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OIB/seamount recycling as a possible process for E-MORB genesis

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[1] This paper deals with the origin of enriched MORB independent from any hot spot activity. Indeed, MORB enrichment was readily attributed to a ridge/hot spot interaction and in absence of identified neighboring hot spot, to more questionable processes (e.g., incipient plume or plume activity residue). More recently, the existence of enriched MORB away from any identifiable hot spot was attributed to different origins (i.e., recycled oceanic crust and/or enriched mantle after subduction metasomatism). Within this frame, we present here a new set of geochemical analyses of major and trace elements and Sr, Nd and Pb isotopes on samples collected by submersible on both intersections of the 15°20'N fracture zone with the spreading axis of the Mid-Atlantic Ridge. This area is characterized by an increasing enrichment of the lava compositions from north to south through the fracture zone. Results show that the geochemical enrichment observed with a different intensity on both sides of the fracture zone is linked to the 14°N topographic and geochemical anomaly. Our modeling shows that both trace element and isotopic compositions are consistent with a binary mixing between the regional depleted MORB mantle source and a recycled OIB/seamount, as previously proposed to explain the observed enrichment at 14°N. This model can also account for other enriched MORB i.e., the 18°–20°S region of the Central Indian Ridge, illustrating that it does not represent an isolated and local process. On the basis of our results and on the DMM isotopic evolution, the age of the recycled OIB/seamount is estimated to be ~250 Ma, suggesting a recycling within the upper mantle. Considering the huge number of ocean islands and seamounts upon the ocean floor, their recycling into the upper mantle is a plausible process to produce enriched MORB.

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Theme: Geochemical Heterogeneities in Oceanic Island Basalt and Mid-ocean Ridge Basalt Sources: Implications for Melting Processes and Mantle Dynamics

1. Introduction

[2] Over the last 3 decades, the increasing number of oceanographic expeditions together with the development of analytical techniques and the understanding of the mantle-derived basalt composition have allowed us to evaluate the chemical structure of the Earth. In particular, the numerous studies made on mid-ocean ridges basalts (MORB) have led to a classification in groups mainly based on trace element enrichment. Historically, the most abundant ones are depleted and were named “normal” MORB (N-MORB). Compared to basalts from other tectonic environments, N-MORB are characterized by low abundances in incompatible elements, low radiogenic Sr and Pb and high radiogenic Nd (and Hf) isotope ratios [e.g., Hofmann, 1997]. Conversely, enriched MORB (E-MORB [Sun *et al.*, 1979; Schilling *et al.*, 1985]) retains their tholeiitic composition but are characterized by more elevated incompatible trace element abundances, trace element ratios and more radiogenic compositions in Sr and Pb isotopes than N-MORB [Saunders *et al.*, 1988]. E-MORBs have long been considered as a rare expression of MORB, but numerous spreading ridge investigations have demonstrated that MORB enrichment does not represent isolated anomalies [Macdougall and Lugmair, 1986; Zindler and Hart, 1986; Leroex *et al.*, 1992; Cousens *et al.*, 1995; Castillo *et al.*, 1998; Niu *et al.*, 1999]. Therefore, these observations highlight the nature and heterogeneity of the upper mantle beneath mid-ocean ridges.

[3] Although the existence of E-MORB has been known for a long time [Schilling *et al.*, 1985], their origin proves difficult to determine within the binary model of enriched hot spot mantle versus depleted MORB upper mantle. Although some E-MORBs are easily explained by hot spot activity in the vicinity [Schilling, 1973; Schilling *et al.*, 1985], the origin of numerous E-MORB found far away from any obvious plume still remains enigmatic. Along the Mid-Atlantic Ridge, several examples of hot spot-unrelated E-MORB have been described, among others at 23°N [Donnelly *et al.*, 2004], 14°N [Bougault *et al.*, 1988; Staudacher *et al.*, 1989; Dosso *et al.*, 1991, 1993; Bonatti *et al.*,

1992; Hémond *et al.*, 2006], or 33°S [Michael *et al.*, 1994]. To explain such enrichment, most of the authors have brought up the presence of an enriched component drowned in the “normal” depleted MORB mantle source (DMM), and several possible origins were proposed: old oceanic crust or sediments [Staudacher *et al.*, 1989], an embryonic mantle plume associated with the triple plate junction [Bougault *et al.* 1988; Dosso *et al.*, 1991], relics of subcontinental mantle [Bonatti *et al.*, 1992], or an unidentified passively embedded chemical heterogeneity in the mantle [Michael *et al.*, 1994]. Nevertheless, none of these have obtained unanimous agreement to explain hot spot-unrelated E-MORB genesis. To account for the origin of E-MORB along the Mid-Atlantic Ridge, Agranier *et al.* [2005] suggested the dispersion of the “South Hemisphere anomalous mantle” (i.e., the DUPAL anomaly [Dupré and Allègre, 1983; Hart, 1984]). Donnelly *et al.* [2004] explained that the enrichment observed in one basalt from the Mid-Atlantic Ridge near Kane fracture zone (23°N MARK area) could result from the presence beneath the ridge of a peridotite previously metasomatised by crust-derived enriched melts during subduction. Finally, using trace element concentrations, Hémond *et al.* [2006] justified the presence of E-MORB at 14°N by recycling beneath oceanic ridges of previously subducted alkali basalts.

[4] In this paper, we present a study of samples recovered on both northern and southern sides of the 15°20'N fracture zone (FZ) during the cruise MODE 98 (R/V Yokosuka, 1998). These samples gave us the opportunity to better constrain the extent of the topographic and geochemical anomaly described at 14°N by Dosso *et al.* [1991, 1993], Bonatti *et al.* [1992], and Hémond *et al.* [2006]. By using binary mixing modeling on new trace element and radiogenic isotope (Sr, Nd and Pb) analyzes, this study shows the plausible implication of recycled OIB-type material to the formation of E-MORB not only at 14°N, but also at a global scale.

2. Geological Background

[5] The 15°20'N FZ is one of the largest transform faults of the Mid-Atlantic Ridge (Figure 1). The

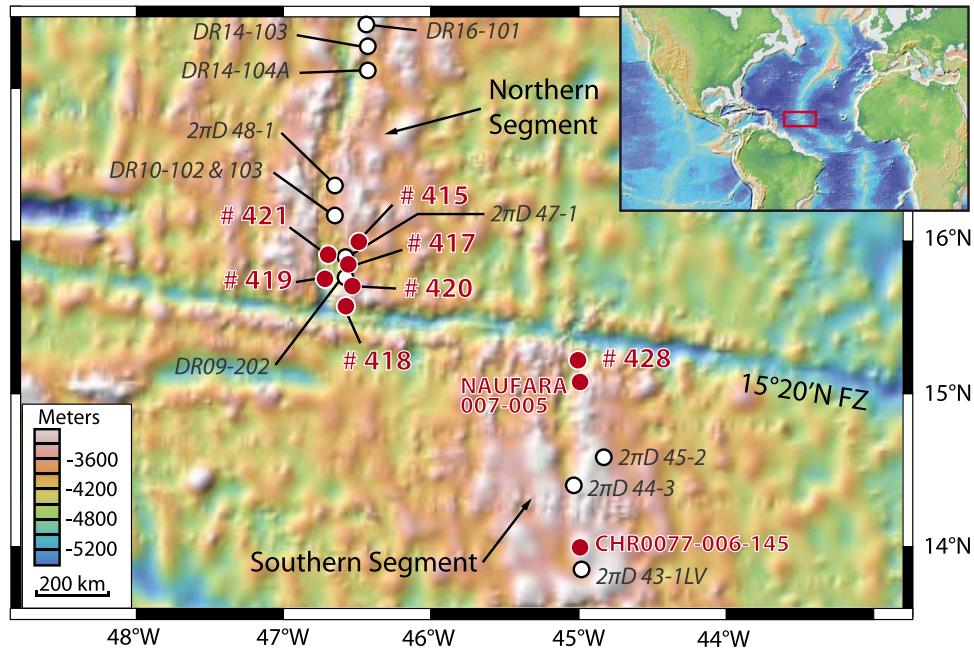


Figure 1. Bathymetry of the study area around the Fifteen–Twenty Fracture Zone, from *Smith and Sandwell* [1997]. Red circles localize the U.S.-Japan MODE 98 dive cruise on board R.V. Yokosuka (dives # 4XX) and complementary samples analyzed during this study (see Table 1). White circles localize additional samples from the 15°20'N FZ area published by *Hémond et al.* [2006] and used in the next figures.

offset is of about 200 km, and the fracture zone 4500 m deep in average. The dextral movement is estimated to 3 cm.yr^{-1} [*Charlou et al.*, 1991; *Castillo et al.*, 1998]. This region is characterized by the presence of enriched MORB [*Dosso et al.*, 1991; *Bonatti et al.*, 1992; *Dosso et al.*, 1993; *Hémond et al.*, 2006] and large but discontinuous ultramafic rock outcrops [*Cannat et al.*, 1997]. These outcrops are not limited to ridge/fracture intersection but extend from 25 km to 35 km on each ridge flank.

[6] The northern segment of the 15°20'N FZ is a complex area with numerous fractures and discrepancies. This segment globally shows slow spreading ridge characteristics, such as a deep axial valley ranging in depth from 3900 m to 4700 m at the intersection with the fracture zone. The ridge flanks are markedly asymmetrical [*Cannat et al.*, 1997]. Ultramafic rocks outcropping north of the 15°20'N FZ are serpentinized peridotites. They have extremely depleted harzburgite-like compositions and represent a high partial melting degree mantle residue [*Cannat et al.*, 1992]. Peridotitic rocks are covered with a thin basaltic layer ($<100 \text{ m}$) locally missing. Basalts crop out in the ridge axis as pillow lavas, tubes or breccia and present fresh glass rims, which are covered by a thin sedimentary

layer ($<10 \text{ cm}$). This attests a fairly recent volcanic activity [*Cannat et al.*, 1997].

[7] South of the 15°20'N FZ, harzburgites tectonically crop out in the internal ridge flanks. Basalts are also erupted along the ridge axis, but no sediments are observed, suggesting a higher magmatic activity compared to the northern segment. They are crosscut by decimeter to meter large fissures revealing recent extensive tectonic activity [*Cannat et al.*, 1997]. The entire segment south of the 15°20'N FZ is characterized by a high topographic bulge centered at 14°N [*Dosso et al.*, 1993].

[8] In summary, basalts are located near the center of the ridge and seem to result from recent eruptions whereas peridotites and gabbros are exposed on the ridge flanks by a conjugate fault set. Uneven topography is attributed to peridotitic massif uplift across the oceanic crust, associated with intrusive gabbroic bodies, suggesting highly heterogeneous lithosphere, both compositionally and mechanically, near the 15°20'N FZ ($<60 \text{ km}$ [*Escartín and Cannat*, 1999]). Away from the fracture zone, the crust is thicker, associated with axis parallel faults and abyssal hills. No peridotite has been dredged, which is consistent with a more homogenous and magmatic crust [*Escartín and Cannat*, 1999].

[9] Samples were picked up using the Shinkai submersible during the Japanese oceanographic cruise MODE 98 on board the R/V Yokosuka (1998). The samples were collected from the spreading axis and from the intersection massifs between the ridge and the transform fault of both northern (dives 415, 417, 418, 419, 420 and 421) and southern (dive 428; Figure 1) segments.

3. Analytical Techniques

[10] Major element contents were obtained by in situ analyses on basaltic glasses by electron probe Camebax SX-50 in Brest, and trace element concentrations using laser ablation sector-field ICP mass spectrometer (laser 213 nm New Wave coupled with Element 2 ICPMS) at the Max-Planck-Institut für Chemie of Mainz (methodology by Jochum *et al.* [2007]). Sr and Nd fractions were separated using standard ion exchange techniques adapted from Richard *et al.* [1976] and White and Patchett [1984]. Analyses were performed using a Triton T1 thermal ionization mass spectrometer (TIMS) at the Institut Universitaire Européen de la Mer of the University of Brest; Pb fractions were separated following Manhès *et al.* [1978] and were analyzed using a Nu Instrument HR multicollector ICP-MS at the École Normale Supérieure of Lyon.

[11] Analyses were done on small basaltic glass chips (2–5 mm) carefully selected under a binocular microscope. All chips with apparent sign of alteration or Mn crust were removed. Selected glass chips were leached with 10 mL of HCl 6 N at 50°C during 24 h. Then, the samples were rinsed twice with distilled water and leachates were saved and analyzed to quantify the effects of the leaching process. Average element concentrations in leachates relative to those of samples do not exceed 8% for trace elements.

[12] Blanks for Sr, Nd and Pb isotope compositions are respectively 250, 90 and 80 pg and are thus negligible compared to the amount of Sr, Nd and Pb concentrations measured here. Sr (NIST-987), Nd (JNd-1) and Pb (NIST-981) isotope ratios of reference materials were run with each sample batch. Average results are 0.710241 ± 9 ($n = 5, \pm 2\sigma$) for $^{87}\text{Sr}/^{86}\text{Sr}$ ratio in NBS-987 reference material, 0.512101 ± 9 ($n = 4, \pm 2\sigma$) for $^{143}\text{Nd}/^{144}\text{Nd}$ ratio in JNd-1 reference material and $16.9327 \pm 5, 15.4865 \pm 6$ and 36.684 ± 2 ($n = 10, \pm 2\sigma$) respectively for $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios in NIST-981 reference material. All these results are consistent with those previously published and

available online in the *Geological and Environmental Reference Materials* (GeoReM, <http://georem.mpch-mainz.gwdg.de/>) database.

4. Results

4.1. Major Elements

[13] Major element data show that our samples are all tholeiitic in composition. It was suggested that major elements in E-MORBs may reveal some variations relative to N-MORBs (e.g., increasing content in Al_2O_3 observed in E-MORBs from the MARK area [Donnelly *et al.*, 2004]). This is not the case for the 15°20'N FZ area. No systematic variations can be identified in major elements versus parameters such as latitude or distance to the 15°20'N FZ (e.g., Mg# ranges from 41 to 47), except for $\text{Na}_{8.0}$ average which decreases from 2.8 for the northern segment to 2.3 for the southern segment.

4.2. Trace Elements

[14] Extended trace element patterns (normalized to primitive mantle) [McDonough and Sun, 1995] are shown by Figure 2. Basalts from the northern segment are slightly enriched in LREE compared to HREE, with average $(\text{La}/\text{Sm})_{\text{N}}$ and $(\text{Dy}/\text{Yb})_{\text{N}}$ ratios of 1.10 and 1.07 respectively. 418R3 is the only sample that presents higher ratios, with $(\text{La}/\text{Sm})_{\text{N}} \sim 1.23$ and $(\text{Dy}/\text{Yb})_{\text{N}} \sim 1.15$. Enrichment for LREE is about 20 to 30 times C1 chondrite contents and less than 20 times for HREE. These samples are therefore intermediate in compositions between N-MORB and more enriched MORB such as basalts from the MARK area or from the MAR 14°N anomaly. Extended trace element patterns show slight enrichment in the most incompatible elements with normalized concentrations ranging from 6 to 10 times primitive mantle contents [McDonough and Sun, 1995]. All samples present negative Pb anomalies, which are characteristic of suboceanic mantle [Hofmann, 1988], but also high positive anomaly in Nb-Ta, and negative anomaly in Sr. Generally speaking, the northernmost samples (e.g., 415R4) are less enriched than samples near the fracture zone (e.g., 418R3; Figure 3), and follow the geochemical gradient described by Dosso *et al.* [1991] from 17°10'N to 14°N (Figure 3).

[15] Samples patterns from the south of the 15°20'N FZ are sub-identical, indicating that samples probably come from the same magmatic body. These basalts are significantly enriched in LREE with average $(\text{La}/\text{Sm})_{\text{N}}$ ratios of ~ 2 . $(\text{Dy}/\text{Yb})_{\text{N}}$ ratios are

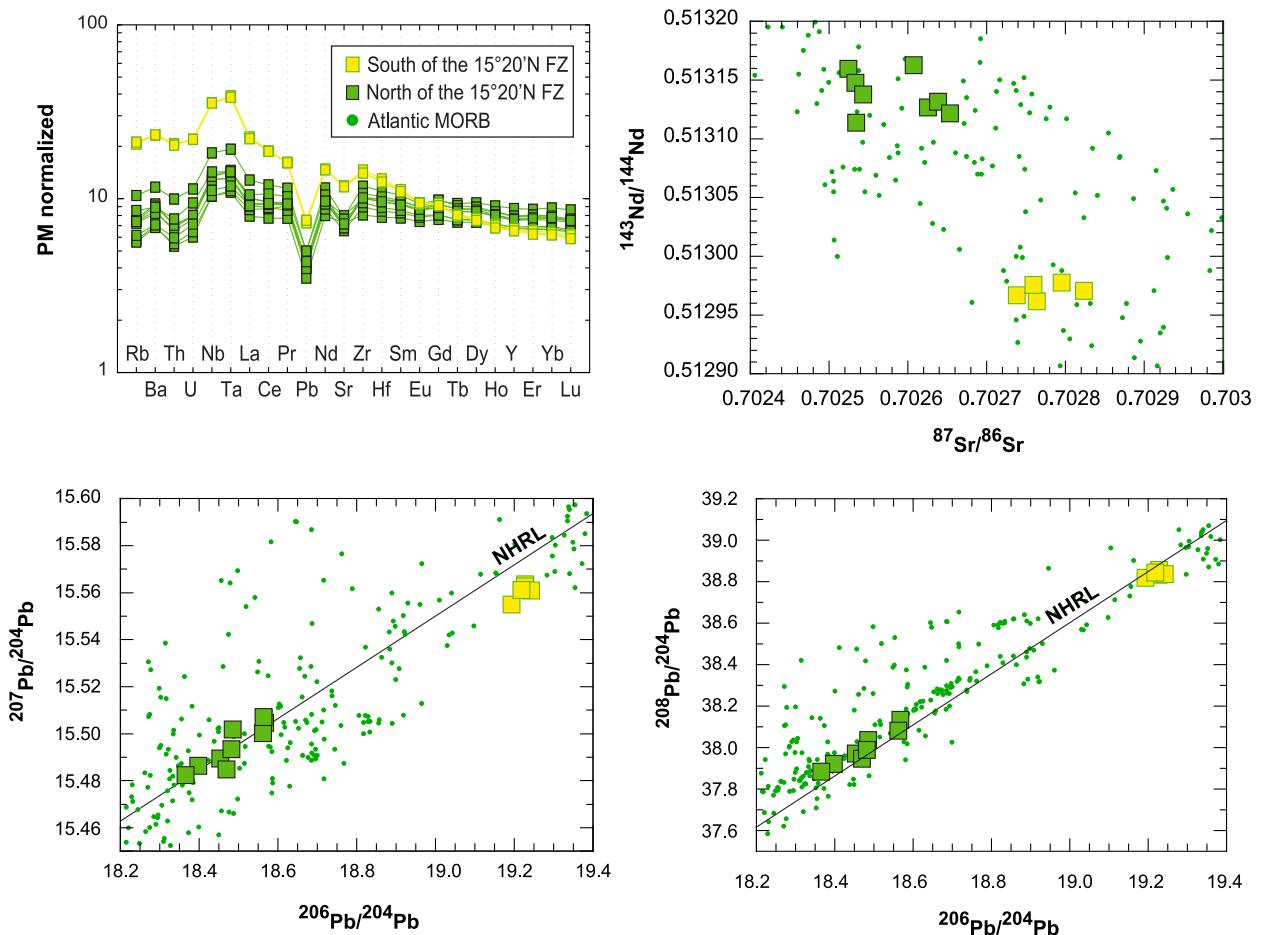


Figure 2. Primitive mantle normalized trace element abundances [McDonough and Sun, 1995] and Sr, Nd and Pb isotopes of samples from the 15°20'N FZ area. Atlantic MORB data are from Agranier *et al.* [2005]. NHRL: Northern Hemisphere Reference Line [Hart, 1984].

slightly higher than those from the northernmost samples (1.25 versus 1.13). Enrichments are also described for strongly incompatible element. Nb-Ta concentrations are 40 times that found in primitive mantle, twice the contents of the most enriched sample from the north segment (418R3).

4.3. Sr, Nd and Pb Isotope Data

[16] New isotope data presented in this study are plotted in Figure 2 together with isotope data from Atlantic MORB [Agranier *et al.*, 2005]. Considering average isotopic compositions for N-MORB from the North Atlantic of 0.70240, 0.51322, 18.116, 15.463 and 37.68 respectively for Sr, Nd and Pb isotopic ratios (averages calculated using MORB away from any hot spot; data are from Agranier *et al.* [2005]), all of our samples presented in this study are

more radiogenic in Sr and Pb (and less in Nd) than those of North Atlantic N-MORB. MORB Sr ratios from the northern segment increase from 0.70249 to 0.70265 toward the 15°20'N FZ. The most radiogenic values (~0.70276) are measured on E-MORBs located south of the fracture zone. The dispersion observed in $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for 415R4, 415R5, 417R1 and 419R4 while they are on the mantle array in Nd-Pb and Pb-Pb spaces is likely to be related to seawater interaction. The fact that NAUFARA-007-005 and CHR0077-006-145 samples from Agranier *et al.* [2005] behave like these four samples also display good evidence for seawater contamination as well. $^{143}\text{Nd}/^{144}\text{Nd}$ ratios from northern segment samples decrease from the 15°20'N FZ to the north. Lower values (~0.512950) are observed in the southern segment samples with isotopic ratio

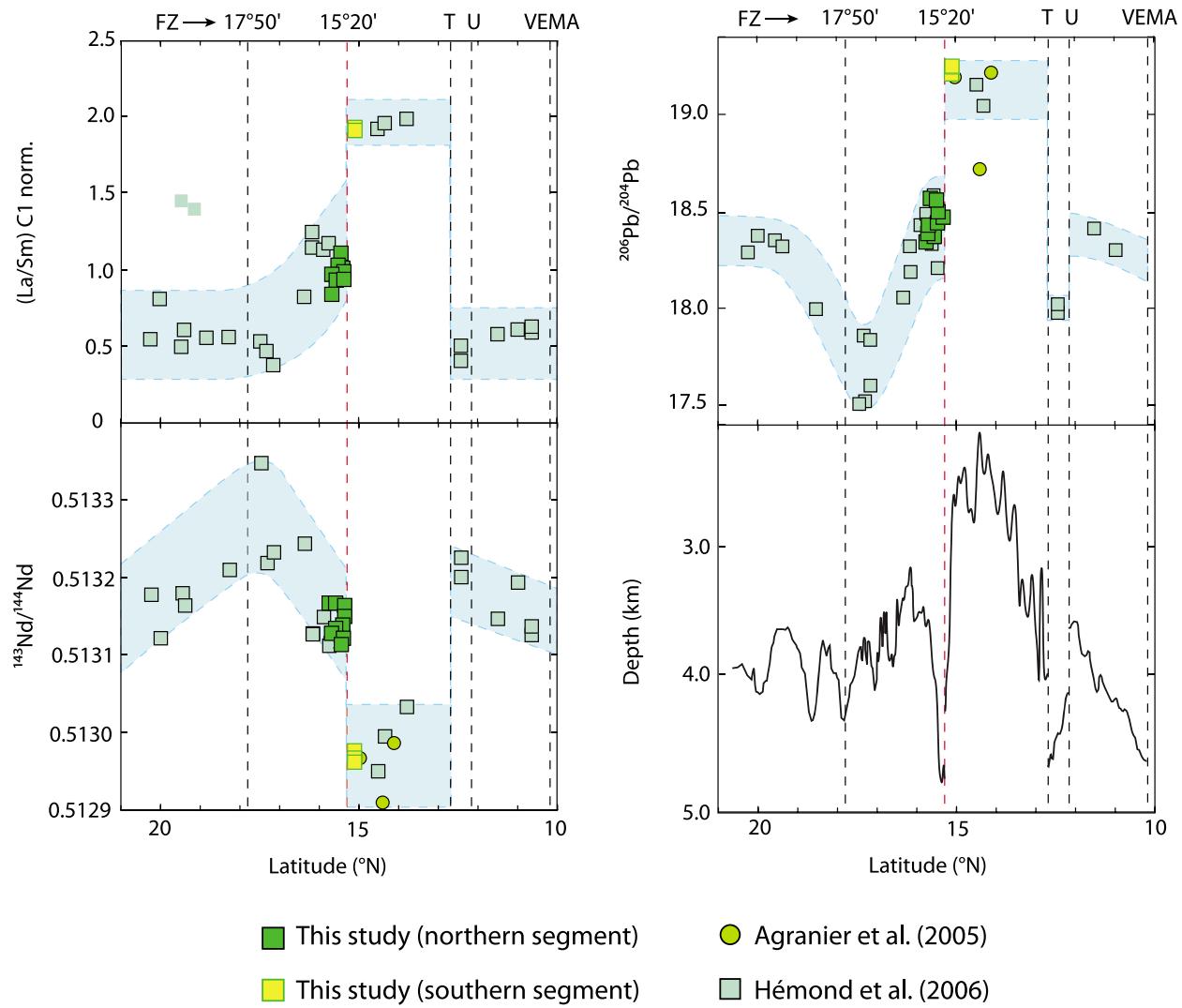


Figure 3. $(\text{La}/\text{Sm})\text{C1}$ normalized ratios, Nd isotope ratios, Pb isotope ratios and along-axis cross-section versus latitude. Green and yellow squares are data from this study (respectively from the northern and the southern segments), blue squares represent data from Hémond et al. [2006], and green dots correspond to additional isotopic data from Agranier et al. [2005]. The along-axis cross-section highlighting the important topographic high (~ 1 km max) centered at 14°N and extending to the overall southern segment (data from Smith and Sandwell [1997]). T: Marathon fracture zone; U: Mercurius fracture zone.

comparable to those of hot spot-related E-MORB (35°N – 45°N region related to the Azores hot spot [Turner et al., 1997]; Equatorial MAR contaminated by the Sierra-Leone mantle plume [Schilling et al., 1994]).

[17] MORBs from the northern segment range from 18.3647 to 18.5668 for $^{206}\text{Pb}/^{204}\text{Pb}$ ratio, from 15.4828 to 15.5075 for $^{207}\text{Pb}/^{204}\text{Pb}$ ratio and from 37.889 to 38.138 for $^{208}\text{Pb}/^{204}\text{Pb}$ ratio. Conversely, E-MORBs from the southern segment are characterized by higher isotopic lead ratios, with values about 19.23, 15.56 and 38.86 respectively for $^{206}\text{Pb}/^{204}\text{Pb}$,

$^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$. The global increase observed in lead isotope ratios from north to south is in agreement with the progressive source enrichment through the $15^{\circ}20'\text{N}$ FZ, as determined by previous trace element and isotope composition analyses (Figure 3).

5. Discussion

5.1. Enrichment Origin

[18] The statistical study published by Arevalo and McDonough [2010] shows that Atlantic MORBs

compared to MORBs from Indian and Pacific oceans are in average more enriched in very incompatible element (e.g., LREE, Ti, Ta, Nb, ...) as well as less incompatible elements (e.g., HREE). To account for this observation, the authors suggested that this enrichment is likely related to smaller degrees of partial melting and/or greater extends of fractional crystallization due to smaller spreading rates, even though the presence of prominent recycled source component or variable proportion of pyroxenite in the Atlantic mantle source is also envisaged. Nevertheless, for the most enriched MORB, it is now commonly accepted that the genesis of E-MORB does not only occur because of smaller degrees of partial melting but requires an enriched source contribution [Dosso *et al.*, 1991, 1993; Michael *et al.*, 1994; Niu and Batiza, 1997; Donnelly *et al.*, 2004; Hémond *et al.*, 2006; Nauret *et al.*, 2006]. Recent studies on MAR basalts propose that Atlantic MORB heterogeneities have two origins: (1) a long-wavelength character revealing the influence of hot spot mantle (e.g., Iceland), and (2) a pervasive shorter wavelength character explained by MORB source contamination with old subducted recycled lithosphere (i.e., the dispersion of the DUPAL anomaly [Dupré and Allègre, 1983; Agranier *et al.*, 2005]). The idea of a hot spot interaction to account for the 14°N anomaly was rapidly abandoned, the nearest hot spot, the Cape Verde islands, being located 1800 km away. Bougault *et al.* [1988] followed by Dosso *et al.* [1991] proposed that the 14°N E-MORB may reflect an interaction with an embryonic mantle plume associated with the triple plate junction, but alternative origins of the enriched mantle source was suggested by the same authors shortly after [Dosso *et al.*, 1993]. It was also demonstrated that E-MORBs from the East Pacific Rise (EPR) could be attributed to small discrete blobs of variably fractionated material, with more radiogenic signatures, dispersed as heterogeneities with a size and distribution related to spreading rate at fast spreading ridge [Allègre *et al.*, 1984; Niu *et al.*, 1999]. Thus, it appears that E-MORBs do occur far from any hot spot and without any relationship with deep mantle plumes [Donnelly *et al.*, 2004]. Therefore, when no hot spot is involved, mixing processes between some N-MORB source and some enriched component is the most adequate explanation accounting for the MORB enrichment variability [Staudacher *et al.*, 1989; Dosso *et al.*, 1991, 1993; Michael *et al.*, 1994; Donnelly *et al.*, 2004; Hémond *et al.*, 2006]. This hypothesis is first suggested by the $\text{Na}_{8.0}$ variation. $\text{Na}_{8.0}$ refers to values of Na_2O corrected from low-pressure fractionation to a

common MgO content of 8 wt.% and is considered as an index of the partial meting degree [Klein and Langmuir, 1987]. This parameter is also known to increase with the axial depth, explaining why samples north to the 15°20'N FZ have higher values than those located to the region of the 14°N topographic high previously described (2.8 versus 2.3 in average; Table 1). As proposed by Klein and Langmuir [1987], such depth/chemistry relationship may reflect source heterogeneity in major element composition. The mixing hypothesis is also suggested by the isotope data. The correlations described between our samples isotope ratios indicate that the observed MORB compositional variability probably derived from a binary mixing. Even if the mixing hypothesis was early proposed [Bougault *et al.*, 1988; Staudacher *et al.*, 1989; Dosso *et al.*, 1991, 1993], the petrological and geochemical processes leading to the formation of enriched sources were not really discussed until recently by Donnelly *et al.* [2004].

[19] These authors underlined that such enrichment requires a low partial melting degree stage to account for the trace element ratios whereas oceanic ridges are mostly characterized by high melting degrees. To account for the enrichment of a mildly enriched basalt (MARK area), authors have developed a model where the MORB mantle source is contaminated by a metasomatised peridotite, which was previously enriched by low degree partial melts during subduction. Without refuting this model, Hémond *et al.* [2006] proposed another explanation for E-MORB genesis. Following the observation of McKenzie *et al.* [2004], they argued that the recycling of alkali basalts composing the great majority of ocean islands and submarine seamounts beneath oceanic ridges could represent an alternative to the metasomatic model proposed by Donnelly *et al.* [2004]. Both models were developed based on trace element characteristics of E-MORB, and particularly the recurrent occurrence of Nb-Ta positive anomaly. Niu *et al.* [1999] estimated that Nb-Ta positive anomaly in oceanic basalts indicates that the sources of these basalts are recycled oceanic lithosphere that had previously undergone subduction zone dehydration which preferentially transferred the Th and U (versus Nb and Ta) to the mantle wedge, leaving a “residual” lithosphere enriched in Nb and Ta. Melting of such enriched lithosphere could therefore enrich the surrounding mantle peridotites and finally produce E-MORB at spreading ridge axis, corroborating the metasomatism model proposed by Donnelly *et al.* [2004]. Nevertheless, Nb-Ta positive anomaly is

Table 1 (Sample). Localization, Major Elements, Trace Elements and Isotopic Data of Samples From the 15°20'N Fz Area Analyzed in This Study^a [The full Table 1 is available in the HTML version of this article]

	415R4	415R5	417R1	418R1A	418R3	419R4	419R11
Latitude (°N)	15°39.001	15°39.001	15°39.001	15°20.589	15°20.589	15°25.121	15°25.479
Longitude (°W)	46°38,673	46°38,630	46°34,714	46°37,460	46°38,214	46°40,209	46°40,959
SiO ₂	50.51	50.82	51.26	50.72	51.08	51.22	50.95
Al ₂ O ₃	15.39	15.85	15.04	14.85	14.56	15.09	15.10
FeO	9.27	9.65	9.59	9.31	9.67	9.25	8.95
MnO	0.17	0.20	0.16	0.18	0.18	0.17	0.21
MgO	8.15	6.61	7.85	8.22	7.90	8.08	8.20
CaO	11.14	11.08	10.28	10.67	10.44	10.81	10.73
Na ₂ O	2.63	3.02	2.85	2.86	2.96	2.94	2.84
K ₂ O	0.19	0.18	0.23	0.24	0.28	0.26	0.26
TiO ₂	1.24	1.46	1.52	1.55	1.69	1.58	1.45
P ₂ O ₅	0.14	0.14	0.20	0.17	0.20	0.16	0.16
Cr ₂ O ₃	0.04	0.06	0.06	0.06	0.04	0.04	0.03
Total	98.88	99.11	98.97	98.83	99.01	99.61	98.88
Mg#	0.47	0.41	0.45	0.47	0.45	0.47	0.48
Nas ₀	2.69	2.50	2.80	2.94	2.92	2.98	2.92
Ca	79608	79222	73451	76284	74599	77267	76682
Co	41.6	43.2	40.0	40.5	38.9	41.1	39.3
Ni	141	106	138	137	119	149	137
Cu	87.4	84.3	70.0	65.8	59.9	66.7	65.1
Ga	15.2	16.1	15.3	15.7	15.7	15.0	14.4
Ge	1.24	1.33	1.22	1.27	1.25	1.23	1.22
Rb	3.47	3.38	4.41	4.34	6.26	4.55	4.43
Sr	144	139	130	135	156	159	152
Y	27.5	36.0	28.6	32.8	33.4	31.4	28.0
Zr	84.7	106	94.2	107	124	105	99.6
Nb	6.81	6.82	8.49	8.41	12.1	8.90	8.83
Mo	0.436	0.447	0.517	0.553	0.715	0.527	0.529
Cd	0.166	0.180	0.175	0.164	0.190	0.187	0.154
In	0.074	0.092	0.082	0.083	0.086	0.079	0.074
Sn	1.30	1.47	1.37	1.48	1.53	1.41	1.35
Sb	0.050	0.051	0.057	0.046	0.090	0.113	0.094
Cs	0.037	0.034	0.040	0.041	0.061	0.038	0.037
Ba	46.0	45.0	56.3	54.3	76.8	61.6	60.0
La	5.14	5.61	5.89	6.13	8.29	6.53	6.48
Ce	13.0	14.7	14.9	15.5	20.1	16.0	16.0
Pr	1.96	2.30	2.18	2.34	2.92	2.41	2.33
Nd	10.0	12.0	10.9	12.0	14.4	12.2	11.7
Sm	3.15	3.91	3.37	3.81	4.25	3.75	3.48
Eu	1.14	1.38	1.20	1.33	1.44	1.32	1.23
Gd	4.13	5.27	4.38	5.02	5.34	4.84	4.39
Tb	0.724	0.929	0.752	0.865	0.902	0.833	0.748
Dy	4.93	6.38	5.11	5.90	6.03	5.65	5.07
Ho	1.03	1.35	1.06	1.24	1.25	1.18	1.06
Er	3.01	3.99	3.12	3.58	3.61	3.41	3.07
Tm	0.432	0.576	0.446	0.512	0.520	0.490	0.440
Yb	2.93	3.90	3.05	3.44	3.50	3.32	3.01
Lu	0.433	0.582	0.446	0.507	0.515	0.494	0.437
Hf	2.21	2.81	2.40	2.81	3.13	2.73	2.52
Ta	0.403	0.411	0.483	0.488	0.712	0.538	0.528
W	0.079	0.077	0.102	0.101	0.136	0.106	0.101
Tl	0.0116	0.018	0.017	0.012	0.019	0.018	0.013
Pb	0.524	0.598	0.613	0.600	0.743	0.645	0.595

^aSee the text for more details on analyses. Additional data in italic are from Melson et al. [2002] for the major elements and from Agranier et al. [2005] for the lead isotopes.

also characteristic in overall OIB [e.g., *Willbold and Stracke*, 2006], probably inherited from the process described by *Niu et al.* [1999], as recycled oceanic lithosphere is thought to be the major source of OIB [*Willbold and Stracke*, 2006]. In addition, based on the calculation of trace element variability of the 14°N MAR samples, *Hémond et al.* [2006] suggested that seamounts can pass through the subduction (in spite of the fractionation related to subduction) without being significantly modified in their trace element composition, even for the more “mobile” element such as Rb and U. Thus, recycling OIB and/or seamounts beneath mid-ocean ridge would also be consistent with the genesis of E-MORB.

[20] The main limitation of the latter model is that *Hémond et al.* [2006] did not extrapolate it to the isotopes. Conversely, *Donnelly et al.* [2004] used them to demonstrate that the recycling of the oceanic crust cannot explain the isotopic composition of E-MORB, since recycled oceanic crust is expected to have a DMM isotopic composition, which is not radiogenic enough to constitute a suitable source of E-MORB. Based on this assumption, they developed a metasomatic model described above which creates the enrichment necessary to account for the trace element and the isotopic compositions of E-MORB. Similarly to *Donnelly et al.* [2004], *Cooper et al.* [2004] considered the origin of Azores E-MORB as related to the melting of a peridotite previously contaminated by melts deriving from the recycling of an old oceanic crust. They suggested that the highly radiogenic Sr composition of E-MORB could be inherited from the hydrothermal alteration of the recycled oceanic crust prior to its subduction. Authors also explain the unradiogenic Nd values by aging effects, but such effects are not only restricted to Nd isotope ratios, and would also affect other isotopic systems. Moreover, this model does not explain the highly radiogenic Pb composition usually observed in E-MORB. Although Pb isotopic compositions can increase due to the seawater alteration, this assumption is far from systematic. Indeed, even Fe-Mn crust contributes to increase Pb isotope ratios [*von Blanckenburg et al.*, 1996; *Abouchami et al.*, 1997], *Vidal and Clauer* [1981] demonstrated that Sr isotopes can show evidences of seawater contamination whereas Pb isotopes do not (this is also the case in this study for some of our samples, Figure 3). The major argument against *Donnelly et al.*'s [2004] model (and also the one of *Cooper et al.* [2004]) concerns the depth of the first melting stage. In both models, low F melting of the subducting slab is believed to occur at greater depths than those usually considered

for arc-magma genesis (i.e., ~110 km [e.g., *Hattori and Guillot*, 2003; *Rüpk et al.*, 2004]). This assumption may be problematic for two reasons: First, melting of subducting slab requires particular conditions that are far from systematic in subduction zones. *Iwamori* [1998] demonstrates that slab melting is restricted to young slab (<10 My), mainly producing particular arc-magmas such as adakites. Such melting generally occurs in the forearc and is thus localized at low depth (<80 km [*Iwamori*, 1998]). Second, melting is unlikely to occur at depths greater than ~150 km. Melting in subduction is associated to the breakoff of hydrous minerals (i.e., serpentine) composing the slab and/or the overlying mantle wedge that release an important amount of water leading to its partial melting [e.g., *Hattori and Guillot*, 2003; *Rüpk et al.*, 2004]. Even small amount of water can be retained in the slab at depth greater than 150 km, the pressure is likely too high and the amount of water too low to produce slab-derived melts. Therefore, while these observations are not sufficient to rule out the metasomatism model of *Donnelly et al.* [2004], they allow us to consider further the OIB recycling model of *Hémond et al.* [2006]. Nevertheless, to validate the model of *Hémond et al.* [2006], isotopic compositions must be taken into account, particularly considering that even if OIB can be estimated as homogeneous in trace element composition, this assumption is not true for isotopes.

5.2. OIB Recycling Model

5.2.1. Formation of Seamounts: On-Axis Versus Off-Axis Origin

[21] The major feature that has to be taken into account before considering the model of seamount recycling is the origin of the seamounts. Indeed, the viability of the model critically depends on how do the seamounts form and do all of them display enriched composition that could contribute to produce E-MORB after their recycling. Basically, seamounts can be related to: (1) a hot spot activity; (2) a mid-ocean ridge activity; (3) an intraoceanic subduction zone. Although seamounts deriving from arc magmatism are clearly different in composition compared to the DMM, their number is very limited [*Koppers and Watts*, 2010] and thus their contribution to the mantle heterogeneity can be considered as negligible.

[22] Volcanic islands and seamounts formed away from tectonic boundaries (so-called intraplate volcanism, e.g., Hawaii-Emperor chain, Society Islands,

Réunion, Cape Verde) are commonly considered as the surface expression of deep mantle plume activity [Morgan, 1971]. In this “off-ridge” setting [Watts *et al.*, 2006; Koppers and Watts, 2010], such intraplate magmatism leads to the formation of large volcanic islands (locating the present-day hot spot activity) which become seamounts as the plate motions move them away from the “stationary” parent plume, forming linear age-progressive seamount trails. Basalts related to hot spots (so-called OIB for oceanic island basalts) differ from MORB in their chemical composition. Unlike the MORB, which are tholeiites thought to be formed by high degrees of partial melting (>10%), OIB are mainly silica-undersaturated alkali basalts deriving from low degrees of partial melting and which are enriched in incompatible elements and radiogenic isotope ratios relative to MORB [e.g., White and Hofmann, 1982; Zindler and Hart, 1986]. As they were believed to derive from a deep source (i.e., the D” layer at the Core-Mantle boundary [Morgan, 1971]), OIB offered a unique opportunity to understand deep mantle processes. Therefore numerous studies focused on them over the last three decades [e.g., White and Hofmann, 1982, White, 1985; Zindler and Hart, 1986; Hart *et al.*, 1992; Stracke *et al.*, 2005; Willbold and Stracke, 2006, 2010]. These studies have demonstrated that OIB do not derive from a unique source and that they arise from the mixing of approximately five mantle reservoirs (DMM, EM-1, EM-2, HIMU and FOZO/PREMA/PHEM/C) defined by their isotope ratios [Zindler and Hart, 1986], leading to a scheme nicknamed “the mantle zoo” [Stracke *et al.*, 2005]. Compared with the DMM, presumed to approximate the source of the MORB, EM-1 (Enriched-Mantle-1) is characterized by high $^{87}\text{Sr}/^{86}\text{Sr}$, very low $^{144}\text{Nd}/^{143}\text{Nd}$ and higher $^{208}\text{Pb}/^{206}\text{Pb}$ at a given $^{206}\text{Pb}/^{204}\text{Pb}$, properties that are thought to be inherited from the recycling of pelagic sediments in the source of magmas. OIB that fall into this category are Pitcairn, Kerguelen, Walvis Ridge, Tristan da Cunha and Gough. OIB from EM-2 source (Enriched-Mantle-2) are more radiogenic in Sr and Pb and less radiogenic in Nd than EM-1 and are associated to the recycling of terrigenous sediments eroded from the continental crust and deposited on the seafloor. Society, Samoa and Marquesas Islands are representative of EM-2 type OIB. HIMU (for High μ , i.e., high $^{238}\text{U}/^{204}\text{Pb}$) type OIB includes St Helena and some of the Cook-Austral chain basalts (others belong to the FOZO type, see below) and are characterized by extremely high Pb isotope ratios. They are also less radiogenic in Sr compared

to EM-types and are likely to be formed by the recycling of old, subducted, dehydrated oceanic crust. Compared to HIMU, the last classification, called FOZO (for FOcal Zone [Hart *et al.*, 1992; Stracke *et al.*, 2005]) was attributed to OIB that were less radiogenic in Pb at a given Sr and Nd isotopic composition. FOZO is mainly represented in basalts from Cook-Austral chain [see Stracke *et al.*, 2005], and it is also considered to be common to most basalts. The idea of a common component mixed in the various sources of OIB, first called PREMA (for PREvalent MAntle [Zindler and Hart, 1986]), has evolved to FOZO, PHEM (for Primitive HElium Mantle [Farley *et al.*, 1992]) and C (Common Component [Hanen and Graham, 1996]), but all of them differ from each other only in detail [Hofmann, 2003]. As the name and the origin of this component have no influence on the proposed model, we will simply refer to FOZO hereafter.

[23] As their chemistry strongly differs from the DMM, the recycling of OIB may introduce significant amount of enriched material in the mantle. Nevertheless, intraplate oceanic seamounts do not all originate from plumes. Even, several studies have shown that this type of seamount is predominant in the seafloor compared to OIB related seamounts [Batiza, 1982; Hofmann, 1997; Watts *et al.*, 2006; Koppers and Watts, 2010]. They usually occur as isolated edifices (near-ridge seamounts, e.g., along the EPR), clusters (Rano-Rahi seamount field, near the EPR) or lines (the Cameroon Line in the Atlantic ocean or the Line Islands and the Pukapuka ridge in the Pacific Plate). As only few data are available on this type of seamounts, their origin remains poorly constrained, Hofmann [1997] arguing that most of these are melting anomalies likely coming from the upper mantle. Following this idea, the hypothesis of small-scale sublithospheric convection [Ballmer *et al.*, 2007] has been developed to explain linear seamount chains that do not display age progression (such as Pukapuka ridge or Line Islands) or seamount clusters (Rano-Rahi seamount field). Recently, on the basis of the elastic thickness, a parameter that is sensitive to whether a seamount is formed nearby or faraway from a mid-ocean ridge, Watts *et al.* [2006] and Koppers and Watts [2010] estimated that the largest part of the world’s seamounts are not related to plume and was formed close to mid-ocean ridge (i.e., “on-ridge” type seamount). Studies focused on these “on-ridge” seamounts have shown that they are geochemically different from N-MORB, carrying slightly enriched

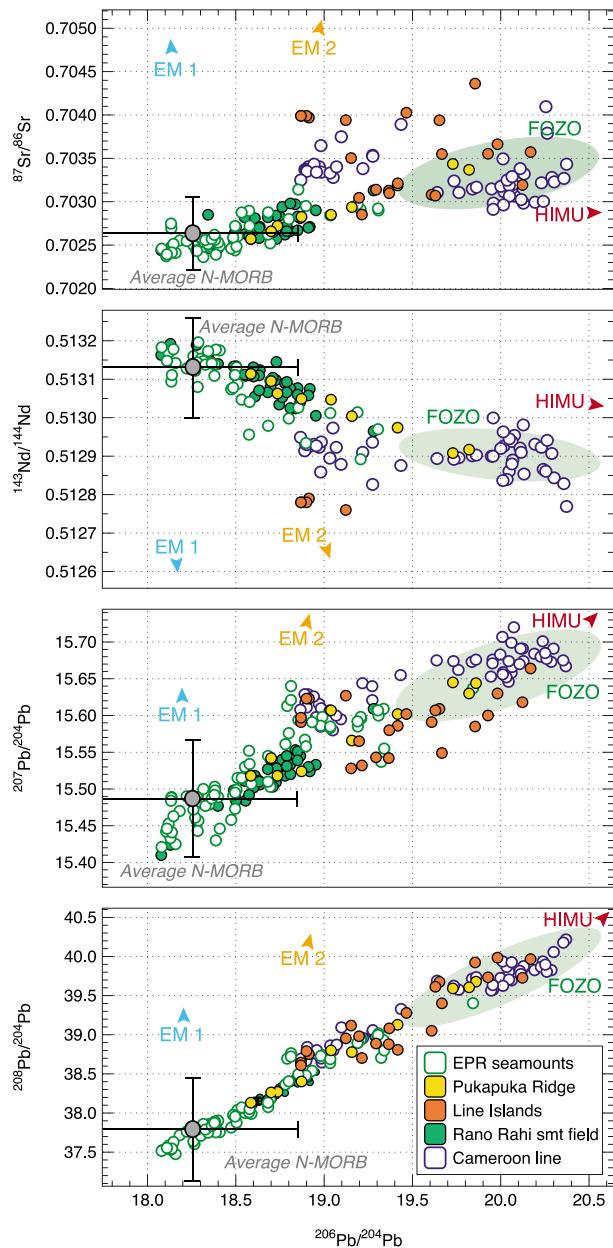


Figure 4. Sr, Nd and Pb isotopic compositions of seamounts presumed unrelated to hot spot (so-called “on-ridge” seamounts [Watts et al., 2006; Koppers and Watts, 2010]): Line Islands [Hart, 1988; Garcia et al., 1993], Marshall Islands [Staudigel et al., 1991; Koppers et al., 2003], Cameroon line [Halliday et al., 1988; Lee et al., 1994], Rano Rahi seamount field [Hall et al., 2006], EPR seamounts [Graham et al., 1988; Niu et al., 2002]. DMM and FOZO field are respectively from Workman and Hart [2005] and Stracke et al. [2005].

MORB-like signatures [Batiza and Vanko, 1984; Zindler et al., 1984; Graham et al., 1988; Niu et al., 1997, 2002]. In reality, non-OIB seamounts display large variations in composition, ranging from

N-MORB-like signature to OIB-like signatures (Figure 4). Particularly, numerous EPR seamounts have isotopic compositions that are highly comparable with those assumed for the DMM. However, most of the “on-ridge” seamounts exhibit slightly to strongly enriched compositions and plot along the DMM-FOZO array. This indicates that such type of seamounts derives from the mixing of small-scale heterogeneities with the surrounding depleted upper mantle and thus also incorporates a significant amount of enriched material in the mantle when they are recycled.

[24] Nevertheless, although we show that “on-ridge” type seamounts are slightly enriched compared to the DMM, both types of seamounts do not have the same potential to produce E-MORB after their recycling. As illustrated above, “on-ridge” seamounts differ from “off-ridge” seamounts mainly on their number, their size and their chemistry. Hillier and Watts [2007] estimate a total number of ~3 million seamounts with a height >0.1 km of which only 200,000 are clearly identified. Considering these numbers, only a few of them are clearly documented enough to constrain their origin. However, Watts et al. [2006] and Koppers and Watts [2010] recently estimated that at least 60% of the identified seamounts were formed in an “on-ridge” setting (estimate made using Wessel’s [2001] database of ~15,000 seamounts with a height >1.5 km). Although “off-ridge” seamounts are less abundant, they are significantly more enriched in incompatible elements and more radiogenic in Sr and Pb isotopes. Among all “on-ridge” seamounts plotted in Figure 4, only a few reaches the radiogenic enrichment usually described in OIB (e.g., the Cameroon line or some samples from the Line Islands). All others more closely resemble E-MORB rather than OIB. “Off-ridge” seamounts are also higher in size and volume [Batiza, 1982; Wessel, 2001]. For instance, the almost totality of the structures that form the Hawaiian-Emperor seamount chains are higher than 4 km, while “on-ridge” seamounts are smaller (likely <1.5 km in average) suggesting that the largest amount of enriched material that derives from the recycling of seamounts is related to OIB.

5.2.2. Trace Element Modeling

[25] The trace element modeling reproduced in this study is nearly the same as developed in Hémond et al. [2006]. This model was used to constrain the Sr, Nd and Pb concentrations in the enriched component, data that are required for isotope modeling. Parameters used in this model

Table 2. References of Data Used in Modeling

Data	Reference
<i>Trace Element Modeling</i>	
Partition coefficient	Niu and Hékinian [1997] ^a Donnelly et al. [2004] ^a
Partial melting degrees (F)	Hémond et al. [2006] ^a
Mixing proportion	Driven by OIB fields (see text for details)
DMM	Regional N-MORB average (PetDB) ^b
EM1, EM2 and HIMU fields	Stracke et al. [2003]; Willbold and Stracke [2006]
FOZO Austral-Cook	Stracke et al. [2003]
FOZO Azores, Cape Verde, Canary, Cameroon Line, Pukapuka Ridge, Rano-Rahi seamounts	GEOROC database http://georoc.mpch-mainz.gwdg.de/georoc/
<i>Isotope Modeling</i>	
Mixing proportion	Driven by OIB fields (see text for details)
Sr, Nd, Pb composition of enriched component	Trace element modeling results ^b
DMM	North Atlantic N-MORB average [Agranier et al., 2005] ^b
EM1, EM2 and HIMU fields	Willbold and Stracke [2010]
EM1, EM2 and HIMU averages (Stars in Figures 6 and 7)	Jackson and Dasgupta [2008] ^a
Elemental Sr, Nd and Pb concentrations in EM1, EM2 and HIMU sources	Willbold and Stracke [2006] ^a
FOZO Austral-Cook	Stracke et al. [2005]
FOZO Azores, Cape Verde, Canary, Cameroon Line, Pukapuka Ridge, Rano-Rahi seamounts	GEOROC database http://georoc.mpch-mainz.gwdg.de/georoc/
<i>Age</i>	
Enriched endmember	Calculated from isotopic modeling
DMM isotope values	Su and Langmuir [2003] ^a
Highly DMM	Workman and Hart [2005] after Su and Langmuir [2003] ^a
μ DMM	Elliott et al. [1999] ^a
κ DMM	

^aValues given in Table S1.^bValues given in Table 3.

and their references are given in Table 2 and 3 (see also Table S1 in the auxiliary material for complete parameter values used for the modeling including partition coefficients, average compositions in trace elements and isotopes for OIB, etc.).¹

[26] The assumption of Hémond et al.'s [2006] modeling is that E-MORB from the 14°N anomaly in the MAR derive from the melting of an enriched source which can be modeled as a solid binary mixing process between the DMM and enriched component (i.e., a recycled OIB/seamount). For purpose of modeling, the trace element composition of the recycled OIB/seamount is calculated using the non-modal batch melting equation from Shaw [1970] and the solid binary mixing model developed by Faure [1977]. The non-modal batch melting equation defined by Shaw [1970] is turned to express the trace element concentration of the E-MORB source (C_0) as a function of the trace element composition

of the E-MORB (C_{EL}), the partial melting degree (F), the source bulk distribution coefficient (D_0) and the melt partition coefficient (P):

$$C_0 = C_{EL}[D_0 + F(1 - P)] \quad (1)$$

The most enriched E-MORB (sample 428R5) was chosen for the melt composition (C_{EL}). It makes sense to use this sample in this purpose as this sample comes from the area where the contamination by the enriched component is the strongest. As postulated by Hémond et al. [2006], a relatively high partial melting degree ($F = 18\%$) was used because of the enhanced fertility caused by the addition of the enriched component within the DMM.

[27] The same expression as (1) is used to constrain the trace element composition of the N-MORB source (i.e., the DMM):

$$C_{DMM} = C_{N-MORB}[D_0 + F(1 - P)] \quad (2)$$

Here, C_{DMM} is the trace element composition of the DMM and C_{N-MORB} corresponds to an average

¹Auxiliary materials are available in the HTML doi:10.1029/2012GC004078.

Table 3. Results of Trace Element and Isotopic Modeling

Elements	N-MORB Average	Mid-Atlantic Ridge (14°N)		Central Indian Ridge (18°–20°S)		
		Modeled Recycled OIB/ Seamount Composition		N-MORB Average	Modeled Recycled OIB/ Seamount Composition	
		0.05:1	0.15:1		0.01:1	0.05:1
Rb	0.571	44.0	14.7	0.613	50.1	30.1
Ba	7.13	541	181	9.99	550	330
Th	0.112	5.67	1.90	0.144	5.48	3.29
U	0.041	1.53	0.512	0.042	1.25	0.749
Nb	2.02	80.4	26.9	1.73	66.7	40.1
Ta	0.123	4.92	1.65	0.112	3.66	2.20
La	2.98	46.8	15.8	2.68	39.2	23.7
Ce	8.38	98.7	33.5	8.84	74.3	45.0
Pr	1.91	11.2	3.87	1.50	8.96	5.44
Pb	0.5	3.07	1.06	0.494	3.84	2.33
Nd	8.13	51.4	17.8	8.35	32.4	19.8
Sr	129	607	212	129	689	419
Zr	68.3	418	145	78.1	280	172
Hf	1.84	9.55	3.34	1.99	5.06	3.14
Sm	2.64	11.1	3.93	2.94	6.26	3.91
Eu	0.92	3.37	1.21	1.12	2.02	1.27
Gd	3.26	11.4	4.10	3.88	6.90	4.36
Tb	0.544	1.74	0.634	0.750	0.926	0.600
Dy	4.04	9.63	3.62	4.71	5.84	3.79
Ho	0.891	1.71	0.66	1.07	1.02	0.68
Y	2.63	4.50	1.78	2.99	3.55	2.33
Er	25.8	39.0	15.7	29.8	39.2	25.4
Yb	2.65	3.84	1.58	2.93	2.35	1.61
Lu	0.409	0.493	0.211	0.450	0.354	0.243
⁸⁷ Sr/ ⁸⁶ Sr	0.70240	0.70293	0.70291	0.70248	0.70554	0.70500
¹⁴³ Nd/ ¹⁴⁴ Nd	0.51322	0.51287	0.51288	0.51320	0.51270	0.51271
²⁰⁶ Pb/ ²⁰⁴ Pb	18.1159	19.588	19.539	18.132	18.959	18.951
²⁰⁷ Pb/ ²⁰⁴ Pb	15.4629	15.597	15.592	15.468	15.619	15.618
²⁰⁸ Pb/ ²⁰⁴ Pb	37.681	39.24	39.19	37.75	39.20	39.18

of N-MORB trace element composition (N-MORB are here defined as MORB with $(\text{La}/\text{Sm})_N < 1$ and $^{87}\text{Sr}/^{86}\text{Sr} < 0.7027$). D_0 and P are the same as in (1) and $F = 10\%$, following the model of Hémond *et al.* [2006]. Then, the composition of the enriched component (i.e., the recycled OIB/seamount; C_{EC}) drowned in the surrounding, more depleted, mantle source (i.e., the DMM; C_{DMM}) is deduced from the solid binary mixing equation given in Faure [1977]:

$$C_0 = C_{EC} \cdot f + C_{DMM} \cdot (1-f) \quad (3)$$

turned into

$$C_{EC} = \frac{C_0 - C_{DMM} \cdot (1-f)}{f} \quad (4)$$

where f corresponds to the proportion of the enriched component in the mixture.

[28] In Figure 5, the trace element spidergrams of two calculated enriched components (red lines) are reported and compared with the OIB fields [Willbold and Stracke, 2006]. The “0.05:1” and “0.15:1” lines represent the PM-normalized trace element compositions of the modeled recycled OIB/seamount calculated for $f = 5\%$ and 15% , respectively (see Table 3 for the compositions). Contrary to Hémond *et al.* [2006] who chose the abundance patterns of the recycled OIB/seamount “so that the final model product matches the observed, average E-MORB at 14–15°N exactly,” we calculate here the trace element composition of the enriched component at different mixing proportions with respect to the OIB fields. In other words, we consider in our calculation that the modeled enriched component is consistent with “natural” OIB when at least 90% of the PM-normalized values plot within OIB field. Even if mixing proportions were chosen to stay within the OIB field, Figure 5 demonstrates

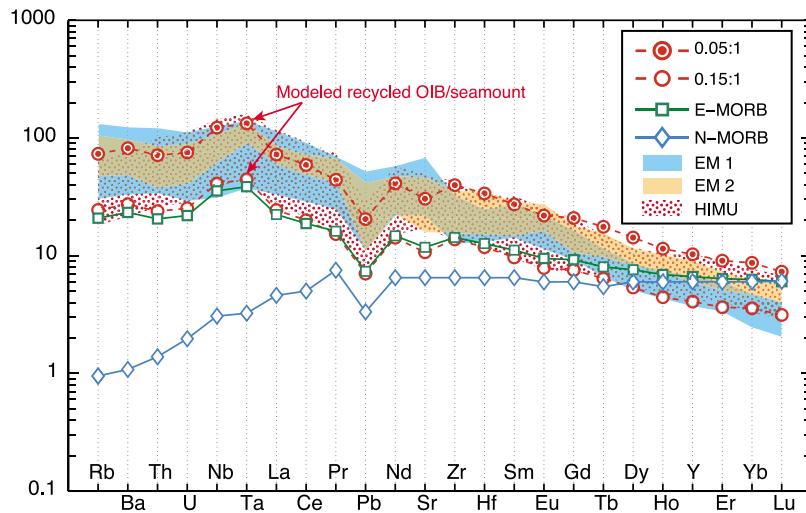


Figure 5. Primitive Mantle (PM [McDonough and Sun, 1995]) normalized trace element compositions of two modeled recycled OIB/seamount (“0.05:1” and “0.15:1” red lines), calculated using the non-modal batch melting equation from *Shaw* [1970] and the solid binary mixing model developed by *Faure* [1977]. Assuming that the E-MORB genesis results from the mixing between the DMM and an enriched component (i.e., the recycled OIB/seamount), the trace element composition of the E-MORB source is first calculated by “inverting” the E-MORB composition (the green line). The DMM composition is modeled from the local N-MORB composition (the blue line) applying 10% of batch melting. Then, two recycled OIB/seamount compositions are deduced from the calculated E-MORB source and the DMM compositions using mixing ratios of 0.05:1 and 0.15:1 (the two red lines). These mixing ratios were chosen so that at least 90% of the PM-normalized trace element contents of the modeled recycled OIB/seamount plot within OIB field. Our accurate inspection of available OIB data allowed us to estimate that OIB fields presented in this study represent an almost complete set of OIB compositions (data from *Willbold and Stracke* [2006]). Results are given in Table 3, see the text for more details.

that chondrite-normalized trace element patterns are consistent with “natural” OIB composition (i.e., incompatible element enrichment, positive Nb-Ta anomaly, HREE depletion). Furthermore, these mixing proportions (5 to 15%) are close to the abundance of recycled material present in north Atlantic MORB source ($\leq 10\%$ [Cooper *et al.*, 2004]). Our estimates can be closer (6 to 11%) to those of *Cooper et al.* [2004] by considering only modeled patterns of which 100% of the chondrite-normalized values plot within OIB field.

5.2.3. Isotope Modeling

[29] Isotope ratios of the enriched component were calculated using the binary mixing equation developed by *Faure* [1977] and *DePaolo and Wasserburg* [1979]:

$$IC_m = \frac{f C_{EC} IC_{EC} + (1-f) C_{DMM} IC_{DMM}}{f C_{EC} + (1-f) C_{DMM}} \quad (5)$$

where IC_m , IC_{EC} and IC_{DMM} are respectively the isotopic compositions of the mixture, the enriched component and the DMM, C_{EC} and C_{DMM} are the elemental concentrations in both components and

f is the proportion of the enriched component in the mixture. Results of trace element modeling were used for Sr, Nd and Pb concentrations, as well as the mixing proportions. Isotope ratios of the mixture and the DMM were determined by the same way as for the trace elements: isotope ratios of the most enriched E-MORB (sample 428R5) were chosen as ratios of the mixture, and DMM isotope ratios correspond to the regional average of N-MORB isotope ratios. Other parameters are summarized in Table 2 and Table S2. Results are presented in Figure 6 and Table 3.

[30] Contrary to the results given by the trace element modeling, variations in the mixing proportions do not lead to significant changes in the modeled enriched component composition. The modeling results are ~ 0.70292 for $^{87}\text{Sr}/^{86}\text{Sr}$, ~ 0.51287 for $^{143}\text{Nd}/^{144}\text{Nd}$ ratio, ~ 19.564 for $^{206}\text{Pb}/^{204}\text{Pb}$ ratio, ~ 15.594 for $^{207}\text{Pb}/^{204}\text{Pb}$ ratio and 39.22 for $^{208}\text{Pb}/^{204}\text{Pb}$ ratio. Compared to OIB fields, the calculated enriched component appears incompatible with both EM end-members. Sr and Nd isotopes are consistent with HIMU, but Pb isotopes are too low to fit with the extremely radiogenic values characterizing HIMU. However, all isotope ratios are

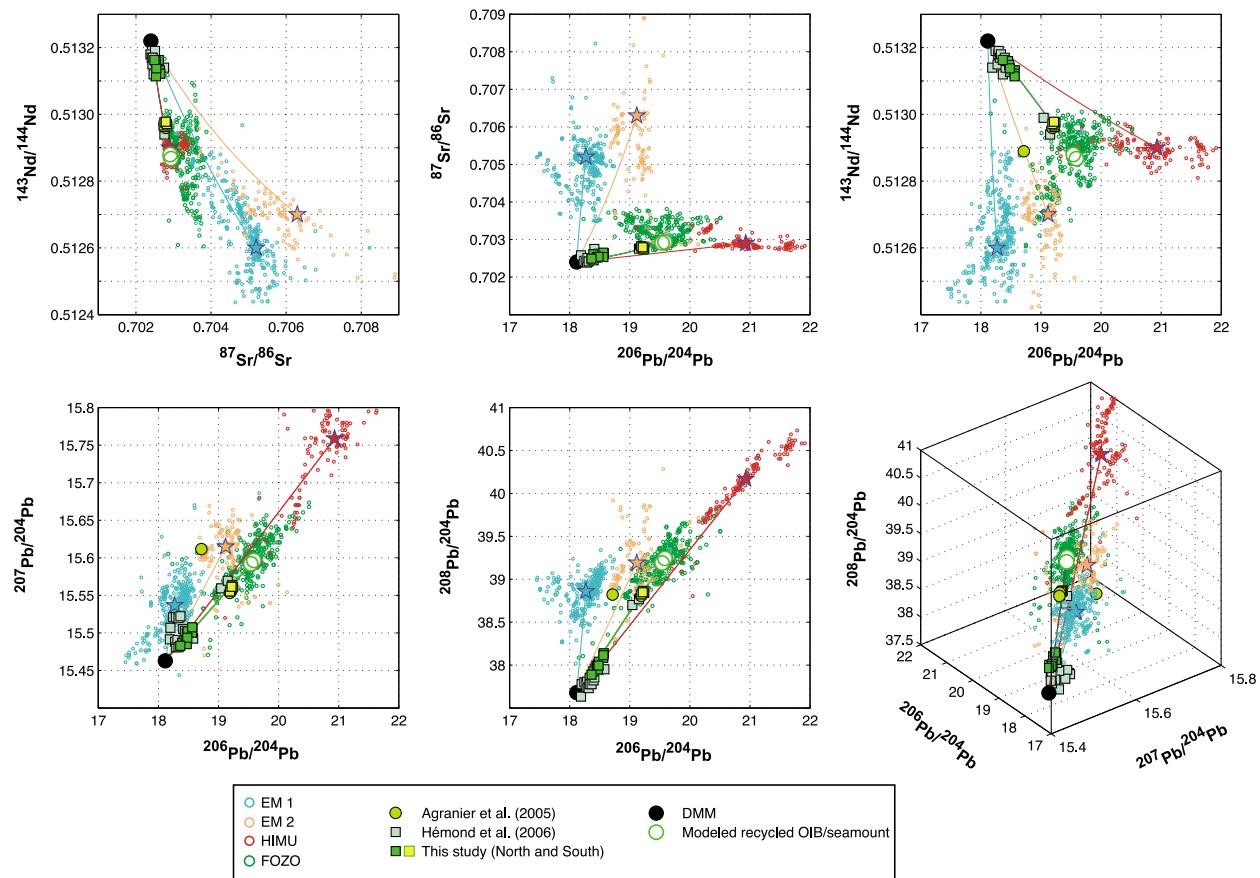


Figure 6. Isotope composition of the enriched component calculated using the binary mixing equation developed by *Faure* [1977]. Isotope ratios of the mixture are those of the most enriched E-MORB (sample 428R5). Isotopes ratios of the DMM correspond to the regional average of N-MORB isotope ratios. Two enriched components (large green circles) are calculated and plotted in each isotope space, depending on the mixing proportions which were constrained by results of the trace element modeling (i.e., 5% and 15%). Other parameters are summarized in Table 2 and Table 3. OIB fields are from *Willbold and Stracke* [2010]: HIMU (in red), EM 1 (in blue), EM 2 (in orange) and FOZO (in green). Colored stars correspond to the average isotopic compositions [*Jackson and Dasgupta*, 2008] for each type of OIB, and colored lines are the mixing trends calculated between the DMM and average OIB isotopic compositions, using the same color codes as for the OIB fields. Sr, Nd and Pb elemental compositions of OIB used to calculate the mixing trends are from *Willbold and Stracke* [2006].

comparable with the FOZO isotope compositions, which are characterized by lower Pb isotopic composition at given Sr and Nd. Considering that both trace element and isotope modeling reach the same results, the recycling of OIB seems to be a reasonable model to explain the E-MORB genesis at 14°N on the MAR.

5.3. Does This Happen Somewhere Else?

[31] Although our model can explain the MORB enrichment in the 14°N region in the MAR, its validity at a global scale (whether this model can account for MORB enrichment in other ridge

environment or not) has to be verified. To test this hypothesis, we used samples from the Central Indian Ridge (CIR) at 18°S [*Nauret et al.*, 2006], as in many points, they resemble to the samples from the 14°N anomaly in the MAR: (1) Samples are enriched in incompatible elements ($(\text{La}/\text{Sm})_N > 1$), (2) they display Nb-Ta positive anomaly, (3) they are highly radiogenic in Sr and Pb and isotopically depleted in Nd compared to the DMM.

[32] *Nauret et al.* [2006] suggested that “the CIR MORB, between 18° and 20°S are generated by partial melting of a heterogeneous source consisting of an enriched component and a normal, depleted

upper-mantle peridotite.” These authors also highlighted that “the nature of the enriched component is a matter of speculation.” They finally noticed that this enrichment is not related to the presence of a hot spot, as the E-MORB composition is distinct from the Réunion plume composition, which is the closest hot spot in the area. Instead, *Nauret et al.* [2006] considered that the enriched component could result from the metasomatic event proposed by *Donnelly et al.* [2004], even if enrichment by recycling ancient alkali basalt is also considered.

[33] The model established on trace elements and isotopes for the samples from the 14°N region in the MAR was extrapolated to the 18°–20°S region of the CIR. Calculations were made using sample GNT-Pl1-1 since it corresponds to the most enriched sample. Results are presented in Figure 7 and Table 3 (all plots are available in Figure S1). Trace element composition of the calculated enriched component is comparable to OIB compositions using mixing proportions varying between 1% and 5%. Applying these results on the isotope modeling gives an enriched component more radiogenic in Sr (~ 0.70525), less radiogenic in Nd and ^{206}Pb (~ 0.51270 and ~ 18.955 respectively) and almost identical in $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ (~ 15.618 and ~ 39.19) compared to the one calculated for the 14°N anomaly in the MAR. As illustrated in Figure 7, this enriched component is consistent with common EM 2 basalts, demonstrating that the model of OIB recycling can also explain MORB enrichment in this region.

[34] More generally, several studies reported the existence of E-MORB along mid-ocean ridges of which at least a part presents identical geochemical characteristics (incompatible element enrichment, Nb-Ta positive anomaly, isotopically enriched in Sr and Pb, depleted in Nd) to E-MORB from 14°N in the MAR. This is the case, for example, of MORB sampled along the MAR at 23°N [*Donnelly et al.*, 2004] and 33°S [*Michael et al.*, 1994], as well as for those located at 11°20'N along the EPR [*Niu et al.*, 1999]. Although our model was not tested in these cases due to incomplete data set, the existence of these MORB with geochemical characteristics as listed above (particularly the Nb-Ta positive anomaly) allows us to suppose that our modeling is expected to work on these regions too.

[35] Finally, we notice that the mixing proportions determined from the CIR samples are notably the same as those calculated for the 14°N anomaly (globally <10%). This result is compatible with previous results obtained by *Cooper et al.* [2004] who estimated that the materials responsible for

the genesis of E-MORB in the North Atlantic are present in abundances of 2–5% in average. However, these abundances are used by Cooper et al. to explain the pollution of upper mantle at a global scale while ours explain smaller heterogeneities along mid-ocean ridges. This means that the percentage of the recycled OIB/seamount in the mantle source of MORB could be greater if required without being in opposition to the model of *Cooper et al.* [2004, 2009]. In any case, the small amount we calculated together with the generally small scale heterogeneity that represent E-MORB we studied along the MAR and the CIR suggest that the enriched component represents a very small volume (maybe a few cubic kilometer scale) drowned in the surrounding “normal” mantle. This statement is compatible with the recycling of OIB or seamount which are below the sea level and thus small sized when they subduct.

5.4. Hyperbolic Versus Straight Mixing Trends

[36] As developed by *Langmuir et al.* [1978], mixing relationships in isotope spaces form linear or hyperbolic trends depending on the denominators in the isotope ratios and on the elemental ratio of both mixing components: if the denominators in isotope ratios are identical (as in the lead isotopes, i.e., ^{204}Pb), the mixing trend is a straight line. Else, the mixing trend depends on the ratio of elemental concentrations in both mixing components:

$$K = \frac{\left(C_x^a/C_y^a\right)}{\left(C_x^b/C_y^b\right)} \quad (6)$$

Choosing for example Sr for x and Nd for y , it becomes

$$K = \frac{(Sr/Nd)_a}{(Sr/Nd)_b} \quad (7)$$

in which a and b are the mixing end-members. If $K = 1$, the mixing trend is also a straight line. Else, it is an hyperbola. In Figures 6 and 7, plotting the model results with the data shows essentially straight lines, even in Nd-Sr, Sr-Pb and Nd-Pb isotopic projections where hyperbolic trends are likely to occur. As described above, this feature can be explained by a K value of ~ 1 , showing that the ratios of elements (i.e., Sr/Nd, Sr/Pb and Nd/Pb) are approximately the same in both end-members. This seems to be paradoxical, since our model involves a mixing between the DMM and an enriched OIB source. Figure 8 presents histograms showing the frequency and variation of element ratios (Sr/Nd,

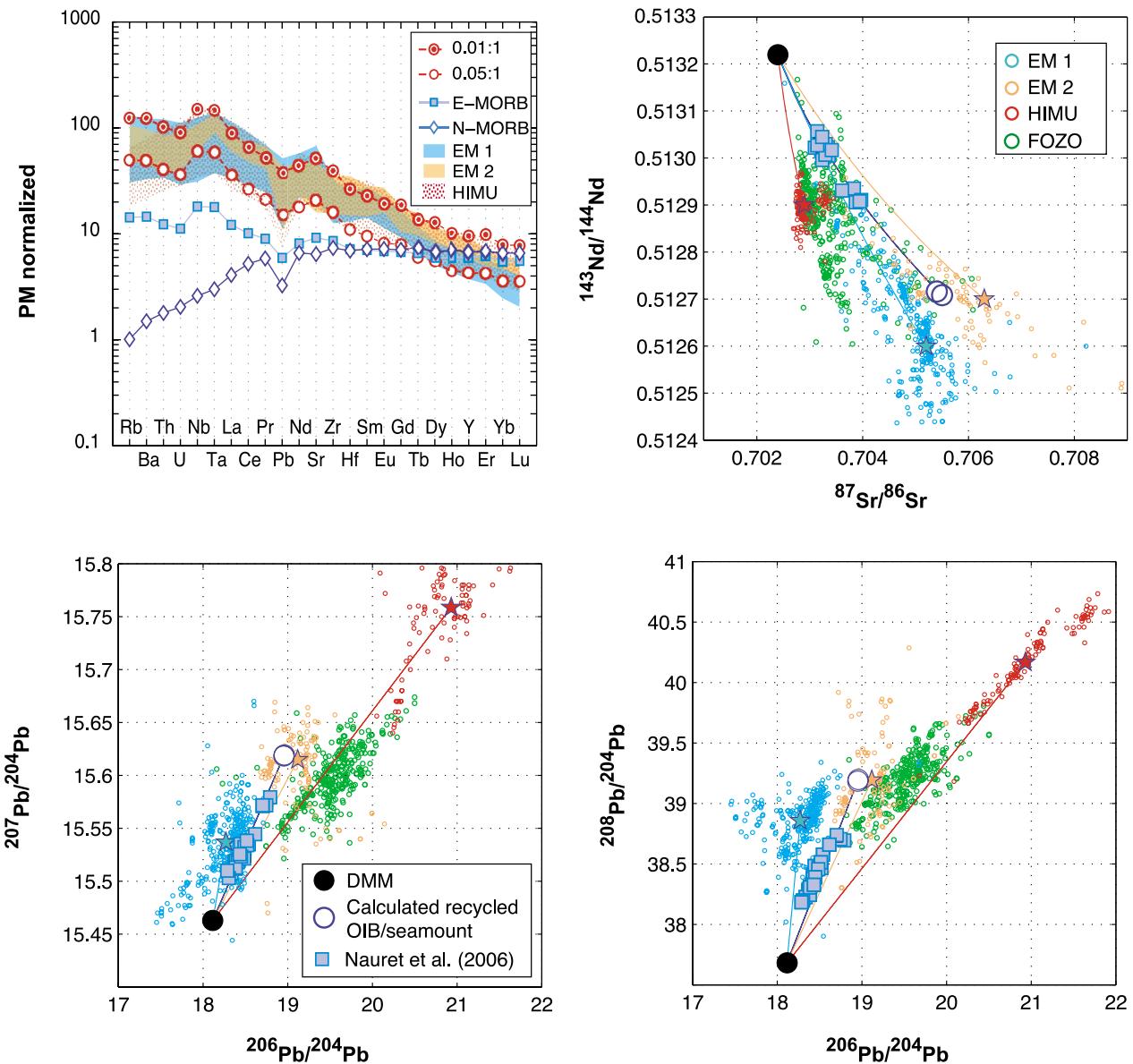


Figure 7. Binary mixing modeling between a recycled OIB/seamount and the N-MORB source applied on the samples from the 18°S region of the CIR [Nauret et al., 2006]. Parameters, given in Tables 2 and 3, were determined by the same way as for the MAR modeling. See the text and the Figure 6 caption for more details about the modeling.

Sr/Pb and Nd/Pb) in MORB and OIB. This figure illustrates the broad overlap in the range of Sr/Nd, Sr/Pb and Nd/Pb ratios in MORB and OIB. We found average ratios of 12.8, 247 and 20.1 for Sr/Nd, Sr/Pb and Nd/Pb ratios in MORB, respectively. Average ratios calculated for the OIB approximate those of MORB, leading to K values close to 1 (calculated here as the ratios in MORB divided by those in OIB). As an example, a virtual mixing between sediments and the DMM leads to a K that can be higher than 20 (calculated using the Nd/Pb ratios of the GLOSS from Plank and

Langmuir [1998]). Therefore, as element ratios in MORB and in OIB are similar, a mixing between these two components will form curvilinear trends rather than hyperbolas in isotope spaces.

5.5. Time Constraints

[37] Several studies try to constrain the age of the heterogeneities responsible for E-MORB genesis [e.g., Dosso et al., 1999; Cooper et al., 2004; Donnelly et al., 2004]. Using mantle isochrons and

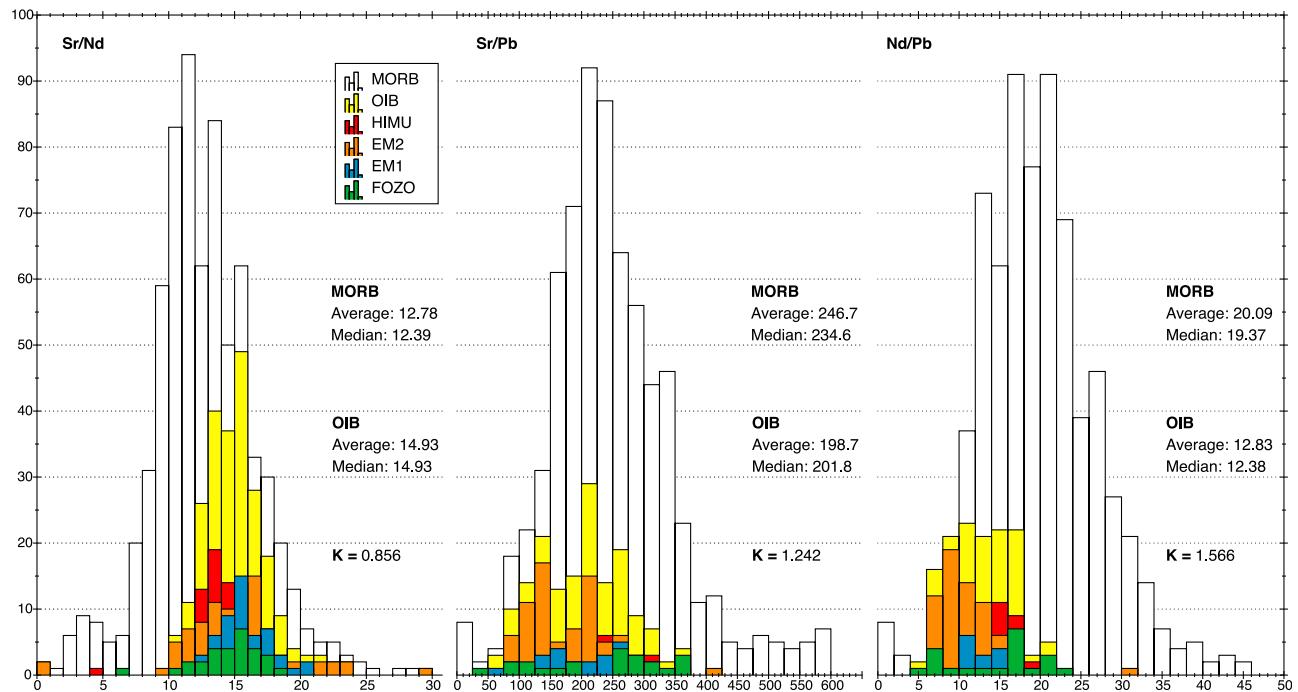


Figure 8. Histograms showing variations and frequencies of Sr/Nd, Sr/Pb and Nd/Pb elemental ratios of MORB and EM-1, EM-2, HIMU and FOZO OIB (the group OIB in yellow corresponds to the addition of the four OIB subgroups). MORB are compiled from the LDEO petrological database at <http://petdb.ldeo.columbia.edu/petdb> (see Tables S4 and S5 for data, selecting criteria and references). OIB data are from *Willbold and Stracke [2006]*. The formula for the K parameter is given in the section 5.4 of the discussion.

Monte-Carlo simulations on Sr, Nd and Pb isotope systems, *Dosso et al. [1999]* give an age of 250 Ma for the heterogeneity beneath the 14°N region. The same age is also given for the enriched component beneath the Azores [*Dosso et al., 1999; Cooper et al., 2004*]. *Cooper et al. [2004]* extrapolate this age to all enriched components present in the North-Atlantic MORB source (excluding deep hot spot), arguing that all of them represent recycled products subducted along the western margin of Pangea. In their study, *Donnelly et al. [2004]* also calculated an age of ~300 Ma based on mantle isochrons, but also showed that this age does not necessarily represent a specific event. Instead, they assume that this age results from the continuous formation and destruction of the enriched mantle source and depends on multiple parameters such as the decay constant of the radioactive decay system or the incompatibility of the daughter element during melting. They finally demonstrated that the mantle isochron ages could approximate the residence time of the enriched reservoir in the mantle if its mass represents a few percent of the system.

[38] Figure 9 shows the evolution of the $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ compositions with time, calculated

for both modeled recycled OIB/seamount (14°N in the MAR and 18°–20°S region of the CIR) and compared to the evolution of the DMM. Considering that the Pb isotopic composition of the recycled OIB/seamount cannot be more depleted than the DMM, a maximum age is given by the intersection between both evolution lines. Pb isotope systems give apparent ages ranging from 266 Ma to 295 Ma for the recycled OIB/seamount beneath the 14°N anomaly in the MAR, which is close to the ~250 Ma estimated by *Dosso et al. [1999]* and *Cooper et al. [2004]*. Age calculated for the recycled OIB/seamount in the CIR region ranges from 298 Ma to 332 Ma. Using extremely depleted isotopic compositions for the DMM (D-DMM in Figure 9), ages measured for recycling OIB in both regions are ~400–550 Ma. *Tackley [2000]* demonstrated that the time scale of complete mantle overturn (i.e., travel down to core-mantle boundary and back to surface) approximates 400–500 Myr. This suggests that recycling OIB underwent a maximum of one “cycle” within the mantle before reappearing beneath the ridge. Considering that the 400–500 Myr age is obtained for an extremely (anomalously) depleted mantle composition, an age of ~250 Ma seems

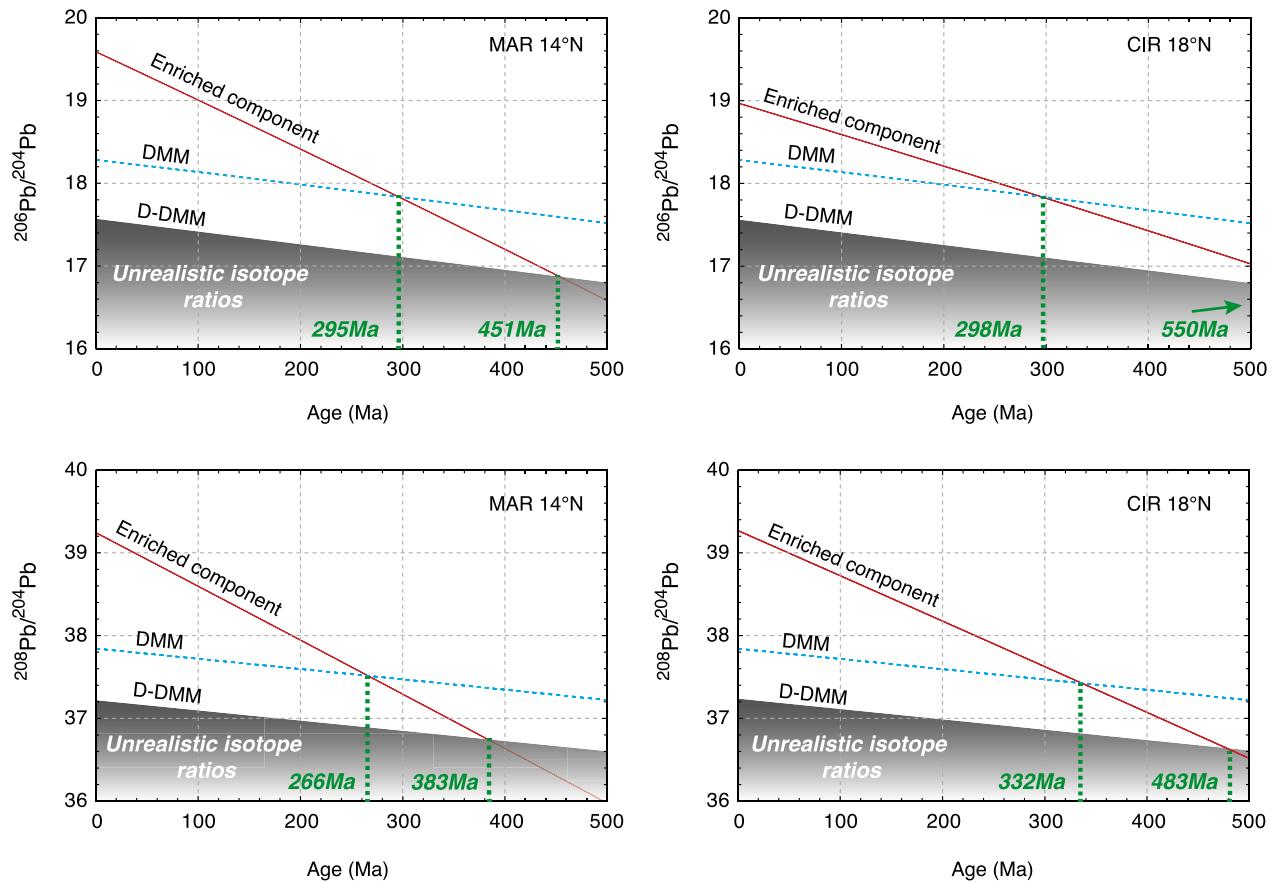


Figure 9. Evolution of $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ composition with time of modeled recycled OIB/seamount, the DMM and the “D-DMM,” using the isotopic decay equations ($^{206}\text{Pb}/^{204}\text{Pb}$) = ($^{206}\text{Pb}/^{204}\text{Pb}$)_t + $\mu(e^{\lambda^{238}t} - 1)$ and ($^{208}\text{Pb}/^{204}\text{Pb}$) = ($^{208}\text{Pb}/^{204}\text{Pb}$)_t + $\kappa(e^{\lambda^{232}t} - 1)$, where t is the time, μ is the $^{238}\text{U}/^{204}\text{Pb}$ ratio, κ is the $^{232}\text{Th}/^{238}\text{U}$ ratio, and λ^{238} and λ^{232} are the decay constants respectively for ^{238}U and ^{232}Th . Parameters are given in Table 2 and in Table S3. DMM Pb isotope compositions are from *Su and Langmuir* [2003] and correspond to a global average of N-MORB isotopic compositions from the MAR, CIR and EPR. D-DMM is calculated based on the isotopes that are 2σ depleted from the average, as calculated by *Workman and Hart* [2005]. Ages given by line intersections are interpreted as the maximum age of the recycled OIB/seamount, considering that the recycled OIB/seamount cannot be more depleted than the DMM.

more reasonable, suggesting that the recycled OIB/seamount likely never entered the deep mantle (Figure 10).

5.6. Reconciling the OIB/Seamount Recycling Model and the Oxygen Isotope Composition of MORB

[39] Recently, *Cooper et al.* [2009] have published a compilation of available laser fluorination $\delta^{18}\text{O}$ data for MORB from the Atlantic, Indian and Pacific ridges and have used it to constrain the nature and percentage of enriched material within the upper mantle globally. They revealed that the $\delta^{18}\text{O}$ compositions of MORB from each ocean basin fit a normal distribution, with identical means and standard deviations ($\delta^{18}\text{O}_{\text{mean}} \sim 5.50\text{\textperthousand}$) and

with a total range of 5.25–5.8%. In addition, they showed that the oxygen isotope data correlate with radiogenic isotope and trace element. Based on this, *Cooper et al.* [2009] suggested the DMM to have a $\delta^{18}\text{O}$ composition of $\sim 5.25\text{\textperthousand}$ and the average measured value of $5.5\text{\textperthousand}$ to reflect a global contamination of the upper mantle by recycling crustal material in proportion ranging from 5 to 10% (taking into account that these estimates may vary depending on the sedimentary contents in the recycled material). This model is much in favor of *Donnelly et al.*'s [2004] model rather than ours, since ones can assume that the OIB/seamount recycling model cannot account for the apparent correlation between oxygen isotope data and radiogenic isotope or trace element, because OIBs display the same range in $\delta^{18}\text{O}$ composition as

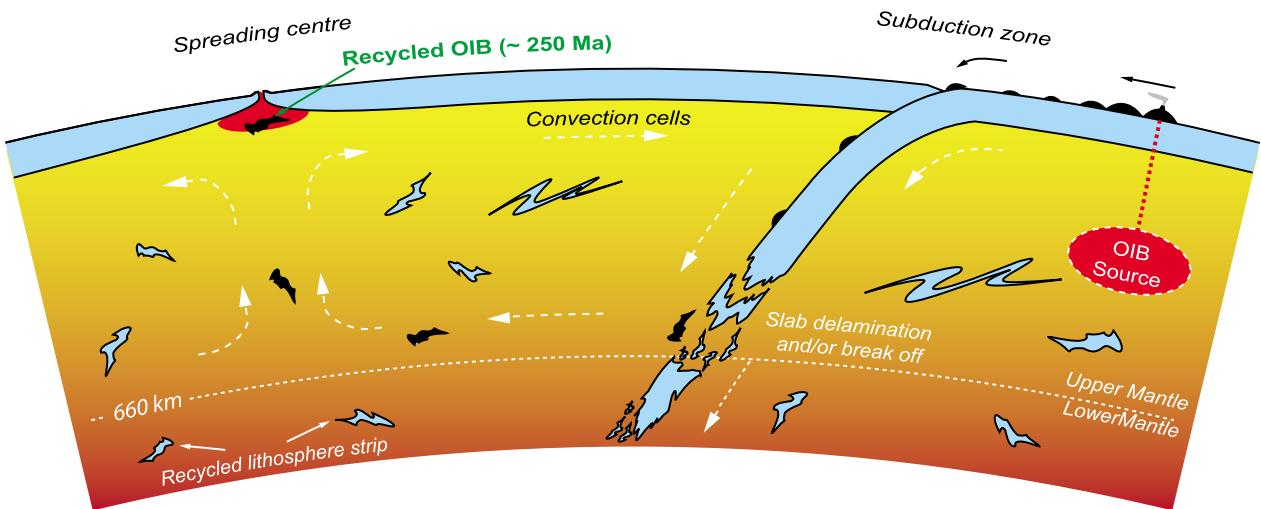


Figure 10. Sketch of E-MORB genesis (not at scale). The seamount is subducted and recycled with some of the oceanic crust into the upper mantle following the mantle convection to finish below a ridge segment to form E-MORB. As a complete mantle overturn (transit to core-mantle boundary and back to surface) is expected to be 400–500 Myr [Tackley, 2000], the ~250 Myr age calculated from isotope data suggests that parts of OIB are likely recycled within the upper mantle.

MORB [Eiler *et al.*, 1996; Eiler, 2001]. Eiler *et al.* [1996] have reported the $\delta^{18}\text{O}$ composition of olivine from basalts belonging to each subtype of OIB (i.e., HIMU, EM 1, EM 2 and others including some of the FOZO group). Their results show no evidence for $\delta^{18}\text{O}$ values outside the range for MORB, except for EM 2 type OIB, of which anomalously high $\delta^{18}\text{O}$ compositions reflect the presence of subducted sediments in their mantle source.

[40] Therefore, taking into account the results of Cooper *et al.* [2009], trends in $\delta^{18}\text{O}$ plotted versus trace elements or isotope data are not consistent with the OIB/seamount recycling model unless the OIB that are recycled are all EM 2. However, this assessment can be addressed by at least two reasons: First, although $\delta^{18}\text{O}$ data are clearly of a certain interest in the study of mantle heterogeneities, the correlations proposed by Cooper *et al.* [2009] are not always well defined (this is particularly true for Pacific samples which exhibit a large range of $\delta^{18}\text{O}$ compositions at a given Sr and Pb isotopic composition). Although the weakness of the correlations may be related to the lack of high resolution oxygen data such as those published by Cooper *et al.* [2004, 2009], these correlations have to be considered with caution and are not robust enough to rule out the OIB/seamount recycling model. Moreover, as oxygen isotope data are available neither in the 14°N region of the MAR nor in the 18–20°S of the CIR, there is no evidence for a

possible application of Cooper *et al.*'s [2009] model in these regions. That is not ruling out the model of Cooper *et al.* [2009], but without a denser database in oxygen isotopes for MORB, one could admit that correlations may occur at large scale (ocean scale) while they do not at more restricted scale (not more than few hundred kilometers). Second, the model defended by Cooper *et al.* [2009] is not incompatible with ours as both could actually happen synchronously in different localities, i.e., a recycled seamount could represent the enriched material in some areas while other types of enriched materials could contribute to form E-MORB in others. The main difference between the two models resides in that the recycling of the oceanic crust may account for large scale heterogeneities (e.g., the global enrichment in oxygen isotopes of the upper mantle at ocean scale, as postulated by Cooper *et al.* [2009]) while the recycling of OIB/seamounts are more likely to form very restricted MORB enrichments (no more than hundred kilometers).

5.7. Recycled Seamounts: More Than a Process for Local MORB Enrichment?

[41] We show in this paper that recycling OIB can be a viable process to explain the genesis of E-MORB at 14°N in the MAR, but it is also susceptible to explain E-MORB genesis along other ridges. As demonstrated above, E-MORB from the 18°S region of the CIR, and in regions potentially presenting similar geochemical features (MORB at

23°N and 33°S in the MAR and those from 11°20'N in the EPR) could be explained by this process. This suggests that the recycling of OIB is not an isolated process. *Wessel* [2001] and *Hillier and Watts* [2007] identified slightly less than 15,000 seamounts with a height >1.5 km, and almost a half of them resides on the Pacific plate. Although the size of a seamount is negligible compared to the volume of the mantle, calculations made by *Hillier and Watts* [2007] show that overall identified seamounts cover more than 18% of the oceanic seafloor area. A virtual addition of these numerous seamounts allows a thickening of the oceanic crust of about 180 m. Considering these estimates and those of *Hofmann* [1997] which showed that 20 km³ of oceanic crust are recycled per year, we can calculate that seamounts represent approximately 2% of the recycled material. Taking into account that the values given by *Hillier and Watts* [2007] are minima estimates, i.e., using only identified and measured edifices and that the authors expect around 3 million seamounts with a height >0.1 km, the proportion of seamounts in the recycled material can significantly increase. Thus, considering the trace element budget, the isotopic composition of the mantle and the 5–10% estimation for the total recycled material in the upper mantle with seamounts representing ~2%, it is clear that the recycling of seamount can hardly be considered as a significant contributor to the upper mantle heterogeneity. Nevertheless, it has to be considered as a plausible mechanism to generate (local) MORB enrichment.

6. Conclusions

[42] New trace element and isotope data have demonstrated that MORB enrichment from north and south of the 15°20'N FZ can find their origin in the geochemical anomaly centered at 14°N and associated with a high topographic bulge.

[43] Isotope ratios confirm and validate the enrichment model established on trace element by *Hémond et al.* [2006]: the E-MORB origin can be attributed to a binary mixing between N-MORB source and OIB/seamount material recycled through subduction. The age of the recycled OIB/seamount is estimated to 250 Ma.

[44] Finally, we show that the model described here can be applied to E-MORB from other ridge segments, and is thus unlikely to be an isolated process. Considering the great number of seamounts on the

ocean floor, their recycling may represent a significant process to account for MORB enrichment.

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References

- Abouchami, W., S. L. Goldstein, S. J. G. Gazer, A. Eisenhauer, and A. Mangini (1997), Secular changes of lead and neodymium in central Pacific seawater recorded by a Fe-Mn crust, *Geochim. Cosmochim. Acta*, 61(18), 3957–3974, doi:10.1016/S0016-7037(97)00218-4.
- Agranier, A., J. Blichert-Toft, D. Graham, V. Debaille, P. Schiano, and F. Albarède (2005), The spectra of isotopic heterogeneities along the Mid-Atlantic Ridge, *Earth Planet. Sci. Lett.*, 238(1–2), 96–109, doi:10.1016/j.epsl.2005.07.011.
- Allègre, C. J., B. Hamelin, and B. Dupré (1984), Statistical analysis of isotopic ratios in MORB—The Mantle Blob Cluster model and the convection regime of the mantle, *Earth Planet. Sci. Lett.*, 71(1), 71–84, doi:10.1016/0012-821X(84)90053-0.
- Arevalo, R., Jr., and W. F. McDonough (2010), Chemical variations and regional diversity observed in MORB, *Chem. Geol.*, 271, 70–85, doi:10.1016/j.chemgeo.2009.12.013.
- Ballmer, M. D., J. van Hunen, G. Ito, P. J. Tackley, and T. A. Bianco (2007), Non-hotspot volcano chains originating from small-scale sublithospheric convection, *Geophys. Res. Lett.*, 34, L23310, doi:10.1029/2007GL031636.
- Batiza, R. (1982), Abundances, distribution and sizes of volcanoes in the Pacific Ocean and implications for the origin of non-hotspot volcanoes, *Earth Planet. Sci. Lett.*, 60, 195–206, doi:10.1016/0012-821X(82)90003-6.
- Batiza, R., and D. Vanko (1984), Petrology of young Pacific seamounts, *J. Geophys. Res.*, 89, 11,235–11,260, doi:10.1029/JB089iB13p11235.
- Bonatti, E., A. Peyve, P. Kepezhinskas, N. Kurentsova, M. Seyler, S. Skolotnev, and G. Uditsev (1992), Upper mantle heterogeneity below the Mid-Atlantic Ridge, 0° to 15°N, *J. Geophys. Res.*, 97(B4), 4461–4476, doi:10.1029/91JB02838.
- Bougault, H., L. Dmitriev, J. G. Schilling, A. Sobolev, J. L. Joron, and H. D. Needham (1988), Mantle heterogeneity from trace elements: MAR triple junction near 14°N, *Earth Planet. Sci. Lett.*, 88(1–2), 27–36, doi:10.1016/0012-821X(88)90043-X.
- Cannat, M., D. Bideau, and H. Bougault (1992), Serpentized peridotites and gabbros in the Mid-Atlantic Ridge axial valley at 15°37'N and 16°52'N, *Earth Planet. Sci. Lett.*, 109(1–2), 87–106, doi:10.1016/0012-821X(92)90076-8.

- Cannat, M., Y. Lagabrielle, H. Bougault, J. Casey, N. de Coutures, L. Dmitriev, and Y. Fouquet (1997), Ultramafic and gabbroic exposures at the Mid-Atlantic Ridge: Geological mapping in the 15°N region, *Tectonophysics*, 279(1–4), 193–213, doi:10.1016/S0040-1951(97)00113-3.
- Castillo, P. R., J. H. Natland, Y. L. Niu, and P. F. Lonsdale (1998), Sr, Nd and Pb isotopic variation along the Pacific-Antarctic risecrest, 53–57°S: Implications for the composition and dynamics of the South Pacific upper mantle, *Earth Planet. Sci. Lett.*, 154(1–4), 109–125, doi:10.1016/S0012-821X(97)00172-6.
- Charlou, J. L., H. Bougault, P. Appriou, T. Nelsen, and P. Rona (1991), Different TDM/CH 4 hydrothermal plume signatures: TAG site at 26°N and serpentinized ultrabasic diapir at 15°05'N on the Mid-Atlantic Ridge, *Geochim. Cosmochim. Acta*, 55(11), 3209–3222, doi:10.1016/0016-7037(91)90484-M.
- Cooper, K. M., J. M. Eiler, P. D. Asimow, and C. H. Langmuir (2004), Oxygen isotope evidence for the origin of enriched mantle beneath the mid-Atlantic ridge, *Earth Planet. Sci. Lett.*, 220(3–4), 297–316, doi:10.1016/S0012-821X(04)00058-5.
- Cooper, K. M., J. M. Eiler, K. W. W. Sims, and C. H. Langmuir (2009), Distribution of recycled crust within the upper mantle: Insights from the oxygen isotope composition of MORB from the Australian-Antarctic Discordance, *Geochim. Geophys. Geosyst.*, 10, Q12004, doi:10.1029/2009GC002728.
- Cousens, B. L., J. F. Allan, M. I. Leybourne, R. L. Chase, and N. Vanwaggoner (1995), Mixing of magmas from enriched and depleted mantle sources: The northeast Pacific-West Valley segment, Juan de Fuca Ridge, *Contrib. Mineral. Petrol.*, 120(3–4), 337–357, doi:10.1007/BF00306512.
- DePaolo, D. J., and G. J. Wasserburg (1979), Petrogenetic mixing models and Nd-Sr isotopic patterns, *Geochim. Cosmochim. Acta*, 43, 615–627, doi:10.1016/0016-7037(79)90169-8.
- Donnelly, K. E., S. L. Goldstein, C. H. Langmuir, and M. Spiegelman (2004), Origin of enriched ocean ridge basalts and implications for mantle dynamics, *Earth Planet. Sci. Lett.*, 226(3–4), 347–366, doi:10.1016/j.epsl.2004.07.019.
- Dosso, L., B. B. Hanan, H. Bougault, J.-G. Schilling, and J.-L. Joron (1991), Sr-Nd-Pb geochemical morphology between 10° and 17°N on the Mid-Atlantic Ridge: A new MORB isotope signature, *Earth Planet. Sci. Lett.*, 106(1–4), 29–43, doi:10.1016/0012-821X(91)90061-L.
- Dosso, L., H. Bougault, and J.-L. Joron (1993), Geochemical morphology of the North Mid-Atlantic Ridge, 10°–24°N: Trace element-isotope complementarity, *Earth Planet. Sci. Lett.*, 120(3–4), 443–462, doi:10.1016/0012-821X(93)90256-9.
- Dosso, L., H. Bougault, C. H. Langmuir, C. Bollinger, O. Bonnier, and J. Etoubleau (1999), The age and distribution of mantle heterogeneity along the Mid-Atlantic Ridge (31–41°N), *Earth Planet. Sci. Lett.*, 170, 269–286.
- Dupré, B., and C. J. Allègre (1983), Pb-Sr isotope variation in Indian Ocean basalts and mixing phenomena, *Nature*, 303(5913), 142–146, doi:10.1038/303142a0.
- Eiler, J. M. (2001), Oxygen isotope variations of basaltic lavas and upper mantle rocks, *Rev. Mineral. Geochem.*, 43, 319–364, doi:10.2138/gsrmg.43.1.319.
- Eiler, J. M., K. A. Farley, J. W. Valley, A. W. Hofmann, and E. M. Stolper (1996), Oxygen isotope constraints on the sources of Hawaiian volcanism, *Earth Planet. Sci. Lett.*, 144, 453–467, doi:10.1016/S0012-821X(96)00170-7.
- Elliott, T., A. Zindler, and B. Bourdon (1999), Exploring the kappa conundrum: The role of recycling in the lead isotope evolution of the mantle, *Earth Planet. Sci. Lett.*, 169, 129–145.
- Escartín, J., and M. Cannat (1999), Ultramafic exposures and the gravity signature of the lithosphere near the Fifteen-Twenty Fracture Zone (Mid-Atlantic Ridge, 14°–16.5°N), *Earth Planet. Sci. Lett.*, 171(3), 411–424, doi:10.1016/S0012-821X(99)00169-7.
- Farley, K. A., J. H. Natland, and H. Craig (1992), Binary mixing of enriched and undegassed (primitive-questionable) mantle components (He, Sr, Nd, Pb) in Samoan lavas, *Earth Planet. Sci. Lett.*, 111, 183–199, doi:10.1016/0012-821X(92)90178-X.
- Faure, G. (1977), *Principles of Isotope Geology*, 464 pp., Wiley, New York.
- Garcia, M. O., B. Jorgenson, J. J. Mahoney, E. Ito, and A. Irving (1993), An evaluation of temporal geochemical evolution of Loihi seamount lavas: Results of Alvin submersible dives, *J. Geophys. Res.*, 98, 537–550, doi:10.1029/92JB01707.
- Graham, D. W., A. Zindler, M. D. Kurz, W. J. Jenkins, R. Batiza, and H. Staudigel (1988), He, Pb, Sr and Nd isotope constraints on magma genesis and mantle heterogeneity beneath young Pacific seamounts, *Contrib. Mineral. Petrol.*, 99, 446–463, doi:10.1007/BF00371936.
- Hall, L. S., J. J. Mahoney, J. M. Sinton, and R. A. Duncan (2006), Spatial and temporal distribution of a C-like asthenospheric component in the Rano Rahi Seamount Field, East Pacific Rise, 15°–19°S, *Geochim. Geophys. Geosyst.*, 7, Q03009, doi:10.1029/2005GC000994.
- Halliday, A., A. Dickin, A. Fallick, and J. Fitton (1988), Mantle dynamics—A Nd, Sr, Pb and Os isotopic study of the Cameroon Line Volcanic Chain, *J. Petrol.*, 29, 181–211.
- Hanan, B. B., and D. W. Graham (1996), Lead and helium isotope evidence from oceanic basalts for a common deep source of mantle plumes, *Science*, 272, 991–995, doi:10.1126/science.272.5264.991.
- Hart, S. (1984), A large-scale isotope anomaly in the Southern Hemisphere mantle, *Nature*, 309, 753–757, doi:10.1038/309753a0.
- Hart, S. (1988), Heterogeneous mantle domains: Signatures, genesis and mixing chronologies, *Earth Planet. Sci. Lett.*, 90, 273–296, doi:10.1016/0012-821X(88)90131-8.
- Hart, S., E. Hauri, L. Oschman, and J. Whitehead (1992), Mantle plumes and entrainment—Isotopic evidence, *Science*, 256, 517–520, doi:10.1126/science.256.5056.517.
- Hattori, K., and S. Guillot (2003), Volcanic fronts form as a consequence of serpentinite dehydration in the forearc mantle wedge, *Geology*, 31, 525–528, doi:10.1130/0091-7613(2003)031<0525:VFFAAC>2.0.CO;2.
- Hémond, C., A. W. Hofmann, I. Vlastelic, and F. Nauret (2006), Origin of MORB enrichment and relative trace element compatibilities along the Mid-Atlantic Ridge between 10° and 24°N, *Geochim. Geophys. Geosyst.*, 7, Q12010, doi:10.1029/2006GC001317.
- Hillier, J. K., and A. B. Watts (2007), Global distribution of seamounts from ship-track bathymetry data, *Geophys. Res. Lett.*, 34, L13304, doi:10.1029/2007GL029874.
- Hofmann, A. W. (1988), Chemical differentiation of the Earth: The relationship between mantle, continental crust, and oceanic crust, *Earth Planet. Sci. Lett.*, 90(3), 297–314, doi:10.1016/0012-821X(88)90132-X.
- Hofmann, A. W. (1997), Mantle geochemistry: The message from oceanic volcanism, *Nature*, 385(6613), 219–229, doi:10.1038/385219a0.

- Hofmann, A. W. (2003), Sampling mantle heterogeneity through oceanic basalts: Isotopes and trace elements, in *Treatise on Geochemistry*, pp. 61–97, Pergamon, Oxford, U.K.
- Iwamori, H. (1998), Transportation of H₂O and melting in subduction zones, *Earth Planet. Sci. Lett.*, 160, 65–80, doi:10.1016/S0012-821X(98)00080-6.
- Jackson, M. G., and R. Dasgupta (2008), Compositions of HIMU, EM1, and EM2 from global trends between radiogenic isotopes and major elements in ocean island basalts, *Earth Planet. Sci. Lett.*, 276(1–2), 175–186, doi:10.1016/j.epsl.2008.09.023.
- Jochum, K. P., B. Stoll, K. Herwig, and M. Willbold (2007), Validation of LA-ICP-MS trace element analysis of geological glasses using a new solid-state 193 nm Nd: YAG laser and matrix-matched calibration, *J. Anal. At. Spectrom.*, 22, 112–121, doi:10.1039/b609547j.
- Klein, E. M., and C. H. Langmuir (1987), Global correlations of ocean ridge basalt chemistry with axial depth and crustal thickness, *J. Geophys. Res.*, 92(B8), 8089–8115, doi:10.1029/JB092iB08p08089.
- Koppers, A. A. P., and A. B. Watts (2010), Intraplate seamounts as a window into deep Earth processes, *Oceanography*, 23, 42–57, doi:10.5670/oceanog.2010.61.
- Koppers, A. A. P., H. Staudigel, M. S. Pringle, and J. R. Wijbrans (2003), Short-lived and discontinuous intraplate volcanism in the South Pacific: Hotspots or extensional volcanism?, *Geochem. Geophys. Geosyst.*, 4(10), 1089, doi:10.1029/2003GC000533.
- Langmuir, C. H., R. Vocke, G. Hanson, and S. Hart (1978), A general mixing equation with applications to Icelandic basalts, *Earth Planet. Sci. Lett.*, 37, 380–392, doi:10.1016/0012-821X(78)90053-5.
- Lee, D., A. Halliday, J. Fitton, and P. Giampiero (1994), Isotopic variations with distance and time in the volcanic islands of the Cameroon line: Evidence for a mantle plume origin, *Earth Planet. Sci. Lett.*, 123, 119–138, doi:10.1016/0012-821X(94)90262-3.
- Leroex, A. P., H. J. B. Dick, and R. T. Watkins (1992), Petrogenesis of anomalous K-enriched MORB from the Southwest Indian Ridge 11°53'E to 14°38'E, *Contrib. Mineral. Petrol.*, 110(2–3), 253–268.
- Macdougall, J. D., and G. W. Lugmair (1986), Sr and Nd isotopes in basalts from the East Pacific Rise—Significance for mantle heterogeneity, *Earth Planet. Sci. Lett.*, 77(3–4), 273–284, doi:10.1016/0012-821X(86)90139-1.
- Manhès, G., J. F. Minster, and C. J. Allègre (1978), Comparative uranium-thorium-lead and rubidium-strontium study of the Saint Severin amphoteric: Consequences for early solar system chronology, *Earth Planet. Sci. Lett.*, 39(1), 14–24, doi:10.1016/0012-821X(78)90137-1.
- McDonough, W. F., and S. S. Sun (1995), The composition of the Earth, *Chem. Geol.*, 120(3–4), 223–253, doi:10.1016/0009-2541(94)00140-4.
- McKenzie, D., A. Stracke, J. Blachert-Toft, F. Albarède, K. Grönvold, and R. K. O'Nions (2004), Source enrichment processes responsible for isotopic anomalies in oceanic island basalts, *Geochim. Cosmochim. Acta*, 68(12), 2699–2724, doi:10.1016/j.gca.2003.10.029.
- Melson, W., T. O'Hearn, and E. Jarosewich (2002), A data brief on the Smithsonian Abyssal Volcanic Glass Data file, *Geochem. Geophys. Geosyst.*, 3(4), 1023, doi:10.1029/2001GC000249.
- Michael, P. J., et al. (1994), Mantle control of a dynamically evolving spreading center: Mid-Atlantic Ridge 31–34°S, *Earth Planet. Sci. Lett.*, 121(3–4), 451–468, doi:10.1016/0012-821X(94)90083-3.
- Morgan, W. J. (1971), Convection plumes in the lower mantle, *Nature*, 230, 42–43, doi:10.1038/230042a0.
- Nauret, F., W. Abouchami, S. J. G. Galer, A. W. Hofmann, C. Hémond, C. Chauvel, and J. Dymnt (2006), Correlated trace element-Ph isotope enrichments in Indian MORB along 18–20°S, Central Indian Ridge, *Earth Planet. Sci. Lett.*, 245(1–2), 137–152, doi:10.1016/j.epsl.2006.03.015.
- Niu, Y. L., and R. Batiza (1997), Trace element evidence from seamounts for recycled oceanic crust in the eastern Pacific mantle, *Earth Planet. Sci. Lett.*, 148(3–4), 471–483, doi:10.1016/S0012-821X(97)00048-4.
- Niu, Y., and R. Hékinian (1997), Basaltic liquids and harzburgitic residues in the Garrett Transform: A case study at fast-spreading ridges, *Earth Planet. Sci. Lett.*, 146, 243–258.
- Niu, Y., C. H. Langmuir, and R. Kinzler (1997), The origin of abyssal peridotites: A new perspective, *Earth Planet. Sci. Lett.*, 152, 251–265.
- Niu, Y. L., K. D. Collerson, R. Batiza, J. I. Wendt, and M. Regelous (1999), Origin of enriched-type mid-ocean ridge basalt at ridges far from mantle plumes: The East Pacific Rise at 11°20'N, *J. Geophys. Res.*, 104(B4), 7067–7087, doi:10.1029/1998JB900037.
- Niu, Y. L., M. Regelous, I. J. Wendt, R. Batiza, and M. J. O'Hara (2002), Geochemistry of near-EPR seamounts: Importance of source vs. process and the origin of enriched mantle component, *Earth Planet. Sci. Lett.*, 199(3–4), 327–345, doi:10.1016/S0012-821X(02)00591-5.
- Plank, T., and C. H. Langmuir (1998), The chemical composition of subducting sediment and its consequences for the crust and mantle, *Chem. Geol.*, 145, 325–394, doi:10.1016/S0009-2541(97)00150-2.
- Richard, P., N. Shimizu, and C. J. Allègre (1976), ¹⁴³Nd/¹⁴⁶Nd, a natural tracer: An application to oceanic basalts, *Earth Planet. Sci. Lett.*, 31(2), 269–278, doi:10.1016/0012-821X(76)90219-3.
- Rüpke, L., J. Morgan, M. Hort, and J. Connolly (2004), Serpentine and the subduction zone water cycle, *Earth Planet. Sci. Lett.*, 223, 17–34, doi:10.1016/j.epsl.2004.04.018.
- Saunders, A., M. Norry, and J. Tarney (1988), Origin of MORB and chemically-depleted mantle reservoirs: Trace element constraints, *J. Petrol.*, 1, 415–445.
- Schilling, J. G. (1973), Iceland mantle plume: Geochemical study of Reykjanes ridge, *Nature*, 242(5400), 565–571, doi:10.1038/242565a0.
- Schilling, J. G., G. Thompson, R. Kingsley, and S. Humphris (1985), Hotspot-migrating ridge interaction in the South Atlantic, *Nature* 313, 187–191.
- Schilling, J. G., B. B. Hanan, B. McCully, R. H. Kingsley, and D. Fontignie (1994), Influence of the Sierra Leone mantle plume on the equatorial Mid-Atlantic Ridge: A Nd-Sr-Pb isotopic study, *J. Geophys. Res.*, 99(B6), 12,005–12,028, doi:10.1029/94JB00337.
- Shaw, D. (1970), Trace element fractionation during anatexis, *Geochim. Cosmochim. Acta*, 34, 331–340.
- Smith, W. H. F., and D. T. Sandwell (1997), Global sea floor topography from satellite altimetry and ship depth soundings, *Science*, 277, 1956–1962.
- Staudacher, T., P. Sarda, S. H. Richardson, C. J. Allègre, I. Sagna, and L. V. Dmitriev (1989), Noble gases in basalt glasses from a Mid-Atlantic Ridge topographic high at 14°N: Geodynamic consequences, *Earth Planet. Sci. Lett.*, 96(1–2), 119–133, doi:10.1016/0012-821X(89)90127-1.

- Staudigel, H., K. Park, M. Pringle, J. Rubenstone, W. Smith, and A. Zindler (1991), The longevity of the South-Pacific isotopic and thermal anomaly, *Earth Planet. Sci. Lett.*, **102**, 24–44, doi:10.1016/0012-821X(91)90015-A.
- Stracke, A., M. Bizimis, and V. J. M. Salters (2003), Recycling oceanic crust: Quantitative constraints, *Geochem. Geophys. Geosyst.*, **4**(3), 8003, doi:10.1029/2001GC000223.
- Stracke, A., A. Hofmann, and S. Hart (2005), FOZO, HIMU, and the rest of the mantle zoo, *Geochem. Geophys. Geosyst.*, **6**, Q05007, doi:10.1029/2004GC000824.
- Su, Y., and C. H. Langmuir (2003), Global MORB chemistry compilation at the segment scale, PhD thesis, Dep. of Earth and Environ. Sci., Columbia Univ., New York.
- Sun, S. S., R. W. Nesbitt, and A. Y. Sharaskin (1979), Geochemical characteristics of mid-ocean ridge basalts, *Earth Planet. Sci. Lett.*, **44**(1), 119–138, doi:10.1016/0012-821X(79)90013-X.
- Tackley, P. J. (2000), Mantle convection and plate tectonics: Toward an integrated physical and chemical theory, *Science*, **288**, 2002–2007.
- Turner, S., C. Hawkesworth, N. Rogers, and P. King (1997), U-Th isotope disequilibria and ocean island basalt generation in the Azores, *Chem. Geol.*, **139**(1–4), 145–164, doi:10.1016/S0009-2541(97)00031-4.
- Vidal, P., and N. Clauer (1981), Pb and Sr isotopic systematics of some basalts and sulfides from the East Pacific Rise at 21°N (project RITA), *Earth Planet. Sci. Lett.*, **55**(2), 237–246, doi:10.1016/0012-821X(81)90103-5.
- von Blanckenburg, F., R. K. O’Nions, and J. R. Heinz (1996), Distribution and sources of pre-anthropogenic lead isotopes in deep ocean water from Fe-Mn crusts, *Geochim. Cosmochim. Acta*, **60**(24), 4957–4963, doi:10.1016/S0016-7037(96)00310-9.
- Watts, A. B., D. T. Sandwell, W. H. F. Smith, and P. Wessel (2006), Global gravity, bathymetry, and the distribution of submarine volcanism through space and time, *J. Geophys. Res.*, **111**, B08408, doi:10.1029/2005JB004083.
- Wessel, P. (2001), Global distribution of seamounts inferred from gridded Geosat/ERS-1 altimetry, *J. Geophys. Res.*, **106**, 19,431–19,441, doi:10.1029/2000JB000083.
- White, W. (1985), Sources of oceanic basalts—Radiogenic isotopic evidence, *Geology*, **13**, 115–118, doi:10.1130/0091-7613(1985)13<115:SOOBRI>2.0.CO;2.
- White, W. M., and A. W. Hofmann (1982), Sr and Nd isotope geochemistry of oceanic basalts and mantle evolution, *Nature*, **296**(5860), 821–825, doi:10.1038/296821a0.
- White, W. M., and J. Patchett (1984), Hf-Nd-Sr isotopes and incompatible element abundances in island arcs: Implications for magma origins and crust-mantle evolution, *Earth Planet. Sci. Lett.*, **67**(2), 167–185, doi:10.1016/0012-821X(84)90112-2.
- Willbold, M., and A. Stracke (2006), Trace element composition of mantle end-members: Implications for recycling of oceanic and upper and lower continental crust, *Geochem. Geophys. Geosyst.*, **7**, Q04004, doi:10.1029/2005GC001005.
- Willbold, M., and A. Stracke (2010), Formation of enriched mantle components by recycling of upper and lower continental crust, *Chem. Geol.*, **276**, 188–197.
- Workman, R. K., and S. R. Hart (2005), Major and trace element composition of the depleted MORB mantle (DMM), *Earth Planet. Sci. Lett.*, **231**, 53–72.
- Zindler, A., and S. Hart (1986), Chemical Geodynamics, *Annu. Rev. Earth Planet. Sci.*, **14**, 493–571, doi:10.1146/annurev.earth.14.050186.002425.
- Zindler, A., H. Staudigel, and R. Batiza (1984), Isotope and trace element geochemistry of young Pacific seamounts: Implications for the scale of upper mantle heterogeneity, *Earth Planet. Sci. Lett.*, **70**, 175–195, doi:10.1016/0012-821X(84)90004-9.