (65-DS4)
508 from Adams Seamount at the leading edge of the Pitcairn hotspot, and the
509 tholeiitic composition of this sample complements tholeiites previously de-
510 scribed from the Pitcairn hotspot. The new tholeiitic Pitcairn seamount
511 sample not only has the highest olivine-corrected SiO_2 abundance in the Pit-
512 cairn dataset, but it also extends the Pitcairn field closer to the Hawaiian
513 Koolau field in Pb-isotopic space than any previously examined Pitcairn lava.
514 The geochemical affinities suggest there may be broad similarities including
515 geochemically-enriched Sr and Nd isotopic compositions and elevated SiO_2 in
516 the mantle sources sampled by tholeiitic Pitcairn lavas and the EM-1 com-
517 ponent in Koolau lavas. The origin of the SiO_2 -rich nature of Koolau lavas
518 from Hawaii is argued to be the result of eclogite melting in the Hawaiian
519 mantle plume. The isotopic and major element affinities between the Pit-
520 cairn EM-1 tholeiitic sample 65DS-4 and the EM-1 component sampled by
521 tholeiitic Koolau lavas from Hawaii suggests a common origin, and we argue
522 that a silica-saturated mafic protolith is in the source of the EM-1 mantle at
523 both hotspots. We also present the highest $^3\text{He}/^4\text{He}$ (12.6 Ra) reported for
524 Pitcairn lavas, and it is associated with a third component with the highest

525 $^{206}\text{Pb}/^{204}\text{Pb}$ and the most geochemically-depleted $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$
 526 ratios in the Pitcairn suite. We also show that EM-1 end-member lavas from
 527 Pitcairn have $^3\text{He}/^4\text{He}$ that is indistinguishable from the MORB mantle. If
 528 the recycled sedimentary and mafic components in the Pitcairn EM-1 oc-
 529 cur at relatively short length scales, distributed in a peridotite matrix with
 530 $^3\text{He}/^4\text{He}$ of 8 Ra, diffusive exchange of helium over timescales > 1.5 Ga will
 531 permit the $^3\text{He}/^4\text{He}$ of the Pitcairn EM-1 mantle source to evolve toward
 532 values that typify the ambient upper mantle sampled by MORB. Alterna-
 533 tively, exchange of helium between EM-1 melts and melt conduits in the
 534 upper mantle host rock may result in EM-1 helium isotopic ratios evolving
 535 toward values typical of the upper mantle.

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Figure 1: Map of Pitcairn island with sample locations from this study. The subaerial samples, and the single submarine sample, were collected prior to GPS availability. The submarine sample 65DS-4 was obtained from volcano 1 located ~90 km ESE of Pitcairn.

Figure 2: TAS diagram shows that Pitcairn lavas are prevailingly alkalic, but some seamount samples are slightly tholeiitic and cross the alkali-tholeiite divide. The new data for Pitcairn seamount sample 65DS-4 show that it is a tholeiite (whole rock and matrix data are shown for this sample). The previously published data shown in this figure (and in all subsequent figures) were obtained from the GEOROC database (<http://georoc.mpch-mainz.gwdg.de/georoc>) on October 1, 2013 and verified with the original references.

Figure 3

Variation diagrams of major element oxides for all Pitcairn lavas. The data are normalized to 100 wt. % on a dry basis. Olivine dominates the major element variations for bulk rock composition where $\text{MgO} > 8\text{-}10$ wt. %. At $\text{MgO} < 8\text{-}10$ wt. %, clinopyroxene fractionation is apparent from reduced $\text{CaO}/\text{Al}_2\text{O}_3$ ratios. Fractionation of Fe- and Ti-rich phases is evident in lavas with $\text{MgO} < 5$ wt. %. For some of the seamount samples, both whole rock and glass data are plotted when it was available on the same sample. For 65DS-4 whole rock (65DS-4 wr) and matrix (65DS-4 m) are plotted separately.

Figure 4: Isotopic data for Pitcairn lavas reveal a typical EM-1 signature: low $^{143}\text{Nd}/^{144}\text{Nd}$, moderately high $^{87}\text{Sr}/^{86}\text{Sr}$ and highly unradiogenic $^{206}\text{Pb}/^{204}\text{Pb}$. The plot of $^{208}\text{Pb}/^{204}\text{Pb}$ against $^{206}\text{Pb}/^{204}\text{Pb}$ reveals the presence of a third component in Pitcairn lavas, which is the second component that is highlighted in the dashed triangle. The third component (called low- κ EM-1) is anchored by geochemically-more depleted rejuvenated lavas. Data fields for Samoa and Hawaii (including Koolau volcano) are shown with the global OIB database, and these data are taken from Georoc database and verified with the original references. The Pitcairn isotopic data are from the following sources: (Woodhead and McCulloch, 1989; Woodhead and Devey, 1993; Woodhead et al., 1993; Eisele et al., 2002; Honda and Woodhead, 2005).

Figure 5: $^3\text{He}/^4\text{He}$ ratios and He concentrations in olivine (in subaerial samples, reported here for the first time) and seamount glasses (reported in Honda and Woodhead, 2005). Only the data for glasses with $^4\text{He} > 10^{-7}$ cc STP/g are plotted in the right panel. The line in the left panel connects two replicate measurements on coarse and fine olivine fractions (sample PIT-16). The diagram to the right shows that the sample with the highest $^3\text{He}/^4\text{He}$ ratio (which is shown as the average of two measurements of this sample, 11.8 Ra and 13.3 Ra, measured on coarse and fine olivine fractions, respectively) also has the weakest EM-1 signature and instead trends towards the depleted mantle component (FOZO?) sampled by Pitcairn lavas. The two lavas with the lowest $^{206}\text{Pb}/^{204}\text{Pb}$ have the lowest $^3\text{He}/^4\text{He}$ ratios, 7.7 Ra (measured in a glass) and 8.0 Ra (measured in olivine), indicating MORB-like $^3\text{He}/^4\text{He}$ signature in the EM-1 mantle.

Figure 6: The new olivine $^{187}\text{Os}/^{188}\text{Os}$ measurements in this study fall in the same range as identified in the most Os-rich whole rocks (> 50 pg/g; Eisele et al., 2002) and support the previous observations that the EM-1 end-member has radiogenic Os, which is consistent with a recycled origin for EM-1 lavas in Pitcairn (Eisele et al., 2002).

Figure 7: $^{206}\text{Pb}/^{204}\text{Pb}$ plotted against the major element compositions of Pitcairn lavas, the Hawaiian dataset is from Jackson et al. (2012) and the global OIB dataset from Jackson and Dasgupta (2008). With the exception of sample 65-DS-4 matrix, the displayed Pitcairn data shows only data with $\text{MgO} > 9\%$, and all lavas are olivine fractionation corrected by the addition or subtraction of equilibrium olivine so that they were in equilibrium with mantle olivine (forsterite 90). The data shows that, together with Hawaii, Pitcairn seamount lavas anchor the high SiO_2 -rich portion of the Pitcairn array.

Figure 8: FeO_T plotted against SiO_2 of Pitcairn lavas and Mauna Loa, Koolau and Loihi from Hawaiian dataset (obtained from GEOROC database on March 13, 2014.). All lavas are olivine fractionation corrected by the addition or subtraction of equilibrium olivine so that they were in equilibrium with mantle olivine (forsterite 90). With the exception of sample 65-DS-4 matrix, the displayed Pitcairn data have $\text{MgO} > 9\%$, and Hawaiian data have $\text{MgO} > 6.5\%$.

Table 1: Whole rock major, trace and isotope analyses for Pitcairn samples from this study, and Os-isotopic analyses on olivines.

	65DS-4 (wr)	65DS-4 (m)	PIT-1	PIT-3	PIT-4A	PIT-6	PIT-7	PIT-8	PIT-11	PIT-12	PIT-13	PIT-16*
SiO ₂ (wt.%)	48.59	51.15	48.10	48.73	48.91	49.54	48.15	46.32	46.45	49.22	47.47	48.71
TiO ₂	1.88	2.53	3.50	3.27	4.34	2.99	3.55	2.77	2.74	3.34	3.75	4.08
Al ₂ O ₃	10.14	13.58	14.30	14.76	14.97	15.96	15.28	13.35	13.24	15.14	14.42	15.82
FeO _T	10.74	9.79	9.99	10.39	10.68	11.35	10.52	11.52	11.53	10.31	10.94	10.46
CaO	7.55	10.05	7.20	7.82	8.53	7.50	8.69	8.31	8.21	7.87	8.12	8.45
MgO	17.09	8.39	7.10	7.77	4.25	3.83	6.39	11.91	12.03	6.27	8.06	4.45
MnO	0.30	0.27	0.15	0.15	0.17	0.21	0.16	0.17	0.17	0.16	0.16	0.17
K ₂ O	0.63	0.88	1.70	1.70	2.13	2.00	1.59	0.98	0.99	1.88	1.59	2.00
Na ₂ O	2.04	2.71	3.90	3.71	4.23	4.42	3.84	3.02	2.94	4.04	3.31	3.92
P ₂ O ₅	0.26	0.36	0.90	0.62	0.82	1.17	0.60	0.43	0.42	0.73	0.61	0.79
Total	99.22	99.71	96.84	98.92	99.03	98.97	98.77	98.78	98.72	98.96	98.43	98.85
olivine X _{Fe}	84.7		80.1	80.2	-		76.5	82	80.6	81.3	79.5	75.7
plagioclase X _{An}			NA	-	NA		58.6	-	-	62.7	72.1	57.8
XRF												
Rb (ppm)	6.0		30.8	32.5	41.7	42.5	28.2	18.0	18.1	34.8	26.7	37.0
Sr	343		669		616	593	595	528	519	666	656	562
Zn	103		111	113	140	154	15	123	123	124	124	121
Ni	556		180	180	47	14	105	311	312	102	165	54
Cr	962		221	231	15	1	141	434	428	108	201	56
V	163		177	171	251	116	226	199	198	207	222	238
Ba	134		474	487	560	302	434	288	289	541	474	568
Y	20		28	30	38	47	30	22	22	33	29	43
Nb	13		44	46	67	76	48	34	34	54	50	61
Zr	144		308	318	439	504	325	220	218	364	328	397
ICP-MS												
Cs (ppm)	0.08		0.16	0.14	0.63	0.33	0.13	0.06	0.10	0.25	0.11	0.13
Rb	7.4		32.5	31.4	44.2	42.5	29.4	17.7	19.5	34.5	27.5	37.5
Ba	125		467	460	567	283	412	273	285	492	468	544
Th	1.39		6.59	7.64	8.44	9.47	6.24	4.35	4.40	8.12	7.02	7.60
U	0.38		1.15	1.31	1.58	2.00	0.88	0.66	0.78	0.97	0.80	1.42
Nb	11.7		50.6	47.1	66.8	75.1	51.1	34.3	34.8	55.3	51.3	61.4
Ta	0.80		2.35	2.46	3.38	4.00	2.58	1.76	1.81	2.91	2.72	3.15
La	15.1		51.1	52.2	62.1	67.8	46.7	30.5	31.7	56.1	50.9	62.2
Ce	34.5		98.4	104.2	124.3	139.5	94.7	63.7	63.8	114.2	100.8	123.9
Pb	3.69		4.87	5.43	6.53	6.63	4.49	2.73	2.91	5.52	4.72	6.39
Pr	4.6		12.9	13.6	15.4	17.7	11.6	8.2	8.0	14.1	12.4	16.3
Nd	20.1		48.6	49.9	59.8	70	46.4	32.5	32.2	54.2	49.2	63.8
Sr	348		653	312	614	580	588	523	513	632	631	658
Zr	143											
Hf	3.57		7.44	7.26	8.73	10.37	6.9	4.7	4.89	7.68	7.32	8.13
Sm	4.76		8.89	9.16	11.42	13.77	9.10	6.36	6.72	10.36	9.42	12.23
Eu	1.65		2.70	2.77	3.52	3.98	2.86	2.16	2.17	3.24	2.98	3.75
Gd	4.79		8.54	8.80	10.41	13.16	8.20	6.24	6.10	9.32	8.21	11.38
Tb	0.74		1.11	1.19	1.43	1.78	1.14	0.86	0.85	1.29	1.13	1.64
Dy	4.07		6.10	6.15	8.20	9.96	6.59	4.84	4.90	7.13	6.42	9.08
Ho	0.74		1.06	1.09	1.44	1.73	1.16	0.85	0.87	1.26	1.14	1.61
Y	18.4		29.8	30.1	37.6	46.8	30.4	22.4	21.9	32.8	28.1	42.3
Er	1.83		2.85	2.99	3.71	4.52	2.92	2.15	2.25	3.18	2.88	4.22
Tm	0.24		0.34	0.36	0.47	0.57	0.36	0.27	0.27	0.41	0.35	0.55
Yb	1.37		2.2	2.27	3.04	3.73	2.34	1.73	1.73	2.57	2.22	3.58
Lu	0.2		0.32	0.33	0.42	0.52	0.34	0.25	0.25	0.36	0.32	0.51
⁴ He ccSTP/g			1.36E-08	5.56E-09			6.24E-09	1.35E-09	6.06E-08		9.78E-09	coarse: 4.3E-09 fine: 5.5E-10
³ He/ ⁴ He (Ra)			8.6	7.7			9.3	8.0	8.2		9.0	coarse: 11.8 fine: 13.3
Os, ppt			48.7	53.2			31.1	41.6	95.0	85.1		
¹⁸⁷ Os/ ¹⁸⁸ Os			0.1364	0.1432			0.1377	0.1393	0.1408	0.1342		
¹⁴³ Nd/ ¹⁴⁴ Nd	0.512585		0.512447	0.512435	0.512533	0.512716	0.512589	0.512538	0.512521	0.512518	0.512518	0.512590
⁸⁷ Sr/ ⁸⁶ Sr	0.704485		0.705243	0.705220	0.704748	0.703702	0.704824	0.704633	0.704639	0.704975	0.704999	0.704687
²⁰⁶ Pb/ ²⁰⁴ Pb	17.732		17.630	17.611	17.789	18.054	17.833	17.900	17.858	17.789	17.735	18.131
²⁰⁷ Pb/ ²⁰⁴ Pb	15.474		15.460	15.461	15.492	15.506	15.494	15.494	15.478	15.487	15.474	15.515
²⁰⁸ Pb/ ²⁰⁴ Pb	38.438		38.936	38.912	38.970	39.064	38.982	38.976	38.939	38.975	38.943	38.926

Data in *italics* are previously published in Eiler et al. (1995) where NA indicates "present but not analyzed".

* ³He/⁴He on sample PIT-16, was measured twice, on coarse (11.8 Ra; 4.3*10⁻⁹ cc STP/g) and fine (13.3 Ra; 5.5*10⁻¹⁰) olivine fractions, to verify the elevated ³He/⁴He ratio.

Table 2: Olivine analyses for 65DS-4.

	SiO₂ (wt.%)	FeO_T (wt.%)	CaO (wt.%)	MgO (wt.%)	MnO (wt.%)	NiO (wt.%)	Total	X_{Fe}
65DS04-ol1-1	40.20	14.50	0.24	45.50	0.18	0.24	100.87	84.8
65DS04-ol1-2	40.56	14.21	0.26	45.18	0.21	0.26	100.69	85.0
65DS04-ol1-3	40.82	14.12	0.26	45.18	0.16	0.24	100.77	85.1
65DS04-ol1-4	40.47	14.41	0.25	45.25	0.22	0.28	100.89	84.8
65DS04-ol1-5	40.20	14.96	0.25	44.71	0.18	0.29	100.58	84.2
65DS04-ol1-6	40.45	14.89	0.24	44.73	0.18	0.22	100.71	84.3
65DS04-ol1-7	40.11	14.45	0.25	45.03	0.21	0.22	100.27	84.8
65DS04-ol1-8	40.51	14.56	0.26	44.88	0.18	0.25	100.63	84.6
65DS04-ol2-1	40.44	14.53	0.26	45.10	0.17	0.23	100.74	84.7
65DS04-ol2-2	40.14	14.47	0.26	44.96	0.20	0.25	100.28	84.7
65DS04-ol2-3	39.94	14.46	0.28	44.66	0.18	0.27	99.78	84.6
65DS04-ol2-4	40.32	14.72	0.27	45.18	0.20	0.24	100.93	84.6

$X_{Fe} = (n(MgO)/(n(MgO) + n(FeO))) * 100$, where n = number of moles;

Multiple spots were measured on two olivine grains. The grains were designated as ol1 (n = 8 spots) and ol2 (n = 4 spots).