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Large iron isotope variation in the eastern Pacific mantle as a consequence of ancient low-degree melt metasomatism

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Abstract

Studies of mid-ocean ridge basalts (MORB) have revealed a heterogeneous asthenospheric mantle in chemical elements and radiogenic isotopes. Here we report that MORB mantle is also heterogeneous in Fe isotopes through studying the glass samples from seamounts flanking the northern East Pacific Rise between 5° and 15°N. These samples show large Fe isotope variation with δ^{56} Fe values (+0.03% to +0.36%) exceeding the known range of MORB (+0.05% to +0.17%). Such highly varied δ^{56} Fe values cannot be well explained by seafloor alteration, fractional crystallization or partial melting processes, but instead require a source mantle significantly heterogeneous in Fe isotope compositions. Importantly, the δ^{56} Fe values of these basalts correlate significantly with major and trace elements and Sr-Nd-Pb-Hf radiogenic isotopes, reflecting melting-induced mixing of a two-component mantle with the enriched component having heavy Fe isotope compositions dispersed as physically distinct domains in the depleted mantle matrix. The major and trace element characteristics of the enriched mantle component, as inferred from these basalts, are consistent with a low-degree melting origin. Such low-degree melts with heavy Fe isotope compositions most likely formed at sites such as the lithosphere-asthenosphere boundary beneath ocean basins, which can metasomatize the overlying oceanic lithosphere by crystallizing dikes/veins of garnet pyroxenite lithologies. Recycling of these dikes/veins with isotopically heavier Fe can readily contribute to the Fe isotope heterogeneity in the MORB mantle. However, the extremely high primitive δ^{56} Fe values of the two alkali basalts (up to 0.34%) require an enriched source component with unusually high δ^{56} Fe values. We suggest that partial melts from the recycled dikes/veins of garnet pyroxenite lithologies can react with the ambient peridotitic mantle and generate a secondary garnet pyroxenite with heavier Fe isotope compositions than, but similar radiogenic isotope compositions as its precursor. Melting-induced mixing between these garnet pyroxenites (recycled and newly formed) and depleted mantle matrix can readily explain the compositional variations in elements, radiogenic isotopes and Fe isotopes observed in these seamount lavas. These new data and correlated variations offer a new dimension for understanding the origin of mantle chemical and isotopic heterogeneity in the context of chemical differentiation of the Earth.

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Keywords: East Pacific Rise; Mid-ocean ridge basalts; Fe isotope; Mantle heterogeneity; Seamounts; Mantle metasomatism

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1. INTRODUCTION

The Earth's upper mantle has been inferred to be broadly heterogeneous at different length scales with respect to its incompatible element abundances and radiogenic isotope compositions through studies of mid-ocean ridge basalts (MORB) (Zindler et al., 1984; Allègre and Turcotte, 1986; Mahoney et al., 1994; Niu and Batiza, 1997; Niu et al., 2002a; Donnelly et al., 2004; Waters et al., 2011). The origin of these chemical and isotopic heterogeneities has long been attributed to the process of subduction, which can reiniect enriched materials such as terrigenous or pelagic sediments (e.g., Cohen and O'Nions, 1982; Weaver et al., 1986; Wright and White, 1987), metasomatized oceanic lithospheric mantle (e.g., Galer and O'Nions, 1986; Niu et al., 1999, Niu et al., 2002a) or oceanic crust materials (e.g., Allègre and Turcotte, 1986; Donnelly et al., 2004; Hirschmann and Stolper, 1996; Stracke and Bourdon, 2009) to the depleted asthenospheric mantle. However, whether the subduction process can introduce heterogeneity of non-radiogenic isotopes in the MORB mantle remains poorly constrained (e.g., Elliott et al., 2006).

Iron is by mass the most abundant metal of the Earth, and studies have observed varied Fe isotope compositions in mantle and mantle-derived rocks (e.g., Williams et al., 2004; Weyer and Ionov, 2007; Teng et al., 2008; Dauphas et al., 2009; Sossi et al., 2012, 2016; Williams and Bizimis, 2014; Konter et al., 2016; Nebel et al., 2018, 2019), which may provide new insights into the origin of chemical and isotopic differentiation of the Earth. As the most abundant terrestrial magmas, the mid-ocean ridge basalts (MORB) have been studied to show globally homogeneous Fe isotopic compositions (δ^{56} Fe = +0.11 ± 0.04‰, 2SD; Teng et al., 2013). This has been interpreted as reflecting a homogeneous asthenospheric mantle source (Beard and Johnson, 2007; Craddock et al., 2013; Poitrasson et al., 2013; Teng et al., 2013) although mantle melting (Williams et al., 2004, 2005; Weyer and Ionov, 2007; Dauphas et al., 2009, 2014) and magma differentiation (Teng et al., 2008; Schuessler et al., 2009; Sossi et al., 2012, 2016; Chen et al., 2019) have been shown to cause measurable Fe isotope fractionation. On the other hand, significant Fe isotope variability observed in lithospheric mantle rocks is thought to reflect an overall heterogeneous upper mantle (Weyer and Ionov, 2007; Huang et al., 2011; Poitrasson et al., 2013; Williams and Bizimis, 2014), leaving the homogeneous Fe isotope composition of MORB explained as resulting from homogenization during melting and meltperidotite reaction in the mantle and further homogenization in crustal magma chambers (Beard and Johnson, 2004; Weyer and Ionov, 2007). Therefore, the fundamental question whether the MORB mantle source has homogeneous or heterogeneous Fe isotope composition remains unanswered.

To address this question, we studied and show here that the glass samples from seamounts flanking the East Pacific Rise (EPR) have significant Fe isotope variation $(\delta^{56}\text{Fe} = +0.03\% \text{ to } +0.36\%)$ that correlates with the abundances and ratios of major elements, trace elements and Sr-Nd-Pb-Hf radiogenic isotopes. The heavier Fe isotope compositions (higher δ^{56} Fe values) in these EPR seamount basalts are associated with the progressively more enriched mantle component. This observation is a straightforward manifestation of Fe isotope heterogeneity at least in the eastern Pacific mantle.

2. SAMPLES AND METHODS

2.1. Sample description

The seamounts were sampled by dredging during the 1988 Raitt 02 expedition aboard R/V Thomas Washington (Batiza et al., 1990). They were located within 1 Ma isochron (<60 km) of the EPR axis between 5° and 15°N on both Pacific and Cocos Plates (Fig. 1). These samples have been previously studied for major and trace elements (Niu and Batiza, 1997; Table S1) and Sr-Nd-Pb-Hf radiogenic isotopes (Niu et al., 2002a; Zhang et al., 2016; Table 1). This sample suite is globally ideal to examine the MORB mantle Fe isotope heterogeneity for important reasons: (1) the near-ridge seamount volcanism shares the common asthenospheric source with the EPR axial volcanism and is far away from any known mantle plume influence; (2) possible mantle source heterogeneity can be preserved in seamount lavas because of small melt volumes tapped locally beneath individual seamounts, which is obscured in EPR axial lavas because of the effective homogenization processes during melt focusing towards the very narrow axial zone in the mantle and also in the long-lived crustal magma chambers (O'Hara, 1977; Zindler et al., 1984; Macdougall and Lugmair, 1985); (3) these seamount lavas vary from extremely depleted tholeiites to highly enriched alkali basalts with the extent of depletion and enrichment exceeding the known range of seafloor lavas in terms of the abundances and ratios of incompatible elements, showing with great fidelity the mantle source heterogeneity (Niu and Batiza, 1997; Niu et al., 2002a; Zhang et al., 2016).

2.2. Analytical methods

Fe isotope compositions of 21 seamount samples with large compositional variations were measured in the Laboratory of Ocean Lithosphere and Mantle Dynamics, Institute of Oceanology, Chinese Academy of Sciences (IOCAS). About 25 mg fresh glasses were hand-picked and then washed ultrasonically in Milli-Q water before digestion in HF-HNO₃-HCl on a hotplate. After repeated re-flux using aqua regia to obtain full digestion, the samples were finally dissolved in 1 ml 9 N HCl ready for chromatographic separation for Fe. Fe was purified using 1 ml anionexchange resin (Bio-Rad AG MP-1M 200-400 mesh) conditioned with 9 N HCl, following the procedure in Gong et al. (2020). Matrix elements including Cr and Ni were removed by washing with 5 ml 9 N HCl. The columns were then washed with 5 ml 6 N HCl to remove Cu and possible residual Cr and Ni. Fe was eluted using 2 ml 1 N HCl. The eluted Fe solutions were analyzed using ICP-OES to ensure purity and full recovery. The total procedural blank for Fe was ~ 80 ng, which is less than 0.01% of the processed



Fig. 1. (A) Tectonic framework of the northern $(5-15^{\circ}N)$ EPR and the vicinity. (B) Simplified map of the study area showing the locations of the near-ridge seamounts. The size of the circles approximates the relative size of seamounts. Seamounts with HIMU-like and alkali basalts are indicated. Modified from Niu et al. (2002a).

samples and is thus considered negligible. Prior to measurements, Fe solutions were diluted to 14 ppm, and 19.6 ppm GSB Ni standard (an ultrapure single elemental standard solution from the China Iron and Steel Research Institute) was added as an internal mass bias monitor to each diluted sample (Ni:Fe = 1.4:1).

Fe isotopes were analyzed by a Nu Plasma MC-ICP-MS with wet nebulization. Backgrounds were measured and

Table 1	
Fe-Sr-Nd-Hf-Pb isotopic data of samples from near-EPR seamounts between 5	and 15°N.

Sample	Type	Latitute (°N) Longitude (°W	Depth (m)	$\delta^{56} Fe$	$2sd \delta^{57}Fe$	$2sd \delta^{56}Fe_{Prim}$	Δ^{56} Fe	e^{87} Sr/ ⁸⁶ Sr (±2 σ)	143 Nd/ 144 Nd ($\pm 2\sigma$) $^{176}\text{Hf}/^{177}\text{Hf}(\pm 2\sigma)$) ²⁰⁶ Pb/ ²⁰⁴ Pt	o ²⁰⁷ Pb/ ²⁰⁴ Pt	o ²⁰⁸ Pb/ ²⁰⁴ Pb
R1-14	Ν	5.77	102.18	1834	0.11	0.04 0.15	0.06 0.11	0.00	0.702393 ± 11	0.513148 ± 14	0.283187 ± 12	18.11	15.45	37.55
R13-1	Alkali	8.40	104.07	2140	0.36	0.06 0.54	0.12 0.34	0.02	0.702895 ± 07	0.512956 ± 09	0.282974 ± 07	18.58	15.55	38.01
R15-1	Alkali	8.76	104.54	1682	0.34	0.06 0.51	0.11 0.32	0.02	0.702877 ± 10	0.512979 ± 12	0.282966 ± 08	18.68	15.55	38.08
Replicate					0.33	0.03 0.46	0.09							
R62-7	E	10.03	104.19	2320	0.05	$0.07 \ 0.08$	0.13 0.05	0.01	0.702420 ± 10	0.513153 ± 08	0.283167 ± 15	18.30	15.48	37.71
R18-3	E	8.93	104.46	2720	0.10	0.05 0.19	0.07 0.08	0.02	0.702722 ± 14	0.513041 ± 11	0.283061 ± 11	18.58	15.53	38.05
R25-1	Ν	8.88	103.79	1980	0.12	0.04 0.17	0.11 0.11	0.01	0.702432 ± 11	0.513144 ± 08	0.283158 ± 13	18.28	15.49	37.79
R3-1	Ν	5.78	102.21	1773	0.09	0.06 0.12	0.06 0.08	0.00	0.702458 ± 13	0.513183 ± 14	0.283194 ± 14	18.08	15.42	37.52
R31-1	E	9.09	105.02	2366	0.06	0.02 0.08	0.06 0.05	0.01	0.702362 ± 11	0.513147 ± 08	-	18.32	15.50	37.86
R32-1	E	9.09	104.92	3025	0.17	0.02 0.24	0.03 0.16	0.01	0.702876 ± 15	0.513047 ± 09	0.283084 ± 12	18.53	15.52	38.07
R3-4	E	5.78	102.21	1773	0.05	0.02 0.10	0.05 0.03	0.02	0.702420 ± 11	0.513169 ± 36	0.283212 ± 13	18.12	15.45	37.58
Replicate					0.06	0.02 0.07	0.05							
R65-1	Ν	10.13	103.41	2074	0.03	0.03 0.08	0.05 0.02	0.01	0.702414 ± 10	0.513111 ± 09	0.283189 ± 13	18.14	15.44	37.55
R66-1	Ν	10.14	103.34	2600	0.07	0.04 0.11	0.16 0.06	0.01	0.702513 ± 13	0.513178 ± 11	0.283189 ± 13	18.15	15.46	37.63
R7-13	Ν	8.14	103.19	2020	0.05	$0.06 \ 0.08$	0.05 0.04	0.01	0.702457 ± 11	0.513174 ± 10	0.283189 ± 15	18.13	15.43	37.53
R73-1	E	10.38	103.92	2547	0.04	0.02 0.07	0.06 0.03	0.01	0.702694 ± 10	0.513094 ± 07	0.283124 ± 10	18.14	15.49	37.55
R78-6	E	11.22	103.58	2450	0.15	0.02 0.21	0.04 0.14	0.01	0.702922 ± 10	0.512970 ± 09	0.283043 ± 09	19.31	15.61	39.00
R79-1	E	11.79	103.25	1620	0.11	0.05 0.19	0.05 0.10	0.01	0.702900 ± 14	-	0.283047 ± 11	19.31	15.60	39.02
R80-1	E	11.80	103.25	1619	0.19	0.06 0.29	0.11 0.18	0.01	0.702529 ± 08	-	0.283144 ± 08	18.40	15.49	37.90
R8-8	Ν	8.34	103.06	3180	0.06	0.06 0.08	0.08 0.04	0.02	0.702484 ± 10	0.513131 ± 10	0.283153 ± 13	18.50	15.50	37.88
R96-24	Е	13.07	103.45	2577	0.06	0.03 0.10	0.11 0.05	0.01	0.702551 ± 12	0.513143 ± 07	0.283209 ± 11	18.34	15.49	37.84
R17-6	E	8.91	104.57	2715	0.13	0.02 0.20	0.07 0.10	0.03	0.702937 ± 11	0.512993 ± 08	0.283027 ± 07	_	_	_
R4-7	E	5.60	103.02	2263	0.11	0.03 0.18	0.03 0.09	0.02	0.702589 ± 10	0.513078 ± 14	0.283110 ± 09	18.49	15.49	37.97

The Sr-Nd-Pb isotope data are from Niu et al. (2002a). The Hf isotope data are from Zhang et al. (2016).

subtracted using electrostatic analyzer (ESA) deflection. Contributions from isobaric interferences (⁴⁰Ar¹⁴N on ⁵⁴Fe and ⁴⁰Ar¹⁶O on ⁵⁶Fe) were eliminated by measuring in pseudo-high resolution mode with $M/\Delta M$ of >8000. The mass bias fractionation during analysis was corrected using Ni doping method by 60Ni/58Ni with the 58Fe interference on ⁵⁸Ni corrected based on ⁵⁶Fe (Oeser et al., 2014; Chen et al., 2017, 2019). The Fe isotope data are expressed using the standard notation δ^{56} Fe (=[(56 Fe/ 54 Fe)_{sample}/ $({}^{56}\text{Fe}/{}^{54}\text{Fe})_{IRMM-014} - 1] \times 1000$). Because the commonly used reference material IRMM-014 is no longer available, this study used a new reference material GSB Fe, which has been reported relative to IRMM014 (δ^{57} Fe_{IRMM014} = δ^{57} Fe_{GSB} + 1.073; δ^{56} Fe_{IRMM014} = δ^{56} Fe_{GSB} + 0.729; He et al., 2015). During analysis, every two sample solutions were bracketed with 14 ppm GSB Fe standard solution that was also doped with the GSB Ni solution with Ni:Fe of 1.4:1. Every sample solution was repeatedly analyzed for four times, with the average δ^{57} Fe and δ^{56} Fe values of each sample given in Table 1. Long-term analyses of an in-house Alfa Fe standard give an average δ^{56} Fe value of 0.52 $\pm 0.03\%$ (2SD). The δ^{56} Fe value of the USGS standard BCR-2 analyzed together with our samples was 0.10 $\pm 0.05\%$ (2SD, n = 13), consistent with the recommended literature values (Craddock and Dauphas, 2011; He et al., 2015). Duplicate digestion, chemical separation and analyses of two samples (R15-1 & R3-4) show good reproducibility (Table 1). Detailed methods for Fe elemental purification and isotope analysis are in Gong et al. (2020).

2.3. Correction of Fe isotopes for fractional crystallization

Olivine is a major liquidus phase during early differentiation of MORB melt, which can subtract Fe²⁺ and light Fe isotopes from the melt (Teng et al., 2008). Compilation of major elements of >400 MORB glasses from the East Pacific Rise and model results showed that olivine is the major mafic liquidus phase in these magmas with $Mg^{\#} > \sim 0.58$ (Niu, 2005). The EPR seamount samples analyzed for Fe isotope compositions in this study have $Mg^{\#}$ of 0.57–0.72 (Table 1), consistent with olivine as the major mafic liquidus phase in the melts. Therefore, in order to investigate the Fe isotope composition of primitive EPR seamount lavas, we corrected the Fe isotopes for the effect of fractional crystallization of olivine, following the method described in Sossi et al. (2016). Olivine composition in equilibrium with each EPR seamount sample is calculated, using a Fe-Mg exchange partition coefficient ($K_{DOI-Melt}^{Fe-Mg}$) of 0.30, and added incrementally into the melt with the olivine composition re-iterated at each increment until melt $Mg^{\#} = 0.72$, in equilibrium with mantle olivine of Fo = 89.6 (Roeder and Emslie, 1970). For the correction, we use a starting $Fe^{3+}/\Sigma Fe = 0.15$ and an isotope fractionation factor between olivine and melt Δ^{56} Fe_{Ol-melt} = 2853 × $(\langle F \rangle_{Ol} - \langle F \rangle_{melt})/T^2$ after Dauphas et al. (2014), where < F > is the average force constant of Fe-O bonds $< F >_{Fe^{2+} in melt} = 199 N/m,$ $(< F >_{Ol} = 197 \text{ N/m},$ $< F >_{Fe^{3+} in melt} = 351 \text{ N/m}, \text{ and } < F >_{melt} = Fe^{3+} / \sum Fe^{3+}$ $\times \langle F \rangle_{Fe^{3+} \text{ in melt}} + (1 - Fe^{3+} / \sum Fe) \times \langle F \rangle_{Fe^{2+} \text{ in melt}}$ and the melt temperature (T) at each increment is calculated using an equation of T (°C) = $1026 \times e^{[0.01894 \times MgO}$ ^(wt.%)] (Niu et al., 2002b). The $\Delta^{56}Fe_{OI-melt}$ value thus depends on temperature and melt Fe³⁺/ Σ Fe and is incrementally adjusted with every 1% addition of olivine until melt Mg[#] = 0.72. However, because the $\Delta^{56}Fe_{mantle}$ peridotite-melt value calculated using the above force constants can only account for 1/3 of the isotopic difference between MORBs and mantle peridotites (Dauphas et al., 2014), we multiply all calculated $\Delta^{56}Fe_{OI-melt}$ by 3, following He et al. (2019). The final calculated $\Delta^{56}Fe_{OI-melt}$ values range from -0.11% to -0.07%, which are consistent with those estimated from natural samples (Williams et al., 2005; Weyer and Ionov, 2007; Weyer and Seitz, 2012). The calculated primitive Fe isotope compositions ($\delta^{56}Fe_{Prim}$) are used to reflect mantle melting processes and/or mantle source characteristics.

3. RESULTS

The measured Fe isotope data of the seamount lavas are given in Table 1. Different from the globally homogeneous MORB with δ^{56} Fe = +0.11 ± 0.04‰ (2SD; Teng et al., 2013), the seamount lavas show a much larger δ^{56} Fe range from +0.03‰ to +0.36‰. Most of these seamount lavas have δ^{56} Fe values ranging from +0.03‰ to +0.19‰, similar to published MORB data (Nebel et al., 2013; Teng et al., 2013; Chen et al., 2019), except for two alkali basalts (R13-1 & R15-1) which have the highest δ^{56} Fe values of 0.36‰ and 0.34‰, respectively, exceeding the known range of MORB (Fig. 2a).

4. DISCUSSION

4.1. Limited effect of seafloor alteration on the Fe isotope compositions of the EPR seamount lavas

Rouxel et al. (2003) revealed significant influence of seafloor alteration on the Fe isotope compositions of oceanic basalts, with altered basalts showing elevated δ^{56} Fe values, which was explained by the preferential leaching of Fe²⁺ and light Fe isotopes during alteration. Therefore, it is necessary to discuss the influence of seafloor alteration on the Fe isotope compositions of these EPR seamount basalts, especially the two alkali basalts with extremely high δ^{56} Fe values (R13-1 & R15-1). However, there are several evidence proving that the seafloor alteration has no influence on the chemical and Fe isotopic compositions of these seamount basalts. First, weathered faces were filtered out and only fresh glass chips were selected during the sample preparation. Second, electron microprobe analyses gave the totals of major elements > 99% for the EPR seamount basalts (Table S1; Batiza et al., 1990), indicating limited seafloor alteration on these samples. Third, the δ^{56} Fe values of these basalts show both positive correlations with alkali elements (e.g., K) and high field strength elements (HFSE; e.g., Nb) (Fig. A1), with the two alkali basalts with highest δ^{56} Fe values showing highest K and Nb contents, which further exclude the possible influence of seafloor alteration on the Fe isotope compositions of these basalts. This is because K and Nb are both incompatible during magmatic



Fig. 2. (a) Measured δ^{56} Fe values (δ^{56} Fe_{meas}) vs. Mg[#] of the near-EPR seamounts. There is no correlation between δ^{56} Fe_{meas} and Mg[#]. (b) Primitive δ^{56} Fe values (δ^{56} Fe_{prim}) corrected for the effect of olivine fractionation. For comparison, the raw δ^{56} Fe and Mg[#] values of various types of MORB from literature (Nebel et al., 2013; Teng et al., 2013; Chen et al., 2019) and those compiled in Sossi et al. (2016) are plotted in figure (a) (E = enriched; T = transitional; N = normal). The corrected δ^{56} Fe values of various types of MORB are plotted for comparison in figure (b). Because the MORB samples in Chen et al. (2019) were highly evolved with Mg[#] of 0.22–0.61, we did not correct the δ^{56} Fe values of these samples for the effect of fractional crystallization of olivine. The error bars on each data point represent two standard deviation. Green field represents the primary N-MORB average.

processes, whereas K is mobile and Nb is immobile during seafloor alteration processes (e.g., Hart et al., 1974). Therefore, the positive correlations of δ^{56} Fe values with both K and Nb contents contradict the influence of seafloor alteration on these samples.

4.2. Iron isotope signature of primitive EPR seamount lavas

Studies on basalts from Kilauea Iki lava lake, Hawaii presented evidence that Fe isotopes can be fractionated during fractional crystallization with heavy Fe isotopes being more enriched in the more differentiated (lower MgO) melt (Teng et al., 2008). This was explained that olivine phenocrysts have lower δ^{56} Fe values than their parental melts because of both equilibrium and kinetic Fe isotope fractionation, and fractional crystallization of olivine results in progressively isotopically heavier Fe in the residual melt (Teng et al., 2008). Similar trends of Fe isotope fractionation during magma differentiation have also been observed in MORB (Chen et al., 2019) and magmatic rocks in various settings (Schuessler et al., 2009; Sossi et al., 2012; Teng et al., 2013; Zambardi et al., 2014; Williams et al., 2018). These studies provide solid evidence for mineral (olivine and clinopyroxene) fractional crystallization to produce variably heavy Fe isotope compositions in the evolved melt. Even though there is no correlation between δ^{56} Fe values with indices of magma differentiation (e.g., Mg[#]) observed in these EPR seamount basalts (Fig. 2a), indicating that fractional crystallization is less significant in explaining these highly varied δ^{56} Fe values, correction for the effect of fractional crystallization is still necessary to explore the Fe isotope composition of primitive EPR seamount lavas.

After correction for the effect of fractional crystallization of olivine (see Section 2.3), the δ^{56} Fe_{prim} values of these EPR seamount lavas still show a large range from +0.02‰ to +0.34‰ (see Table 1 for corrected δ^{56} Fe_{prim}). It should be noted that the samples selected for Fe isotope analysis are relatively less evolved (Mg[#] = 0.57–0.72), and so the difference between measured and corrected δ^{56} Fe values (Δ^{56} Fe) are limited within 0.03‰, which is within the analytical errors. The EPR seamount lavas show δ^{56} Fe_{prim} values ranging from average N-MORB like (δ^{56} Fe_{prim} N-MORB = 0.09 ± 0.04‰) to higher values similar to reported enriched MORB (E-MORB; δ^{56} Fe_{prim} E-MORB = 0.12 ± 0.06‰) with the two alkali basalts showing highest δ^{56} . Fe_{prim} values of 0.34‰ and 0.32‰, respectively (Fig. 2b).

4.3. Can partial melting process explain the highly varied δ^{56} Fe_{prim} values?

Studies have found that mantle peridotites from different localities and tectonic settings display a negative correlation between δ^{56} Fe and Mg[#], with mantle peridotites that had experienced the highest extent of melt extraction (highest Mg[#]) showing the lightest Fe isotope compositions (e.g., Weyer and Ionov, 2007; Williams and Bizimis, 2014). This implies that Fe isotopes could be fractionated during partial melting, with heavier Fe preferentially entering the melt, leaving an isotopically light residue. This conclusion was further supported by the studies of abyssal peridotites (Craddock et al., 2013), which represent the melting residue

0.45

0.40

0.35

0.30

0.15

0.10

0.05

0.00

-0.05

0

1

2

3

[Sm/Yb]_N

8 0.25

8⁵⁶Fe_{prim} 1 0.20

for MORB. The abyssal peridotites have on average the same Fe isotope compositions as chondrites with δ^{56} -Fe_{abyssal peridotites} = 0.01‰ (Craddock et al., 2013), systematically lower than that of the average MORB $(\delta^{56} \text{Fe}_{\text{MORB}} = 0.11\%; \text{Teng et al., 2013})$. The preferential residence of heavy Fe isotopes in the melt indicates that the heavier Fe (e.g., 56 Fe) behaves more incompatible than the lighter Fe (e.g., 54 Fe). This is because the heavier Fe (i.e., ⁵⁶Fe vs. ⁵⁴Fe) is preferentially incorporated in Fe-O bonds predominated by higher valent Fe (i.e., Fe^{3+} vs. Fe^{2+}) at magmatic conditions (Polyakov and Mineev, 2000; Polyakov et al., 2007; Schuessler et al., 2007; Shahar et al., 2008; Dauphas et al., 2009, 2014; Schoenberg et al., 2009), i.e., there is a physiochemical 56 Fe-Fe³⁺ affinity, and because Fe³⁺ (vs. Fe²⁺) is more incompatible during mantle melting (Canil et al., 1994; Mallmann and O'Neill, 2009), the melt is thus expected to have higher Fe^{3+}/Fe^{2+} and high $\delta^{56}Fe$ (high ${}^{56}Fe/{}^{54}Fe$). The measurement of the force constant of iron bonds in olivine and basaltic glasses under various redox conditions demonstrated significant equilibrium Fe isotopic fractionation between Fe^{3+} and Fe^{2+} at mantle temperatures (Dauphas et al., 2014), which allows quantitative modelling of Fe isotopic fractionation during mantle melting (e.g., Dauphas et al., 2014; He et al., 2019).

On the other hand, melts derived from different source lithologies (peridotite versus garnet pyroxenite) were suggested to show different extent of Fe isotopic fractionation relative to their sources (Williams and Bizimis, 2014; Sossi and O'Neil, 2017). Partial melting modelling for peridotitic and garnet pyroxenitic mineralogies with the same initial Fe isotope compositions showed that at the same degree of partial melting, melts from the garnet pyroxenite have heavier Fe isotope compositions than melts from the peridotite (Williams and Bizimis, 2014). This is attributed to the greater contribution of clinopyroxene to the melt during partial melting of the garnet pyroxenite and the large Fe isotope fractionation between clinopyroxene and garnet $(\delta^{56} \text{Fe})$ values of major mantle minerals: garnet < olivine < opx < cpx < spinel; Beard and Johnson, 2004; Williams et al., 2004, 2005, 2009; Zhao et al., 2010, 2017; Johnson et al., 2020). In addition, the natural pyroxenites show heavier Fe isotope compositions than the peridotites (Williams and Bizimis, 2014), which can further enhance the enrichment of heavy Fe isotopes in the derived melts. Therefore, mantle pyroxenite components have been popularly invoked to explain the heavy Fe isotope compositions observed in the worldwide OIB and MORB influenced by mantle plume materials (Williams and Bizimis, 2014; Konter et al., 2016; Sossi and O'Neil, 2017; Nebel et al., 2019; Gleeson et al., 2020).

To investigate whether partial melting process could produce the highly varied Fe isotope compositions of the EPR seamount lavas, we modelled the melt δ^{56} Fe variation during partial melting in the peridotite and garnet pyroxenite facies, using the methods described in He et al. (2019) and Williams and Bizimis (2014), respectively (Fig. 3) (Detailed modelling parameters and results are in Table S2). The samples with $\delta^{56}Fe_{prim} < \sim 0.20\%$ and $[Sm/Yb]_N < \sim 1.5$ values can be quantitatively modelled

Fig. 3. Comparison of the δ^{56} Fe_{prim} and $[Sm/Yb]_N$ (primitive mantle normalized Sm/Yb) of the EPR seamount samples with those of modelled melts from different source lithologies (spinel peridotite and garnet pyroxenite). Modellings of the melt δ^{56} Fe variation during partial melting in the spinel peridotite and garnet pyroxenite facies follow the methods described in He et al. (2019) and Williams and Bizimis (2014), respectively. For partial melting in the spinel peridotite facies, three different source compositions are applied, ranging from oxidized ($Fe^{3+}/Fe^{2+} = 0.5$) with high δ^{56} Fe value (0.05‰), normal with primitive mantle-like Fe³⁺/Fe²⁻ (0.037; Canil et al., 1994) and $\delta^{5\hat{6}}Fe$ (0.02‰; Weyer and Ionov, 2007), to reduced (Fe³⁺/Fe²⁺ = 0.2) with low δ^{56} Fe value (0.00‰), similar as the modelling in Sossi and O'Neil (2017). ~10-15% partial melting of a normal peridotitic mantle generates melts with δ^{56} Fe similar to the average primary N-MORB (black diamond; δ^{56} Fe_{prim N-MORB} = 0.09 ± 0.04‰). For partial melting in the garnet pyroxenite facies, Δ^{56} Fe_{garnet-cpx} of -0.15‰ and Δ^{56} Fe_{melt-cpx} of 0.00% are used. A garnet pyroxenite source with average MORB-like δ^{56} Fe (~0.11‰; Teng et al., 2013) cannot explain the high δ^{56} Fe_{prim} values of the two alkali basalts which instead require high source δ^{56} Fe value of at least ~0.25‰. Modelling of melt [Sm/Yb]_N variation by batch melting of a depleted mantle (Salters and Stracke, 2004) under spinel peridotite facies uses initial source modes and incongruent melting relationship from Niu (1997) and partition coefficients of Sm and Yb from the compilation in Niu et al. (1996). Modelling of melt [Sm/Yb]_N variation by batch melting in the garnet pyroxenite facies uses average N-MORB like source compositions (Sun and McDonough, 1989), initial source modes and melting relationship from Williams and Bizimis (2014) and partition coefficients of Sm and Yb from Pertermann and Hirschmann (2003). See Table S2 for detailed modelling parameters and results.

by partial melting of a spinel peridotite with varied $Fe^{3+}/$ Fe²⁺ (0.02–0.05) and initial δ^{56} Fe values (0.00–0.05‰). However, this model is apparently inadequate to explain the two alkali basalts with extreme $\delta^{56}Fe_{Prim}$ values (up to 0.34%) and strong garnet signature, which comprise the highest $[Sm/Yb]_N$ and lowest Sc contents (18 and 24 ppm, respectively) among all the northern EPR samples (Niu et al., 2002a). Only if unreasonably high source Fe^{3+}/Fe^{2+} ratios (e.g., >0.2) were used in the modelling can such elevated melt δ^{56} Fe values be achieved. Partial melting in the garnet peridotite facies may explain the garnet signature of the two alkali basalts (not shown) but is still

Garnet pyroxenite facies

(A56Fe_

4

= - 0.15%)

5%

5

6

inadequate to explain their high δ^{56} Fe_{Prim} values, because compared with melting in the spinel peridotite facies, melting in the garnet peridotite facies can produce limited increase in melt δ^{56} Fe ($\leq 0.01\%$) (Sossi and O'Neil, 2017).

On the other hand, partial melting in the garnet pyroxenite facies can readily explain the high [Sm/Yb]_N ratios of the two alkali basalts (Fig. 3) and also has the potential to explain their high δ^{56} Fe_{Prim} values, considering that melts from garnet pyroxenites have heavier Fe isotope compositions than those from peridotites (Williams and Bizimis, 2014: Sossi and O'Neil, 2017). With certain initial and melting modes, the δ^{56} Fe values of melts from garnet pyroxenites depend mainly on the source δ^{56} Fe values, the fractionation factor between garnet and clinopyroxene $(\Delta^{56} \text{Fe}_{\text{garnet-cpx}})$ and the fractionation factor between melt and clinopyroxene (Δ^{56} Fe_{melt-cpx}) (Williams and Bizimis, 2014; He et al., 2017; Sossi and O'Neil, 2017). Following He et al. (2017), we use Δ^{56} Fe_{garnet-cpx} (\sim -0.38 × 10⁶/T²) measured in natural samples (Beard and Johnson, 2004; Williams et al., 2009; Williams and Bizimis, 2014). At the temperature of mantle melting beneath ocean ridges (~1300 °C; McKenzie and Bickle, 1988), Δ⁵⁶Fegarnet-cpx is $\sim -0.15\%$. Δ^{56} Fe_{melt-cpx} is assumed to be zero in our modelling, following Williams and Bizimis (2014), which is consistent with the observation in natural samples (Weyer and Seitz, 2012). The modelling results show that assuming an average MORB-like source δ^{56} Fe values (0.11‰; Teng et al., 2013), partial melting in the garnet pyroxenite facies is still inadequate to explain the high δ^{56} Fe_{Prim} values of the two alkali basalts, which instead require source δ^{56} Fe_{Prim} values of at least $\sim 0.25\%$ (Fig. 3).

Based on the above modelling and discussion, we conclude that (1) partial melting in the spinel peridotite facies can explain neither the high $[Sm/Yb]_N$ nor the high δ^{56} -Fe_{Prim} values of the EPR seamount basalts, and (2) partial melting in the garnet pyroxenite facies can explain the high [Sm/Yb]_N of the two alkali basalts, whereas abnormally high source δ^{56} Fe values are required to explain their high δ^{56} Fe_{Prim} values. Therefore, significant Fe isotope heterogeneity must exist in the mantle source of the EPR seamount lavas. The positive correlation between δ^{56} Fe_{Prim} and $[Sm/Yb]_N$ in Fig. 3 can be best explained by mixing between melts from a spinel peridotitic source and those from a garnet pyroxenitic source pre-enriched in heavy Fe isotopes prior to the major melting event. This is in fact consistent with the conclusion of previous studies that significant chemical and lithological heterogeneities exist in the northern EPR mantle source (Niu and Batiza, 1997; Castillo et al., 2000; Niu et al., 1999, 2002a; Zhang et al., 2016; also see below).

4.4. Mantle heterogeneity beneath the northern EPR in element, radiogenic and Fe isotope compositions

Studies on the northern EPR axial lavas (e.g., Niu et al., 1999) and seamount lavas on the ridge flanks (Niu and Batiza, 1997; Niu et al., 2002a; Zhang et al., 2016) have revealed a mantle source with heterogeneous element and Sr-Nd-Pb-Hf radiogenic isotope compositions. Compared with the axial lavas, the seamount lavas preserve more

information of the source compositional heterogeneity, showing larger variations in Sr-Nd-Pb-Hf radiogenic isotopes (Fig. 4). This is because the axial lavas experienced significant extent of melt mixing and homogenization within the mantle during melt aggregation towards the ridges and in the crustal magma chambers before their eruptions (O'Hara, 1977, 1985; Macdougall and Lugmair, 1985; Sinton and Detrick, 1992; Batiza et al., 1996; Niu, 1997). In contrast, the seamount lavas represent much smaller volumes of melts tapped locally by volcanos lacking steady-state magma chambers and thus avoid efficient mixing (Batiza and Vanko, 1984; Zindler et al., 1984; Niu et al., 2002a).

The EPR seamount lavas define significant correlations among Sr-Nd-Pb-Hf isotopes (Fig. 4). Two samples with highest Pb isotope ratios which were described as HIMUlike in Niu et al. (2002a) and two alkali basalts show most enriched radiogenic isotope compositions. These correlations can in first order be explained as reflecting meltinginduced mixing of a two-component mantle with the enriched component dispersed as physically distinct domains in the depleted mantle matrix having radiogenic isotope compositions similar to the ultra-depleted Garrett transform lavas (Fig. 4) (Niu et al., 2002a). The coupled correlations of radiogenic isotopes with incompatible elements (e.g., Ba, Nb, Rb, La) and their ratios (e.g., Rb/Sr, Ba/Zr, [La/Sm]_N, [Sm/Yb]_N) indicate that both the enriched and depleted mantle components are ancient and have independently developed their isotopic characteristics (Fig. A2). However, it should be noted neither of the two components is necessarily homogeneous in chemical and isotopic compositions. For example, the HIMU-like samples and two alkali basalts show similar Sr whereas distinct Pb isotope compositions (Fig. 4d), and the samples with similarly enriched Sr isotope compositions (⁸⁷Sr/⁸⁶Sr values of ~ 0.7029 in Fig. A2) show large variations in element abundances and their ratios (e.g., Ba/Zr = 0.67-1.97, $[La/Sm]_N = 1.57 - 3.73).$

As we discussed above, large Fe isotope heterogeneity should exist in the mantle source beneath the northern EPR, which is corroborated by the significant correlations between δ^{56} Fe_{prim} values and Sr-Nd-Pb-Hf isotope ratios (Fig. 5). The samples with enriched radiogenic isotope compositions also show higher δ^{56} Fe_{prim} values. This is in contrast with our recent Fe isotope study on the MORB lavas at 10° 30'N, whose Fe isotope variation is mainly controlled by varying extent of fractional crystallization, instead of mantle heterogeneity (Chen et al., 2019). After correction for the effect of fractional fractionation, these MORB lavas show a uniform δ^{56} Fe_{prim} value of ~0.04‰ (Chen et al., 2019). However, as we discussed above, MORB lavas experienced higher extent of melt mixing and homogenization within the mantle and in the crustal magma chambers. During this process, information of mantle heterogeneity in Fe isotopes and radiogenic isotopes could be erased. This is consistent with the limited variations in Sr-Nd-Pb isotopes of the MORB lavas at 10° 30'N (Fig. 4). It is therefore not surprising that these MORB lavas with homogeneous radiogenic isotope compositions show uniform δ^{56} Fe_{prim} values.



Fig. 4. Plots of the EPR seamount lavas in the Sr-Nd-Pb-Hf isotopic spaces. For comparison, EPR axial basalts at 10° 30'N (Regelous et al., 1999) and 11° 20'N (Niu et al., 1999) and lavas from Garrett Transform Fault (Wendt et al., 1999) are also plotted. The Sr-Nd-Pb isotope data of the EPR seamount lavas are from Niu et al. (2002a). The Hf isotope data are from Zhang et al. (2016).

It should be noted that similar to the abundances of incompatible elements and their ratios, the δ^{56} Fe_{prim} values also show large variation (+0.10‰ to +0.34‰) in the samples with similar ⁸⁷Sr/⁸⁶Sr value of ~0.7029 (Fig. 5). This reflects heterogeneous Fe isotope compositions in the enriched mantle component and more importantly implies a common process that results in the fractionation of Fe isotopes and variation of incompatible element abundances in the enriched mantle component.

²⁰⁶Ph/²⁰⁴Ph

Indeed, the large δ^{56} Fe_{prim} variation of the EPR seamount lavas correlates significantly with the major and trace element abundances (Fig. 6). The subscript 72 in Fig. 6 refers to the corresponding oxides (i.e., Na₂O, TiO₂, Al₂O₃ and FeO) corrected for the effects of crustal level differentiation to constant Mg[#] (Mg/[Mg + Fe²⁺] × 100) = 72, following the method in Niu et al. (1999) and Niu and O'Hara (2008). The corrected major element compositions thus reflect the effects of mantle sources and processes because basaltic melts with Mg[#] \geq 72 are in equilibrium with mantle olivine (the corrected major element data are in Table S3). Because their significant correlations with Sr-Nd-Pb-Hf radiogenic isotopes (Fig. A3), the

corrected major element compositions reflect, to a great extent, the compositional variation of fertile mantle sources (Niu et al., 2002a). Apparently, heavier Fe isotope composition (higher δ^{56} Fe_{prim}) is associated with melts derived from an enriched source endmember with high Na₂O, TiO₂, Al₂O₃ (also high P₂O₅; Fig. A4), low FeO (also low MgO, CaO and CaO/Al₂O₃; Fig. A4), and high incompatible element abundances. It follows that the origin of the Fe isotope heterogeneity in the source of these lavas must have been caused by the same processes that have led to the mantle major and trace element heterogeneities at least in the northern EPR mantle.

⁸⁷Sr/⁸⁶Sr

4.5. Origin of the enriched mantle component with heavy Fe isotope composition

4.5.1. A low-degree melting origin of the enriched mantle component

As shown in Fig. 7, δ^{56} Fe_{prim}, together with Sr-Nd-Pb-Hf radiogenic isotopes, correlates increasingly better with the abundances of the progressively more incompatible elements. This indicates that the enriched component must



Fig. 5. Correlated variations of δ^{56} Fe_{prim} (δ^{56} Fe values corrected for the effect of fractional crystallization of olivine) with Sr-Nd-Pb-Hf radiogenic isotopes of the EPR seamount lavas.

have a magmatic origin. Elemental ratios such as Zr/Hf, Nb/Ta, Rb/Cs and Ce/Pb are often considered as constant in most magmatic processes because of the closely similar incompatibility between the element in the numerator and the element in the denominator (Hofmann and White, 1983; Hofmann et al., 1986; Newsom et al., 1986), but it is the seamount data that demonstrate for the first time that the element in the numerator is actually more incompatible than the element in the denominator (Niu and Batiza, 1997). The samples having progressively sub-chondritic Zr/Hf (<~36.3) and Nb/Ta (<~17.6), and Rb/Cs (<~80) and Ce/Pb (<~25) lower than mean oceanic basalt (Sun and McDonough, 1989) reflect their mantle source with prior melt extraction/depletion. The samples having higher of these ratios reflect their source with prior low-degree (low-F) melt enrichment/metasomatism. This is a straightforward concept because the effects of elemental fractionation between elements with subtle incompatibility difference can be readily seen in the melting residues, but not detectable in the melt unless the extent of melting is rather small (i.e., the low-F effect) (Niu et al., 2002a). Hence, the positive correlations of δ^{56} Fe_{prim} values (also Sr-Nd-Pb-Hf isotopes; Fig. A5) with elemental ratios of Zr/Hf, Nb/Ta, Rb/Cs and Ce/Pb (Fig. 8) reflect that the enriched component with

heavy Fe isotope composition must be of a low-F melt origin. Indeed, the major element characteristics of this low-F melt (high Na₂O, TiO₂, Al₂O₃ and low CaO, FeO, MgO and thus CaO/Al₂O₃), as inferred from the data trends in Fig. 3 and Fig. A4, are in agreement with experimental results (Baker et al., 1995) and glass inclusions in mantle xenoliths (Schiano et al., 1998; Schiano and Bourdon, 1999).

As mentioned above, because Fe^{3+} (vs. Fe^{2+}) is more incompatible during mantle melting (Canil et al., 1994; Mallmann and O'Neill, 2009) and because heavy Fe isotopes (e.g., ⁵⁶Fe) favor Fe³⁺ complexes (Polyakov and Mineev, 2000; Polyakov et al., 2007), the heavier Fe isotopes (e.g., ⁵⁶Fe) behave more incompatible than the lighter Fe isotopes (e.g., ⁵⁴Fe). Therefore, melt is expected to have heavier Fe isotope compositions than the source mantle with melt from lower degree of partial melting having higher δ^{56} Fe. We thus suggest that the mantle source region of the EPR seamount lavas must have been pre-enriched/ metasomatized by such a low-F melt component with enriched incompatible element and heavy Fe isotope compositions prior to the major melting event. The enriched (easily melted) low-F melt component with heavy Fe isotope compositions were most likely dispersed as physically



Fig. 6. Correlated variations of δ^{56} Fe p_{rim} (δ^{56} Fe values corrected for the effect of fractional crystallization of olivine) with the abundances of major element oxides corrected for the effects of crustal level fractionation to Mg[#] = 72 (100 × Mg/[Mg + Fe²⁺], i.e., the subscript 72) and incompatible trace elements.



Increasing incompatibility

Fig. 7. Correlation coefficients (R-values) of δ^{56} Fe_{prim} (δ^{56} Fe corrected for the effect of fractional crystallization of olivine) and Sr-Nd-Pb-Hf isotopic ratios with incompatible element abundances of the EPR seamount lavas in the order of increasing relative incompatibility from right to the left. The δ^{56} Fe_{prim} values show progressively better correlations with the more incompatible elements, which suggests that the heavy Fe isotope enriched component is of magmatic origin. The significant coupling between Sr-Nd-Pb-Hf isotopic ratios and incompatible element abundances demonstrates the ancient nature of both the depleted and enriched source materials.

distinct domains in the more depleted (refractory) matrix prior to major melting events (Fig. 9a). Varying extent of melting of such a two-component mantle (Fig. 9a) will produce mixing relations in geochemical diagrams as illustrated in Figs. 3–6.

The chemically and physically most likely localities for the low-F melt metasomatism lie within the seismic low velocity zone (LVZ) beneath ocean basins where incipient melt forms with the low-F melt geochemical characteristics enriched in volatiles (e.g., H₂O and CO₂) and incompatible elements (Fig. 9b) (Niu and O'Hara, 2009; Green et al., 2010). Because of ridge-suction induced asthenospheric flow towards ridges (Niu and Hékinian, 2004), these low-F melts can be extracted and directly contribute to the magmatism beneath oceanic ridges (Fig. 9b). In this case, the low-F melts would be young features with enriched incompatible elements but depleted Sr-Nd-Pb-Hf isotopes (Halliday et al., 1995). However, as the low-F melt components in the mantle source of these EPR seamount lavas are ancient (Niu et al., 2002a; see above), there must be a process which kept these low-F melt components chemically isolated to develop long-time integrated radiogenic Sr and Pb isotopes and unradiogenic Nd and Hf isotopes.

One reasonable explanation is that after their generation within the LVZ, these low-F melts can concentrate towards the lithosphere-asthenosphere boundary (LAB) and forms a melt-rich layer beneath the LAB (Fig. 9b). Such a meltrich layer beneath ocean basins has been seismically detected (Kawakatsu et al., 2009), which can metasomatize the overlying growing/thickening lithosphere by crystallizing dikes and veins of garnet pyroxenite lithologies embedded in the mature lithosphere (Fig. 9b) (Niu and O'Hara, 2003; Niu, 2008; Niu and Green, 2018). The high Fe^{3+} and δ^{56} Fe signature of the low-F melts will inevitably be preserved in these dikes and veins. Indeed, studies have revealed high $Fe^{3+}/\Sigma Fe$ (up to 0.40) in clinopyroxene megacrysts that were crystallized from basaltic melts under high pressures (McGuire et al., 1989; Dyar et al., 1996; Yang and McCammon, 2012). Recent studies on garnet pyroxenites from Oahu, Hawaii with a high-pressure cumulate origin showed high δ^{56} Fe values (0.05–0.16‰) of these mantle lithologies (Williams and Bizimis, 2014). Such a metasomatized oceanic lithosphere will finally be subducted into the mantle at subduction zones and reside there for a long geologic history before being recycled into the mantle source region and contributing to the incompatible element, Sr-Nd-Pb-Hf isotopes and heavy Fe isotope enriched components in oceanic basalts such as ocean island basalts and seamount lavas in this study (Niu and O'Hara, 2003; Niu et al., 2002a, 2012; Williams and Bizimis, 2014).

4.5.2. Recycled oceanic crust material or carbonated peridotites are not suitable to be the enriched source components of the EPR seamount lavas

Recycled oceanic crust material has long been invoked to be the enriched source endmember of E-MORB (e.g.,



Fig. 8. Correlated variations of δ^{56} Fe_{prim} (δ^{56} Fe values corrected for the effect of fractional crystallization of olivine) with element ratios with elements on the numerator and denominator of each ratio pair having similar incompatibilities (Zr/Hf, Nb/Ta, Rb/Cs and Ce/Pb). The element pairs Zr/Hf, Nb/Ta, Rb/Cs and Ce/Pb do not fractionate from each other in most magmatic processes except in the case of very low-degree (low-F) melting (Niu et al., 2002a), which manifests low-F melt action in causing the observed Fe isotope fractionation.

0.4

15

10

-0.1

Allègre and Turcotte, 1986; Donnelly et al., 2004; Hirschmann and Stolper, 1996; Stracke and Bourdon, 2009). Recently, studies on the OIB lavas also use recycled oceanic crust materials as possible enriched source endmember with heavy Fe isotope compositions (Konter et al., 2016; Nebel et al., 2019). As a represent of partial melt from the depleted sub-ridge mantle, oceanic crust is expected to have heavier Fe isotope compositions than the source mantle (see above). However, in the case of the EPR seamount lavas in this study, the oceanic crust material is not suitable to be the enriched source component because: (1) the oceanic crust endures significant dehydration during subduction with the residual crust highly depleted in water-soluble incompatible elements (e.g., Ba, Rb, Cs, K, Sr, Pb etc.) (Niu et al., 2002a; Niu and O'Hara, 2003; Workman et al., 2004). If the recycled oceanic crust was geochemically responsible for those enriched EPR seamount lavas, the latter should be highly depleted

0.2

 $\delta^{56} \mathrm{Fe}_{\mathrm{prim}}$ (%)

0.3

52

47

42

37

32

27 20

18

16

14

12

10

-0.1

0.0

0.1

in these elements, which is not observed (Figs. 6 and 7); (2) the oceanic crust is on average product from high degree (>10%; Niu, 1997) of mantle melting with essentially constant Zr/Hf (~36.3), Nb/Ta (~17.6), Rb/Cs (~80) and Ce/Pb (~25) ratios (Hofmann and White, 1983; Hofmann et al., 1986; Newsom et al., 1986; Sun and McDonough, 1989). However, the enriched endmember of these EPR seamount lavas has higher values of these element ratios (Fig. 8), indicating an apparent low-F (vs. high-F) origin (see above). Therefore, although recycled oceanic crust in the mantle source region could explain the heavy Fe isotope composition of mantle-derived melts, it cannot explain the incompatible element characteristics of the EPR seamount lavas.

0.2

 $\delta^{56} Fe_{prim}$ (%)

0.3

0.4

0.1

0.0

The Cenozoic basalts in eastern China were reported to show variably heavy Fe isotope compositions with δ^{56} Fe values of 0.10–0.29‰, which were explained as melts from an extremely oxidized peridotitic source (He et al., 2019). (a)



Heating from basal thermal boundary layer - CMB?

Together with evidence from Mg and Zn isotopes, such an oxidized peridotitic source was argued to be formed by the metasomatism of recycled carbonates whose reduction to diamonds in the deep mantle caused the oxidization of surrounding mantle (He et al., 2019). Partial melting experiments and studies on natural samples revealed that melts from a carbonated mantle source show depletion in HFSEs (e.g., Ti, Zr, Hf) (e.g., Yaxley et al., 1991; Baker and Wyllie, 1992; Hauri et al., 1993; Dasgupta et al., 2009). Indeed, we have observed more prominent negative Ti anomalies (Ti/Ti* < 1; Ti/Ti* = $2 \times Ti_{PM}/[Sm_{PM} + Eu_{PM}]$) in the EPR seamount samples with heavier Fe isotope compositions (Fig. A6), which is consistent with their derivation from a carbonated mantle source. However, the seamount samples with heavier Fe isotope compositions show more prominent positive Zr anomalies $(Zr/Zr^* > 1; Zr/Zr^* =$ $2 \times Zr_{PM}/[Nd_{PM} + Sm_{PM}])$, which is in contrast with their derivation from a carbonated mantle source (Fig. A6). In fact, modelling of Ti/Ti* and Zr/Zr* variations during partial melting under spinel peridotite facies shows that the low-degree melts are characterized by negative Ti but positive Zr anomalies (Fig. A7), consistent with the modelling results in Johnson (1998). Therefore, the negative Ti but positive Zr anomalies observed in the EPR seamount lavas with heavy Fe isotope compositions are consistent with their derivation from an enriched source component having a low-F melt origin, instead of a carbonate metasomatic origin.

4.5.3. A three-stage model to generate the EPR seamount lavas with elevated $\delta^{56} Fe_{prim}$ values

The major and trace element characteristics of the EPR seamount lavas with heavy Fe isotope compositions all indicate an enriched mantle source component generated by low-F melting, which most likely happened in the LVZ beneath the oceanic lithosphere. Modelling of the Fe isotopic fractionation during melting of a normal peridotitic source with initial δ^{56} Fe of 0.02‰ (Weyer and Ionov,

2007) and Fe^{3+}/Fe^{2+} of 0.037 (Canil et al., 1994) showed that the low-F (e.g., 1%) melt has δ^{56} Fe of ~0.15% (Fig. 3). Such a δ^{56} Fe value of the enriched source component, although can explain the Fe isotope compositions of most seamount samples, still cannot satisfy the high source δ^{56} Fe value (at least ~0.25‰; Fig. 3) required by the two alkali basalts (Fig. 10). Considering the two alkali basalts have δ^{56} Fe_{prim} values higher than, whereas 87 Sr/ 86 Sr values similar to the other samples with ⁸⁷Sr/⁸⁶Sr values of ~ 0.7029 (Fig. 10), a recent secondary enrichment process in the source region prior to the major melting event is required. Such a process is expected to cause further enrichment of heavy Fe isotopes and incompatible elements with little change of the radiogenic isotope compositions in the enriched source component, which is also required to explain the large variations of element abundances and their ratios in the samples with similar ⁸⁷Sr/⁸⁶Sr values of ~0.7029 (Fig. A3).

To explain the Fe isotope compositions of the Pitcairn OIB lavas, Nebel et al. (2019) provided a two-stage model in which low-F melts from the recycled crustal eclogite react with the ambient peridotitic mantle to form EM1pyroxenite with heavier Fe isotope compositions in stage 1 and low-F melting of this pyroxenite component generates the OIB lavas with high δ^{56} Fe values in stage 2. In this study, the recycled low-F melt component in the garnet pyroxenite lithologies can partially melt and react with the ambient mantle to produce a secondary garnet pyroxenite with higher δ^{56} Fe value (~0.25‰; Fig. 10) and higher incompatible element abundances than, but similar radiogenic isotope compositions as its precursor (Yaxley and Green, 1998; Herzberg, 2006; Sobolev et al., 2007). During the major melting event, because the pyroxenite components (secondary and its precursor) have lower solidus temperature (T_{solidus}), they will melt preferentially and generate enriched lavas represented by the two alkali basalts and other samples with high ⁸⁷Sr/⁸⁶Sr values (Fig. 10). Increasing extent of melting of this two-component mantle (garnet

Fig. 9. (a) Schematic illustration of the concept of a two-component heterogeneous mantle source with enriched endmember dispersed in the refractory and predominantly depleted peridotite matrix. Fig. 9a[I] and 9a[II] illustrate the enriched component with different sizes and shapes. Fig. 9a[III] illustrates the enriched component in the heterogeneous domains. Because the enriched dikes/veins have lower solidus temperature (T_{solidus}), they will melt preferentially during melting of such a two-component mantle. As a result, the enriched component dominates the composition of melt produced in the early stages and decreases with further melting as a result of dilution. Concurrently, the source region is progressively more depleted in the enriched dikes/veins, and further melting of this depleted source material can only produce melts progressively depleted in volatiles and incompatible elements with decreasing δ^{56} Fe, 87 Sr/ 86 Sr and 206 Pb/ 204 Pb (Fig. 9a[IV]). (b) Schematic representations of our preferred model for the origin of the enriched component in the sources of oceanic basalts. The green layer (Fig. 9b) beneath the LAB indicates the presence of a melt-rich layer (supplied by the rising incipient melts denoted by the green arrowed wavy lines). These incipient melts are originated in the low velocity zone (LVZ) and have low-degree (low-F) melt characteristics with enrichment in incompatible elements and volatiles. During the oceanic lithosphere thickening with age due to heat loss to the surface, these low-F melts can be incorporated at the base of the growing lithosphere, where they crystallize and form fine dikes and veins of garnet pyroxenite lithology enriched in volatiles and incompatible elements with high δ^{56} Fe. This low-F metasomatism is taking place today and likely also in Earth's history. Therefore, the deep portions of the oceanic lithosphere are important reservoirs of volatiles, heavy Fe isotopes and incompatible elements with high Rb/Sr, U/Pb, Nd/Sm and Hf/Lu responsible for their future radiogenic Sr and Pb and unradiogenic Nd and Hf isotopes. Recycling of such metasomatized mantle lithosphere into the asthenospheric mantle source regions of oceanic basalts explains the MORB mantle compositional heterogeneity. The blue-arrowed dash lines indicate ridge-suction induced asthenosphere flow towards ridges (Niu and Hékinian, 2004) with small arrows indicating sub-ridge extraction of dominantly depleted (red) and minor enriched (green) melts parental to MORBs. Modified from Niu and Green (2018). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)



Fig. 10. A three-stage model to explain the varied $\delta^{56}Fe_{prim}$ and ${}^{87}Sr/{}^{86}Sr$ of the EPR seamount lavas. Stage 1: low-F (1%) melting of a peridotitic mantle in the LVZ with present-day DMM-like Rb and Sr values (Salters and Stracke, 2004), ${}^{87}Sr/{}^{86}Sr$ of (~0.7020) and $\delta^{56}Fe$ of 0.02% (Weyer and Ionov, 2007), 0.5 Ga ago. The low-F melt that was preserved as dikes/veins of garnet pyroxenite lithologies at the bottom of oceanic lithosphere would have $\delta^{56}Fe$ of ~0.15% and present-day ${}^{87}Sr/{}^{86}Sr$ of ~0.70303. The DMM ${}^{87}Sr/{}^{86}Sr$ value 0.5 Ga ago (~0.7020) is calculated by using the present-day DMM ${}^{87}Rb/{}^{86}Sr$ (Salters and Stracke, 2004) and ${}^{87}Sr/{}^{86}Sr$ estimated from the depleted Garrett Transform lavas (~0.7022; Wendt et al., 1999). Stage 2: low-degree melting of the recycled garnet pyroxenite and melt reaction with the ambient periodotite prior to the major melting event to form a secondary garnet pyroxenite with highe $\delta^{56}Fe$ of ~0.25%. Stage 3: partial melting of the secondary garnet pyroxenite generated melts with high $\delta^{56}Fe$ (e.g., the two alkali basalts). See detailed modelling parameters and results in Table S4.

pyroxenite and depleted mantle matrix) cause the variations of elements, Fe isotope and Sr-Nd-Pb-Hf isotope compositions observed in the EPR seamount lavas (Fig. 9a).

We therefore suggest a three-stage model (Fig. 10) to explain the varied δ^{56} Fe_{prim} values of the EPR seamount lavas: Stage 1, the low-F melts with enriched incompatible elements and heavy Fe isotope compositions (δ^{56} Fe $\sim 0.15\%$) were formed in the LVZ and metasomatized the overlying oceanic lithosphere by crystallizing dikes and veins of garnet pyroxenite lithologies. Such a metasomatized oceanic lithosphere was finally subducted into the mantle at subduction zones and resided there for a long geologic history before being recycled into the mantle source region of the EPR seamount lavas; Stage 2, partial (low-F) melts from the recycled ancient (~ 0.5 Ga) garnet pyroxenite component reacted with the ambient mantle to form a secondary garnet pyroxenite with higher δ^{56} Fe values (~0.25‰) than its precursor; Stage 3, melting induced mixing of the secondary garnet pyroxenite and its precursor with depleted mantle matrix generated the EPR seamount lavas with varied δ^{56} Fe_{prim} values (0.02–0.34‰).

We conclude that Fe isotope heterogeneity exists in a two-component EPR mantle, with enriched garnet pyroxenites having variably heavy Fe isotope compositions dispersed in the depleted mantle matrix (Fig. 9). Because Fe is one of the major elements in seafloor basalts and in Earth's mantle, the discovery of the correlated variation of Fe isotopes with other major elements, trace elements and Sr-Nd-Pb-Hf radiogenic isotopes in the EPR seamount lavas is informative towards resolving the fundamental question on whether global MORB major element systematics reflects mantle melting conditions (Gale et al., 2014) or mantle source compositional variation (Niu and O'Hara, 2008; Niu, 2016) or both. Furthermore, Fe is by mass the most abundant metal of the Earth and the new data and understanding presented here provide new insights into the origin of mantle chemical and isotopic heterogeneity in the grand context of chemical differentiation of the Earth.

5. CONCLUSIONS

We report non-radiogenic Fe isotope analyses from <1 m.y. seamount lavas taken from the flanks of the northern East Pacific Rise between 5° and 15°N. These samples show large Fe isotope variation with δ^{56} Fe values (+0.03% to +0.36%) exceeding the known range of MORB (Nebel et al., 2013; Teng et al., 2013; Chen et al., 2019). Such highly varied Fe isotope compositions must reflect mantle source heterogeneity and cannot be explained by seafloor alteration, magma crystallization evolution or mantle partial melting processes.

After correction for the effect of fractional crystallization of olivine, the large δ^{56} Fe_{prim} variation (+0.02‰ to +0.34‰) of the EPR seamount lavas correlates significantly with the abundances of major and trace elements and Sr-Nd-Pb-Hf radiogenic isotopes, with heavier Fe isotope compositions (higher δ^{56} Fe) being associated with melts derived from a more enriched source endmember (e.g., high Na₇₂, Ti₇₂, Al₇₂, ⁸⁷Sr/⁸⁶Sr and ²⁰⁶Pb/²⁰⁴Pb, and low Ca₇₂, Fe₇₂, Mg₇₂, ¹⁴³Nd/¹⁴⁴Nd, ¹⁷⁶Hf/¹⁷⁷Hf and thus Ca₇₂/Al₇₂), showing significant source heterogeneity. δ^{56} Fe_{prim}, together with Sr-Nd-Pb-Hf radiogenic isotopes, correlates increasingly better with the abundances of the progressively more incompatible elements, indicating a magmatic origin of the heavy Fe isotope enriched component. In addition, higher δ^{56} Fe is associated with samples with high Zr/Hf, Nb/Ta, Rb/Cs and Ce/Pb, indicating that the enriched endmember must be of low-degree (low-F) melting origin, because only low-F melting can effectively fractionate elements with subtle differences in incompatibility in each ratio pair (Niu et al., 2002a).

We thus suggest that a low-F melt metasomatism at sites such as the lithosphere-asthenosphere boundary beneath ocean basins can effectively cause the Fe isotope fractionation. Such low-F metasomatic melt will freeze at the base of the growing/thickening lithosphere as dikes and veins of garnet pyroxenite lithologies embedded in the mature lithosphere, and recycling of such metasomatized mantle lithosphere can readily contribute to the Fe isotope heterogeneity in the MORB mantle.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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APPENDIX A. SUPPLEMENTARY MATERIAL

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