

On the origin of EM–I end-member

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With 3 figures

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Abstract: On the basis of geochemical data available for oceanic basalts distinct mantle reservoirs have been identified during the last decades. Among these reservoirs, the EM–I end-member (Enriched Mantle type I) shows almost unique geochemical characteristics. The definition of the EM–I end member and the petrogenetic processes related to its origin are still debated: the geochemical composition of EM–I basalts has been interpreted in terms of crust-mantle interaction during 1) mantle recycling of subducted crustal material; 2) detachment of lithospheric keels in over-thickened regions; 3) thermal erosion by anomalously hot mantle. Key features of the EM–I end-member are: low uraniumogenic Pb ratios ($^{206}\text{Pb}/^{204}\text{Pb} < 17$), slightly radiogenic Sr isotopes ($^{87}\text{Sr}/^{86}\text{Sr} \sim 0.706$), unradiogenic Nd ($^{143}\text{Nd}/^{144}\text{Nd} \sim 0.5121$), unradiogenic Hf ($^{176}\text{Hf}/^{177}\text{Hf} \sim 0.2826$), slightly radiogenic Os ($^{187}\text{Os}/^{188}\text{Os} \sim 0.135\text{--}0.145$), high $^{207}\text{Pb}/^{206}\text{Pb}$ (>0.85), and $^{208}\text{Pb}/^{206}\text{Pb}$ (>2.08). EM–I basalts show apparent or inexistent Ba, Pb, Eu or Sr anomalies in primitive mantle-normalized diagrams and have Ba/Nb ratios ranging from 3.5 to 47.4, Ce/Pb from 1.2 to 24.6, Nb/U from 10.5 to 71.8, Sr/Nd from 6.2 to 36.4, and Eu/Eu* from 0.83 to 1.25. On these basis, the origin and evolution of the EM–I end-member cannot be explained by a single petrological model, and even the definition of elemental and isotopic parameters is controversial.

Introduction

The description of geochemical heterogeneities in the Earth's mantle, and the proposal of models to explain these anomalies have been some of the most important goals of the modern igneous petrology (e.g., ZINDLER & HART 1986, HOFMANN 1997). Many studies carried out during the last decades have evidenced chemical anomalies deviating from the relatively homogeneous upper asthenospheric depleted mantle composition. In particular, incompatible trace element investigations coupled with radiogenic and stable isotopic (Sr, Nd, Pb, Os, O, Hf, He and Ne) studies of oceanic basalts allowed for the identification of distinct geochemical reservoirs in the Earth's mantle.

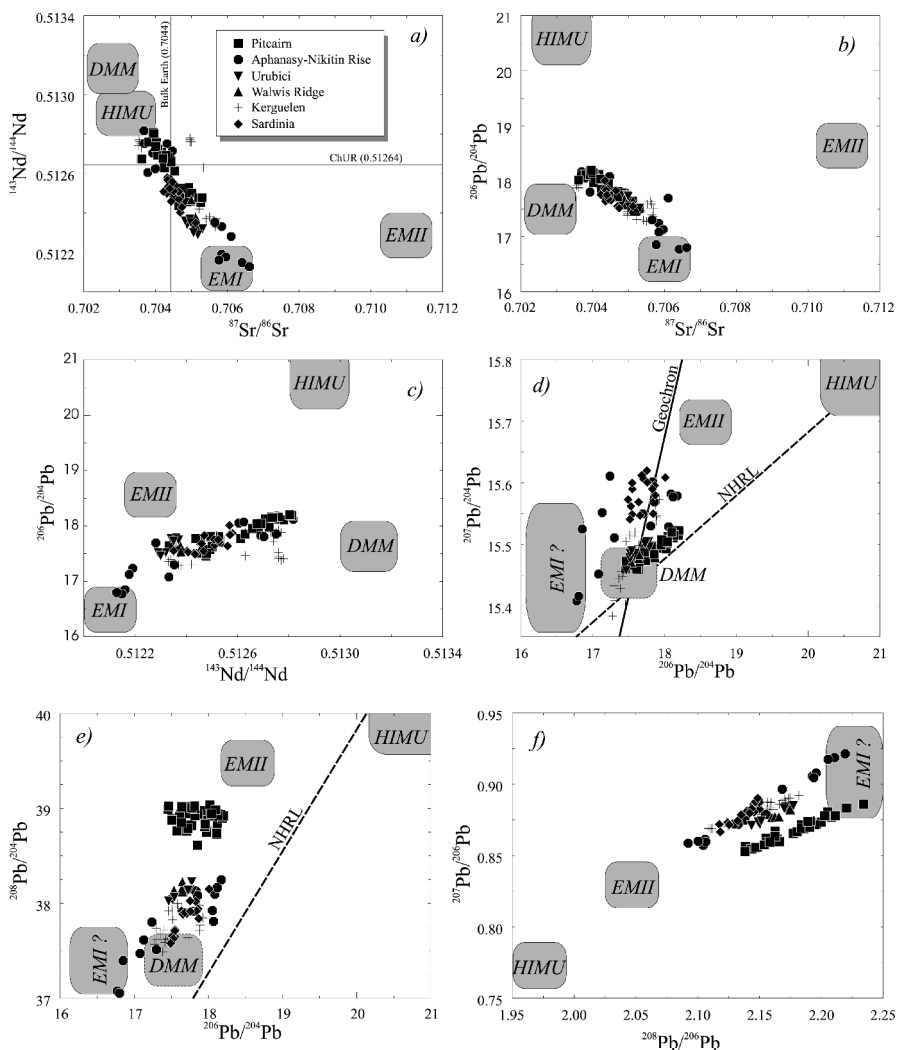


Fig. 1. Sr–Nd–Pb isotopic systematics of EM–I basalts from continental and oceanic settings: Pitcairn islands and seamounts (Pacific Ocean; WOODHEAD & DEVEY 1993, EILER et al. 1995, EISELE et al. 2002); Aphanasy–Nikitin Rise (Indian Ocean; MAHONEY et al. 1996, BORISOVA et al. 2001); Urubici magma-type (Paraná, Brazil; PEATE et al. 1999); Walvis Ridge (Atlantic Ocean; RICHARDSON et al. 1984); Kerguelen plateau (Indian Ocean; FREY et al. 2002); Sardinia (Italy; GASPERINI et al. 2000, LUSTRINO et al. 2000, 2002). DMM = Depleted Morb Mantle; HIMU = High μ ; EM–I and EM–II = Enriched Mantle I and II. NHRL = Northern Hemisphere Reference Line; HART 1984). Contour of Pb isotopic field of DMM is a broken line because of the extreme variability of Atlantic, Pacific and Indian MORBs in terms of $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$.

Among oceanic island basalts (OIBs), at least three geochemical end-members have been identified (ZINDLER & HART 1986): HIMU (high μ , where μ is $^{238}\text{U}/^{204}\text{Pb}$ ratio), EM–I and EM–II (Enriched Mantle I and II). Mixing of these three end-members together with the DMM (Depleted MORB Mantle) component can explain many of the geochemical features of oceanic basalts (e.g., CARLSON 1995). Together with these mantle end-members, other terms such as EM–III (HEAMAN 1989), PREMA (PREvailing MAnTle; STEIN & HOFMANN 1994), FOZO (FOcus ZOne; HART et al. 1992), PHEM (Primitive He Mantle; FARLEY & CRAIG 1992) and C component (HANAN & GRAHAM 1996) have been proposed as other possible reservoirs.

Originally thought for basalts from oceanic islands, these terms have been transposed and widely used also to explain the petrogenesis of basalts occurring in continental settings (CHUNG et al. 1995, MILNER & LE ROEX 1996, PEATE et al. 1999, LUSTRINO et al. 2000, 2002). Among these reservoirs, the definition of the EM–I end-member is probably the most puzzling. Since the pioneering work of ZINDLER & HART (1986), the type locality for EM–I has changed from Walvis Ridge (ODP Site 525A) in Atlantic Ocean (RICHARDSON et al. 1984, ZINDLER & HART 1986) to Pitcairn islands and seamounts in Pacific Ocean (WOODHEAD & DEVEY 1993, EISELE et al. 2002), Tristan and Inaccessible islands in Atlantic Ocean (HOFMANN 1997), Aphanasy Nikitin Rise (MAHONEY et al. 1996, BORISOVA et al. 2001) and eventually Kerguelen islands in the Indian Ocean (FREY et al. 2002). Trace element and isotopic compositions have been interpreted in terms of mixing of various components such as pelagic and/or terrigenous sediments, sub-continental lithospheric mantle (SCLM), ancient oceanic plateaux, oceanic crust, lower continental crust. These components could be recycled either via 1) subduction, 2) delamination and detachment processes, 3) thermal erosion by anomalously hot mantle in a variably depleted asthenosphere. Metasomatism of mantle peridotite by interaction with low V/Pb, low Rb/Sr, LREE-rich carbonatite melts has been proposed as an alternative model to explain the EM–I end-member Features (CARLSON 1995). Up to now, it has not been possible to define a common mechanism (or combination of processes) which can account for the geochemical features of oceanic and continental EM–I basalts from different geological settings.

Radiogenic and stable isotopes constraints

The key features of the EM–I end-member can be summarized as follow (Fig. 1): low uranogenic Pb ratios (e.g., $^{206}\text{Pb}/^{204}\text{Pb} < 17$), coupled with slightly radiogenic Sr isotopes ($^{87}\text{Sr}/^{86}\text{Sr} \sim 0.706$), unradiogenic Nd ($^{143}\text{Nd}/^{144}\text{Nd} \sim 0.5121$), unradiogenic Hf ($^{176}\text{Hf}/^{177}\text{Hf} \sim 0.2826$), and slightly radiogenic Os

($^{187}\text{Os}/^{188}\text{Os} \sim 0.135\text{--}0.145$). Hf and Os isotopes of EM-I have not been plotted in Fig. 1 because of the paucity of data available in literature. Typical features of EM-I basalts are also high $^{207}\text{Pb}/^{206}\text{Pb}$ (>0.85) and high $^{208}\text{Pb}/^{206}\text{Pb}$ (>2.08 ; LUSTRINO 2000). With regards to the Sr-, Nd-, and Pb-isotopic compositions, many of the EM-I basalts define broad compositional fields (Fig. 1 a–c), which can be interpreted in terms of mixing processes between a pure EM-I reservoir and the HIMU-DMM component, the DMM prevailing over HIMU. Still, this simple mixing model is not sufficient to explain the geochemical characteristics of several EM-I basalts (Fig. 1 d–f). Indeed, the strong variations of the $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ (and $^{207}\text{Pb}/^{206}\text{Pb}$, and $^{208}\text{Pb}/^{206}\text{Pb}$) ratios unlikely reflect a single EM-I reservoir. In particular, the $^{207}\text{Pb}/^{204}\text{Pb}$ ratios of the Aphanasy–Nikitin Rise basalts range from ~ 15.4 to 15.6 , with a nearly constant $^{206}\text{Pb}/^{204}\text{Pb}$ (~ 17). Similarly, decoupling of the $^{207}\text{Pb}/^{204}\text{Pb}$ ratio from the $^{206}\text{Pb}/^{204}\text{Pb}$ ratio is envisaged for Kerguelen and, to a minor extent, for Sardinian EM-I basalts. The $^{208}\text{Pb}/^{204}\text{Pb}$ (thorogenic) ratios are even more puzzling: the Pitcairn basalts show constant and high ratios for these isotopes ($^{208}\text{Pb}/^{204}\text{Pb} \sim 38.9$), above those of other EM-I basalts with comparable $^{206}\text{Pb}/^{204}\text{Pb}$ ratios ($^{208}\text{Pb}/^{204}\text{Pb} \sim 37.5\text{--}38.2$ with $^{206}\text{Pb}/^{204}\text{Pb} \sim 17.5\text{--}18.0$).

Proposed models

The Pitcairn islands and seamounts are the type-locality for EM-I basalts (WOODHEAD & DEVEY 1993), so that a wealth of data have been produced for these lavas. On the basis of trace element and Sr–Nd–Pb–Os–Hf isotopic ratios, EISELE et al. (2002) inferred that 1) the geochemical characteristics of the EM-I component are strongly influenced by a small fraction of pelagic and terrigenous sediments (generally $< 5\%$ in fraction); 2) even a consistent amount (up to 60%) of recycled SCLM does not contribute significantly to the Os isotopic budget. Because the addition of Os-rich and Re-poor SCLM results in low (subchondritic) $^{187}\text{Os}/^{188}\text{Os}$ ratios (<0.127), the relatively radiogenic Os composition of EM-I ($^{187}\text{Os}/^{188}\text{Os} > 0.135$) requires the involvement of some Re-rich component with high time-integrated $^{187}\text{Re}/^{188}\text{Os}$ ratios.

Addition of ancient SCLM, and/or sediment recycling have been called upon for the origin of the EM-I signature of continental basalts (e.g., CHUNG et al. 1995, MILNER & LE ROEX 1996, SMITH 1998); the involvement of lower continental crust may also produce the trace element and isotopic compositions of EM-I-like oceanic and continental igneous products (e.g., HART & ZINDLER 1989, CARLSON 1995, LUSTRINO et al. 2000, TATSUMI 2000, BORISOVA et al. 2001, FREY et al. 2002).

The lower crust-like signature of the EM–I basalts may result from a source process, rather than magma contamination en route to the surface (e.g., LUSTRINO 2000), as consequence of delamination and/or detachment of dense granulitic to eclogitic lower crust (density up to 3.5 g/cm^3 ; WOLF & WYLLIE 1994). Amphibolite to granulite (and eclogite) metamorphic reactions in the lower crust are associated with partial melting and density increase (with increasing garnet content). LILE and LREE are readily mobilized during partial melting, and low Rb/Sr and U/Pb ratios can be produced (e.g., RUDNICK & TAYLOR 1987). The Sm/Nd ratio is virtually unaffected by this process because REEs are less mobile under these metamorphic conditions (e.g., FARMER 1992, DOSTAL et al. 1998). Lower crust has lower Rb/Sr (0.02–0.14) and lower U/Pb (0.05–0.11) ratios (and lower time-integrated $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$) than upper crust (Rb/Sr = 0.26–0.35; U/Pb = 0.08–0.15; RUDNICK & FOUNTAIN 1995, WEDEPOHL 1995, GAO et al. 1998, LIU et al. 2001); conversely, Sm/Nd (0.120 for average lower continental crust and 0.121 for average upper crust) and Hf–Nd isotopic ratios of these two reservoirs overlap (VERVOORT et al. 2000 and references therein). These characteristics can be used to stress the higher mobility of ^{87}Rb and ^{238}U isotopes relative to ^{147}Sm , by considering the relations of $^{87}\text{Sr}/^{86}\text{Sr}$, $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ with Hf isotopic ratios. The upper and lower crustal lithologies have generally ϵ_{Hf} and $\epsilon_{\text{Nd}} < 0$, whereas mantle rocks have positive values. Therefore, mantle sources contaminated with lower crustal lithologies can evolve toward slightly radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$, unradiogenic $^{143}\text{Nd}/^{144}\text{Nd}$ and unradiogenic $^{206}\text{Pb}/^{204}\text{Pb}$ compositions. In particular, the Sr–Nd isotopic field of EM–I basalts is offset to low Nd from mantle array, as evidenced by the low ΔNd (down to -50) (ZINDLER et al. 1982, HART et al. 1986).

The Os isotopic ratios of lower crust can vary from subchondritic to suprachondritic values in the case of cumulitic/restitic or differentiate compositions, respectively (e.g., ESSER et al. 1993, REISBERG et al. 1993). No correlation between Os and Pb isotopic ratios has been evidenced (EISELE et al. 2002); this is likely due to the influence of the recycled material (oceanic crust vs. sediments) on the Pb isotopic composition. Overall, the covariations among Sr, Nd, Pb and Hf isotope compositions can be interpreted as resulting from similar factors, whereas the Os isotopic composition reflects the overall amount of recycled material and the degree of peridotite melting. Similarly, the Hf isotopic composition of the lower crust is variable from subchondritic values ($\epsilon_{\text{Hf}} < -30$ in the case of cumulitic composition and/or with metamorphic garnet) to radiogenic ($\epsilon_{\text{Hf}} > 20$ in the case of differentiation processes in the presence of garnet residual from melting or differentiation events (e.g., VERVOORT et al. 2000).

The study of oxygen isotope composition of EM–I basalts and their main phases (e.g. olivine, clinopyroxene) can be considered a potential tool to trace the addition of crustal material to their mantle source. This because of the strong fractionation between ^{18}O and ^{16}O isotopes in upper crustal and mantle reservoirs (JAMES 1981, ITO et al. 1987, BALDRIDGE et al. 1996). High $\delta^{18}\text{O}$ values (up to 25‰) are typical of terrigenous and pelagic sediments, and crustal rocks variably exposed to low temperatures (<300 °C) water-rock interaction (e.g., EILER et al. 2000 and references therein). On the other hand, high temperature (>350 °C) hydrothermal alteration shift original $\delta^{18}\text{O}$ toward very low values (down to –10‰; GREGORY & TAYLOR 1981). Peridotitic mantle and MORBs have a rather uniform $\delta^{18}\text{O}$ values (5.5–5.9‰; ITO et al. 1987, MATTEY et al. 1994, EILER et al. 1997) and the measurements that deviate sensibly from this range (down to 4.6‰ and up to 7.5‰; HARMON & HOEFS 1995) have been related to different degrees of alteration of the rocks and/or their MORB glass fraction (EILER et al. 1997). Investigation performed via laser fluorination of MORB olivine (EILER et al. 1997) evidenced uniform $\delta^{18}\text{O}$ values ($5.16 \pm 0.09\%$); these values are consistent with the $\Delta_{\text{basalt-olivine}} = 0.4\text{--}0.5\%$ inferred on the basis of several studies (ANDERSON et al. 1971, MATTHEWS et al. 1998, APPORA et al. 2003). Thus, olivine $\delta^{18}\text{O}$ values $\sim 5.2 \pm 0.2\%$ can be considered the baseline to check isotopic anomalies of the source.

WOODHEAD et al. (1993) interpreted the high $\delta^{18}\text{O}$ values (5.9–7.4‰) measured for basaltic glasses of Pitcairn seamounts in terms of source contamination by subducted crustal material with high $\delta^{18}\text{O}$ (>15‰). Accordingly, the ^{18}O -enriched compositions of Pitcairn glasses would result from the incorporation of significant amounts (up to 9 wt %) of crustal rocks. A substantial contribution of recycled sediments to Pitcairn mantle source was ruled out on the basis of homogeneous and low $\delta^{18}\text{O}$ values of olivine phenocrysts ($5.21\% \pm 0.08$), similar to the $\delta^{18}\text{O}$ values of olivine from MORBs (EILER et al. 1995). The high whole rocks $\delta^{18}\text{O}$ values were interpreted as due mostly to sample alteration, and not to anomalous mantle sources (EILER et al. 1995). The $^{18}\text{O}/^{16}\text{O}$ ratios of olivine from Pitcairn limit the possible amount of crustal material involved in EM–I source to be 1–2 wt % (EILER et al. 1995). More data are needed to constrain the role of crustal recycling in EM–I basalts, and to evaluate the $\delta^{18}\text{O}$ variations as a function of different degrees of partial melting (EILER et al. 2000).

Further constraints on the genesis of the EM–I mantle source could be obtained by noble gas (He–Ne) isotope systematics. The $^3\text{He}/^4\text{He}$ (R/R_A) ratios of Polynesian EM–I–OIBs (~ 9.5 at Rarotonga; HANYU et al. 1999; 8–13 at Pitcairn; EISELE et al. 1997, and references therein) and MORB values ($\sim 8\text{--}9$; KURZ et al. 1982, HILTON et al. 1993, NIEDERMAN et al. 1997) overlap. These values are lower than those reported for Loihi lavas ($\sim 20\text{--}30$; HONDA et al.

1991, 1993), and suggest that EM–I basalts (and MORB) were derived from a source with high $(U+Th)/^3\text{He}$ ratios. These latter ratios may result from the interaction of subducted crustal rocks with very low ^3He (primitive He), and high ^4He (derived by radioactive decay of U and Th) contents. Thus the $^3\text{He}/^4\text{He}$ ratio of EM–I basalt would arise from the recycling of oceanic crust (very low $^3\text{He}/^4\text{He}$) in the upper mantle, whereas the Loihi lavas would represent melts of undegassed mantle. If this holds true, the higher ^3He concentration of MORB (between 10^{-11} and 10^{-9} cm^3 STP/g) relative to Loihi and Kilaueha (between 10^{-13} and 10^{-10} cm^3 STP/g; HIYAGON et al. 1992, HONDA et al. 1993), is the opposite of what expected for a depleted (and therefore degassed) MORB source (HONDA et al. 1993, OZIMA & IGARASHI 2000). Moreover, HIMU basalts show $^3\text{He}/^4\text{He}$ ratios higher (Hawaii and Iceland) and lower (St. Helena, Mangaia, Tubuai) than MORB, thus requiring different origins for the HIMU source, at least from the He point of view (e.g., see HANYU et al. 1999, 2001).

Neon isotopic data are virtually absent for EM–I basalts and the available analyses are mostly confined to MORB and HIMU basalts. These results indicate at least two components distinct from that of the present atmosphere: a $^{20}\text{Ne}/^{22}\text{Ne}$ -rich solar (primitive) component, and a $^{21}\text{Ne}/^{22}\text{Ne}$ -rich nucleogenic (U- and Th-derived) component (HONDA et al. 1991, 1993, MATSUMOTO et al. 1997, HANYU et al. 2001). The possible origins of the radiogenic/nucleogenic component (^{21}Ne derived from U and Th) is still uncertain, due to similar $^{21}\text{Ne}/^4\text{He}$ production rates ($\sim 1 * 10^{-7}$) in mantle and crust (OZIMA & PODOSEK 2002). Therefore, the $^{21}\text{Ne}/^{22}\text{Ne}$ -rich component cannot be automatically interpreted in terms of recycling of subducted crust (HONDA et al. 1993).

Trace elements constraints

The trace element compositions of EM–I oceanic basalts (e.g., Ba, Nb, Pb, Sr, Eu, Nd) are variable, resulting in a wide range of inter-elemental ratios (e.g., Ba/Nb ranging from 3.5 to 47.4, Ce/Pb from 1.2 to 24.6, Nb/U from 10.5 to 71.8, Sr/Nd from 6.2 to 36.4 and Eu/Eu* from 0.83 to 1.25; Fig. 2). The same inter-elemental variability has been documented also for EM–I-like basalts emplaced on continental settings from Taiwan (CHUNG et al. 1995, SMITH 1998), Urubici (Serra Geral formation, Brazil; PEATE et al. 1999) and Sardinia (Italy; GASPERINI et al. 2000, LUSTRINO et al. 2000, 2002): Ba/Nb = 10.9–35.0, Ce/Pb = 3.3–21.1, Nb/U = 13.4–50.8, Sr/Nd = 8.7–35.0 and Eu/Eu* = 1.01–1.25 (Fig. 2). EM–I basalts show primitive mantle-normalized patterns (Fig. 3) with positive anomalies for barium (Sardinia, Aphanasy-Nikitin Rise, Kerguelen, Urubici), lead (Sardinia, Aphanasy-Nikitin Rise, Kerguelen), and strontium (Sardinia, Kerguelen), together with patterns where these anomalies are virtually absent (Pitcairn). Similarly, troughs at Nb can be envisaged only

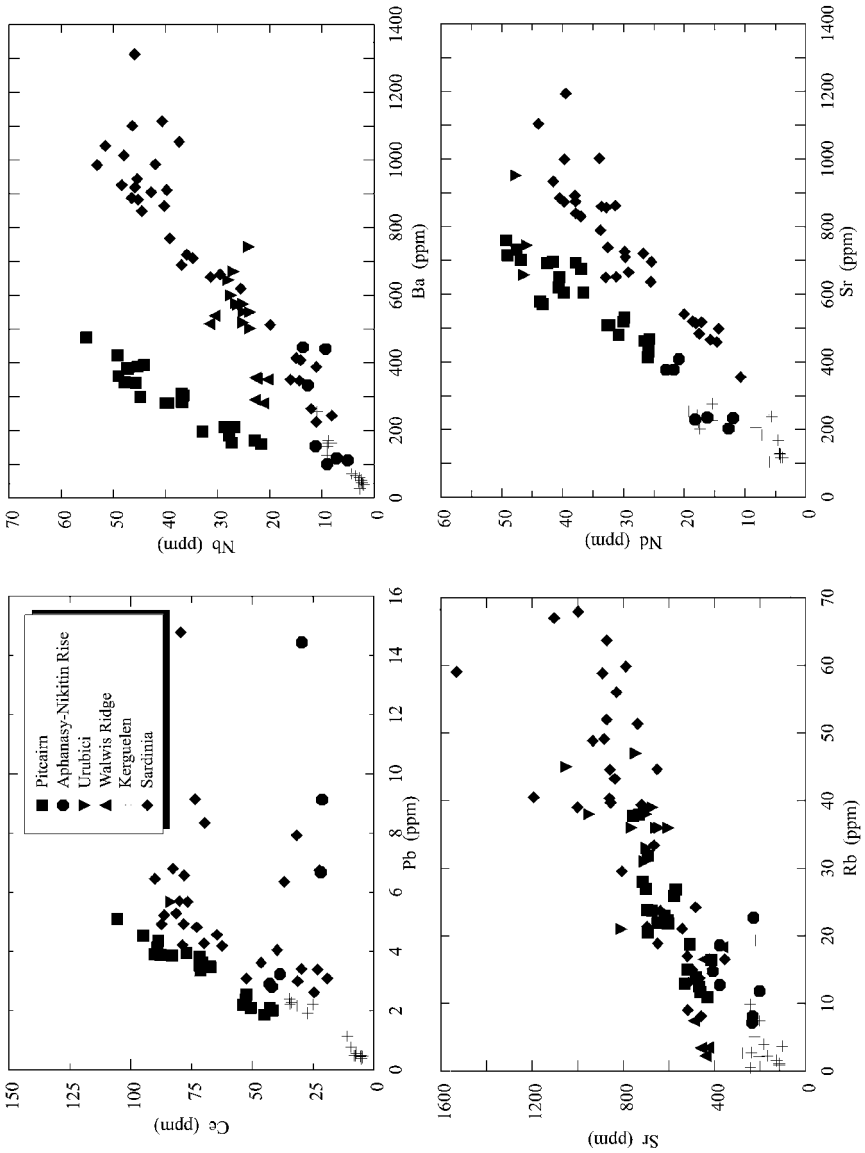


Fig. 2.

in some EM-I suites (e.g., Sardinia, Aphanasy-Nikitin Rise, Urubici; Fig. 3). The involvement of lower continental crust in the genesis of EM-I basalts would change the original trace element ratios because this reservoir has high and variable Ba/Nb (16–224) and low and variable Ce/Pb (2.8–8.5) and Nb/U

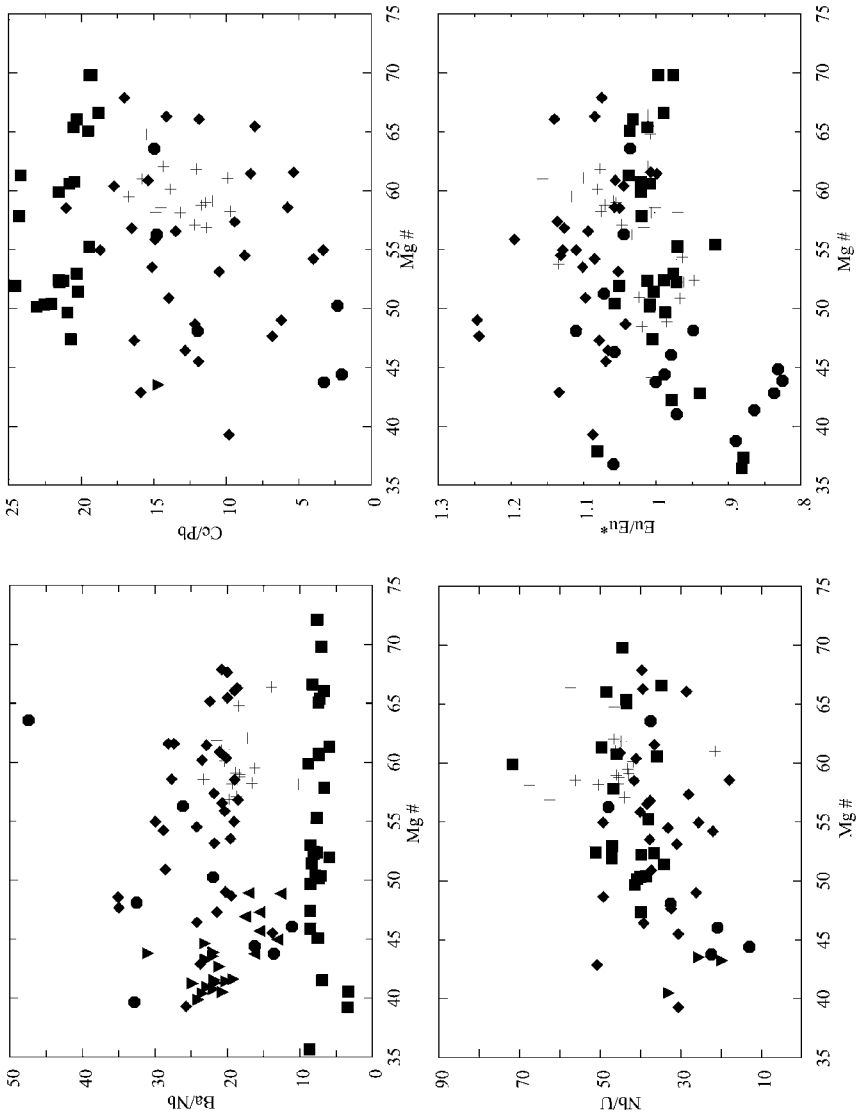


Fig. 2. Interelemental ratios for EM-I basalts from continental and oceanic settings. References as in Fig. 1. Note the great variability of key trace element ratios and the impossibility to propose unique values for the EM-I end member.

(8–25; GAO et al. 1998, LIU et al. 2001, and references therein). Eu/Eu* and Sr/Nd ratios of lower crust are dependent on the amount of cumulate plagioclase in the lower crust (see discussion in FREY et al. 2002). On the basis of the above mentioned trace element data, it is possible to distinguish EM-I

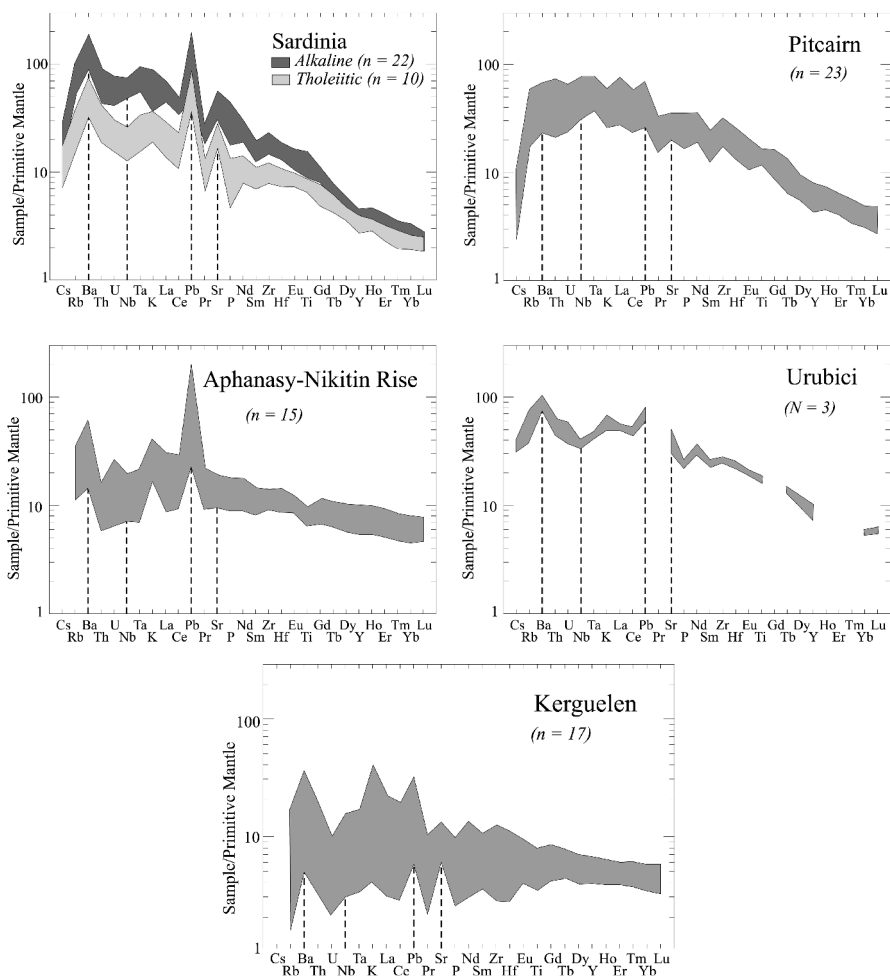


Fig. 3. Primitive mantle-normalized diagrams (normalizing factors after SUN & McDONOUGH 1989) of the most mafic EM-I lavas; references as in Fig. 1. n = number of samples. Sardinia = hawaiite, alkali basalt, basanite, tholeiitic basalt, basaltic andesite (Mg# 0.71–0.46); Pitcairn = tholeiitic and transitional basalt, hawaiite (Mg# 0.72–0.51); Aphanasy-Nikitin Rise = tholeiitic and transitional basalt (Mg# 0.66–0.47); Urubici = tholeiitic basalt and basaltic andesite (Mg# 0.48–0.44); Kerguelen = tholeiitic and transitional basalt (Mg# 0.69–0.60).

from HIMU and DMM reservoirs (WEAVER 1991, CARLSON 1995, DOSTAL et al. 1998). In contrast, trace element content cannot be effectively used to distinguish EM-I from EM-II basalts, because many of their chemical characteristics are shared (e.g., HEMOND et al. 1994, CARLSON 1995).

Concluding remarks

EM–I end-member is one of the most puzzling aspects of modern isotopic topology. The following conclusions arise from the data produced in recent years:

- EM–I basalts are characterized by relatively high Ba, Nb, Pb, Sr, Eu and Nd contents, high but variable Ba/Nb and Sr/Nd, and low and variable Ce/Pb and Nb/U ratios.
- EM–I end-member is likely characterized by relatively low time-integrated Rb/Sr, and low to very low Sm/Nd and U/Pb ratios resulting in time-integrated slightly radiogenic Sr-, unradiogenic Nd-, and unradiogenic Pb-isotopic compositions. These compositions are likely derived from source contamination by crustal rocks with low Rb/Sr, Sm/Nd and U/Pb ratios. Possible mechanisms for EM–I origin include origin from ancient lithosphere, interaction with pelagic–terrigenous sediments, and/or digestion of subducted plagioclase-rich large igneous provinces or lower crustal lithologies.
- The EM–I basalts occur in oceanic and continental settings. Proposed models suggest that the SCLM likely represents the locus where source contamination and/or metasomatism occur due to the effects of lithospheric keel delamination within the asthenospheric depleted mantle.
- The EM–I end-member, although unique, is unlikely to represent a single component. The marked differences in trace element contents and radiogenic isotope compositions of EM–I basalts cast for a heterogeneous reservoir resulting from distinct processes (addiction of crustal material, recycling of oceanic sediments); each of these petrogenetic processes is likely to prevail over the others according to specific geodynamic settings.

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