The OIB paradox

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ABSTRACT

Ocean-island basalt (OIB) and OIB-like basalt are widespread in oceanic and continental settings and, contrary to popular belief, most occur in situations where mantle plumes cannot provide a plausible explanation. They are readily distinguished from normal mid-ocean ridge basalt (N-MORB) through ΔNb, a parameter that expresses the deviation from a reference line (ΔNb=0) separating parallel Icelandic and N-MORB arrays on a logarithmic plot of Nb/Y vs. Zr/Y (Fitton et al., Earth and Planetary Science Letters, v. 153, p. 197–208). Icelandic basalts provide a useful reference set because (1) they are by definition both enriched mid-ocean ridge basalt (E-MORB) and OIB, and (2) they represent a larger range of mantle melt fractions than do intraplate OIBs. Virtually all N-MORB has ΔNb<0 while all Icelandic basalts have ΔNb>0. E-MORB with ΔNb>0 is abundant on other sections of ridge, notably in the South Atlantic and South Indian Oceans. E-MORB and N-MORB from this region form strongly bimodal populations in ΔNb, separated at ΔNb=0, suggesting that mixing between their respective mantle sources is very limited. Most OIBs and basalts from many small seamounts, especially those formed on old lithosphere, also have ΔNb>0. HIMU OIB (OIB with high $^{206}Pb/^{204}Pb$ and therefore a high-μ (U/Pb) source) has higher ΔNb on average than does EM (enriched mantle) OIB, consistent with the presence of recycled continental crust (which has ΔNb<0) in the EM source. Although EM OIBs tend to have the lowest values, most still have ΔNb>0 suggesting that a relatively Nb-rich component (probably subducted ocean crust) is present in all OIB sources. The OIB source components seem to be present on all scales, from small streaks or blobs of enriched material (with positive ΔNb) carried in the upper mantle

convective flow and responsible for small ocean islands, some seamounts, and most E-MORB, to large mantle upwellings (plumes), inferred to be present beneath Hawaii, Iceland, Réunion and Galápagos. It is not possible to identify a point on this continuum at which mantle plumes (if they exist) become involved, and it follows that OIB cannot be a diagnostic feature of plumes. The geochemical similarity of allegedly plume-related OIB and manifestly non-plume OIB is the first part of the OIB paradox. Continental intraplate transitional and alkali basalt in both rift and non-rift (e.g. Cameroon line) settings usually has positive ΔNb and is geochemically indistinguishable from OIB. Continental volcanic rift systems erupt OIB-like basalt irrespective of whether they are apparently plume-driven (e.g. East Africa; Basin and Range), passive (e.g. Scottish Midland Valley) or somewhere between (e.g. North Sea Basin). Magma erupted in passive rifts must have its source in the upper mantle and yet it is always OIB-like. N-MORB-like magma is only erupted when rifting progresses to continental break-up and the onset of seafloor spreading. Continental OIB-like magma is frequently erupted almost continuously in the same place on a moving lithospheric plate for tens of millions of years, suggesting that its source is coupled in some way to the plate, and yet the Cameroon line (where continental and oceanic basalts are geochemically indistinguishable) suggests that the source is sublithospheric. The causes and sources of continental OIB-like magma remain enigmatic and form the second part of the OIB paradox.

INTRODUCTION

Plate tectonics can account for the volume and composition of magmas erupted at plate boundaries (mid-ocean ridges and above subduction zones) but not those erupted within plates. Intraplate magmatism is responsible for ocean islands and some continental volcanoes, and also frequently accompanies continental rifting. The observation that some ocean islands and seamounts form time-progressive chains led to the hypothesis that they are the products of convective plumes originating in the lower mantle (Morgan, 1971). Ocean island basalt (OIB) is chemically and isotopically distinct from basalt (N-MORB) erupted at normal segments of mid-ocean ridges, and this is frequently cited as evidence that the two basalt types originate in different parts of the mantle. The N-MORB source has been depleted (compared to primitive mantle) in those elements that are incompatible in mantle phases, largely through the extraction of the continental crust. By contrast, the OIB source is consistently less depleted in incompatible elements, and the corresponding isotopic differences require around 1-2 Ga to develop (e.g. Hofmann, 1997).

Ocean crust forms at mid-ocean ridges through the passive upwelling of the asthenosphere, and so the composition of N-MORB is taken to represent that of the upper mantle. OIB, on the other hand, is the characteristic basalt type erupted at hotspots and is often assumed to originate in lower mantle brought to the surface in mantle plumes. A hotspot, in this context, is any localised occurrence of anomalous (usually, but not always, intraplate) magmatism not easily explained by plate-tectonic processes. The common assumption that hotspots are synonymous with mantle plumes and that OIB is therefore diagnostic of plumes is clearly questionable (e.g. Anderson and Schramm, 2005). It is not easy, for example, to reconcile the large number of isolated islands and seamounts (all of them hotspots) with a mantle-plume origin. Wessel and Lyons (1997) estimate that there are around 70,000 seamounts with heights >1 km in the Pacific Ocean alone, and most of these are not aligned in chains. A significant proportion of these seamounts are likely to be made of OIB (e.g. Niu and Batiza, 1997). OIB-like basalt is frequently erupted in continental rift systems, and this is also difficult to explain through mantle plumes, especially when rifting appears to be a passive response to plate stresses.

Hypotheses for OIB formation fall into three categories: (1) mantle plumes; (2) dispersed blobs or streaks of incompatible-element-enriched material in the depleted upper mantle; and (3) a layer of shallow mantle (the perisphere; Anderson, 1995) that is enriched in incompatible elements compared to the deeper parts of the upper mantle. In the ***plume*** hypothesis, subducted and dehydrated ocean crust is stored in the deep mantle and eventually returned to the surface in mantle plumes (Hofmann and White, 1982). The core-mantle boundary, the 660 km discontinuity, and discontinuities within the lower mantle have all been suggested as possible storage sites. The resulting mixture of peridotite and eclogite or pyroxenite melts on decompression because it is both hot and fertile. In earlier versions of the plume hypothesis, OIB were thought to be generated from the melting of primitive (undepleted) mantle but the wide range of OIB compositions makes this untenable (see the review of this subject by Hofmann, 1997). The second (***blobs or streaks***) hypothesis also appeals to mantle enrichment through the addition of subducted ocean crust, but here the enriched components are dispersed in a depleted matrix to form the convecting upper mantle. Small degrees of melting will tend to sample the enriched and more easily fusible parts preferentially, leading to the formation of OIB, whereas higher-degree melting under mid-ocean ridges will homogenise the mixture and produce more depleted (N-MORB) magma (e.g. Sleep, 1984; Fitton and James, 1986; Meibom and Anderson, 2003; Ito and Mahoney, 2005). The Anderson (1995) ***perisphere*** hypothesis explains OIB as the product of melting of a global, weak, enriched layer that lies immediately beneath the lithosphere. Seafloor spreading drags this layer aside, allowing deeper and more-depleted upper mantle to rise to the surface under mid-ocean ridges.

The objective of this paper is to address the question of the nature and whereabouts of the OIB source(s) through a survey of the scale and distribution of OIB and OIB-like basalt. To do this, it is necessary to establish reliable quantitative criteria for discriminating between OIB and N-MORB and then to apply these criteria to basalt erupted in oceanic and continental settings. The approach used is that developed by Fitton et al. (1997) to distinguish Icelandic basalt from N-MORB and uses Nb, Zr and Y data.

DATA SOURCES

The usefulness of the discrimination technique applied here is limited by analytical consistency in the data used. Except where otherwise stated, the data used in this study were produced by X-ray fluorescence (XRF) spectrometry in Edinburgh using analytical conditions described in Fitton et al. (1998). The concentration of Nb is particularly critical and this was determined with long count times on samples with Nb contents <5 ppm. Analytical precision in these samples is ±0.1 ppm. Older Edinburgh data were corrected for subsequent changes in the recommended values for international reference standards. Most of the mid-ocean ridge data were taken from the literature but, because of the low concentrations of Nb in N-MORB, only recent (mostly post-1990) data were used, and then only where analytical results for international reference standards were given. Where necessary, the analyses were adjusted to bring them into line with standard values used in Edinburgh. Typical values (in ppm) obtained on four basalt reference standards in the Edinburgh XRF laboratory are given in Table 1.

Table 1. Typical analyses (ppm) of reference standards obtained in the Edinburgh XRF laboratory

Standard	Nb	Zr	Y
BIR-1	0.6	16.2	16.1
BCR-1	13.0	192	38.4
BHVO-1	19.8	175	27.4
BE-N	116	268	30.4

Hitherto unpublished data used in this study are available from the GSA Data Repository (http://www.xxxxxxxxxxxxxxxxxx)

OCEANIC VOLCANISM

Iceland

Iceland sits on the Mid-Atlantic Ridge and, because it is both an ocean island and an anomalous segment of mid-ocean ridge, it provides a link between OIB and MORB. It is widely regarded as the best example of plume-ridge interaction, although Foulger et al. (2005a,b) have proposed a non-plume explanation in which the excess magmatism results from the melting of upper mantle that has been fertilised through the addition of ancient subducted oceanic crust trapped in the Caledonian suture. Variation in Icelandic basalt composition has long been ascribed to mixing between depleted N-MORB and enriched plume-derived OIB magma (e.g. Hanan and Schilling, 1997). Correlated chemical and isotopic variation in Icelandic basalt requires a heterogeneous mantle source with depleted and enriched components

(Hémond et al., 1993; Thirlwall et al., 2004) but there is strong Pb-isotope and trace-element evidence that the depleted component is not derived from the ambient upper mantle but instead forms an intrinsic part of the Iceland plume (Thirlwall, 1995; Fitton et al., 1997, 2003). This being so, Iceland provides a unique area in which to study the nature of the OIB source. Iceland's location allows the mantle source to be sampled through large-degree melting beneath the rift axes and smaller-degree melting off-axis, in contrast to most other ocean islands where only small-degree melts are produced.

Figure 1 (after Fitton et al., 1997) shows the compositional range of Icelandic basalt on a logarithmic plot of Nb/Y *vs*. Zr/Y. The Icelandic data define a linear array, with incompatible-element-depleted basalt at the low-Zr/Y end and enriched transitional to alkaline basalt at the other end. The most enriched basalts are from off-axis locations and from the propagating tip of the eastern rift zone; the rest of the samples are from the actively spreading rift zones, with the most depleted rocks being picrites from the Reykjanes Peninsula and the Northern Rift Zone. Fitton et al. (1997) showed that the Iceland array can be modelled by variable degrees of fractional melting of a heterogeneous mantle source. Mid-ocean ridge basalt sampled at locations away from the influence of hotspots (N-MORB) plots on a parallel array at lower Nb/Zr and clearly cannot be the depleted end-member in the Iceland array. As with the Iceland array, the N-MORB array also reflects degree of melting. The smallest-degree melts (highest Zr/Y) are from the slow-spreading Southwest Indian (Robinson et al., 2001) and Gakkel (Muhe et al., 1997) Ridges. The highest-degree melts (lowest Zr/Y) are from 57.5º–61ºN on the Reykjanes Ridge, south of the Iceland geochemical anomaly, where hotter-than-normal mantle has resulted in oceanic crust 8–10 km thick (Smallwood et al., 1995).

It is clear from Figure 1 that both the depleted and enriched source components of Icelandic basalt must be enriched in Nb compared with the N-MORB source. This is a characteristic feature of OIB, which tends to have positive Nb anomalies on primitive-mantle-normalised diagrams (Hofmann, 1997) and is the reason why relative Nb (and Ta) abundances in basalt have been used to discriminate between ancient tectonic environments (e.g. Meschede, 1986).
The abundance of most elements in the depleted upper mantle (the source of N-MORB) can be modelled adequately by mass-balance calculations in which

average continental crust is subtracted from primitive mantle (Hofmann, 1988). This procedure, however, fails to account for the low abundance of Nb in the upper mantle because both continental crust and N-MORB are depleted in Nb compared with similarly incompatible elements. For example, both the continental crust and N-MORB have lower Nb/La than does primitive mantle (Rudnick et al., 2002; Hofmann, 2004).

Figure 1 shows clearly the relative deficiency in Nb in N-MORB and continental crust; the composition of primitive mantle (PM) lies within the Iceland array, while both N-MORB and average continental crust plot below it. The missing Nb is probably stored in subducted and dehydrated ocean crust, possibly in rutile, and ultimately recycled as the OIB source (Hofmann, 1997; Rudnick et al., 2000). The observation that both N-MORB and average continental crust are deficient in Nb appears to be at odds with the observation that N-MORB is not depleted in Nb compared with U, which has a similar bulk partition coefficient during mantle melting (Hofmann et al., 1986; Hofmann, 2004). Both N-MORB and OIB have Nb/U ≈ 50 (Hofmann, 2004), a value that is significantly higher than the chondritic ratio of 30, while the average continental crust has subchondritic Nb/U (5.6; Barth et al., 2000). The conclusions drawn in the present paper, however, will be based on the empirical observation that Icelandic basalt and N-MORB form distinct arrays on a logarithmic plot of Nb/Y vs. Zr/Y and are therefore unaffected by the Nb/U problem.

The parallel lines on Figure 1 mark the upper and lower limits of the Iceland data array. The lower of these separates the Iceland and N-MORB data and, as will be shown, is a useful reference line when discussing the global distribution of N-MORB and OIB. Fitton et al. (1997) defined a parameter (ΔNb) that expresses the excess or deficiency in Nb relative to this line:

$$\Delta Nb = 1.74 + \log(Nb/Y) - 1.92 \log(Zr/Y)$$

ΔNb is unaffected by degree of melting, which controls the position of a sample along the array but not across it (Fitton et al., 1997). Arrays of data representing variable degrees of melting (such as the Iceland array) would not be linear if this were not so. Position vertically within the array is a function of the relative Nb abundance in the source. ΔNb is also unaffected by fractional crystallisation of olivine and plagioclase since Nb, Zr and Y are highly incompatible in these phases. This is not entirely true of augite, in which Y is moderately, and Zr slightly, compatible. Consequently, ΔNb falls slightly during prolonged fractional crystallisation of augite and so, to eliminate this effect, rock samples with <5 wt.% MgO have been excluded from this study. In the following sections, data from a variety of mid-ocean ridge and intraplate settings are plotted on the Nb/Y-Zr/Y diagram and compared with the Iceland array. Data from Iceland are a useful reference set because Icelandic basalt is generated over a much larger melting range than are OIBs elsewhere.

N-MORB and E-MORB

Iceland, being part of the Mid-Atlantic Ridge, is composed of mid-ocean ridge basalt (MORB), but Icelandic basalt is more enriched in highly incompatible elements than is N-MORB and is also distinct in its radiogenic isotope ratios. Such basalts are described as enriched MORB, or E-MORB. Iceland represents by far the world's most voluminous occurrence of E-MORB, and the excess magmatism there is generally ascribed to plume-ridge interaction. Other examples of anomalous ridge segments composed of E-MORB, though much smaller in volume, tend to be topographically higher than normal segments and are often located close to ocean islands or large seamounts. For this reason they are also often assumed to result from plume-ridge interaction (e.g. Douglass et al., 1999) even though the case for plumes is far less compelling than in the case of Iceland.

There are no universally agreed geochemical criteria in the literature for defining N- and E-MORB. Samples tend to be assigned to one type or the other on an ad hoc basis, and those that are transitional between the two are sometimes described as T-MORB (e.g. Mahoney et al., 1994). The distinction between N- and E-MORB is generally based on enrichment in K and other highly incompatible elements relative to moderately incompatible elements such as Ti or P. Thus Mahoney et al. (1994) define T-MORB as MORBs with Rb/Nd > 0.15, whereas Hall et al. (2006) refer to samples with K_2O/TiO_2 ≥0.1 as E-MORB and comment that this is roughly equivalent to the Mahoney et al. (1994) definition of T-MORB. Niu et al. (2002) use K/Ti > 0.11 to define E-MORB.

Other authors use relative depletion in light rare-earth elements (LREE) to discriminate between N- and E-MORB. Mahoney et al. (2002), for example, use chondrite-normalised La/Sm, $(La/Sm)_n > 0.8$ to define E-MORB but emphasise that a continuum of compositions exists between N- and E-MORB. All of the criteria that have been used suffer from the same inability to distinguish between the effects of source enrichment and degree of melting. A small melt fraction from a depleted (low K/Ti) source could have the same K/Ti as a large melt fraction from an enriched (high K/Ti) source. La/Sm is likewise sensitive to degree of melting though to a lesser extent than is K/Ti. ΔNb is insensitive to degree of melting and, in the following discussion, I show that is a more useful discriminant between N- and E-MORB because it reflects source composition alone.

In order to assess the similarity between Icelandic basalt and much less voluminous E-MORB, I use data from the southern Mid-Atlantic and southwest and central Indian Ridges. These ridges form a continuous spreading centre around the south of Africa and are noted for the abundance of E-MORB. Figure 2 shows the distribution and composition of dredged MORB samples, colour-coded for ΔNb (positive in red, negative in yellow). Samples represented by red points plot with Icelandic basalt (E-MORB) and the yellow points with N-MORB. Note that the samples with positive ΔNb tend to cluster in regions of anomalously shallow ridge, close to ocean islands and large seamounts.

The data points on the Nb/Y-Zr/Y plots in Figure 2 appear to form clusters separated by the reference line (ΔNb=0), and this is confirmed by the histogram of ΔNb for the whole data set. The basalt population is strongly bimodal with positive and negative peaks separated by a trough centred on ΔNb=0. This is an important observation because it suggests that the Iceland reference line is of global significance. Furthermore, E-MORB and N-MORB form discrete populations rather than a continuum, and there appears to be limited mixing between the E-MORB and N-MORB mantle sources. That the E-MORB data plot in the Iceland array (ΔNb>0) suggests that the E-MORB source has an excess of Nb similar that in the source of Icelandic basalt. Recycled subducted ocean crust seems to be involved in both but this need not imply the involvement of mantle plumes in the generation of E-MORB as is often assumed (e.g. Douglass et al., 1999). Passive blobs or streaks of recycled crust carried around in the upper-mantle flow could equally well explain the distribution and composition of E-MORB in Figure 2.

Figure 3 shows ΔNb for the samples used in Figure 2, plotted against (La/Sm)n (Sm/Nd normalised to the ratio in chondrite meteorites or primitive mantle). (La/Sm)n is a measure of relative enrichment or depletion of LREE on chondrite-normalised REE patterns and is often used to discriminate between N- and E-MORB. The data form two distinct clusters, separated at ΔNb ≈ 0 and (La/Sm)n ≈ 0.8. Mahoney et al. (2002) use (La/Sm)n = 0.8 to distinguish between N- and E-MORB on the Southeast Indian Ridge and note that the two types form a continuum. The histogram in Figure 3 shows no bimodality in (La/Sm)n, in contrast to ΔNb, which is strongly bimodal (Figure 2). Thus ΔNb is a much better discriminant between N- and E-MORB than any parameter used previously, probably because it reflects fundamental differences in their mantle sources and is insensitive to degree or depth of melting.

E-MORB tends to occur on elevated ridge segments close to ocean islands or large seamounts, as in the southern Mid-Atlantic and southwest and central Indian Ridges (Figure 2), but this is not always so. The southeast Indian Ridge (SEIR), for example, is remote from ocean islands (Figure 4) and composed mostly of N-MORB, but E-MORB is locally abundant, especially in the eastern part where the water depth is greater (Mahoney et al., 2002; Figure 4). Basalt samples dredged from the SEIR appear to form a continuum on a logarithmic plot of Nb/Y vs. Zr/Y, and the distribution of ΔNb is only weakly bimodal (Figure 4), in contrast to the strongly bimodal distribution seen in data from the southern Mid-Atlantic and southwest and central Indian Ridges (Figure 2). This suggests that mantle is heterogeneous on a smaller scale beneath the SEIR and that the enriched domains are less able to dominate the melt zone. This was also noted by Mahoney et al. (2002) who commented that N- and E-MORB samples were often retrieved in the same dredge haul.

Ocean islands

There is a tendency in the literature to equate ocean islands with mantle plumes. This seems not unreasonable in the case of those islands (e.g. Hawaii and Réunion) that lie on the ends of long, time-progressive chains of islands and seamounts, but is debateable for many if not most islands. The composition of OIB is shown in Figure 5. The OIB data set includes samples from most major islands and island groups and therefore reflects the range of OIB composition. Data from the Galápagos Islands are not included in the data set because, as will be shown later, these islands are unique and not typical OIB. Most of the data plot within and at the high-Zr/Y end of the Iceland array, implying that OIB and Icelandic basalt have similar mantle sources but that the OIB source has generally been melted to a smaller degree than has the mantle under Iceland. This is consistent with Iceland being located on a mid-ocean ridge whereas most ocean islands form over thick lithosphere.

The similarity between OIB and Icelandic basalt is shown in the histograms of ΔNb values in Figure 5. The two populations have very similar distributions but the OIB data extend outside the Iceland array to higher and lower values. Of the 53 samples (out of a total of 768) with negative ΔNb, 24 are from the Hawaiian Islands, seven are from Tahiti, five from Kerguelen, four from the Caroline Islands and four from the Cameroon line. There is no obvious relationship between ΔNb and size of island, nor with whether or not it forms part of a time-progressive chain. The Hawaiian Islands, for example, span the whole range of ΔNb (Figure 6). All of the samples from the Hawaiian Islands that plot around the lower (reference) line (ΔNb = 0) are from the shield-forming stage of their respective islands, while those plotting around the upper line are from the rejuvenated stage, as has been noted by Frey et al. (2005). Figure 6 also shows data from the Emperor Seamounts (from Regelous et al., 2003). These data, representing the composition of the Hawaiian mantle source from 85 to 42 Ma, plot around the lower line along with the shield-forming lavas from the islands, suggesting that this reflects the long-term composition of the Hawaiian plume (Frey et al., 2005).

Zindler and Hart (1986) used radiogenic isotope ratios (Sr, Nd and Pb) to identify three extreme OIB types (HIMU, EM-1 and EM-2) representing mantle reservoirs thought to originate, respectively, through the recycling of subducted ocean crust, lower continental crust, and upper continental crust (or marine sediment). Mixing between these end members and the depleted MORB-source mantle (DMM) can account for the isotopic variation seen in all oceanic basalts. When plotted in $^{87}Sr/^{86}Sr$–$^{143}Nd/^{144}Nd$–$^{206}Pb/^{204}Pb$ space, data from individual islands or island groups form arrays that suggest mixing between the various mantle reservoirs. DMM is rarely at the end of these arrays, however, and they fan out instead from a common or focal zone, which Hart et al. (1992) proposed as a fifth mantle reservoir (FOZO) residing in the lower mantle or at the core-mantle boundary. Stracke et al. (2005) noted that OIBs with the most radiogenic lead isotope ratios fall into two groups; HIMU types and those close to FOZO in isotopic composition. They argued that both the FOZO and HIMU reservoirs represent recycled subducted ocean crust but that the rare HIMU type is not a common mixing end member in other OIB arrays and MORB. Their redefined FOZO, on the other hand, is a ubiquitous component common to both OIB and MORB.

Variation of ΔNb in OIB correlates significantly with the isotopic character of the respective island or island group. Willbold and Stracke (2006) have compiled a set of trace element data representing the three extreme OIB types and have used these, in conjunction with isotope data, to argue that all OIB sources contain subducted ocean crust (the redefined FOZO of Stracke et al., 2005) and that the two EM types contain additional components derived from the continental crust. Figure 7, based on the Wilbold and Stracke (2006) data set, shows histograms of ΔNb in OIB representing the three extreme types. HIMU OIB has generally higher ΔNb and, apart from two samples with extreme values, a more restricted range than EM-1 and EM-2. This is consistent with the HIMU component being subducted and dehydrated ocean crust because this is likely to have excess Nb retained in rutile-bearing eclogite (Rudnick et al., 2000). Recycled continental crust will have negative ΔNb (Figure 1), and addition of this component will therefore lower the ΔNb in OIB and account for the slightly negative values in Figure 5. The positive values of ΔNb in most EM-type OIBs is, however, consistent with the dominance of subducted oceanic crust in their source.

From the histograms in Figure 5 it is clear that ΔNb is an excellent discriminant between N-MORB and OIB; at least as good any other geochemical parameter, including isotope ratios. The near-normal distribution of ΔNb in OIB suggests that little if any mixing takes place between the OIB source and the ambient upper mantle, as was also noted in the case of E-MORB and N-MORB. A similar conclusion was reached by Hart et al. (1992) on the basis of radiogenic isotope ratios in OIB. As was noted earlier, these authors showed that many OIB plot on sublinear arrays in Sr-Nd-Pb isotopic space but the depleted upper mantle is rarely a mixing end-member of these arrays. Only basalts from Iceland and the Galápagos Islands seemed to show mixing with the upper mantle, and both of these sit on or close to mid-ocean ridges. The data in Figure 1 show little, if any, mixing between Icelandic basalt and N-MORB, but what about the Galápagos Islands?

The Galápagos Islands formed through the interaction of a hotspot with the Galápagos spreading centre, currently to the north of the archipelago. Relative motion of the hotspot and spreading centre over the past 20 Ma has resulted in hotspot magmas being erupted preferentially on the Cocos plate to the north and the Nazca plate to the south at different times (Werner et al., 2003). This has resulted in the formation of the Cocos and Carnegie Ridges, respectively north and south of the spreading centre. Because of its long history of magmatism and association with time-progressive aseismic ridges, the Galápagos hotspot is generally interpreted as due to a mantle plume.

Geochemical studies on basalt from the Galápagos Islands (Geist et al., 1988; White et al., 1993) suggest that the plume has a depleted, N-MORB-like, core partly surrounded on the north, west and south by a horseshoe-shaped OIB-like outer zone. Mixing between these two end-member components can account for the chemical and isotopic variation seen in Galápagos basalts. Geist et al. (1988) propose that the depleted core of the plume is ambient upper mantle entrained during ascent. If the depleted basalts are the products of N-MORB-source upper mantle then they should have negative ΔNb. Apparently similar depleted basalt in Iceland has also been interpreted as the result of mixing with ambient upper mantle (Schilling et al., 1982) but these have positive ΔNb (Figure 1) and therefore more likely represent melts from an intrinsic depleted component of the Iceland plume (Fitton et al., 2003).

The suite of Galápagos samples used by White et al. (1993) was analysed for Nb, Zr and Y in Edinburgh and the results plotted in Figure 8(a), along with data from the Cocos and Carnegie Ridges (from Harpp et al., 2005) and East Pacific Rise N-MORB (from Mahoney et al., 1994). The Galápagos samples show a wide scatter extending to negative ΔNb, unlike basalts from Iceland (Figure 1). Figure 8(b) shows that ΔNb correlates negatively with $^{143}Nd/^{144}Nd$, as would be expected from mixing between the Galápagos plume and ambient upper (N-MORB-source) mantle represented by the East Pacific Rise data. Interestingly, the Cocos and Carnegie Ridge data plot almost entirely within the Iceland array on Figure 8(a), suggesting that mixing between plume and upper mantle is a recent phenomenon. Figure 9(a) shows the location of samples on the Cocos and Carnegie Ridges, colour coded for ΔNb. All the samples dredged from the ridges have ΔNb>0 (as in Iceland); the few sample with ΔNb<0 were dredged from the seafloor adjacent to the ridges. Figure 9(b) is an enlargement of part of Figure 9(a) showing ΔNb for each of the Galápagos Islands. The zonation noted by Geist et al. (1988) is clearly reflected in ΔNb, with an N-MORB-like central area surrounded to the north, west and south by a zone with ΔNb>0.

The Galápagos Islands seem to be unique among ocean islands in the involvement of depleted upper mantle in their formation. No other island or island group contains basalt with such strongly negative ΔNb (Figure 5; Galápagos data were not included in the OIB compilation). Basalt from the shield-forming stage on the Hawaiian Islands has ΔNb≈0±0.1 (Figure 6), but isotopic data show that these low values cannot be due to the involvement of the N-MORB source. The rejuvenated-stage basalts (with strongly positive ΔNb) have radiogenic-isotope ratios closer to those in Pacific N-MORB than do the shield-forming basalts (Frey et al., 2005). Blichert-Toft et al. (1999) have shown that Hf-isotope ratios in Hawaiian shield-forming basalts identify a recycled pelagic sediment component in their mantle source, and this is consistent with low ΔNb in these basalts (Frey et al., 2005).

Seamounts

The largest ocean islands, such as Iceland, Hawaii, Réunion and the Galápagos Islands are associated with time-progressive aseismic ridges or seamount chains that often extend back in time to link with large igneous provinces. A mantle-plume origin provides the most plausible explanation for such islands. At the other end of the scale are small islands and seamounts with no obvious connection to mantle plumes. Figure 5 includes many small islands, and these share with the larger islands the OIB characteristic of positive ΔNb, but no seamount data are included in the compilation.

A geochemical study carried out by Niu and Batiza (1997) on samples dredged from small seamounts on the flanks of the East Pacific Rise (EPR) provides ideal data for the present survey of OIB distribution. Niu and Batiza (1997) showed that the seamounts have an extraordinary range in composition from highly depleted N-MORB to enriched alkali basalt and conclude that their mantle source must be heterogeneous on a small scale. Niu and Batiza (1977) and Niu et al. (2002) argue that that the region must be underlain by a two-component mantle with an enriched and easily melted component dispersed as physically distinct domains in a more depleted and refractory matrix. The enriched component, they suggest, is recycled subducted ocean lithosphere (crust and metasomatically enriched lower lithospheric mantle).

Figure 10 shows that seamount data from Niu and Batiza (1997) plot in distinct N-MORB and OIB (or E-MORB) groups separated almost perfectly by the ΔNb=0 reference line and with a weakly bimodal distribution of ΔNb. About 20% of the data plot in the OIB group. The accompanying map shows that OIB-type seamounts are interspersed with N-MORB-type seamounts and show no systematic geographical distribution. Niu and Batiza (1997) note that some individual seamounts contain both OIB- and N-MORB-type basalt (Figure 10) and therefore require a mantle source that is heterogeneous on a scale of several hundred metres.

The northern EPR seamounts studied by Niu and Batiza (1997) were erupted on very young ocean crust, close to the ridge axis, and are therefore part of the ridge itself. Many (possibly most) seamounts, however, formed farther from ridge axes, and data from examples of these are provided by the Hall et al. (2006) study of the Rano Rahi seamount field off the southern EPR. The distribution and composition of these seamounts and of samples dredged from the spreading axes in the region are shown in Figure 11. It is clear that N-MORB dominates both the axes and the seamounts. Mahoney et al. (1994)

identify all but six of the southern EPR samples as N-MORB on the basis of their definition of T-MORB (Rb/Nd > 0.15), but only one sample has ΔNb>0 (Figure 11). Of the eight samples with ΔNb>0 from the spreading axes around the Easter Microplate, seven are from segments adjacent to Easter Island. Four of the 30 samples from the Rano Rahi seamounts have ΔNb>0, and ^{40}Ar-^{39}Ar dating of two of these (one of the cluster of three in the north and the sample in the south) shows that these two seamounts were erupted onto older oceanic crust (3 and 2.5 Ma respectively) than all the other dated seamounts (Hall et al., 2006). The other two samples with ΔNb>0 have not been dated. This relationship between ΔNb and lithosphere age is not apparent in the Pukapuka Ridges; the one sample with ΔNb>0 (Figure 11) was erupted onto younger crust than were the other seamounts.

It is clear from the histograms in Figure 11 that the spread of ΔNb is much larger in the seamount data than in samples from the spreading axes. This is probably because individual seamounts will only sample their underlying mantle whereas axial magma reservoirs will accumulate magma from a larger volume of mantle and then homogenise it, as was suggested by Niu et al. (2002) for the northern EPR seamounts and adjacent ridge axis. The high values and strongly bimodal distribution of ΔNb seen in the southern Mid-Atlantic and southwest and central Indian Ridges (Figure 2) are not seen in the data in Figure 11. This suggests that the enriched domains in the mantle beneath the southern EPR and adjacent areas are smaller and less abundant than they are in the mantle beneath the south Atlantic and Indian Oceans, and are therefore unable to dominate the melt zones beneath ridge axes and seamounts.

The Rano Rahi seamounts were erupted onto crust as old as 3 Ma (Hall et al., 2006) but seamounts elsewhere in the Pacific Ocean were erupted onto much older crust. Seamounts off the California coast, for example, have been shown to be 7-11 Ma younger than the underlying ocean crust (Davis et al., 2002), and volcanoes dated 1 to 8 Ma have been found on 135-Ma ocean crust in the northwest Pacific Ocean off Japan (Hirano et al., 2001, 2006). Alkali basalt is the dominant rock type in both of these examples, and in both cases the magmatism has been ascribed to extension or fracturing of the lithosphere. They have been described respectively as a “different” and a “new” kind of intraplate magmatism, and used to argue that oceanic intraplate magmatism need not be related to mantle plumes. McNutt (2006) has used the Japanese example to argue for a re-examination of the plume hypothesis. Data from these two occurrences plot within the Iceland array on a Nb/Y vs. Zr/Y diagram (Figure12) and are compositionally indistinguishable from OIB. This is an important observation because it shows that very small-degree melting of the upper mantle, as would be expected beneath thick lithosphere, produces OIB-like magma (ΔNb>0) and is, therefore, sampling only the enriched mantle domains and not the depleted matrix (ΔNb<0).

CONTINENTAL VOLCANISM

Comparing continental and oceanic basalt is complicated by the effects of contamination with the continental crust, which could have the effect of blurring the geochemical distinction between N-MORB-like and OIB-like magmas. However, the use of the Nb/Y-Zr/Y plot avoids this problem because both N-MORB and average continental crust have negative ΔNb (Figure 1). Contamination of N-MORB-like magma with continental crust can never produce hybrid magmas with positive values of ΔNb, although OIB-like magma could, if it were contaminated with enough continental crust, have ΔNb < 0. The contaminated basalt would, however, be readily distinguished from N-MORB through its high concentrations of most highly incompatible elements. Another useful feature of the Nb/Y-Zr/Y plot is that Nb, Zr and Y are high field-strength elements and therefore relatively immobile during weathering, alteration and low-grade metamorphism.

The Cameroon line

The Cameroon line in West Africa (Fitton, 1987; Déruelle et al., 1991) provides a valuable and unique link between ocean-island and continental alkaline magmatism. It consists of a Y-shaped chain of volcanoes extending for 1600 km from the Atlantic island of Annobon to the interior of the African continent, and has been essentially continuously active since the late Cretaceous. Magmatic activity from 66 to ~30 Ma is represented by plutonic complexes, and the more recent activity by volcanic edifices. Cameroon line magmatism has been essentially alkaline throughout its history. The plutons are composed of gabbro, syenite and alkali granite, and the volcanic rocks range in composition from alkali basalt to trachyte and alkali rhyolite. There is no consistent age progression among the volcanic and plutonic centres, and the most recent volcanic activity (Mt Cameroon in 2000) was in the middle of the line. Mt Cameroon is composed mostly of alkali basalt and basanite and, at 4070 m, is one of the largest volcanoes in Africa.

The origin of the Cameroon line remains unexplained (see Déruelle et al., 1991 for a review of proposed origins). There is no evidence for extensional faulting on either the oceanic or continental sectors. Its longevity and large volume of magma suggest a plume origin, but the plume would need to have a sheet-like form and have remained stationary with respect to the African plate for the last 66 Ma. The importance of the Cameroon line is that it allows a lithospheric source for the continental alkaline magma to be ruled out. Fitton and Dunlop (1985) showed that basalt from the continental sector is identical to that in the oceanic sector despite the two sectors being formed on lithosphere that differs considerably in age and geological history. Whatever the cause of Cameroon line magmatism, its source must be sublithospheric. Data from Cameroon line basalts are plotted in Figure 13, which shows that the oceanic and continental populations are indistinguishable and that the continental basalts are therefore OIB-like (positive ΔNb).

Rift systems

Continental extension and rifting is often accompanied by the eruption of alkali basalt that is generally OIB-like in composition. This is illustrated in Figure 14 which shows data from four rift systems. In two of these (East Africa and Western USA) rifting has been accompanied by around 1 kilometre of uplift, but the other two rifts remained close to sea level throughout their active phase. The East African rifts and the Western USA both need some means of dynamic support to account for their elevation, and mantle plumes provide a plausible mechanism. The volume of primary basaltic magma needed to account for the observed and inferred volume of basic and evolved volcanic rocks in the Kenya

Rift is far too high to be accounted for solely by decompression melting accompanying extension (Latin et al., 1993). Some means of actively feeding mantle into the melt zone is needed, and this observation is consistent with a plume origin.

Despite their tectonic differences, basalts from the four rift systems are remarkably similar and OIB-like (Figure 14). The only basalts with significantly negative ∆Nb are from the early (16-5 Ma) phase of magmatism in the Basin and Range province of the Western USA. Fitton et al. (1991) and Kempton et al. (1991) have shown that these basalts probably inherited their chemical and isotopic characteristics from a subcontinental lithospheric mantle source enriched by fluids expelled from a subducted slab. The enriched lithospheric mantle component is best represented by the lamproites emplaced in several parts of the Western USA shortly before the onset of Basin and Range magmatism. The composition of some of these is shown in Figure 14. The sudden onset of OIB-like Basin and Range magmatism at ~5 Ma coincided with the beginning of the recent phase of uplift on the Colorado Plateau (Lucchitta, 1979).

Africa has many examples of long-lived alkaline magmatism (e.g. Cameroon line; East African rifts) and this may be due in part to the African plate being virtually stationary in the hotspot reference frame since ~35 Ma (Burke, 1996). Ebinger and Sleep (1998) have suggested that a single mantle plume centred under Afar could be responsible for prolonged magmatism in East Africa and the Cameroon line. A mantle plume, however, cannot explain Permo-Carboniferous rift magmatism in the Scottish Midland Valley, which lasted for ~70 Ma but stayed close to sea level throughout (Read et al., 2002). Basalt erupted over this time interval is OIB-like in its incompatible-element abundances and Sr-Nd-Pb-isotope ratios (Smedley, 1986, 1988; Wallis, 1989; Upton et al., 2004), and also has generally positive ∆Nb (Figure 14). Unlike the African rifts, which formed on a plate that was stationary with respect to hotspots, the Scottish Midland Valley moved ~15º northwards across the Equator during the 70 Ma when it was volcanically active (Smith et al., 1981; Lawver et al., 2002; Figure 15). The ~15º northward motion of Scotland is inferred mostly from paleomagnetic evidence (e.g. Irving, 1977) and must be a minimum estimate since this takes no account of east-west motion.

In contrast to the Scottish Midland Valley, magmatism in the North Sea rifts may not have been entirely a passive response to extension. Late Jurassic to Early Cretaceous extension and volcanism was immediately preceded by a phase of domal uplift which interrupted an earlier (Permian to Early Jurassic) phase of basin subsidence (Underhill and Partington, 1993). These authors propose that this ~1000-km-diameter dome was caused by a “warm, diffuse and transient plume head”. If this uplift really was caused by a mantle plume, then it must have had the form of a buoyant blob of mantle, quite unlike the long-lived features thought to be responsible for Hawaii, Réunion and Iceland. This raises the question of why the transient “plume” should have been located so precisely under the centre of a developing rift basin. Whatever its cause, subsidence of the dome was accompanied by Late Jurassic to earliest Cretaceous extension leading to the formation of three rift valleys radiating from the dome centre (Underhill and Partington, 1993).

The rift-to-drift transition

Rifting is a precursor to continent break-up and the formation of new ocean basins, and at some point in this process OIB-like magmatism, if present, must give way to the eruption of N-MORB. This transition is difficult to investigate because the volcanic rocks involved will be under water and buried beneath thick piles of continental-margin sediments. Several Deep Sea Drilling Project (DSDP) and Ocean Drilling Program (ODP) legs have been devoted to investigating the continent-ocean transition. Two of these (DSDP Leg 80 on Goban Spur in the NE Atlantic Ocean, and ODP Leg 210 on the Newfoundland margin) have recovered samples of the earliest ocean-floor basalts. The location of the drill sites and the composition of the basalt samples recovered are shown in Figure 16. At both locations, basalt erupted at the point of continental separation is clearly N-MORB.

DISCUSSION

The data presented in the previous sections show that Icelandic basalt invariably has more Nb, relative to Zr and Y, than does N-MORB. A reference line separating Icelandic basalt from N-MORB on a plot of Nb/Y vs. Zr/Y

(ΔNb=0) is also an effective discriminant between OIB in general and N-MORB (Figure 5). Basalt with positive ΔNb (OIB) occurs in the ocean basins on all scales, from large islands with time-progressive hotspot trails (e.g. Iceland, Hawaii, Réunion) to tiny seamounts (Figures 10, 11 and 12). The former may be the product of deep-mantle plumes but the latter clearly are not, and yet they seem to have very similar mantle sources. A clear bimodality in ΔNb in basalt from sections of mid-ocean ridge affected by hotspots (Figure 2) suggests only limited mixing between the OIB and N-MORB sources. This appears to be a general feature of OIB (Figure 5) and is supported by isotopic studies (e.g. Hart et al., 1992; Blichert-Toft et al., 1999). Only on the Galápagos Islands (Figure 8) do melts from the OIB and N-MORB sources appear to mix to a significant extent.

The OIB source contains enriched components (HIMU, EM-1 and EM-2) mixed in variable proportions. Additionally, the source of Icelandic basalt contains at last one depleted component. Variable degrees of melting beneath Iceland effectively samples both the enriched and depleted components, and therefore the mantle must be heterogeneous on a scale that is smaller than the melt zone. It is clear from Figure 1 that the depleted component in the source of Icelandic basalt, with positive ΔNb, cannot be the ambient upper mantle, which has negative ΔNb. Frey et al. (2005) showed that the source of Hawaiian basalt contains a similar depleted component. Storage of magma in large reservoirs tends to homogenise magma so that large basaltic flows in Iceland have a fairly uniform composition representing a blend of the depleted and enriched parts of the source (Hardarson and Fitton, 1997). The magma composition will be biased towards the more enriched and fusible components.

The upper mantle source of N-MORB must likewise be heterogeneous, as can be seen from the composition and distribution of near-axis seamounts on the northern East Pacific Rise (Figure 10). Most of the seamounts have N-MORB composition but a few are made of OIB-like basalt, suggesting that the upper mantle contains blobs of OIB-source mantle (positive ΔNb) in a depleted matrix with negative ΔNb. Small-degree melting of the upper mantle beneath old, thick oceanic lithosphere only samples the enriched blobs and not the depleted matrix (Figure 12), consistent with the former being more easily fusible than the latter. The OIB and N-MORB sources, therefore, appear to share a common set of enriched components but, in the case of Iceland and Hawaii, differ in their depleted component.

Enriched components (recycled subducted ocean crust with or without material derived from the continental crust) must be present in the convecting upper mantle and also in the source of deeper-mantle upwellings (plumes) responsible for the larger islands. Ocean islands and seamounts thought to result from mantle plumes and those formed by the melting of fusible blobs carried passively in the upper-mantle flow are indistinguishable on trace element (including ΔNb) or isotopic criteria. E-MORB from segments of the South Atlantic and Indian Ocean Ridges (Figure 2), for example, is at least as likely to result from enriched blobs in the upper mantle as from mantle plumes. Moreover, E-MORB in many other areas (e.g. the eastern Southeast Indian Ridge, Figure 4), thousands of kilometres from the nearest hotspots, is clearly not formed by mantle plumes.

The depleted component in the OIB source is less accessible than the enriched components because it is likely to be more refractory and therefore only sampled at larger degrees of melting than are represented by most OIB. It is not surprising, therefore, that the clearest evidence for its existence is provided by basalt from Iceland (Fitton et al., 1997, 2003) and Hawaii (Frey et al., 2005). Subducted oceanic lithospheric mantle provides a plausible source for the depleted component in the Iceland plume (Skovgaard et al., 2001). The upper mantle depleted component is the principal source of N-MORB and formed through the formation of the continental crust and, crucially for the development of negative ΔNb, the removal of dehydrated subducted ocean crust.

Continental intraplate magmatism is much more difficult to explain. Large, long-lived continental alkaline volcanic provinces can be explained by mantle plumes only in cases where they form on stationary continents such as Africa. A mantle-plume origin is arguably the only plausible explanation for magmatism in the East African rift system, given the volumes of magma produced and the scale of regional uplift (e.g. Latin et al., 1993; Ebinger and Sleep, 1998; Macdonald et al., 2001). Whether the same is true for other African alkaline provinces, however, is debateable. The Cameroon line, for example, has been active for 66 Ma without any systematic shift in the locus of activity. The magma source seems to have been fixed very precisely to the African plate over this period.

Passive extension of continental lithosphere leading to rift formation and decompression melting of the upper mantle might be expected to produce at least some magma with N-MORB composition, but it never does. Contamination of N-MORB-like parental magma with continental crust cannot produce OIB-like hybrids because the contaminated magma would have negative ΔNb (Figure 1) whereas rift basalts have positive ΔNb (Figure 14). The geochemical similarity of continental rift basalts and OIB, most strikingly demonstrated by the Cameroon line (Figure 13), also rules out a source in the continental lithospheric mantle (Fitton and Dunlop, 1985). The source of most rift-zone magmatism must, therefore, be sublithospheric and, in cases of passive rifting, must originate in the upper mantle.

Basaltic magmatism in the Mesozoic North Sea basin was accompanied by doming, which Underhill and Partington (1993) ascribe to a blob-like, transient mantle plume head. A mantle plume could account for the OIB-like composition of the basalts (Figure 14) but it seems an unlikely coincidence for a plume originating deep in the mantle to arrive at the surface under a developing rift basin. Late Jurassic to earliest Cretaceous extension in the North Sea basin resulted in lithospheric thinning by a factor of 1.6 at most (Barton and Wood, 1984) and this is insufficient to cause melting in peridotite with an upper mantle potential temperature of 1300ºC (Latin et al., 1990). Unless the mantle was significantly hotter, passive upwelling would allow only the more fusible (i.e. enriched) parts of the upper mantle to melt, and these, by analogy with seamounts, would melt to produce OIB-like magma. It would take much larger degrees of extension, decompression and melting for the depleted matrix to begin to melt and thereby produce N-MORB-like magmas, as clearly happens at the initiation of seafloor spreading (Figure 16). This may explain the complete absence of N-MORB-like magma from passive continental rifts. Thus basalt erupted in passive rifts, and on most ocean islands and some seamounts may share a common source in small-scale blobs or streaks of enriched material in the depleted upper mantle.

Magmatism in the North Sea Basin might fit a simple model of lithospheric extension leading to melting of enriched components in the upper mantle because magma volumes are small and localised, and melting was associated with extension (Latin et al., 1990). Other rift basins are not so simple. The Permo-Carboniferous Midland Valley of Scotland (Figure 14) is part of a much larger rift system that extends across the North Sea into north Germany and northwards into the Oslo Rift. OIB-like magmatism in the Midland Valley was more voluminous and widespread than in the Mesozoic North Sea Basin and persisted almost continuously for about 70 Ma in a belt across central Scotland (Upton et al., 2004). The volume of igneous rocks (~6000 km^3; Tomkieff, 1937) and duration of magmatism cannot be reconciled with simple stretching and yet there is no evidence for mantle-plume activity. The Midland Valley area remained close to sea level throughout its magmatic history (Read et al., 2002) and, over this period, drifted at least 1700 km northwards (Fig. 15). The magma source appears to be fixed to the lithosphere, as in the case of the Cameroon line, and its origin is equally enigmatic.

The selective sampling of enriched components in the upper mantle during extension of the lithosphere is clearly not a viable explanation for magmatism in all passive continental rifts. Nor can it explain magmatism in the Cameroon line. Non-plume explanations for OIB-like magmatism in the Cameroon line and in some passive rifts require a fortuitous concentration of enriched blobs or streaks in the underlying upper mantle and also some mechanism to keep them supplied and melting for very long periods. This problem is circumvented in the Anderson (1995) perisphere hypothesis which explains OIB as the product of melting of a global, weak, enriched layer that lies immediately beneath the lithosphere. Such a layer provides a potentially inexhaustible supply of OIB-like magma wherever the lithosphere is ruptured. The hypothesis predicts that OIB-like magmatism should persist through the continental rifting stage and into the early stages of seafloor spreading until the perisphere was locally exhausted under the new ocean basin. Continental rifting alone would not be able to exhaust the perisphere. A test of the hypothesis is provided by the composition of basalt from the continent-ocean transition on passive margins. That this is clearly N-MORB (Figure 16) seriously undermines the perisphere hypothesis.

CONCLUSIONS

OIB (*sensu stricto*) occurrences can be explained satisfactorily by a combination of mantle plumes and passive enriched streaks and blobs in the convecting upper mantle. Plumes currently provide the best explanation for large islands with time-progressive ridges and seamount chains (e.g. Hawaii, Réunion, Iceland, Galápagos) but not smaller islands and isolated seamounts. E-MORB is formed from OIB-type mantle when a spreading centre encounters a mantle plume (as in Iceland) or a passive enriched blob in the upper mantle. The so-called “Shona plume” (Figure 2) is probably an example of the latter. E-MORB and N-MORB form bimodal populations (Figure 2) suggesting only limited mixing between enriched and depleted components in the upper mantle. A continuum exists between small seamounts and large islands, and geochemistry seems unable to distinguish between OIB formed from passive blobs and that formed from mantle plumes. The enriched components appear to be the same in both. It is not possible to identify the point on this continuum at which mantle plumes (if they exist) become involved, and it follows that OIB cannot be a diagnostic feature of plumes. The geochemical similarity of allegedly plume-related OIB and manifestly non-plume OIB is the first part of the OIB paradox.

Continental intraplate transitional and alkali basalt is identical in composition to OIB but its origin is even more enigmatic than is the origin of true OIB. It is frequently erupted on the same small area of crust over tens of millions of years on moving plates. Its source appears to move with the plates but is sublithospheric. Passive rift systems, if volcanic, always tap the OIB source and only erupt N-MORB following continental break-up and the onset of seafloor spreading. Some OIB and some continental OIB-like basalt appear to share a common plume origin but this cannot be invoked in the majority of continental rifts, any more than it can be invoked in the majority of ocean islands and seamounts. Why do passive continental rifts never erupt N-MORB? The coupled problems of location of the sub-continental OIB source and the mechanisms by which continental OIB-like magmas are produced remain unresolved. None of the proposed models for OIB generation (***plumes***, enriched ***blobs*** in the upper mantle, or an enriched ***perisphere***) is able, either singly or in combination, to explain the global distribution of OIB and OIB-like basalt. This is the second part of the OIB paradox. OIB and OIB-like basalt are widespread in oceanic and continental settings and, contrary to popular belief, most occurrences are in situations where mantle plumes cannot provide a plausible explanation.

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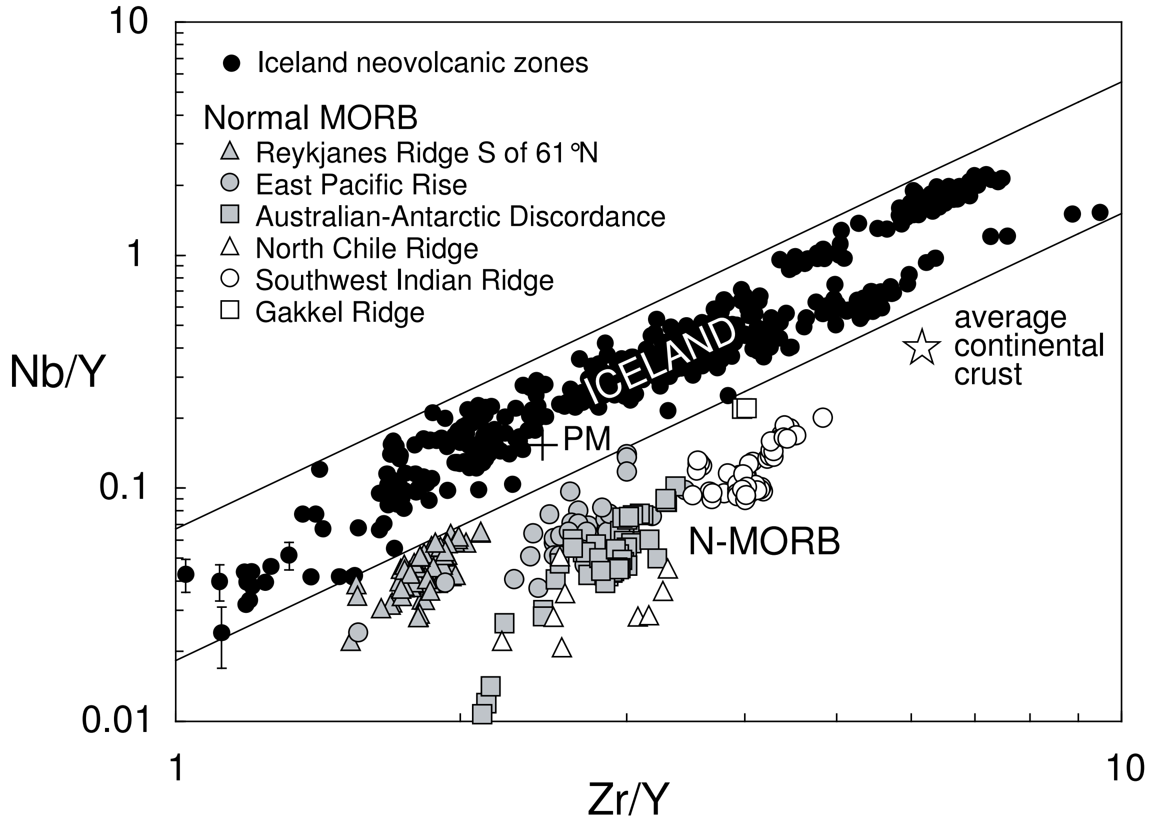

Figure 1. Nb/Y and Zr/Y variation for Icelandic basalt (MgO >5 wt.%) and N-MORB (after Fitton et al., 1997, 2003). Data sources for N-MORB are given in Fitton et al. (1997) with additional data from the Australian-Antarctic Discordance (Kempton et al., 2002), North Chile Ridge (Bach et al., 1996), and Gakkel Ridge (Muhe et al., 1997). The + symbol represents an estimate of primitive mantle composition (McDonough and Sun, 1995). The average crust composition is from Rudnick and Fountain (1995) and Barth et al. (2000). The Southwest Indian Ridge data can be found in GSA Data Repository Item XXXXX.

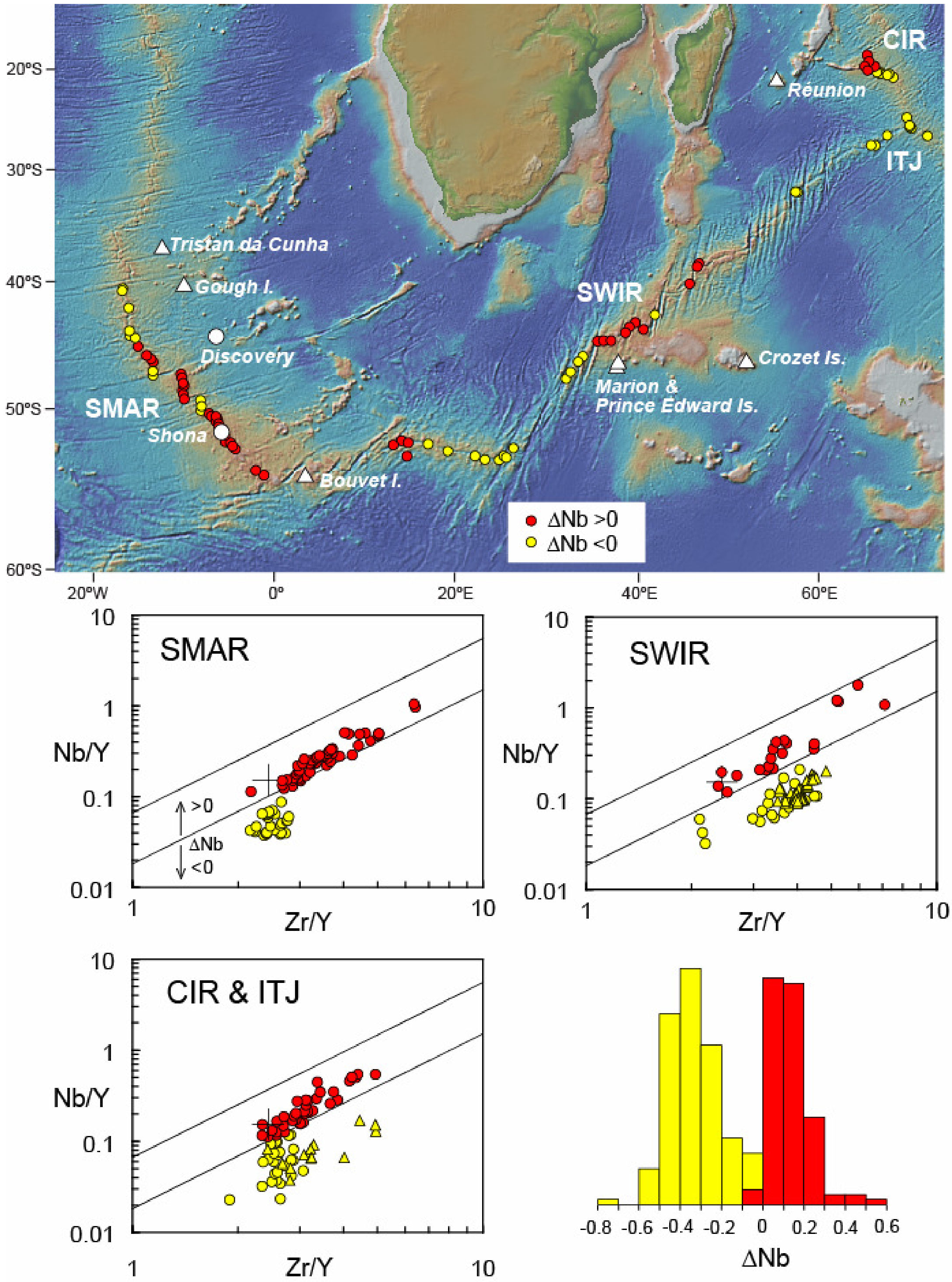

Figure 2. Map of the South Atlantic and South Indian Ocean showing the location of mid-ocean ridge basalt samples for which reliable Nb, Zr and Y data are available. These data are shown on the three plots below the map. SMAR = south Mid-Atlantic Ridge; data from le Roex et al. 2002). SWIR = Southwest Indian Ridge; data from Janney et al., (2005) (circles) and unpublished data from 57ºE and 66ºE (triangles; GSA Data Repository Item XXXXX). CIR (circles) = Central Indian Ridge; data from Murton et al. (2005) and Nauret et al. (2006). ITJ (triangles) = Indian Ocean triple junction; data from Price et al. (1986) and Chauvel and Blichert-Toft (2001). Parallel lines on the plots mark the limits of the Iceland array in Fig. 1; the + symbol represents primitive mantle. ΔNb is the deviation, in log units, from the reference line separating the Iceland array from N-MORB (the lower of the two parallel lines). The points on the map and the data points are colour coded for positive (red) and negative (yellow) values of ΔNb. Three SMAR samples with slightly negative ΔNb (≥-0.02) are included with the positive group because they are clearly different from the negative samples. The histogram shows the distribution of ΔNb in the whole data set (254 analyses). Note that the samples with ΔNb>0 are located on anomalously shallow segments of ridge close to ocean islands or large seamounts. The location of the Discovery and Shona “plumes” is from Douglass et al. (1999).

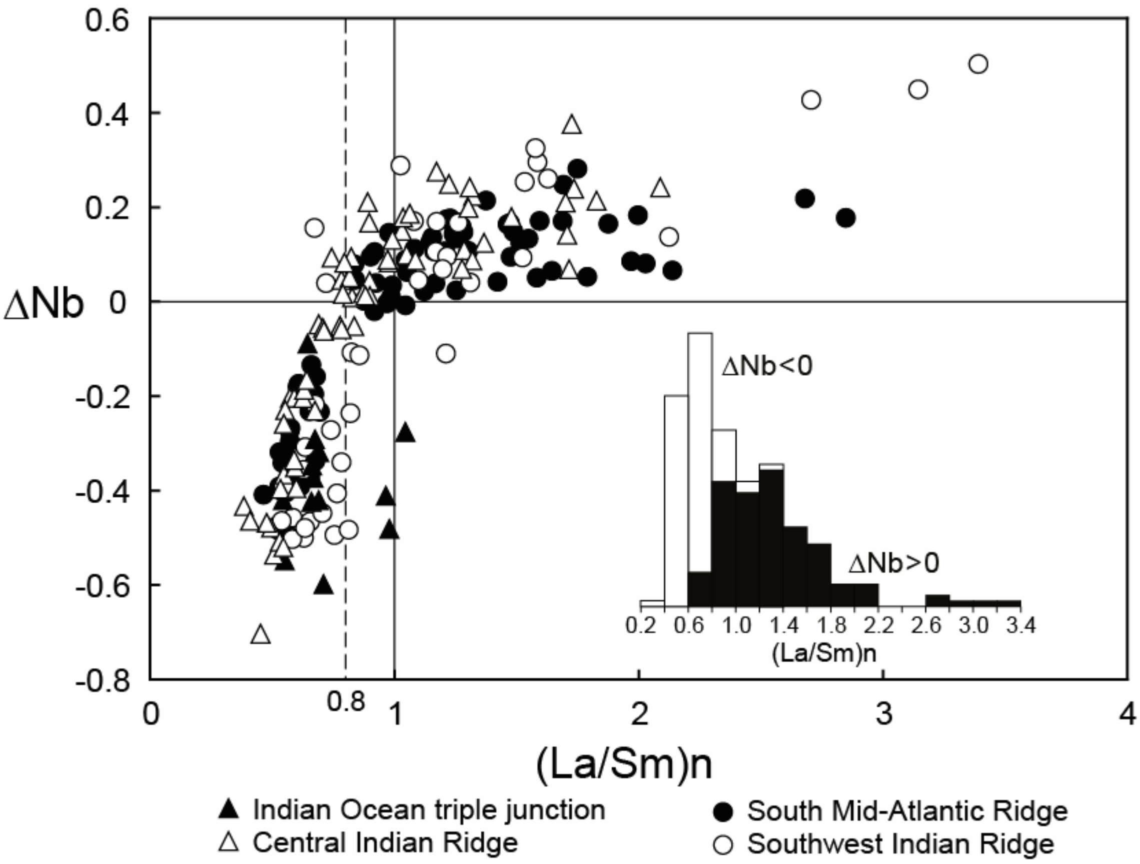

Figure 3. ΔNb plotted against (La/Sm)n for the same basalt samples used Figure 2 (data sources as in Figure 2). (La/Sm)n is primitive-mantle-normalised La/Sm (normalising values from McDonough and Sun, 1995). The vertical line at (La/Sm)n = 1 separates LREE-depleted samples (<1) from LREE-enriched samples (>1); (La/Sm)n = 0.8 (broken vertical line) is the value used by Mahoney et al. (2002) to separate N- and E-MORB on the southeast Indian Ridge. Note that the data form two distinct clusters separated by ΔNb ≈ 0 and (La/Sm)n ≈ 0.8. The inset histogram shows the distribution of (La/Sm)n in the data set (202 analyses); samples with ΔNb<0 (white) and ΔNb>0 (black). (La/Sm)n does not separate the two clusters as effectively as does ΔNb (Figure 2).

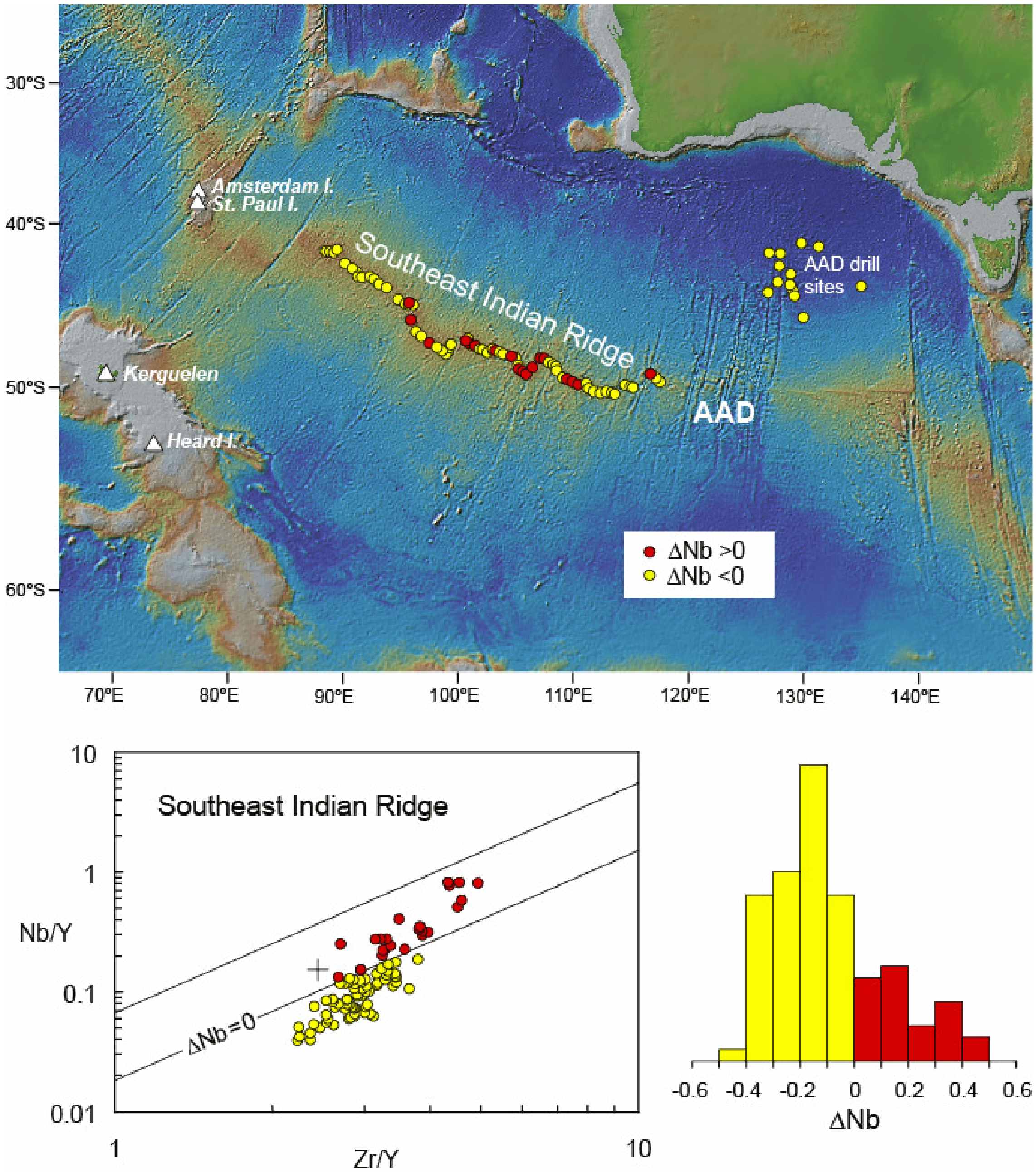

Figure 4. Map of the southeast Indian Ridge showing the distribution and composition of dredged samples. Nb data from Mahoney et al. (2002); Zr and Y data from D.M. Christie (unpublished) All the data are available in GSA Data Repository Item XXXXX. The location and composition (Kempton et al., 2002) of samples from the Australian-Antarctic Discordance (AAD) drill sites (Figure 1) are also shown but these data are not included in Figure 4. Parallel lines on the plots mark the limits of the Iceland array in Fig. 1; the + symbol represents primitive mantle. ΔNb is the deviation, in log units, from the reference line separating the Iceland array from N-MORB (the lower of the two parallel lines). The points on the map and the data points are colour coded for positive (red) and negative (yellow) values of ΔNb. The histogram shows the distribution of ΔNb in the data set (95 samples).

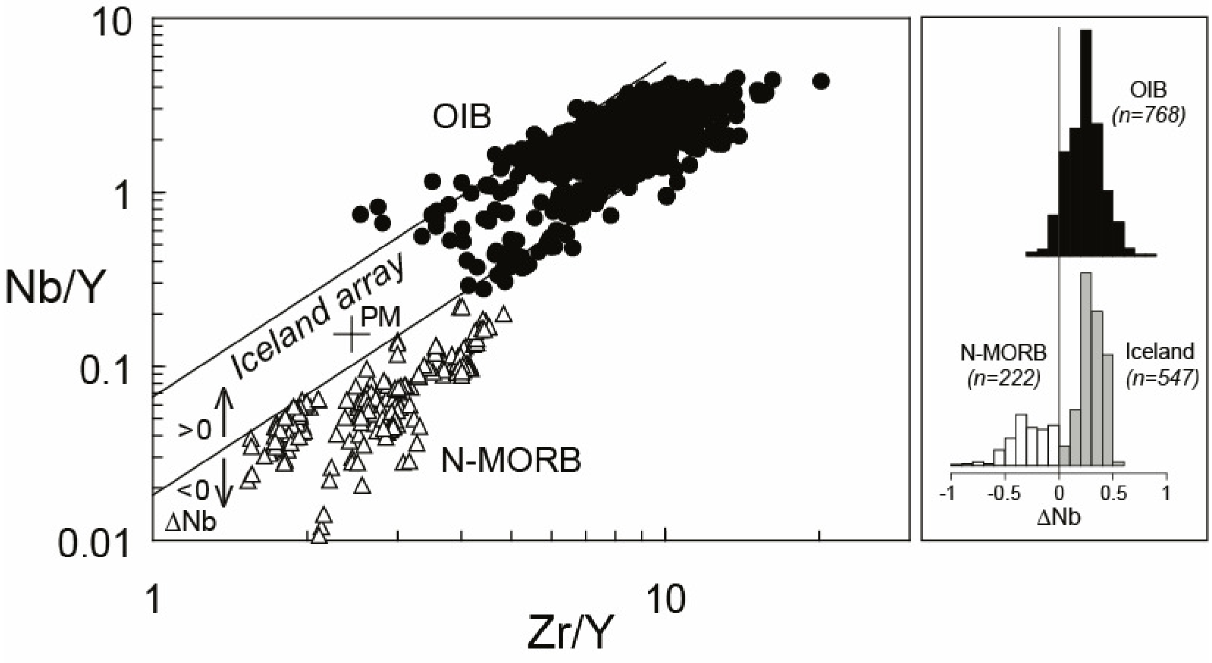

Figure 5. Comparison of ocean island basalt (OIB) data (J.G. Fitton and D. James, GSA Data Repository Item XXXXX) with the Iceland array and N-MORB composition (Fig. 1). The OIB data set includes samples from most major ocean islands and island groups but excludes data from the Galápagos Islands. PM is the composition of primitive mantle (McDonough and Sun, 1995). The histogram compares the distribution of ΔNb in OIB, Icelandic basalt, and N-MORB.

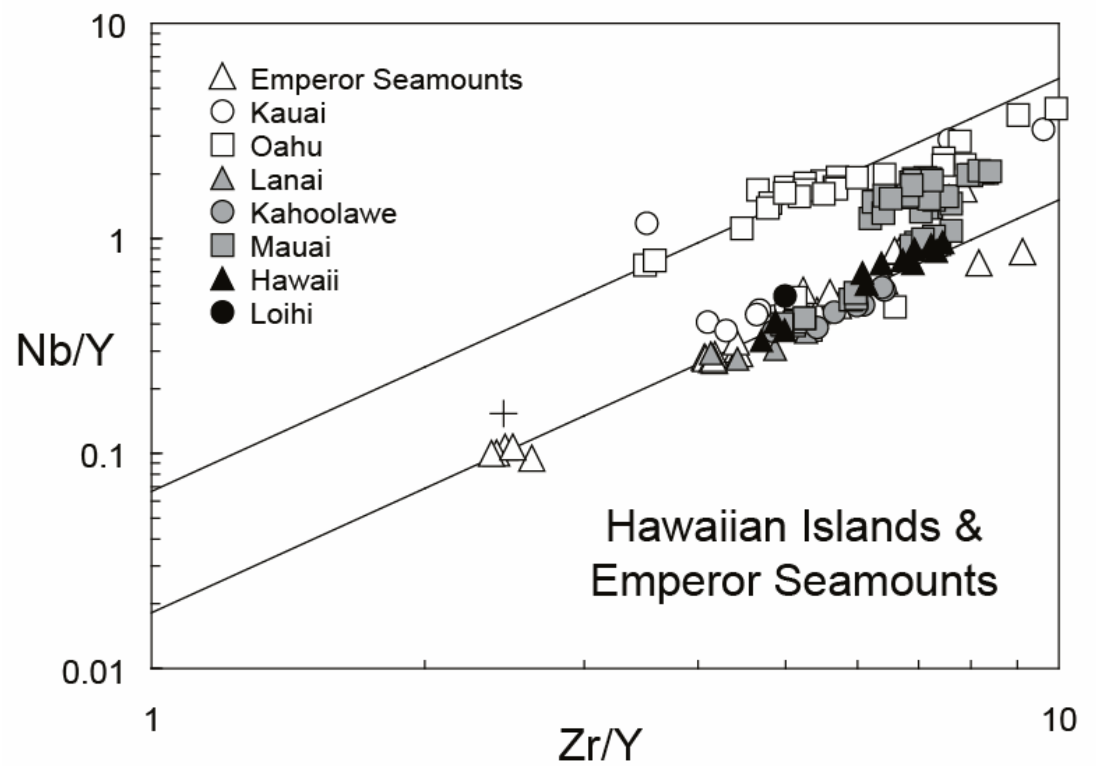

Figure 6. Composition of basalt samples from the Hawaiian islands (J.G. Fitton, GSA Data Repository Item XXXXX) and Emperor Seamounts (data from Regelous et al., 2003). The two parallel lines mark the limits of the Iceland array; the + represents the composition of primitive mantle (see Fig. 1). Hawaiian basalts plotting on the lower line (ΔNb = 0) are from the shield-forming stage of their respective islands; those plotting on the upper line are from the rejuvenated stage (Frey et al., 2005). The Emperor Seamount data, representing the composition of the Hawaiian mantle source from 85 to 42 Ma, plot around the lower line with the shield-forming lavas from the islands.

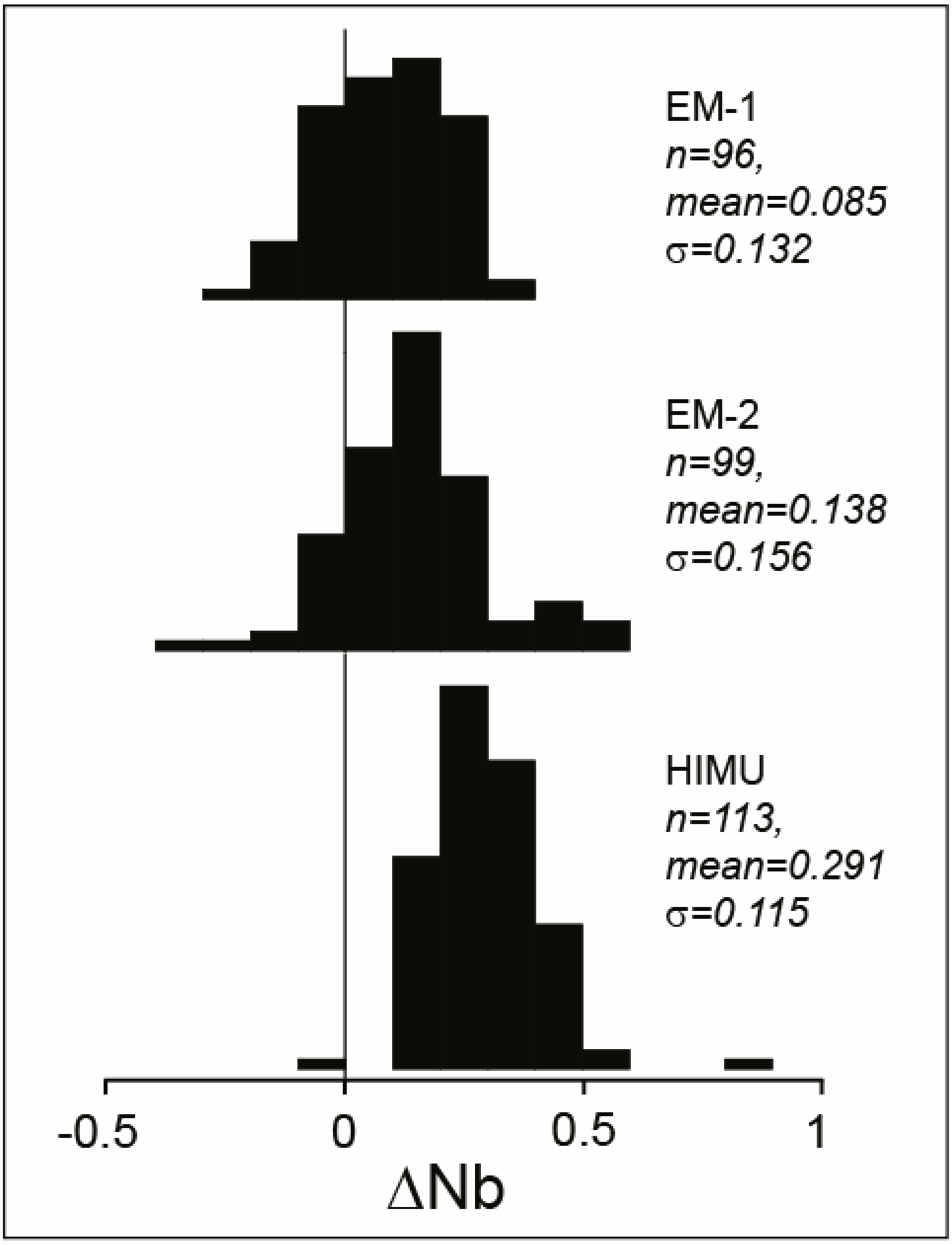

Figure 7. Histograms showing the distribution of ΔNb in extreme examples of HIMU, EM-1 and EM-2 OIB. Data from Willbold and Stracke (2006), filtered to exclude samples with MgO < 5 wt.%.

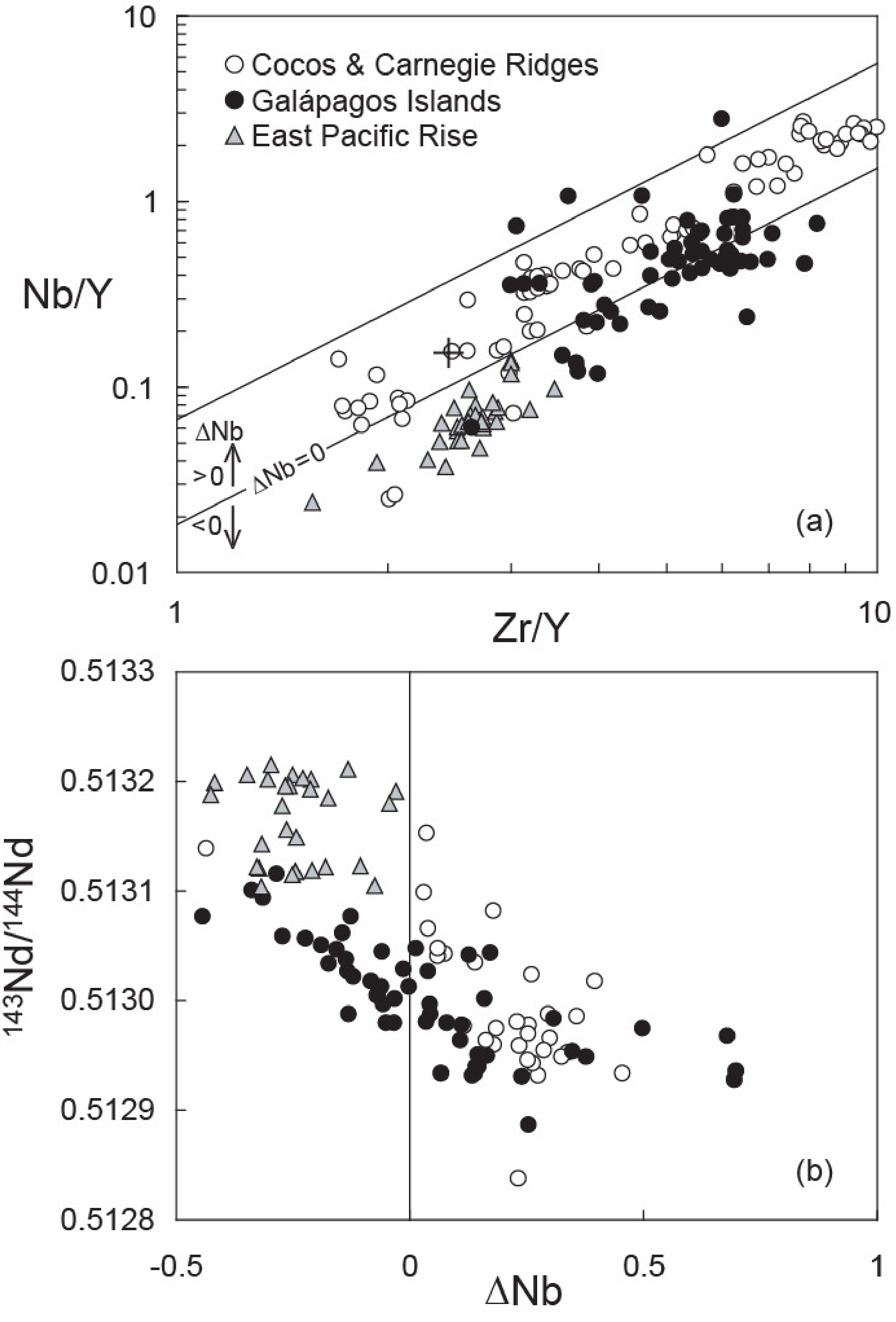

Figure 8. (a) Comparison of Nb/Y and Zr/Y from the Galápagos Islands (Thompson et al, 2003 and J.G. Fitton, GSA Data Repository Item XXXXX) with data from the Cocos and Carnegie Ridges (Harpp et al., 2005) and East Pacific Rise (Mahoney et al., 1994). The two parallel lines mark the limits of the Iceland array; the + represents the composition of primitive mantle (see Fig. 1). Some of the Galápagos basalt samples have significantly negative ΔNb, unlike other OIB (see Fig. 3, which does not include the Galápagos data). These samples probably represent small-degree melts of an N-MORB source. Note that the data from the Cocos and Carnegie Ridges plot within the Iceland array; the samples with ΔNb<0 were dredged from the sea floor adjacent to the ridges (Fig. 7). This suggests that mixing with ambient upper mantle is a recent phenomenon in the Galápagos Islands. **(b)** Variation of $^{143}Nd/^{144}Nd$ with ΔNb in the same suite of samples. Isotope data are from White et al. (1993), Galápagos Islands; Werner et al. (2003), Cocos and Carnegie Ridges; Mahoney et al. (1994), East Pacific Rise. The good negative correlation probably results from mixing between Galápagos plume mantle and ambient upper mantle (Geist et al. 1988).

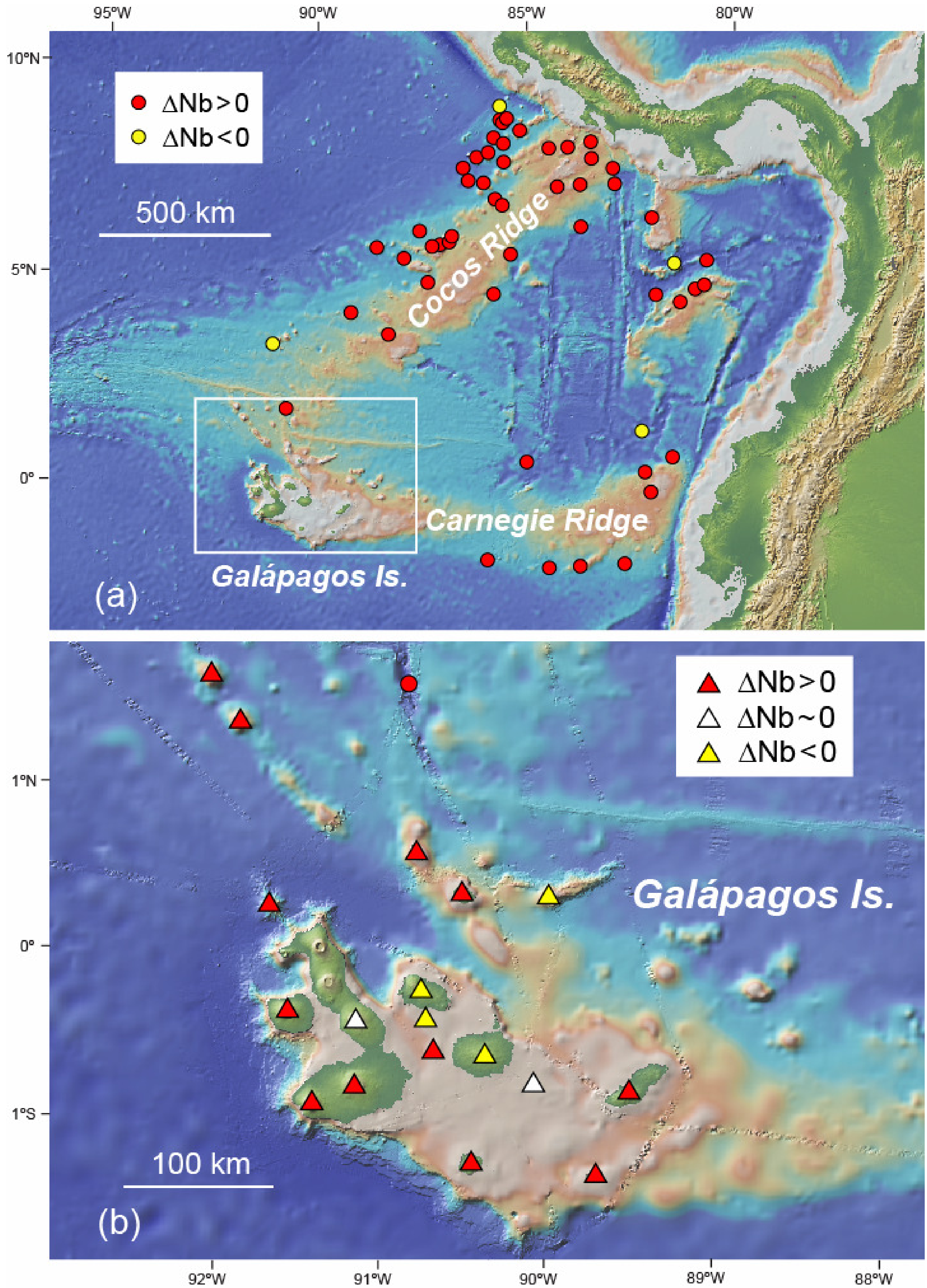

Figure 9. Maps of the Galápagos Islands and adjacent sea floor showing the location of samples in Fig. 6. **(a)** The location and character (ΔNb) of samples dredged on and around the Cocos and Carnegie Ridges. Note that the samples with ΔNb<0 were dredged from the seafloor adjacent to the ridges; samples from the ridges have ΔNb>0 (similar to Iceland). **(b)** An enlargement of the area in the white box in (a) summarising the character of the basalts on each of the Galápagos Islands. Islands in the centre of the archipelago have ΔNb≈0 or <0. This is consistent with the observation of Geist et al. (1988) and White et al. (1993) that the Galápagos hotspot has an N-MORB-like core partly surrounded by a horseshoe-shaped OIB-like outer zone.

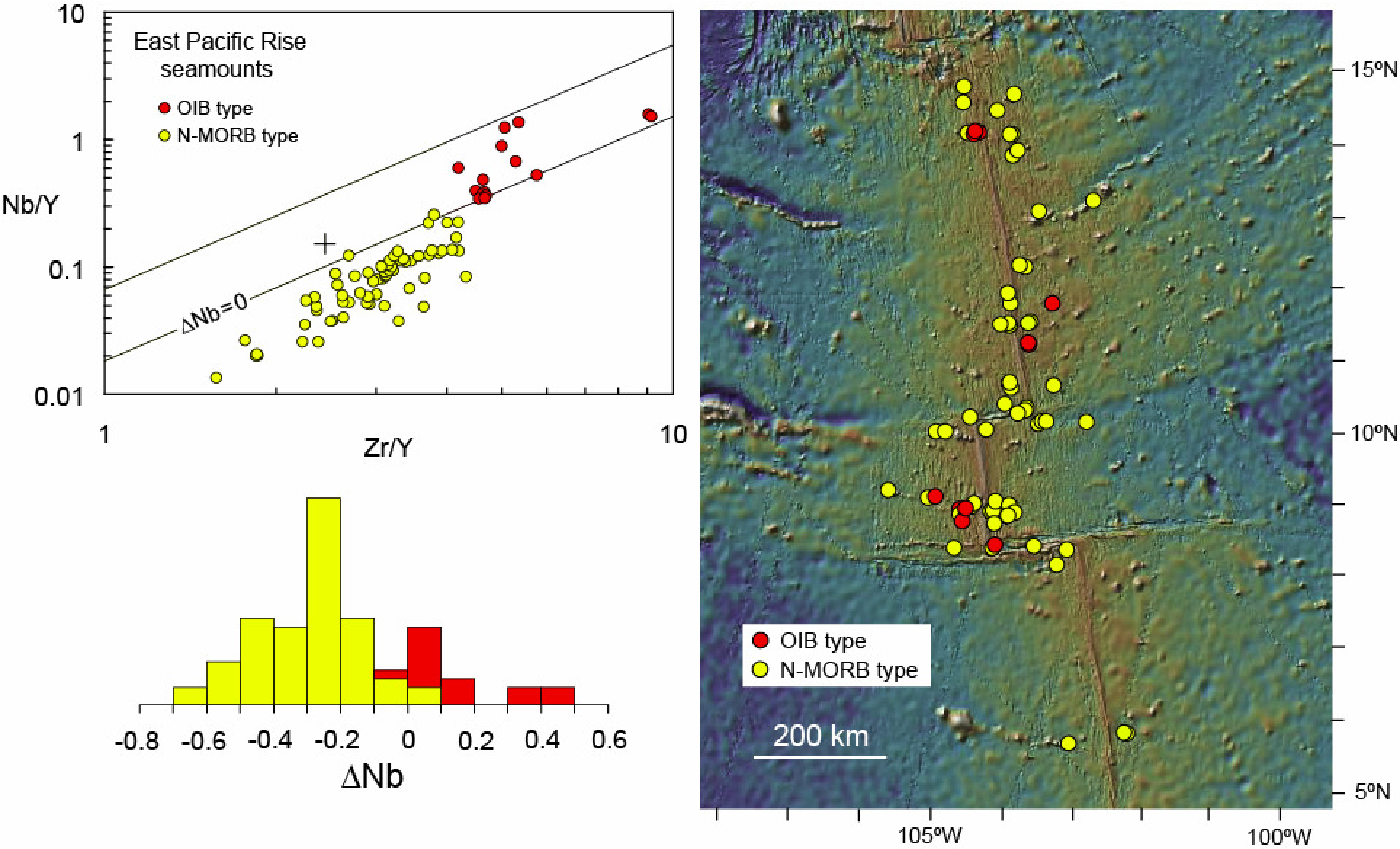

Figure 10. The location and composition of dredged seamounts on the flanks of the northern East Pacific Rise (data from Niu and Batiza, 1997). The two parallel lines mark the limits of the Iceland array; the + represents the composition of primitive mantle (see Fig. 1). The seamounts vary in composition from highly depleted N-MORB to enriched alkali basalt (Niu and Batiza, 1997) and fall into two distinct N-MORB (yellow) and OIB (red) compositional types separated almost perfectly by the ΔNb=0 reference line. The histogram (colour coded as for the data points) shows the weakly bimodal distribution of ΔNb in the data set (80 samples); the distribution of N-MORB- and OIB-type seamounts is shown on the map.

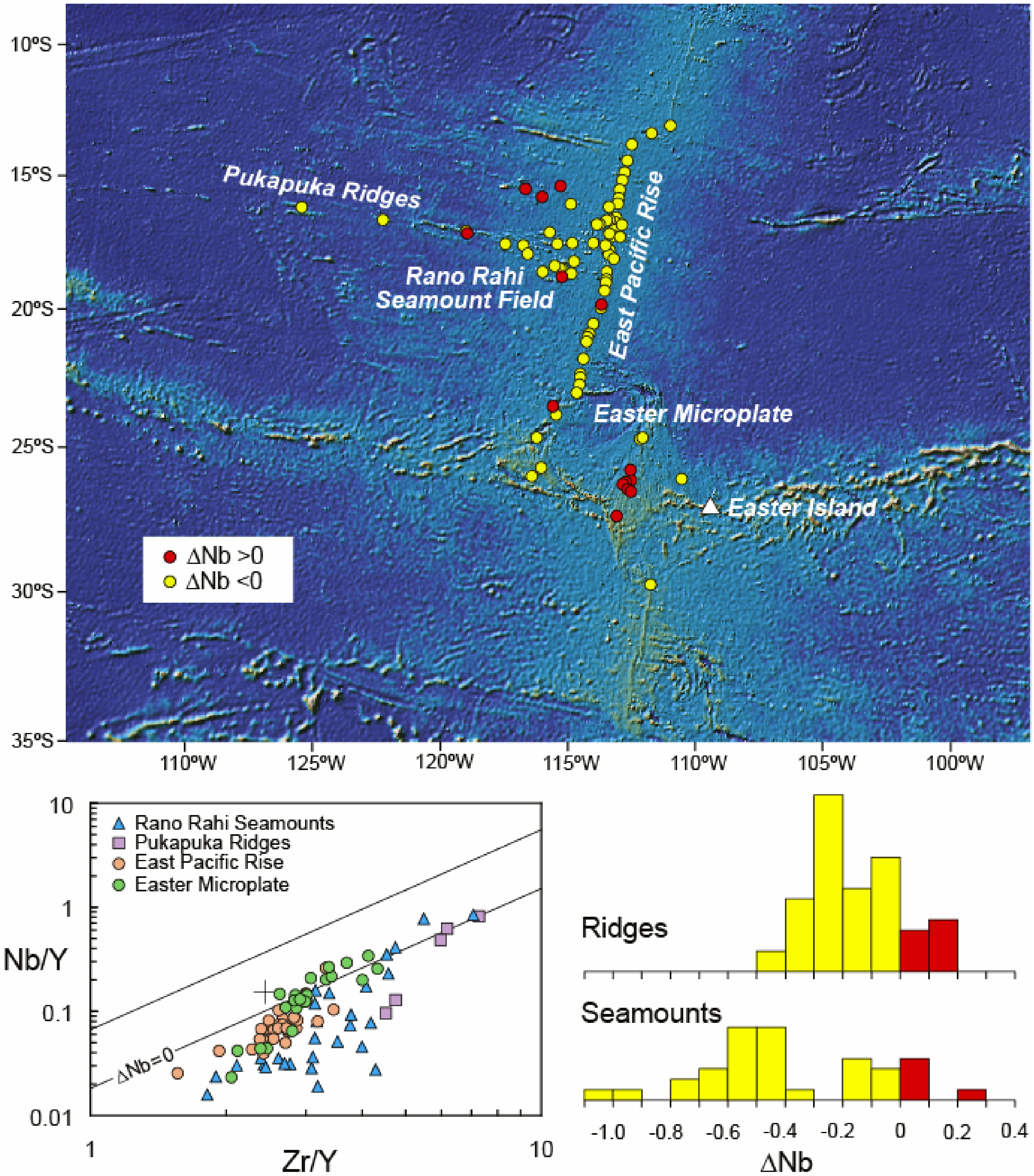

Figure 11. Map of the southern East Pacific Rise and adjacent areas showing the distribution and composition of samples dredged from the Rano Rahi Seamount Field (Hall et al., 2006), the Pukapuka Ridges (Janney et al., 2000), the axis of the southern East Pacific Rise (Mahoney et al., 1994), and the spreading centres around the Easter Microplate (Haase, 2002). Parallel lines on the Nb/Y vs. Zr/Y plot mark the limits of the Iceland array in Fig. 1; the + symbol represents primitive mantle. ΔNb is the deviation, in log units, from the reference line separating the Iceland array from N-MORB (the lower of the two parallel lines). Points on the map and the data points are colour coded for positive (red) and negative (yellow) values of ΔNb. The histograms compare the distribution of ΔNb in seamounts (34 samples) with that in the spreading axes (54 samples); note the greater spread of ΔNb in the seamounts. Most of the axial samples with ΔNb>0 are from parts of the ridge closest to Easter Island.

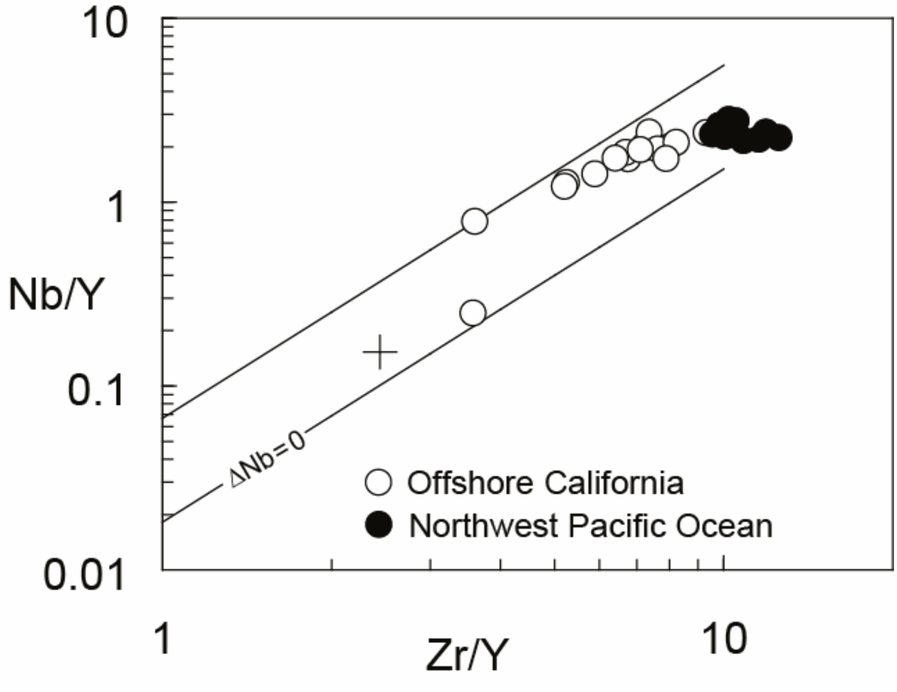

Figure 12. Composition of alkali basalt from two groups of seamounts in the Pacific Ocean thought to originate through lithospheric extension and fracture. The seamounts off the California coast (Davis et al., 2002) are 7–11 Ma older than the underlying ocean crust; those from the northwest Pacific, off Japan (Hirano et al., 2006), formed at 1–8 Ma on 135-Ma ocean crust. Parallel lines on the plot mark the limits of the Iceland array in Fig. 1; the + symbol represents primitive mantle. Basalt samples from each group of seamounts plot within the Iceland array and are compositionally indistinguishable from OIB.

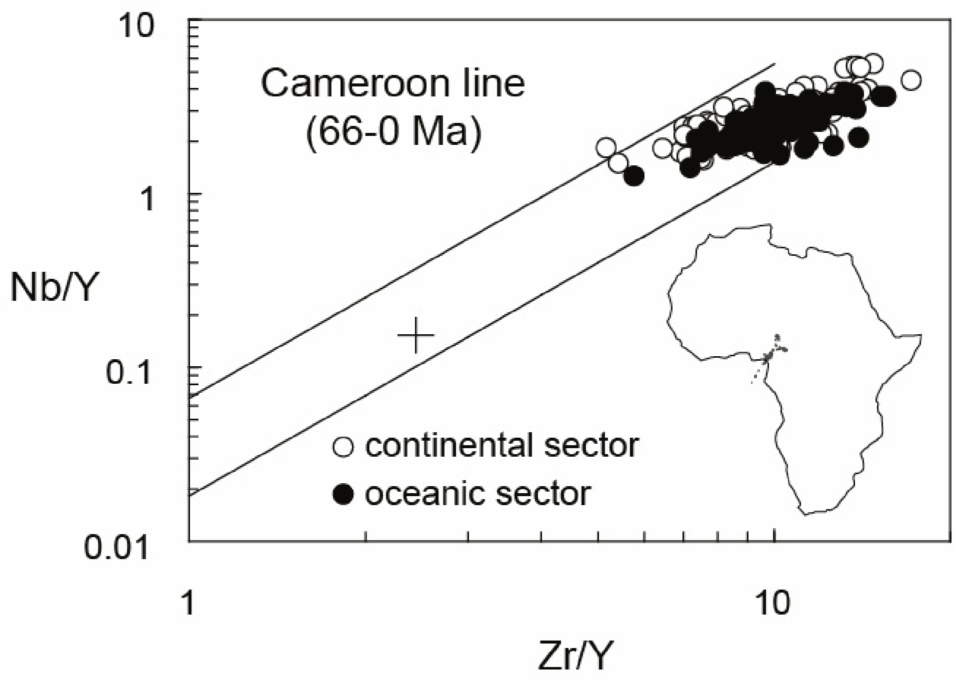

Figure 13. Composition of basalt from the oceanic and continental sectors of the Cameroon line (J.G. Fitton, GSA Data Repository Item XXXXX). The two parallel lines mark the limits of the Iceland array; the + represents the composition of primitive mantle (see Fig. 1). Inset map of Africa shows the outcrop of Cameroon line volcanic rocks. The data plotted are from basalt samples ranging in age from ~30-0 Ma; the older rocks (not plotted) are plutonic. The basalts from the continental sector are indistinguishable in composition from those from the oceanic sector OIB and therefore must have a sublithospheric source (Fitton and Dunlop, 1985).

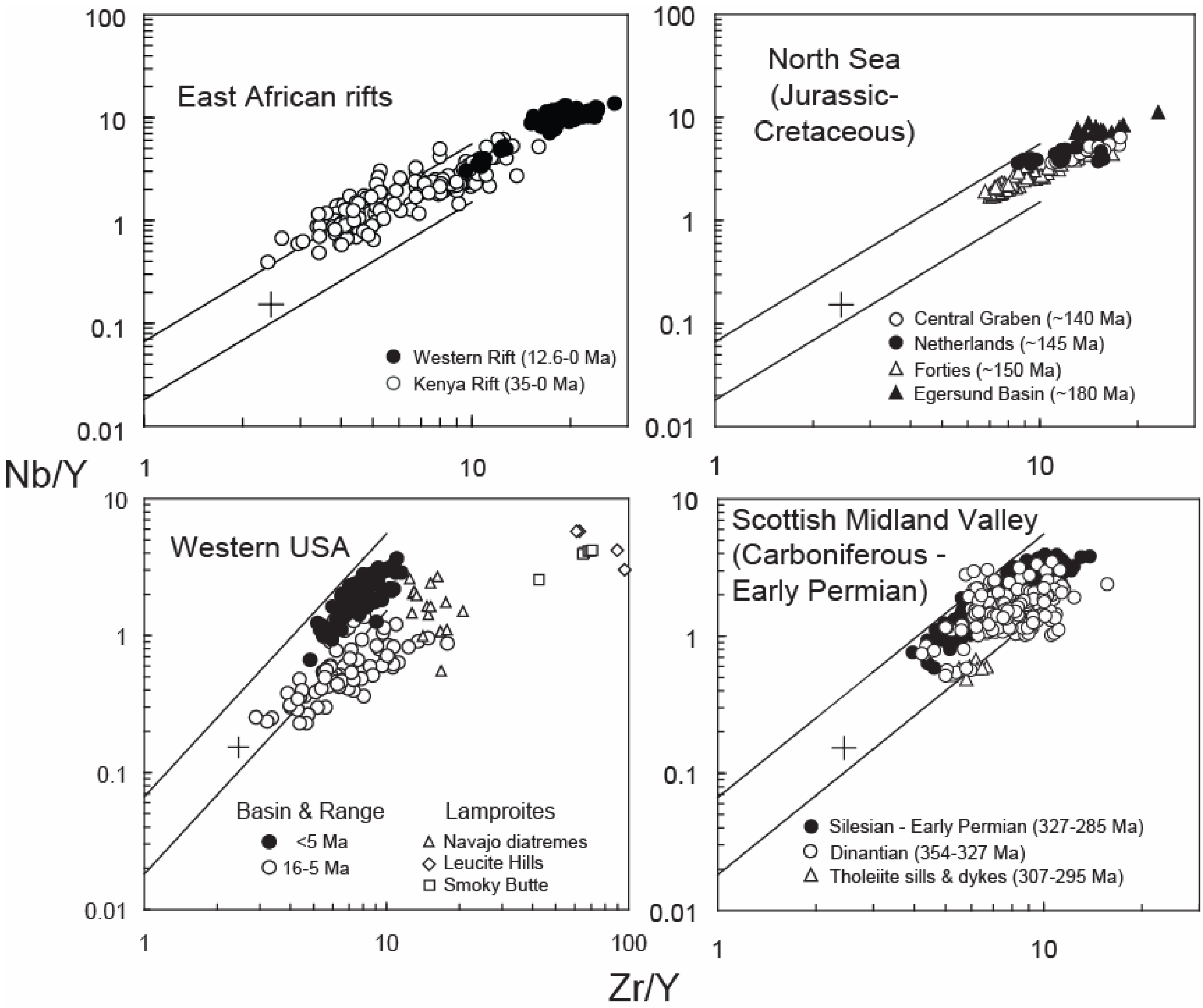

Figure 14. Composition of basalt from four continental rift systems. The two parallel lines mark the limits of the Iceland array; the + represents the composition of primitive mantle (see Fig. 1). Kenya Rift data: Macdonald et al. (2001). Western Rift: James (1995). North Sea data: Latin (1990); Latin et al. (1990). Western USA data: Fitton et al. (1991); Smoky Butte data from Fraser et al. (1985). Scottish Midland Valley data: Smedley (1986); Wallis (1989). Data from PhD theses and other unpublished data are available in GSA Data Repository Item XXXXX. Age of volcanism from: Ebinger (1989) (Western Rift); Macdonald (2003) (Eastern Rift); Latin et al. (1990) (North Sea Basin); Fitton et al. (1991) (Basin and Range); Upton et al. (2004) (Scottish Midland Valley). With the exception of the early (16-5 Ma) Basin and Range basalts, most of the basalts are OIB-like. The composition of the early Basin and Range basalts is best explained by mixing with an enriched component, represented by lamproites, in the subcontinental lithospheric mantle.

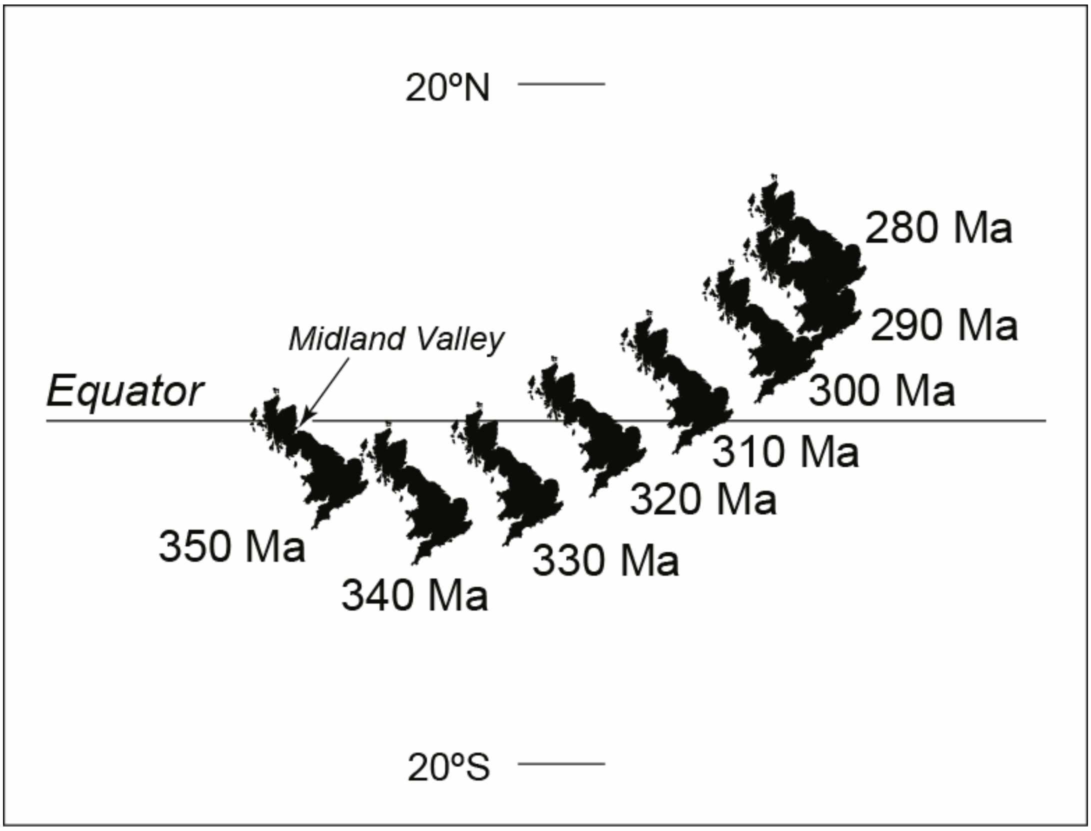

Figure 15. Location of the Scottish Midland Valley relative to the Equator during the period of rifting and magmatism (Early Carboniferous to Early Permian) based on Lawver et al. (2002). The Midland Valley moved around 15º northwards across the Equator during the 70 Ma when it was volcanically active.

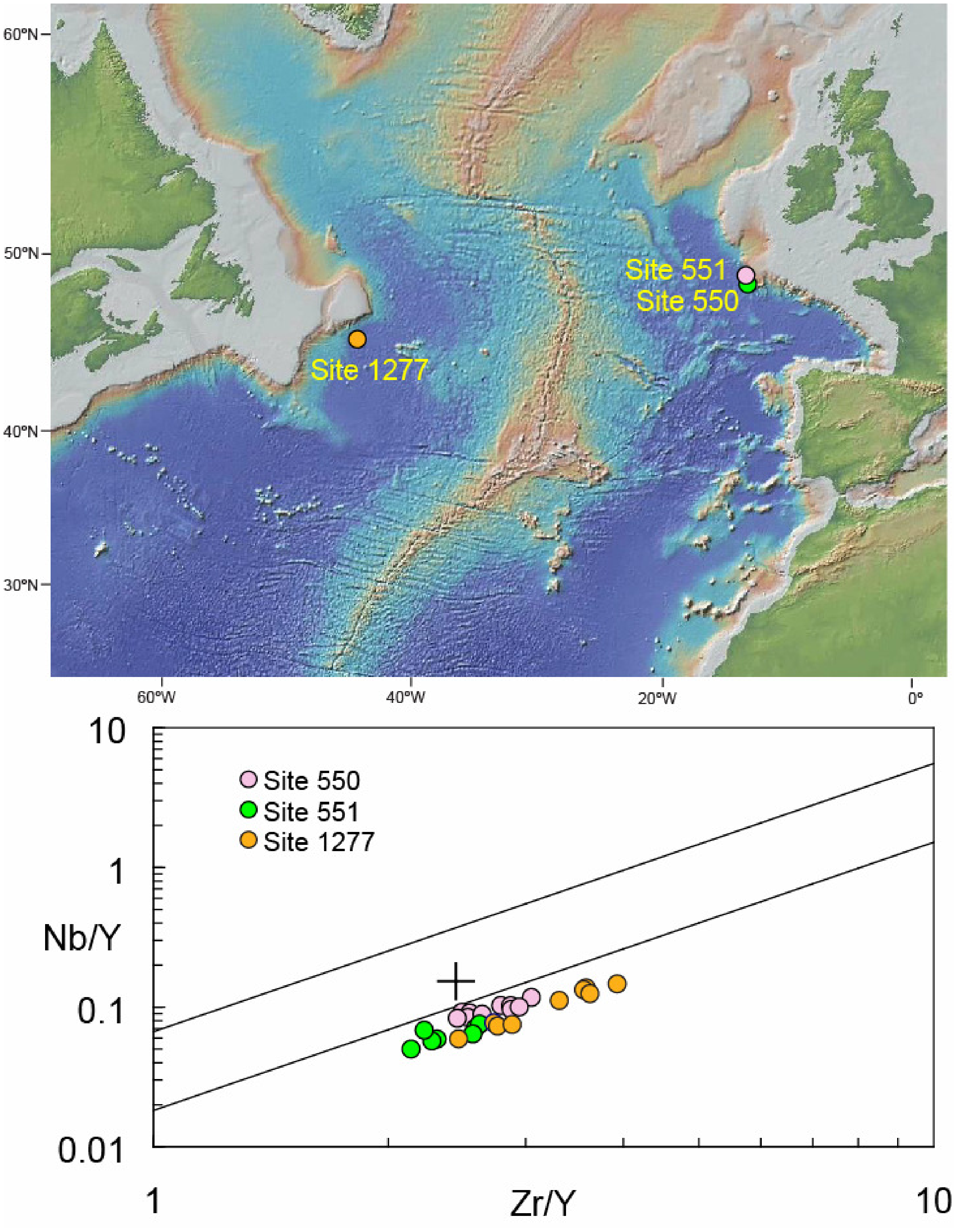

Figure 16. Map showing the location of DSDP and ODP drill sites that have penetrated the earliest ocean crust on non-volcanic ocean margins. The composition of basalt samples recovered at these three sites is shown on the diagram below the map. Data sources: Site 550 and 551, Kempton et al. (2000) and J.G. Fitton, GSA Data Repository Item XXXXX; Site 1277, Robertson (in press). The two parallel lines mark the limits of the Iceland array; the + represents the composition of primitive mantle (see Fig. 1). All the samples have negative ΔNb and appear to be N-MORB.