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Oxygen isotope heterogeneity of the mantle deduced from global ^{18}O systematics of basalts from different geotectonic settings

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Abstract Based upon a compilation and analysis of O-isotope data for Neogene volcanic rocks worldwide, the $\delta^{18}\text{O}$ variation for 743 basalts (historic lavas, submarine glasses, and lavas with < 0.75 wt% H_2O) is $+ 2.9$ to $+ 11.4\%$. Mid-ocean-ridge basalt (MORB) has a uniform O-isotope composition with $\delta^{18}\text{O} = + 5.7 \pm 0.2\%$. Basalts erupted in different tectonic settings have mean $^{18}\text{O}/^{16}\text{O}$ ratios that are both lower and higher than MORB, with continental basalts enriched in ^{18}O by ca. 1% over oceanic basalts. The $\delta^{18}\text{O}$ range for the subset of 88 basalts with $\text{Mg}\# [100 \cdot \text{Mg}/(\text{Mg} + \text{Fe}^{2+})]$ 75–68, considered to be unmodified primary mantle partial melts, is $+ 3.6$ to $+ 8.7\%$. These features are a clear indication that: (1) the Earth's upper mantle is heterogeneous with respect to its O-isotope composition; (2) that both low- ^{18}O and high- ^{18}O reservoirs have contributed to basalt petrogenesis. Large-ion lithophile element-enriched basalts associated with subduction at convergent plate margins are slightly enriched in ^{18}O , a characteristic that is considered to be an intrinsic feature of the subduction process. Intraplate oceanic and continental basalts have highly variable $^{18}\text{O}/^{16}\text{O}$ ratios, with individual localities displaying $\delta^{18}\text{O}$ ranges in excess of 1.5 to 2% . Systematic co-variations between O-, Sr-, Nd-, and Pb- isotope ratios reflect the same principal intramantle end-member isotopic components (DMM, HIMU, EM-I, EM-II) deduced from radiogenic isotope considerations and, therefore, imply that a common process is responsible for the origin of upper mantle stable and radiogenic isotope heterogeneity, namely the recycling of lithospheric material into the mantle.

Introduction

As a result primarily of trace element, radiogenic isotope, and noble gas studies of volcanic rocks and mantle xenoliths over the past two decades, it is now well established that the Earth's mantle is compositionally heterogeneous on a variety of scales and that basaltic magmas generated in different tectonic settings are derived from geochemically distinct mantle sources. The elemental and isotopic heterogeneities observed within the mantle are a time-integrated manifestation of both depletion and enrichment processes that have operated over much of Earth history. The process primarily responsible for this mantle heterogeneity is considered to be the recycling of lithospheric material into the mantle through such mechanisms as subduction of oceanic crust and entrained sediment (Armstrong 1968; Hofmann and White 1982), tectonic erosion of the continental margins during subduction (von Huene and Scholl 1991), and delamination of subcontinental lithosphere (McKenzie and O'Nions 1983; Kay and Kay 1990), although other possible causes such as the entrainment of either metasomatised mantle domains or primitive lower mantle by uprising plumes have also been advocated (Vollmer 1983; Galer and O'Nions 1985; Sun and McDonough 1989).

Oxygen isotope studies of volcanic rocks have the potential to make important contributions to studies of mantle heterogeneity, basalt petrogenesis, and crust-mantle evolution. Oxygen is the most abundant element in silicate rocks, as well as in both the melt and fluid phases considered to be responsible for metasomatic enrichment within the mantle. Importantly, the stable isotopes ^{18}O and ^{16}O are strongly fractionated in the low-temperature geochemical environments prevailing at the Earth's surface. Therefore, tectonic processes that return to the mantle material that has at one time resided in surface or near-surface environments, should provide an effective means of introducing oxygen of variable isotopic composition into the upper mantle.

Mid-ocean ridge basalt (MORB), volumetrically the most important and geochemically coherent lava type

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presently erupted at the Earth's surface, exhibits a restricted range of O-isotope variation (Ito et al. 1987). As a consequence, petrogenetic studies utilizing oxygen isotopes as a geochemical tracer have tended to assume that the MORB $^{18}\text{O}/^{16}\text{O}$ ratio is representative of the mantle source from which all basaltic magmas are derived. Although this is a reasonable generalization, there is evidence for mantle O-isotope heterogeneity on a significant scale from studies of peridotite massifs (Javoy 1980; Pearson et al. 1990), peridotite xenoliths in alkali basalt lavas (Kyser et al. 1981, 1982; Harmon et al. 1987; Kempton et al. 1988), and eclogite xenoliths from kimberlites (Garlick et al. 1971; Ongley et al. 1987; Neal et al. 1990). O-isotope data recently obtained for basaltic volcanism at Hawaii (Garcia et al. 1989, 1993), the Afar Rift (Barrat et al. 1993), and several Atlantic volcanic islands (Javoy et al. 1986; Weis et al. 1987; Halliday et al. 1988; Sigmarsson et al. 1992a) have documented the widespread occurrence of both tholeiitic and alkalic ocean-island basalt (OIB) with low- $\delta^{18}\text{O}$ values. This contrasts sharply with the situation in Polynesia (e.g. Woodhead et al. 1993) and at many continental intraplate locations (e.g. Kyser et al. 1982; Harmon et al. 1987; Stern et al. 1990; Duncker et al. 1991; Embey-Isztin et al. 1993) where basalts are characterized by $\delta^{18}\text{O}$ values which extend from the MORB range to significantly higher values. The strongly ^{18}O -enriched character of mafic potassic lavas of the Italian peninsula has been long recognized (Turi and Taylor 1976; Taylor and Turi 1976; Taylor et al. 1979) and remains a hotly debated topic (Holm and Munksgaard 1982; Rogers et al. 1986; Varenkamp and Kalamarides 1989; Turi et al. 1991).

From this brief overview, as well as previous summaries of O-isotope data for volcanic rocks (Taylor 1968; Kyser et al. 1982; Harmon and Hoefs 1984; Valley et al. 1986; Kyser 1990), it is clear that the O-isotope composition of basalts can be highly variable and that a comprehensive review and evaluation is warranted. Here we present the results of such a study, based upon our compilation of O-isotope data for Neogene volcanic rocks worldwide (Harmon and Hoefs 1993), and consider the ^{18}O - ^{87}Sr - ^{143}Nd - ^{206}Pb systematics of basalts with the aim of better understanding the geochemistry of the upper mantle and magmagenesis in different tectonic settings.

The database

From more than 130 literature sources, plus a large amount of our own unpublished data and that provided by colleagues, we have assembled a database consisting of 2855 O-isotope analyses for Neogene volcanic rocks (Fig. 1). Some 40% of the entries are accompanied by major element chemistry and/or radiogenic isotope data (Harmon and Hoefs 1993). A copy of the database, with its complementary geochemical data and a bibliography listing the literature sources consulted, is available from the authors upon request. In constructing the database, we have followed the classification system for volcanic

rocks recommended by the IUGS Subcommittee on the Systematics of Igneous Rocks (Le Bas et al. 1986). Here, we present and discuss the most important observations and conclusions as they relate to basalts (i.e. volcanic rocks classified as microbasalt, basalt, trachybasalt, and tephrite basanite according to the IUGS scheme); a more comprehensive discussion of our study, as it applies to the petrogenesis of differentiated volcanic rock suites and to individual tectonic settings will be presented elsewhere.

It is well established that the O-isotope composition of volcanic rocks, especially glassy and microphyric lavas, is particularly sensitive to secondary, post-depositional modification; processes such as hydrothermal alteration, weathering, and hydration can produce significant enrichments in ^{18}O (Taylor 1968; Muehlenbachs and Clayton 1972; Matsuhisa and Kurawasa 1983; Cerling et al. 1985; Ferrara et al. 1986; Harmon et al. 1987). Therefore, it was necessary to filter our primary database to account for this fact. It is difficult to know with certainty what the actual water content of a volcanic rock more than a few thousand years' old was at the time of its formation because: (1) basaltic magmas of different composition have variable dissolved water contents, which can exceed 1 wt% H_2O ; (2) degassing of a magma begins at high levels in the upper crust before the magma is erupted and continues past the time a lava has solidified; (3) subaerial lavas are more extensively degassed prior to and during eruption than submarine lavas; (4) the rate of secondary uptake of water after eruption and cooling is dependent upon lava bulk composition, climate, and time (Moore 1970; Friedman 1976, Garcia et al. 1979, 1989; Kyser and O'Neil 1984; Taylor et al. 1983; Cerling et al. 1983; Byers et al. 1985; B. Taylor 1986). Therefore, based upon the $\delta^{18}\text{O}$ - H_2O relationships observed for both individual sample suites and the total sample population, we have restricted the database for this study (Fig. 1b) to include only: (1) historic lavas < 1500 years old; (2) pristine submarine glasses; (3) samples with measured water contents of < 0.75 wt%. Basalts (Fig. 1c) constitute 55% of the total sample population of 1358 in this screened database.

The filtering criteria were chosen to be as objective and conservative as possible, while still maintaining a sufficiently large population of samples to overcome the "signal-to-noise" problem that is inherent in any overview study of this kind. Much of the potentially useful data in the literature had to be omitted from the screened database due to a lack of information about water content or age of the samples analysed. Modification of the screening criteria, for example reducing the water content limit to 0.5 wt% or increasing it to 1 wt%, does not alter the basic picture or conclusions presented here.

Shown in Fig. 1 are histograms which illustrate the $\delta^{18}\text{O}$ frequency distributions for: (a) all of the 2855 samples in the database; (b) the 1358 volcanic rocks of all composition in the screened population; (c) only the 743 basalts from the screened population for which $^{18}\text{O}/^{16}\text{O}$

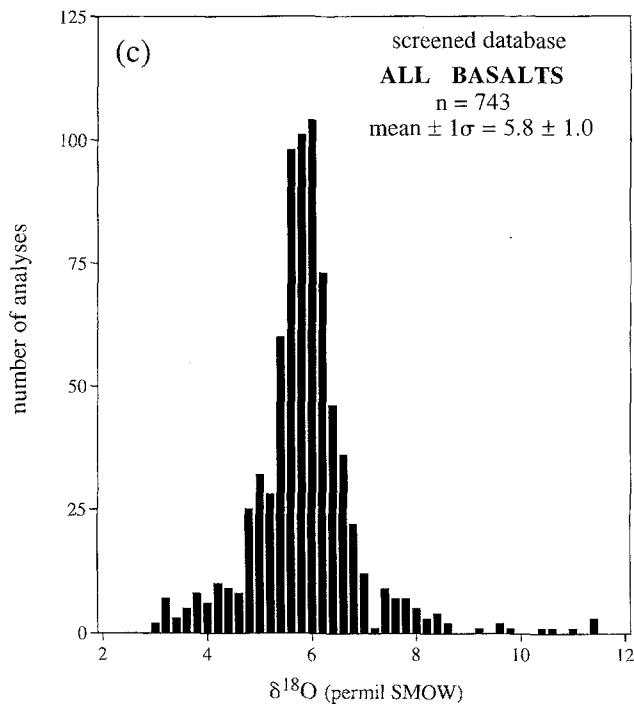
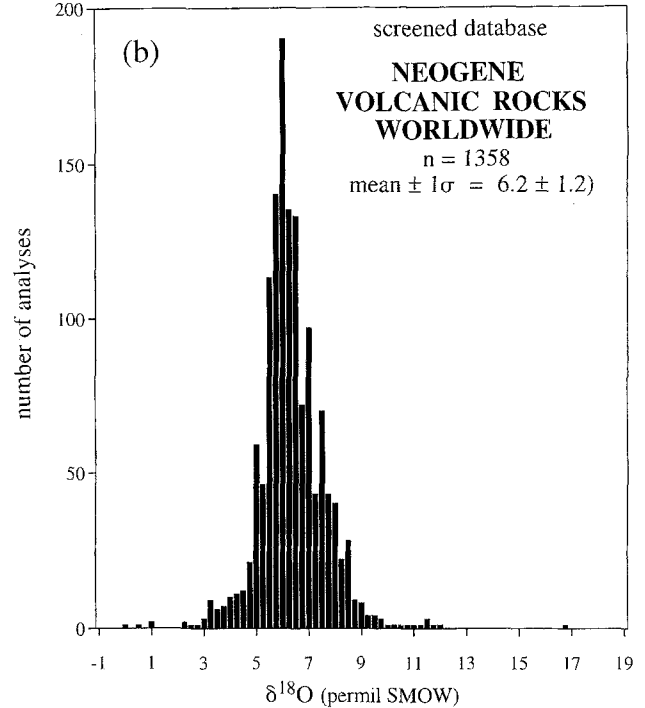
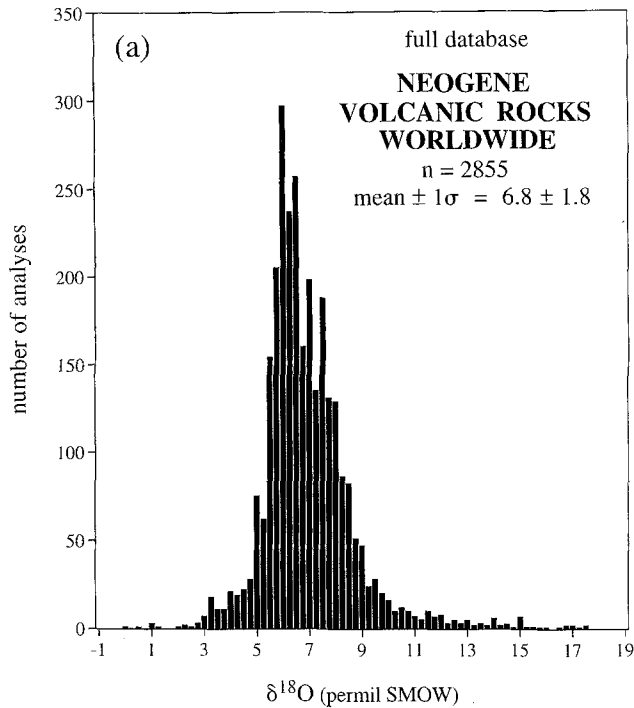


Fig. 1a–c Histograms showing the $\delta^{18}\text{O}$ frequency distributions for: **a** 2855 Neogene volcanic rocks worldwide; **b** the 1358 samples in the screened population of 2855 that are of historic age (< 1500 years old), are pristine submarine glasses, or have water contents < 0.75 wt%; **c** the 743 basalts from the screened population of 1358. See text for discussion

ratios are considered pristine. Two important observations can be drawn from these $\delta^{18}\text{O}$ distributions:

1. The total variation in O-isotope composition for Neogene volcanic rocks worldwide is almost 18‰, extending from $\delta^{18}\text{O} = -0.2$ to $+17.4$ ‰; however, > 90% of samples have $\delta^{18}\text{O}$ values in the restricted interval $+5$ to $+9$ ‰. The reduction in both overall ^{18}O variation and average $\delta^{18}\text{O}$ value between the complete (Fig. 1a) and the screened (Fig. 1b) volcanic rock databases,

$$\delta^{18}\text{O}_{\Sigma\text{DB}} = +6.8 \pm 1.8\text{‰}$$

$$\delta^{18}\text{O}_{\text{Screened DB}} = +6.2 \pm 1.2\text{‰},$$

is a clear reflection of the removal from the sample population of lavas whose O-isotope composition has been affected by secondary, subsolidus alteration.

2. The basalt subset of the screened database (Fig. 1c) has a $\delta^{18}\text{O}$ frequency distribution with the same form as that of the total screened database for lavas of all compositions (Fig. 1b). Although the global average $^{18}\text{O}/^{16}\text{O}$ ratio for all basalts is the same as the MORB average, as defined by this and previous studies (Table 1),

$$\delta^{18}\text{O}_{\Sigma\text{basalts}} = +5.8 \pm 1\text{‰}$$

$$\delta^{18}\text{O}_{\Sigma\text{MORB}} = +5.7 \pm 0.2\text{‰},$$

the > 8‰ range of O-isotope variation exhibited by basalts worldwide, from $\delta^{18}\text{O} = +2.9$ to $+11.4$ ‰ (Fig. 1c), is surprisingly large. This feature indicates that both low- ^{18}O and high- ^{18}O reservoirs have contributed to basalt petrogenesis. This substantial ^{18}O variation must reflect either significant O-isotope heterogeneity in the mantle source regions in which particular basalts are generated or substantial interaction of basaltic magma with isotopically diverse crust prior to eruption – or variable combinations of these two effects.

Basalt O-isotope systematics

Physiographic setting

Figure 2 and Table 1 compare $\delta^{18}\text{O}$ distributions for the 440 oceanic and 303 continental basalts in the screened database. Considered together, the histograms illustrate a distinct difference in O-isotope character for basalts erupted in these two fundamentally different physiographic settings. On average, oceanic basalts are significantly less ^{18}O rich than continental basalts and exhibit substantially less O-isotope variation:

oceanic basalts:

average = $+5.4 \pm 0.7\text{‰}$, median = $+5.5\text{‰}$;

continental basalts:

average = $+6.4 \pm 1.1\text{‰}$, median = $+6.1\text{‰}$.

For oceanic basalts, $\delta^{18}\text{O}$ values vary from $+2.9$ to $+7.5\text{‰}$ and the $\delta^{18}\text{O}$ distribution is skewed toward lower $^{18}\text{O}/^{16}\text{O}$ ratios, whereas for continental basalts the $\delta^{18}\text{O}$ values range from $+4.8$ to $+11.4\text{‰}$ with the distribution strongly biased toward higher ratios.

Primary partial melts

Although all basalts are a consequence of partial melting within the mantle, the $\delta^{18}\text{O}$ value measured for a basaltic lava will reflect that of its source region only to the extent that it is pristine and petrogenetic processes operating during magma segregation, storage, and transit have not significantly modified the $^{18}\text{O}/^{16}\text{O}$ ratio of the magma. Primary basalts can be recognized on the basis of their Mg# [$\text{Mg\#} = 100 \cdot \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$]; Roeder and Emslie (1970), as basalts in equilibrium with a peridotite mantle source are characterized by Mg# values of 0.75–0.68 (Hansen and Langmuir 1978). The variation in basalt $\delta^{18}\text{O}$ values as a function of Mg# for the 512 basalts in our screened database for which chemical analyses are available is illustrated in Figs. 3 and 4, from which two important points emerge.

First, for the basalt population as a whole, there is no overall tendency for $^{18}\text{O}/^{16}\text{O}$ ratios to vary systematically as a function of bulk composition. The $\delta^{18}\text{O}$ distribution for primary basalts with $\text{Mg\#} = 75\text{--}68$, which can be considered primary mantle partial melts, have a $\delta^{18}\text{O}$ range in excess of 5‰ , from $+3.6$ to $+8.7\text{‰}$. Both basalts with $\text{Mg\#} > 75$ that have accumulated olivine and more differentiated basalts with lower Mg# of 67–50 display an equivalent magnitude of ^{18}O variation. Secondly, although the shapes of the histograms in Fig. 3 differ for the three Mg# groups, their mean $\delta^{18}\text{O}$ values are equivalent. This is further evidence that, in general, the effects of petrogenetic processes occurring during the early, closed system stages of basalt differentiation (such as crystal fractionation and accumulation) on the O-isotope composition of basaltic magmas are small, probably no more than a few tenths permil at most.

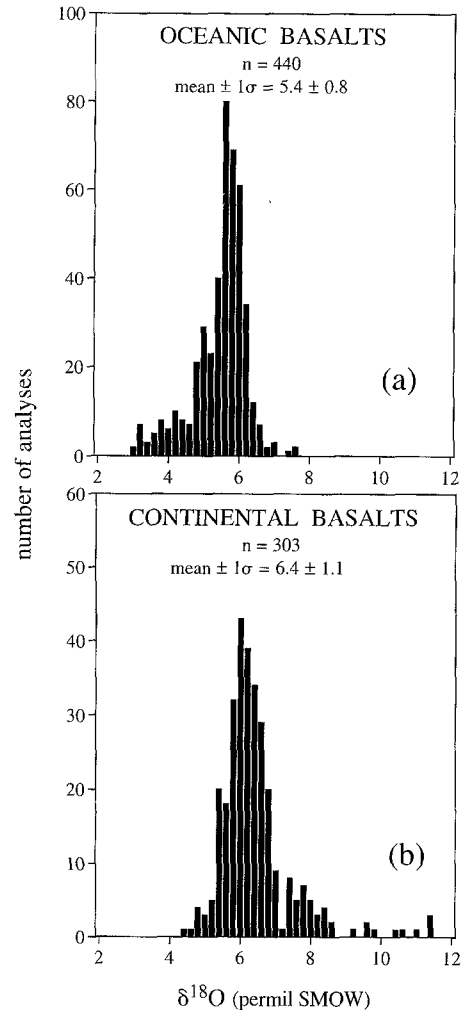


Fig. 2a,b Histograms showing the $\delta^{18}\text{O}$ frequency distributions for: **a** 440 oceanic basalts; **b** 303 continental basalts of the screened population in Fig. 1c. Included in the oceanic group are ocean-floor basalts (MORB), intraplate oceanic island and seamount basalts (OIB), fore-arc trough and back-arc basin basalts (FATB-BABB), and oceanic arc basalts (OAB); comprising the continental group are intraplate continental basalts (CIB), continental arc basalts (CAB), and continental flood basalts (CFB). Although the two groups have distinctly different $\delta^{18}\text{O}$ frequency distributions, basalts with low- and high- $\delta^{18}\text{O}$ values are present within each population. See text for discussion. Note the different vertical scales for the two histograms

Greater degrees of ^{18}O modification are to be expected in particular situations where a differentiated basalt is generated through a combined process of assimilation-crystal fractionation from a primary or less-evolved parental magma (Carlson et al. 1981; Ellam and Harmon 1990; Duncker et al. 1991; Hildreth et al. 1991) involving either basaltic underplating of the lower crust (Hildreth and Moorbath 1988; Arndt and Goldstein 1989) or the injection of a basaltic sheet into the continental crust (Patchett 1980; Huppert and Sparks 1988). Extensive crustal contamination can also occur through magma mixing when a basaltic magma passes through a tectonically over thickened and thermally elevated crustal section that has been subject to partial anatexis (Hildreth

Table 1 O-isotopic characteristics of basalts (<0.75% H₂O plus submarine glasses and historic lavas) erupted in different tectonic settings

Tectonic setting	Number	$\delta^{18}\text{O}$ range	Mean $\delta^{18}\text{O} \pm 1\sigma$
Oceanic basalts			
MORB	127	5.2– 6.4	5.7 ± 0.2
OIB	148	4.6– 7.5	5.5 ± 0.5
Iceland	104	2.9– 6.2	4.5 ± 0.8
FATB and BABB	28	5.5– 6.6	5.9 ± 0.3
OAB	33	5.3– 7.5	6.1 ± 1.1
All oceanic basalts	440	2.9– 7.5	5.4 ± 0.8
Continental basalts			
CIB	171	4.5– 8.1	6.1 ± 0.7
CAB	82	4.8– 7.7	6.2 ± 0.7
CFB	17	4.3– 6.5	5.6 ± 0.6
Italy	33	6.3–11.4	8.5 ± 1.4
All continental basalts	303	4.3–11.4	6.4 ± 1.1
All basalts	743	2.9–11.4	5.8 ± 1.0

and Moorbath 1988; Bohlen and Metzger 1989; Patino Douce et al. 1990; Davidson et al. 1990).

High- and low- $\delta^{18}\text{O}$ values

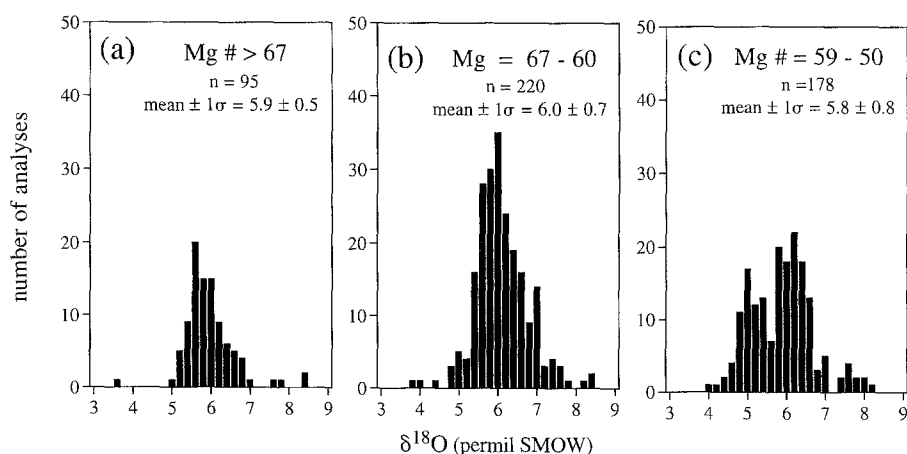
Mid-ocean ridge basalts exhibit the least O-isotope variation of the nine basalt groups; $\delta^{18}\text{O}$ values for the 127 samples in the database range from only + 5.2 to + 6.4‰ and average + 5.7 ± 0.2 ‰ (Table 1, Fig. 4). This standard deviation of 0.2‰ is equivalent to the analytical uncertainty typically reported in the literature for individual O-isotope ratio measurements of silicate rocks and minerals. Therefore, in the discussion which follows, MORB will be considered to be representative of the bulk ^{18}O character of the upper, suboceanic mantle and used as a reference against which the O-isotope composition of basalts erupted in other tectonic settings will be compared.

Of the 743 basalts in the screened population, 354 (48%) have $\delta^{18}\text{O}$ values which fall outside the ± 2 standard deviation range about the MORB mean of + 5.7‰, (i.e. $6.1\text{‰} < \delta^{18}\text{O} < + 5.3\text{‰}$). The low- ^{18}O and high- ^{18}O groups are predominantly associated with, but not restricted to, particular physiographic settings (Fig. 2).

Oceanic basalts comprise 135 of the 155 samples with $\delta^{18}\text{O}$ values $< + 5.3\text{‰}$, whereas 161 of the 199 samples with $\delta^{18}\text{O} > + 6.1\text{‰}$ are continental basalts. However, continental basalts comprise 13% of the low- $\delta^{18}\text{O}$ samples and, similarly 19% of the high- $\delta^{18}\text{O}$ samples are oceanic. Thus, although there is a bias towards particular physiographic settings, both low- $\delta^{18}\text{O}$ and high- $\delta^{18}\text{O}$ signatures are a feature of basalts erupted in both oceanic and continental environments.

The most obvious explanation for this observation is that the extended $\delta^{18}\text{O}$ ranges in Fig. 2 are a direct result of crustal contamination of the basalts prior to eruption, either during transit or while in residence in crustal magma chambers. The low- ^{18}O tails to the $\delta^{18}\text{O}$ distributions would then reflect the interaction of parental basaltic magmas with hydrothermally altered, low- ^{18}O crust. The former explanation is applicable to lavas erupted from the central volcanoes in Iceland (Muehlenbachs et al. 1974; Hémond et al. 1988, 1993; Sigmarsson et al. 1991a, 1992b; Nicholson et al. 1991) and also has been advocated to explain the ^{18}O -depleted character of intraplate basalts of the Yellowstone Plateau (Hildreth et al. 1984, 1991). At these localities, compositionally fractionated tholeiitic basalts with Mg# < 50 typically display low- $\delta^{18}\text{O}$ values, and there is a continuum of $^{18}\text{O}/^{16}\text{O}$ ratios between these tholeiitic basalts and derivative intermediate/evolved lavas. Likewise, it is clear that crustal contamination may be responsible for the elevation of O-isotope ratios for subduction-related basalts above the ca. + 6‰ baseline, particularly in instances where strong covariations of $^{18}\text{O}/^{16}\text{O}$ ratios with bulk chemistry and radiogenic isotope composition are ob-

Fig. 3 Histograms showing the $\delta^{18}\text{O}$ frequency distributions for 512 of the 743 basalts from the screened population for which chemical data are available as a function of Mg#. Basalts considered primary mantle partial melts exhibit a $\delta^{18}\text{O}$ range of + 3.6 to + 8.5‰ and a mean $\delta^{18}\text{O}$ value of + 5.9 ± 0.5 ‰ (a). These parameters are equivalent for more fractionated basalts with Mg# = 67–60 (b) and Mg# = 59–50 (c), suggesting that the effects of differentiation on the O-isotopic composition of basaltic magmas, in general, are small and that the ca. 5‰ variation in $\delta^{18}\text{O}$ values for basalts with Mg# > 50 is an indication of a mantle source that is heterogeneous with respect to its O-isotope composition



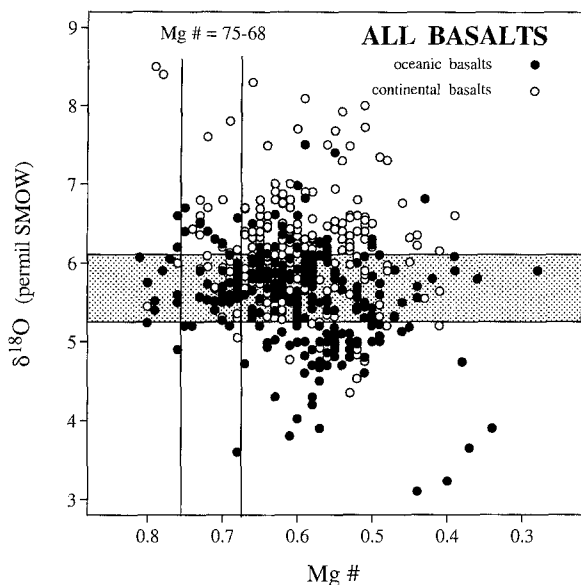


Fig. 4 Plot of $\delta^{18}\text{O}$ versus Mg# for oceanic basalts (filled circles) and continental basalts (open circles). The shaded field across the middle of the diagram denotes the $\pm 2\sigma$ range from + 5.3 to + 6.1‰ about the MORB mean $\delta^{18}\text{O}$ value of + 5.7‰ and the clear vertical field denotes the Mg# range of 75–68 for primary basaltic partial melts in equilibrium with a peridotite source. See text for discussion

served (Harmon et al. 1981; James 1982; Davidson and Harmon 1989; McBirney et al. 1987; Ellam and Harmon 1990). It also applies to continental rift basalts, in cases where compositionally evolved quartz-normative tholeiitic basalts are more ^{18}O rich than associated alkali olivine basalts or olivine-normative tholeiites (Harmon et al. 1987; Duncker et al. 1991).

Although crustal interaction/contamination processes may explain much of the ^{18}O variation observed for both oceanic and continental intraplate basalts (OIB and CIB), it is not the only cause, particularly for unfractionated basalts (Figs. 3 and 4), where the fundamental cause must be intramantle O-isotope heterogeneity. This point is clear when the O-isotope data for intraplate basalts are considered on a locality basis (Table 2).

For OIB, not only do particular islands or seamounts exhibit highly variable O-isotope compositions, but individual localities appear to have a characteristic O-isotope signature, which may be MORB-like or may be quite distinct from MORB. For example, both the Canary Islands in the Atlantic Ocean and Reunion in the Indian Ocean are MORB-like in terms of their $\delta^{18}\text{O}$ ranges and mean $\delta^{18}\text{O}$ values. Iceland, the Cameroon Line islands, Tristan de Cunha, and Ascension in the Atlantic Ocean – as well as Hawaii in the Pacific Ocean – have distinctly lower $^{18}\text{O}/^{16}\text{O}$ ratios, with $\delta^{18}\text{O}$ values extending well below + 5‰. By contrast, Gough Island in the Atlantic Ocean and Polynesian localities of the western Pacific tend to exhibit higher $^{18}\text{O}/^{16}\text{O}$ ratios, with $\delta^{18}\text{O}$ values extending well above 6‰. As noted by Kyser et al. (1981), primarily for the Pacific Ocean, alkalic OIB as a group tend to be slightly more ^{18}O rich than tholeiitic

Table 2 O-isotope characteristics of basalts (<0. 75% H_2O plus submarine glasses and historic lavas) from selected OIB and CIB localities.

Locality	Number	$\delta^{18}\text{O}$ range	Mean $\delta^{18}\text{O} \pm 1\sigma$
OIB			
Iceland	104	2.9–6.2	4.5 ± 0.8
Ascension Island	7	5.2–6.3	5.7 ± 0.4
Gough Island	7	5.8–7.0	6.5 ± 0.4
Tristan de Cunha	2	5.0–5.3	5.4 ± 0.3
Canary Islands	5	5.2–5.9	5.5 ± 0.2
Cameroon Line Islands	4	4.6–5.5	5.2 ± 0.4
Reunion Island	19	5.1–6.2	5.7 ± 0.3
Pitcairn seamounts	8	5.9–6.5	6.1 ± 0.2
Society Islands and seamounts	7	6.0–6.4	6.2 ± 0.1
Samoa Islands	4	5.7–6.1	5.9 ± 0.1
Hawaii	81	4.6–6.4	5.3 ± 0.3
CIB			
Pannonian Basin, Hungary	6	6.0–6.7	6.3 ± 0.3
Rhine Graben, Germany	29	5.3–7.6	6.1 ± 0.5
Massif Central, France	25	5.4–6.5	5.9 ± 0.3
Etna, Italy	3	5.9–6.1	6.0 ± 0.1
Southwestern US	22	5.1–8.1	6.2 ± 0.6
Cameroon Line	4	4.6–5.5	5.2 ± 0.4
Afar Rift, Djibouti	7	5.4–6.0	5.6 ± 0.2
Dead Sea Rift, Turkey	4	5.8–6.5	6.1 ± 0.3
Tariat and Dariganga, Mongolia	5	5.2–5.8	5.5 ± 0.2
Patagonia	30	5.3–7.8	6.1 ± 0.6
Mt. Melbourne, Antarctica	5	5.2–6.2	5.8 ± 0.3

OIB. Importantly, where such basalt types are erupted together this distinction is maintained, with both types being either of a low- $\delta^{18}\text{O}$ or a high- $\delta^{18}\text{O}$ character. For example, all tholeiites erupted on Hawaii have $^{18}\text{O}/^{16}\text{O}$ ratios that are distinctly lower than MORB (Garcia et al. 1989). However, the tholeiitic and alkalic basalts that occur within a single stratigraphic succession of lavas on the summit of Loihi are all of a low- ^{18}O character but display a clear trend of increasing ^{18}O content from tholeiite to alkali olivine basalt, $\delta^{18}\text{O}_{\text{TH}} = + 4.96 \pm 0.17\text{‰}$, $\delta^{18}\text{O}_{\text{AOB}} = + 5.18 \pm 0.13\text{‰}$ (Garcia et al. 1993).

In the case of CIB (Table 1), some volcanic fields have a distinctly high- $\delta^{18}\text{O}$ (e.g. the Pannonian Basin in Hungary) or a low- $\delta^{18}\text{O}$ (e.g. the Cameroon Line) character, whereas others exhibit a wide range of $\delta^{18}\text{O}$ values which extend from well below to well above the MORB $\delta^{18}\text{O}$ range (e.g. the Rhine Graben of Germany and Patagonia). Thus, although individual CIB samples may have MORB-like $\delta^{18}\text{O}$ values, no locality appears to be characterized by $^{18}\text{O}/^{16}\text{O}$ ratios restricted to the MORB range.

To summarize, significant numbers of low- $\delta^{18}\text{O}$ (13) and high- $\delta^{18}\text{O}$ (26) lavas, from both oceanic and continental settings, are included within the group of 88 primary basalts that have a Mg# > 67 (Fig. 4). These 39 samples span a $\delta^{18}\text{O}$ range of nearly 5‰, from + 3.6 to + 8.5‰. A similar range of $\delta^{18}\text{O}$ variation is displayed by samples in the Mg# intervals from 67–60 and 59–50, and the ^{18}O character of individual intraplate localities is highly variable. Thus, we consider this large departure

from the MORB $\delta^{18}\text{O}$ range to be an intrinsic feature of basalts erupted in many tectonic settings and, in the remainder of the discussion, will consider the origin of this large variation in the O-isotopic composition of basalts.

Discussion

Oxygen – radiogenic isotope systematics

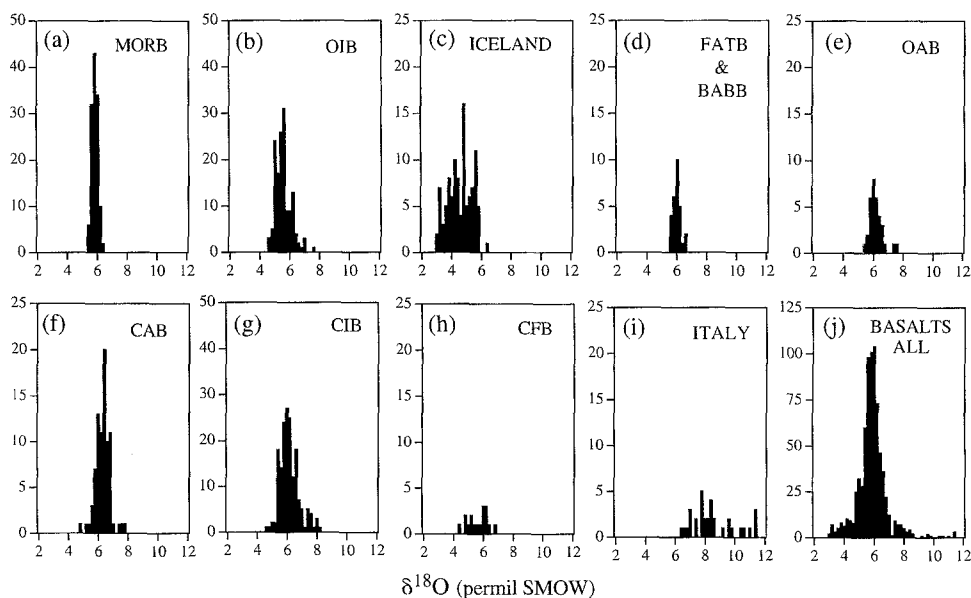
Multiple, geochemically distinct mantle components, generally considered to be mixing end-members, themselves heterogeneous to different extents, have been proposed to explain the Sr-, Nd-, and Pb-isotope systematics observed in oceanic basalts (Zindler et al. 1982; White 1985; Zindler and Hart 1986; Hart 1988). These are: depleted MORB-source mantle (DMM), a component characterized by high-Pb-isotope ratios and a subducted slab geochemical signature (HIMU), and two components characterized by differing styles (and possibly times) of incompatible element enrichment (EM-I and EM-II). Additional components have been advocated by some authors, but the evidence for them is not compelling. The EM and HIMU characteristics are largely restricted to OIB from a belt between 50°S and the Equator in the southern hemisphere – the DUPAL anomaly of Hart (1984) – and there is evidence to indicate that these components can closely be juxtaposed physically in the mantle (Weaver 1991; Chauvel et al. 1992). The situation for continental basalts is more complicated. Not only is there potential for localized trace element-enriched and compositionally distinct domains of various scale to be isolated and stored within the non-convecting subcontinental lithosphere for long periods of time, but there is also the ever-present possibility in continental settings for magma contamination during ascent by lithosphere and crust of extremely variable composition and age.

The basalt database shown in Fig. 1c further has been subdivided and evaluated in terms of seven different tectonic settings (Fig. 5; Table 1): mid-ocean ridges (MORB), oceanic intraplate islands (OIB), fore-arc troughs and back-arc basins (FATB-BABB), oceanic arcs (OAB), continental arcs (CAB), continental intraplate rift-zones (CIB), and continental flood basalt provinces (CFB). Iceland and the potassic magmatism of the Italian peninsula have been considered separately because of their unusual tectonic setting and particular O-isotope characteristics. As it is both difficult and problematic to distinguish source characteristics of continental basalts from those which may have been acquired subsequently as a result of contamination within the deep crust, the discussion which follows will focus initially on the oceanic basalt suites.

1. MORB

Based upon a detailed study of the oxygen and radiogenic isotope geochemistry of 52 carefully selected MORB samples, some 40% of the total sample population shown in Fig. 5a, Ito et al. (1987) concluded that O-isotope variations in MORB were largely independent of the incompatible element depletions and enrichments which respectively characterize N-type and P-type MORB. Overall, the expanded data set considered here supports this view. Not only is the mean value of + 5.7‰ $\delta^{18}\text{O}$ for MORB a global characteristic, it is also a regional attribute of most ridge segments where more than just a few samples have been analysed. However, the larger database MORB considered here reveals some fine structure to this general, large-scale feature (Fig. 6) that supplement the observations and conclusions of Ito et al. (1987). For example, the full 1.2‰ $\delta^{18}\text{O}$ variation for MORB of + 5.2 to + 6.4‰ is present at a single locality, the FAMOUS-AMAR section of the Mid-Atlantic Ridge at 36–37°N. This area is situated between two portions of

Fig. 5a–j Histograms showing the $\delta^{18}\text{O}$ frequency distributions for the basalts from different tectonic settings (MORB, OIB, FATB-BABB, OAB, CAB, CFB, and CIB) plus Iceland and Italy. Note the difference in vertical scale among the ten histograms. Abbreviations are defined in the caption to Fig. 2



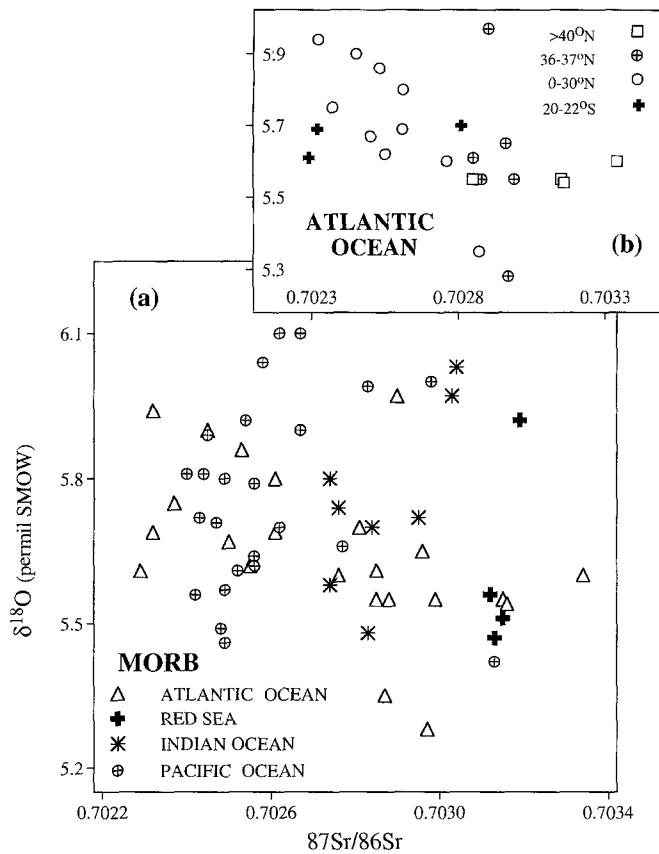


Fig. 6 Plots of $\delta^{18}\text{O}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$ for MORB worldwide (a) and various latitude bands of the Atlantic Ocean (b). See text for discussion

the ridge characterized as plume segments (Schilling et al. 1983) and displays quite uniform radiogenic isotope compositions, but highly variable incompatible element abundance ratios (Frey et al. 1993). By contrast, the segments between 3–6°N along the Mid-Indian Ridge and between 0–3°N on the East Pacific Rise, which lack a plume influence, are quite uniform in terms of both oxygen and radiogenic isotope compositions. On a larger scale, $\delta^{18}\text{O}$ variation for the Mid-Atlantic Ridge between 0–30°N is substantial and exhibits a negative correlation with Sr-isotope ratios (Fig. 6b), a feature which contrasts with the weak positive co-variation of O- and Sr- isotope ratios for the Pacific Ocean (Fig. 6a). These observations support the view of Ito et al. (1987) that the radiogenic isotope – and also the $\delta^{18}\text{O}$ – variations observed for MORB, both between and within different ocean basins, are a manifestation of heterogeneous domains of various spatial scale and magnitude within the depleted, asthenospheric upper mantle.

II. Ocean island basalts

The O-isotope character of OIB (Fig. 5b) presents a striking contrast to that of MORB. Whereas MORB is characterized by its relative O-isotope homogeneity, the 12 OIB localities represented in the database are notable

in terms of their O-isotope variability. The 148 samples comprising the OIB population in the screened database range in $\delta^{18}\text{O}$ value from +4.6 to +7.5‰ and have an average $\delta^{18}\text{O}$ value of $+5.5 \pm 0.5\text{‰}$ (Table 1). More than 45% of the analyses comprising the histogram in Fig. 5b lie outside the $\pm 0.2\text{‰}$ standard deviation range about the MORB mean $\delta^{18}\text{O}$ value of +5.7‰. This substantial $\delta^{18}\text{O}$ range for OIB of ca. 3‰, which extends to both significantly higher and lower $\delta^{18}\text{O}$ values than observed for MORB, cannot be explained in terms of the small ^{18}O fractionations that occur during either partial melting (Muehlenbachs and Kushiro 1974) or the closed system fractional crystallization of a basaltic magma (Matsuhisa 1979; Ito et al. 1987; Woodhead et al. 1987; Harmon and Gerbe 1992). As noted above, alkalic OIB exhibit a general tendency to be more ^{18}O rich than tholeiitic OIB ($\delta^{18}\text{O} = +5.8 \pm 0.5\text{‰}$ for 61 alkalic basalts vs $\delta^{18}\text{O} = +5.3 \pm 0.3\text{‰}$ for 74 tholeiitic basalts). The significance of this observation is not clear; it may be a distinctive feature reflecting isotopically distinct mantle source regions (e.g. plume and lithosphere) which produce these two different basalt types (Garcia et al. 1993), a temperature effect during mantle partial melting (Ionov et al. 1994), a manifestation of more complex intramantle petrogenetic processes (Kyser et al. 1982), or a locality effect that is related to the melting or assimilation of hydrothermally altered oceanic lithosphere by large plumes (Hémond et al. 1993).

Figures 7a and 8a are generalized $^{87}\text{Sr}/^{86}\text{Sr}$ - $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ - $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams for the oceanic basalt localities listed in Table 2 for which both O-isotope and radiogenic isotope analyses are available; Figs. 7b and 8b are analogous plots illustrating the global fields for MORB and OIB, together with the radiogenic isotope mantle components after Zindler and Hart (1986) and Hart (1988). In Fig. 9, $\delta^{18}\text{O}$ values of all oceanic basalts from the database are plotted as a function of their Sr-, Nd-, and Pb-isotope ratios. The radiogenic isotopic compositions for the mantle end-member components are shown after Figs. 7 and 8. The O-isotope positions for the end-member components have been selected arbitrarily to be close to the extreme points of the basalt data arrays. Although not shown here, the data distribution patterns do not change significantly as a function of Mg#; plots for compositionally restricted subsets of samples with $\text{Mg}\# > 67$ and $\text{Mg}\# = 67\text{--}60$ do not differ in form from those shown in Fig. 9, and the O-radiogenic isotope trends observed are the same.

The striking feature of Figs. 7–9 is that the mantle components defined on the basis of their radiogenic isotope compositions can also be delineated in terms of the O-radiogenic isotope systematics of oceanic basalts. The MORB cluster into well-defined and coherent fields on the $\delta^{18}\text{O}$ - $^{87}\text{Sr}/^{86}\text{Sr}$ and $\delta^{18}\text{O}$ - $^{143}\text{Nd}/^{144}\text{Nd}$ plots (Fig. 9a,b). The greater dispersion on the $\delta^{18}\text{O}$ - $^{206}\text{Pb}/^{204}\text{Pb}$ plot (Fig. 9c) reflects the considerable isotopic variation in MORB of Pb relative to Sr and Nd. For OIB (excluding Iceland) distinct mantle components are recognized by the $\delta^{18}\text{O}$ -radiogenic isotope trends away from

Fig. 7a,b Plots of $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$ for: **a** MORB and OIB localities based upon the samples included in the O-isotope database, with $\delta^{18}\text{O}$ ranges indicated; **b** the global MORB and OIB fields after Hart (1988) and Chauvel et al. (1992), with the DMM, HIMU, EM-I, and EM-II mantle source components indicated

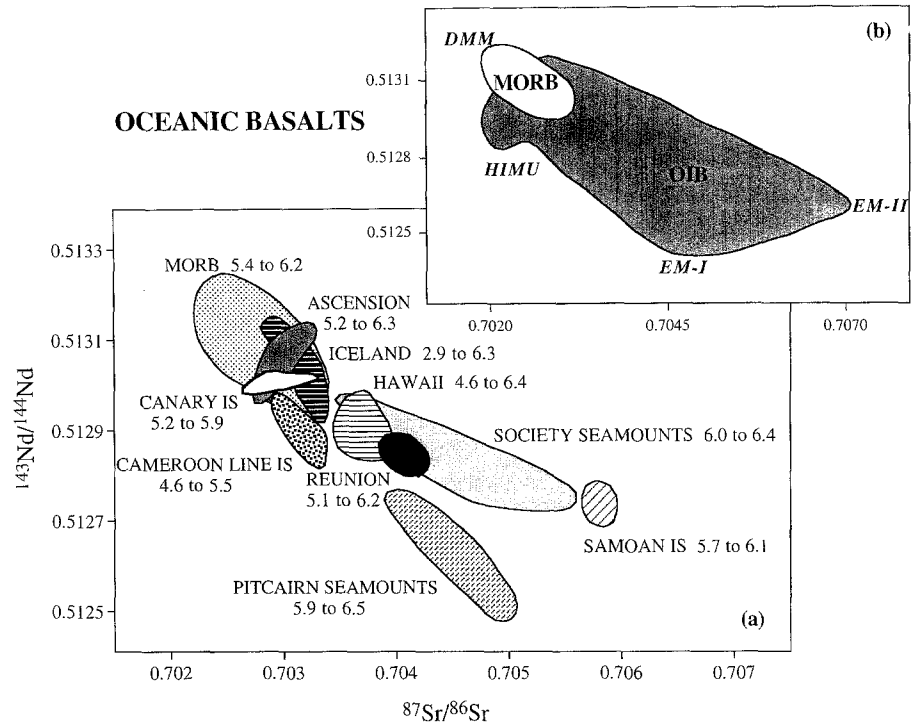
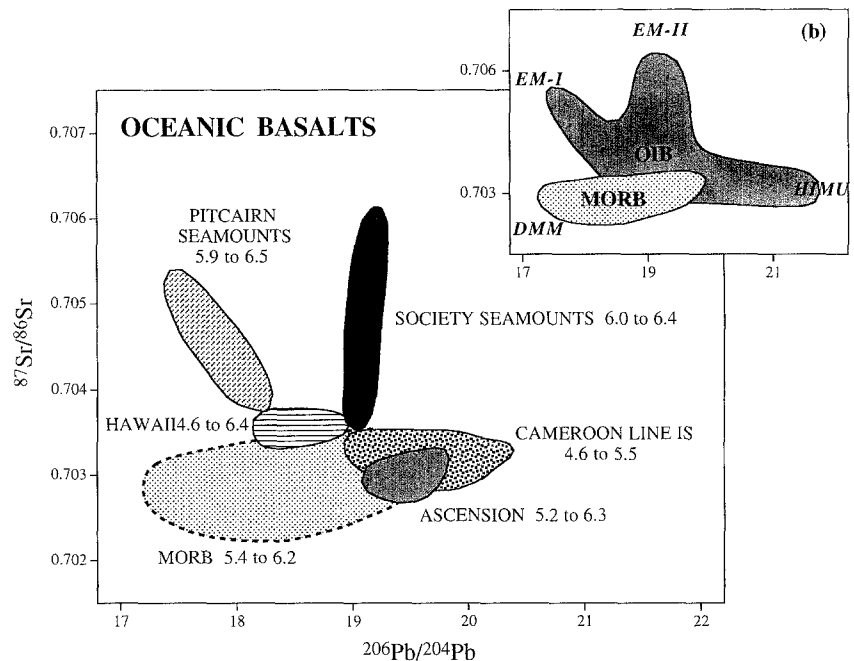


Fig. 8 Plots of $^{206}\text{Pb}/^{204}\text{Pb}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$ for MORB and OIB localities analogous to (a) and (b) of Fig. 7



the MORB cluster in each diagram towards mantle end-member compositions that are both enriched and depleted in ^{18}O relative to depleted upper mantle (DMM). In terms of Sr- and Nd-isotopes, the area of MORB-OIB overlap in Figs. 7a and 9a corresponds to the “Prevalent Mantle” or PREMA composition of Wörner et al. (1986). Interestingly, the OIB localities that fall in this central portion of the Sr-Nd diagram are associated with low- $\delta^{18}\text{O}$ values.

The enriched components EM-I and EM-II, which are defined on the basis of radiogenic Sr and unradiogenic

Nd, are clearly delineated in the different figures, where the basalt data describe distinct $\delta^{18}\text{O}$ - $^{87}\text{Sr}/^{86}\text{Sr}$, $\delta^{18}\text{O}$ - $^{143}\text{Nd}/^{144}\text{Nd}$, and $\delta^{18}\text{O}$ - $^{206}\text{Pb}/^{204}\text{Pb}$ trends. The EM-I trend, defined by OIB of the Pitcairn seamounts, is recognized by highly correlated enrichments in ^{18}O and ^{87}Sr coincident with Nd- and Pb-isotope depletion. By comparison, the EM-II trend, a feature described by OIB from Samoa and the Society seamounts, is defined by ^{18}O , ^{87}Sr , and Pb-isotope enrichments, coincident with ^{143}Nd depletions, that are not strongly interdependent. The EM-I trend has been modelled by Woodhead et al.

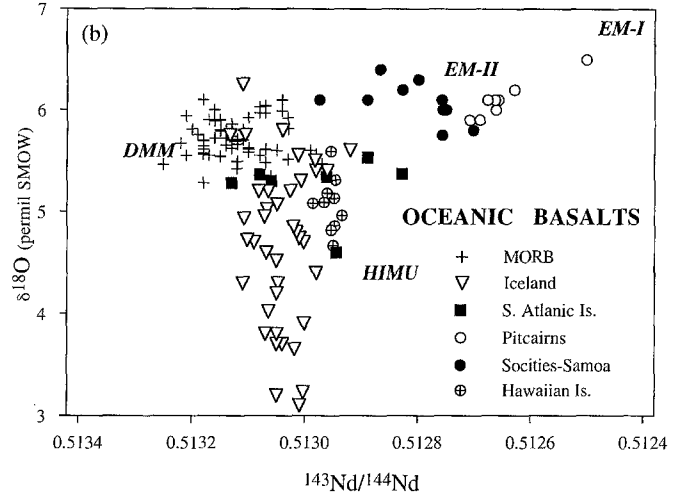
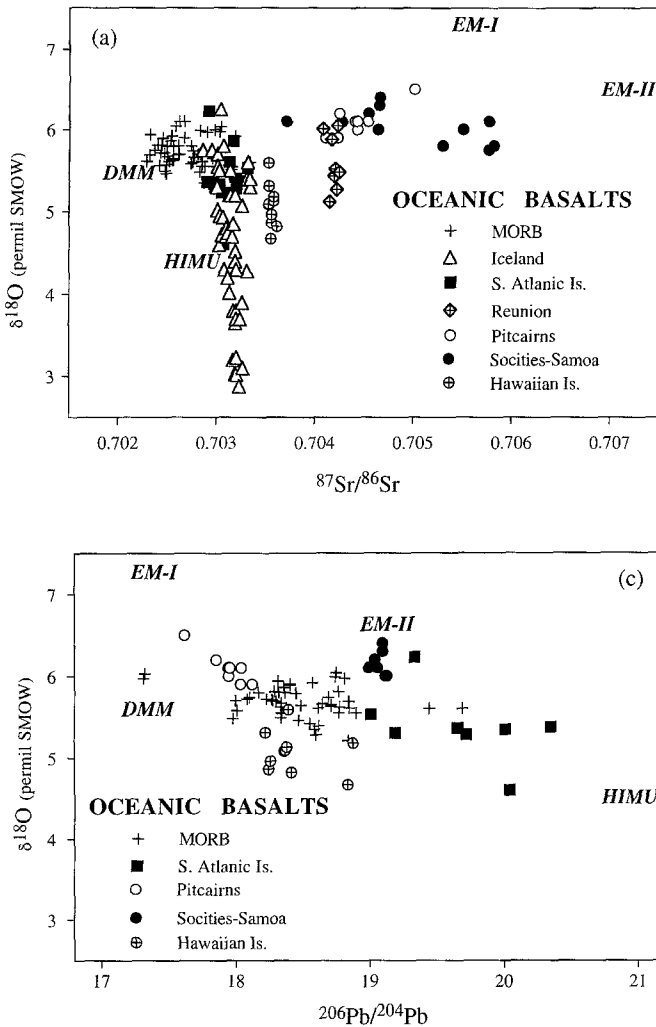


Fig. 9 Plots of $\delta^{18}\text{O}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$ (a), $\delta^{18}\text{O}$ versus $^{143}\text{Nd}/^{144}\text{Nd}$ (b), and $\delta^{18}\text{O}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ (c) for oceanic basalts [Atlantic Ocean: Iceland and S. Atlantic Islands (Canaries, Ascension, Cameroon Line), Indian Ocean: Reunion, and Pacific Ocean: Pitcairns, Societies-Samoa, and Hawaiian Islands] illustrating that the mantle components defined on the basis of radiogenic isotope variations are also reflected in terms of oxygen-radiogenic isotope systematics. For radiogenic isotopes, the positions of the DMM, HIMU, EM-I, and EM-II mantle source components are based upon Figs. 7 and 8; the O-isotope position has been arbitrarily selected to lie close to the extreme points of the basalt data arrays. See text for discussion. The smaller number of localities and data points in (b) and (c) relative to (a) is a reflection of incomplete radiogenic isotopic analyses for the suites of samples in the O-isotope database.

(1993) in terms of 2-component mixing within the mantle between a MORB-like end-member ($^{87}\text{Sr}/^{86}\text{Sr} < 0.704$, $^{143}\text{Nd}/^{144}\text{Nd} > 0.5128$, $^{206}\text{Pb}/^{204}\text{Pb} > 18.25$, and $\delta^{18}\text{O} = + 5.7\text{‰}$) and a plume component ($^{87}\text{Sr}/^{86}\text{Sr} > 0.7055$, $^{143}\text{Nd}/^{144}\text{Nd} < 0.51245$, $^{206}\text{Pb}/^{204}\text{Pb} > 17.5$, and $\delta^{18}\text{O} > + 15\text{‰}$) and is interpreted to reflect the recycling of ancient, ^{18}O -enriched sedimentary crustal material into the mantle. The lack of similar, well-defined O-Sr and O-Pb isotopic co-variations for EM-II basalts may indicate that the mixing process between depleted and enriched mantle components was much older for the EM-II than for the EM-I component (Woodhead et al. 1993). Alternatively, it may reflect a crustal recycling process which does not bear a first-order relationship to subduction, but instead reflects the introduction into the mantle, through a process such as lithospheric delamination, of ^{18}O -enriched crustal material in which both O-isotope and radiogenic isotope ratios are elevated but not strongly coupled (Kempton and Harmon 1992). Or, it may be indicative of a different type of crustal contaminant. Given the large difference in O-isotopic composition between pelagic and terrigenous sediments ($\delta^{18}\text{O}_{\text{PS}} = > + 18\text{‰}$ vs $\delta^{18}\text{O}_{\text{TS}} = \text{ca.} + 8$ to $+ 12\text{‰}$), the difference in $^{18}\text{O}/^{16}\text{O}$ ratio observed for

EM-I OIB relative to EM-II OIB may reflect a pelagic versus terrigenous sediment origin for these enriched components, as recently advocated by Weaver (1991) and Chauvel et al. (1992). Addressing this question through detailed O-isotope studies of minerals from OIB of different type, continental mantle xenoliths, and obducted ophiolite massifs is a high priority for future research.

Highly radiogenic Pb is a characteristic of OIB from such diverse islands as Tubuaii, Mangaia, St. Helena, Ascension, and the Cameroon Line (White 1985; Zindler and Hart 1986; Halliday et al. 1988) and, in conjunction with relatively unradiogenic Sr, is the basis for the mantle component termed HIMU. As seen in Figs. 8 and 9, this component is also characterized by $\delta^{18}\text{O}$ values which extend as low as ca. $+ 4.5$ to $+ 5\text{‰}$.

Of the OIB localities included in the database, Hawaii is atypical, in that its O-radiogenic isotope systematics can not readily be explained in terms of the conventional DMM-EM-HIMU mantle components that circumscribe the data arrays for the other OIB suites (Figs. 7–9). There is strong evidence to indicate that the ^{18}O depletion is a primary feature of the mantle source, rather than a secondary feature reflecting the introduction of altered oceanic or volcanic crust into the basaltic magmas. Be-

cause water is the main hydrogen reservoir in any water-rock system, the introduction of even small amounts of seawater or seawater-equilibrated material into a basaltic magma will cause a major shift in H-isotope composition towards oceanic δD values before any shift in O-isotope composition has occurred. In their study of all of the major Hawaiian volcanoes, Garcia et al. (1989) observed mantle δD values for glass rims on submarine pillow basalt lavas and no relationship between O- and H-isotope compositions, indicating the low $\delta^{18}O$ values measured are primary. This evidence is supported by the temporal trend in O-isotope variation from ca. + 5.5‰ for the older volcanoes (Mauna Kea, Hualalai, and Mauna Loa) to $\leq + 5‰$ for the younger volcanoes (Kilauea and Loihi), which parallels shifts from less radiogenic to more radiogenic Sr- and Pb-isotopic compositions (Staudigel et al. 1984) and is opposite to the $^3\text{He}/^4\text{He}$ trend observed for Hawaii (Kurz et al. 1983; Kurz and Kammer 1991). Garcia et al. (1993) view this latter feature as a reflection of the interplay between plume and non-plume source components to Hawaiian magmatism. In a global context, it implies the existence of multiple sources or reservoirs in the mantle which contribute to the low $^{18}\text{O}/^{16}\text{O}$ ratios observed for some oceanic intraplate magmatic centres, one expressed at Hawaii that has a radiogenic isotope character akin to PREMA (Wörner et al. 1986; Zindler and Hart 1986) and $\delta^{18}O$ values of ca. + 4 to + 4.5‰, and a second exhibiting the HIMU characteristics with slightly higher $\delta^{18}O$ values of ca. + 5‰.

III. Iceland

Iceland is unique in that it is a manifestation of the superposition of a mantle plume upon an active mid-ocean ridge system. Here, the Neovolcanic Zone of Iceland consists of three distinct elements: (1) the rift and fracture zones characterized by tholeiitic basalt swarms which are a sub-aerial expression of the Mid-Atlantic Ridge; (2) the off-rift alkalic and transitional magmatism considered to be a product of the Iceland plume; (3) large central volcanoes that frequently erupt large amounts of silicic lava (Jakobsson 1979).

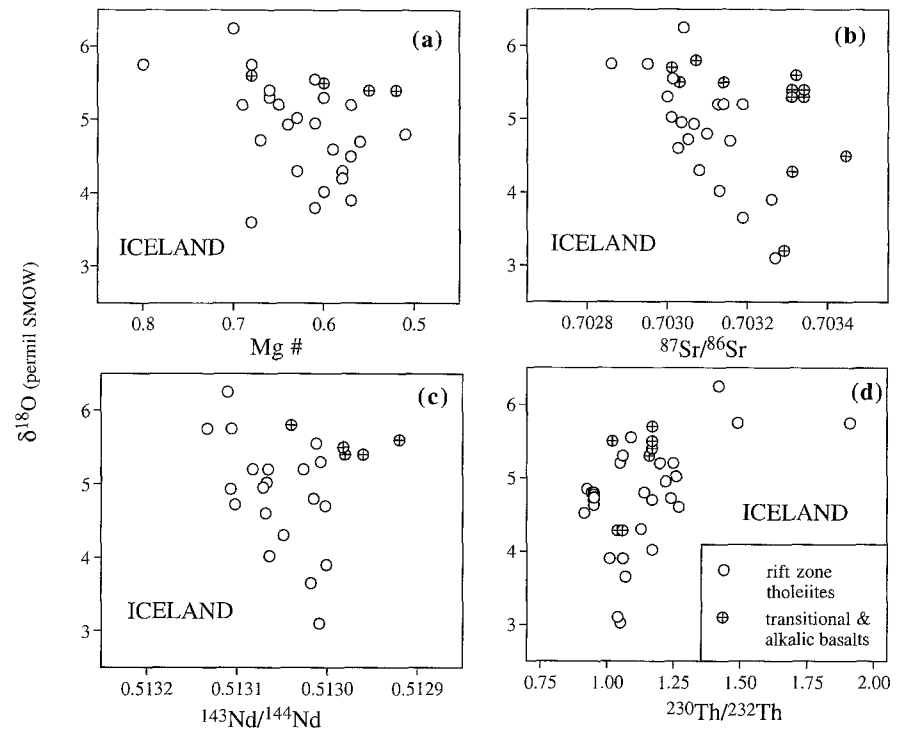
Overall, Iceland is dominated by low $\delta^{18}O$ values. Ranging from + 2.9 to + 6.2‰ (Table 1), $\delta^{18}O$ values for the 104 Icelandic basalt samples in the screened database extend from the upper limit of the MORB range to the lowest $^{18}\text{O}/^{16}\text{O}$ ratios observed for any oceanic locality. The ^{18}O distribution for Iceland (Fig. 5c) is very strongly skewed towards low- $^{18}\text{O}/^{16}\text{O}$ ratios, resulting in the lowest mean value ($\delta^{18}O = + 4.5 \pm 0.8‰$) for the 11 oceanic localities listed in Table 2. Low- $\delta^{18}O$ values, outside the MORB range and extending down to + 3.6‰, are observed for primary basalts with Mg# expected of unfractionated mantle partial melts, as well as for highly fractionated Fe-Ti basalts. In terms of the three major volcanic settings, there is a systematic variation in $\delta^{18}O$ ranges and averages from transitional and alkalic basalts having $\delta^{18}O = + 4.1$ to + 5.7‰ ($\bar{x}_{18} = + 5.1 \pm 0.7‰$),

to rift tholeiites with $\delta^{18}O = + 3.6$ to + 6.2‰ ($\bar{x}_{40} = + 4.8 \pm 0.6‰$), and to basalts from central volcanoes with $\delta^{18}O = + 2.9$ to + 5.1‰ ($\bar{x}_{42} = + 3.9 \pm 0.6‰$). In all three cases, $\delta^{18}O$ values that fall within the MORB range of + 5.4 to + 6.2‰ comprise a minority of the sample population. These details, together with the observations by Muehlenbachs et al. (1974) that: (1) phenocryst-groundmass ^{18}O fractionations are of the magnitude expected at basaltic temperatures; (2) individual basalt flows on Iceland are internally homogeneous with respect to their O-isotopic composition but $\delta^{18}O$ values display a local-scale spatial variability that is uncommon on other ocean islands, raises the fundamental question of whether the sub-Iceland mantle might be of anomalous O-isotope character.

Figure 10 portrays O-isotope variations for Icelandic basalts as a function of bulk chemistry and radiogenic isotope composition. As noted above, basalts with Mg# between 75–68 have $\delta^{18}O$ values which range from + 3.6 to + 6.2‰ (Table 1) and $^{18}\text{O}/^{16}\text{O}$ ratios for these mafic basalts exhibit no systematic co-variation with either bulk composition or radiogenic isotopic ratios. The rough triangular distribution of data on O-Sr and O-Nd plots observed by Hémond et al. (1994) for all Icelandic lavas is preserved in our larger dataset and a similar feature is observed in the $\delta^{18}O$ - $^{232}\text{Th}/^{230}\text{Th}$ plot. These relationships can be explained by petrogenetic models involving mixing between multiple mantle components and ^{18}O -depleted Icelandic crust (Muehlenbachs et al. 1974; Condomines et al. 1983; Hémond et al. 1988; Nicholson et al. 1991; Sigmarsson et al. 1991a,b). The most recent models (Sigmarsson et al. 1992b,c; Hémond et al. 1993) envisage an AFC process in which evolved basalts are produced as the result of contamination of parental mafic tholeiites and alkali basalts by hydrothermally altered basaltic crust within large, convecting magma reservoirs at the crust-mantle boundary region.

As indicated on Fig. 10, the depleted end-member has $\delta^{18}O = \text{ca. } 6.0‰$, $^{87}\text{Sr}/^{86}\text{Sr} = \text{ca. } 0.7028$, $^{143}\text{Nd}/^{144}\text{Nd} = \text{ca. } 0.51322$, and $^{232}\text{Th}/^{230}\text{Th} > 1.80$, whereas the enriched end-member is characterized by $\delta^{18}O = \text{ca. } 5.5‰$, $^{87}\text{Sr}/^{86}\text{Sr} = \text{ca. } 0.7034$, $^{143}\text{Nd}/^{144}\text{Nd} = \text{ca. } 0.51290$, and $^{232}\text{Th}/^{230}\text{Th} = \text{ca. } 1.25$, whereas the source of the low- ϵ_{Nd} , enriched alkali basalt end-member is the Iceland plume (Schilling 1973; Hémond et al. 1994). The origin of the high- ϵ_{Nd} , depleted tholeiitic end-member is less certain; both MORB-source upper mantle (Langmuir et al. 1978) and a second plume component (Elliott et al. 1991) have been advocated. One explanation for the picrites and tholeiitic basalts with high Mg# and very low $\delta^{18}O$ values is a third, more ^{18}O -depleted mantle component having $\delta^{18}O \leq 4.0‰$, which also resides within the Iceland plume. Kurz et al. (1985) have interpreted the high-He concentrations and variable $^3\text{He}/^4\text{He}$ ratios for Icelandic lavas in terms of mixing between an undegassed plume source and local depleted upper mantle considered diagnostic of a plume source. If the plume was also characterized by a low- ^{18}O signature, then it could be viewed as belonging to the

Fig. 10 Plots of $\delta^{18}\text{O}$ versus Mg# (a), $\delta^{18}\text{O}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$ (b), $\delta^{18}\text{O}$ versus $^{143}\text{Nd}/^{144}\text{Nd}$ (c) and $\delta^{18}\text{O}$ versus $^{230}\text{Th}/^{232}\text{Th}$ (d) for basalts from Iceland. See text for discussion



low- ^{18}O branch in Figs. 7 and 9 that is largely defined by Hawaii. However, such an hypothesis is difficult to establish with any degree of certainty because of the pervasive effects of crustal contamination on Iceland. Given that most of the low- ^{18}O mafic tholeiites tend to have $^{232}\text{Th}/^{230}\text{Th}$ ratios less than the ca. 1.2 value considered diagnostic of primary mantle melts on Iceland (Sigmarsson et al. 1992c) and that low- $\delta^{18}\text{O}$ values for basalts tend to be correlated with low- $^3\text{He}/^4\text{He}$ ratios (Condomines et al. 1983), the case for contamination producing the very low $^{18}\text{O}/^{16}\text{O}$ ratios is strong, but a plume source for some of the O-isotope variability in at least the highly magnesian basalts remains a possibility.

IV. Subduction-related basalts

Basaltic lavas erupted in fore-arc trough and back-arc basin settings (FATB-BABB) are considered as a separate group because these are environments related to plate convergence and subduction in which the basalts erupted tend to be transitional in geochemical character between MORB and IAB (Saunders and Tarney 1989). The O-isotope character of FATB-BABB reflects these genetic links. As a group, FATB-BABB are distinguished from MORB by their bias towards higher $^{18}\text{O}/^{16}\text{O}$ ratios, average $\delta^{18}\text{O}$ value = $+5.9 \pm 0.3\text{‰}$ (Table 1), and correlated oxygen - radiogenic isotope variations (Fig. 5d).

Oceanic arc basalts (OAB) are the most ^{18}O -enriched group of the oceanic basalts, with a mean $\delta^{18}\text{O}$ value of $+6.1 \pm 1.0\text{‰}$ (Table 1). However, the large $\delta^{18}\text{O}$ range of $+5.3$ to 7.5‰ for OAB (Fig. 5e) is almost certainly not typical of this class of subduction-associated basalts.

This is because the majority of samples in the database come from the island of Grenada in the Lesser Antilles, which has a peculiar tectonic setting and geochemical character with respect to the remainder of the arc. Both the lowest and highest $\delta^{18}\text{O}$ values for OAB are observed within the "C-series" basalts of Grenada, one of the rare sites where a sodic alkaline magmatic suite erupts in an arc setting (Arculus 1976). Thus, the atypically large $\delta^{18}\text{O}$ variation observed for Grenada basalts is likely a reflection of the unusual geologic situation of this island in a region of hinge faulting at the edge of a subduction zone. Consideration of OAB from only the Mariana-Volcano-Izu, New Hebrides, and West Aleutian arcs of the western Pacific (Ito and Stern 1985/86; Woodhead et al. 1987; Singer et al. 1992; Harmon and Hoefs 1993) yields a slightly lower average $\delta^{18}\text{O}$ value of $+6.0 \pm 0.3\text{‰}$, which is comparable to that for FATB-BABB (Table 1), and a much reduced range of $\delta^{18}\text{O}$ variation from only $+5.6$ to $+6.5\text{‰}$.

As expected from their continental setting, with the enhanced possibilities for magma-crust interaction during ascent, continental arc basalts (CAB), on average, are both more ^{18}O rich in character and more variable in terms of their O-isotope composition than their oceanic counterparts (Fig. 5f). A clear indication of the role of continental crust in the petrogenesis of basalts in destructive plate margin settings is provided by the continuum of subduction-related basalts. The FATB-BABB, which display relatively modest degrees of ^{18}O enrichment, have a mean $\delta^{18}\text{O}$ value of $+5.9\text{‰}$, as compared with $\delta^{18}\text{O} = \text{ca. } +6.0\text{‰}$ for OAB and $\delta^{18}\text{O} = +6.2\text{‰}$ for CAB (Table 1). The overall tendency towards progressive ^{18}O enrichment from FATB-BABB to OAB to

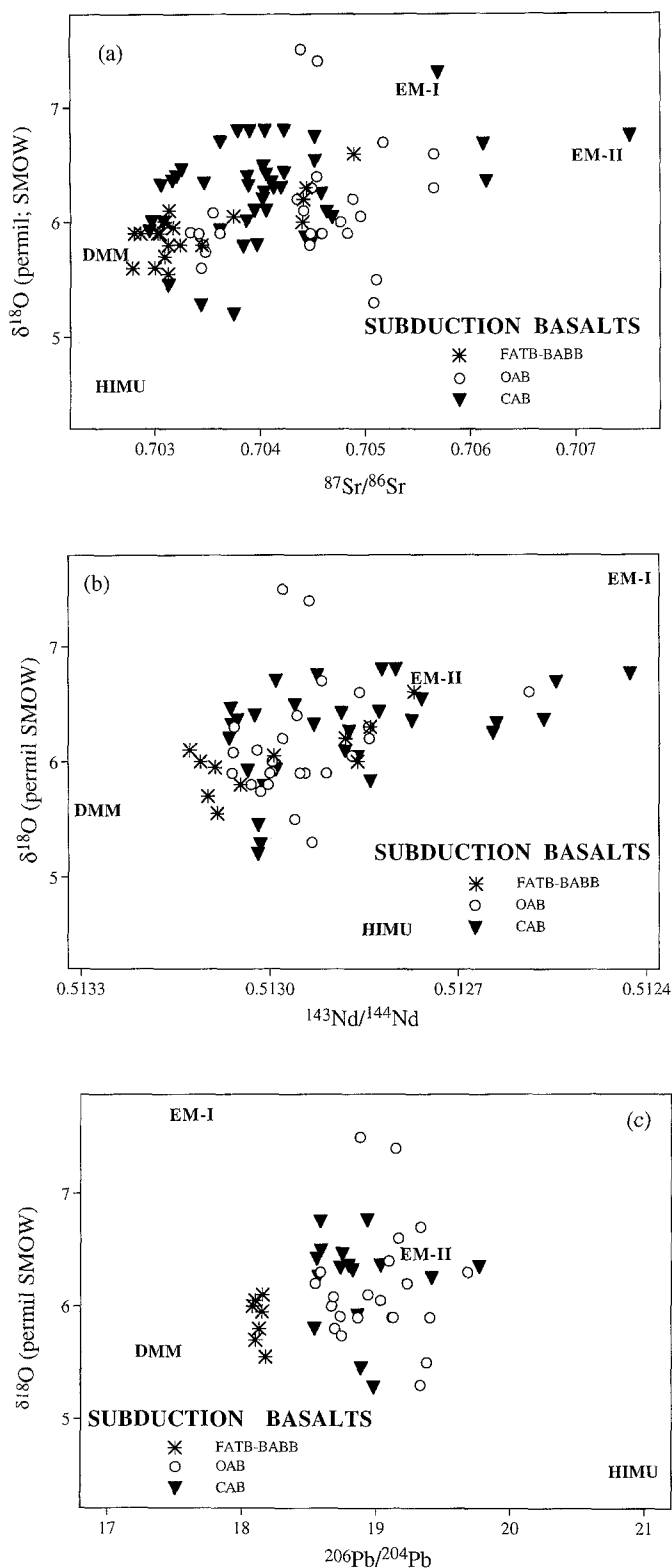


Fig. 11 Plots of $\delta^{18}\text{O}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$ (a), $\delta^{18}\text{O}$ versus $^{143}\text{Nd}/^{144}\text{Nd}$ (b), and $\delta^{18}\text{O}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ (c) analogous to Fig. 9 for basalts associated with subduction. See text for discussion

CAB (Fig. 5d–f) can be attributed to the interaction of parental basalt magmas in oceanic arcs with the volcanic arc basement (Davidson and Harmon 1989) and in continental arcs to interactions within the continental crust (Matsuhisa 1979; Harmon et al. 1981, 1984; Ellam and Harmon 1990).

Figure 11 presents $\delta^{18}\text{O}$ - $^{87}\text{Sr}/^{86}\text{Sr}$, $\delta^{18}\text{O}$ - $^{143}\text{Nd}/^{144}\text{Nd}$, and $\delta^{18}\text{O}$ - $^{206}\text{Pb}/^{204}\text{Pb}$ plots for basalts related to subduction (FATB-BABB, OAB, and CAB). In general, the distribution of data in Fig. 10 is similar to that in Fig. 9 for oceanic basalts, with two exceptions: (1) the tendency for higher $\delta^{18}\text{O}$ values for OAB and CAB relative to FATB-BABB, accompanied by strong ^{87}Sr enrichment and ^{143}Nd depletion; (2) the lack of a discernible HIMU trend for subduction basalts.

We consider the base-level $\delta^{18}\text{O}$ value of ca. + 5.9 to + 6.2‰ for arc basalts to be a direct expression of intramantle enrichment processes, related to the subduction and recycling of hydrothermally altered oceanic crust and entrained pelagic sediments, that lead to arc magmatism (e.g. Nicholls and Ringwood 1973; Sekine and Wylie 1982; Morris et al. 1990). In the same manner that arc basalts approach but rarely attain, the $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of N-type MORB, we consider this small but characteristic ^{18}O enrichment of ca. + 0.2 to 0.3‰ observed for arc basalts worldwide compared to MORB (Table 1) to be an intrinsic signature of magmatism associated with subduction. Striking oxygen-radiogenic isotope co-variations are observed for FATB-BABB in all three plots of Fig. 11. That the $\delta^{18}\text{O}$ - $^{87}\text{Sr}/^{86}\text{Sr}$ and $\delta^{18}\text{O}$ - $^{143}\text{Nd}/^{144}\text{Nd}$ arrays have essentially the same shape and extent as observed for EM-I type OIB (Figs. 6, 7), and that the $\delta^{18}\text{O}$ - $^{206}\text{Pb}/^{204}\text{Pb}$ trends for the two basalt types are superimposed (Fig. 8), implies a common, subduction-related origin for both features. Mass balance considerations require the addition of only small proportions (ca. 1–3%) of an ^{18}O -enriched, sediment-derived fluid to a depleted, MORB-type mantle source region to explain the isotope systematics observed for subduction-related basalts (e.g. Ito and Stern 1985/86; Woodhead et al. 1987).

This selective transfer of an ^{18}O -enriched fluid component from the descending slab to the convecting mantle wedge is a unique aspect of convergent plate margin processes, and it is at this step in the arc magmatogenesis process at which the diagnostic chemical and isotopic signatures which characterize arc magmas are generated (DePaolo and Wasserburg 1977; Hawkesworth et al. 1977; Sun 1980; Gill 1981; McCulloch and Gamble 1991). Arc parental magmas have a relatively uniform ^{18}O signature, but a variably enriched radiogenic isotope character, and it is upon this base-level isotopic signature that any subsequent intra-arc crustal interaction is superimposed. As exemplified by the Andean arc, crustal contamination of primary arc basalt parental magmas can lead to either higher (James 1982; Harmon et al. 1984; Hildreth and Moorbath 1988; Davidson et al. 1990) or lower (Grunder 1987) $^{18}\text{O}/^{16}\text{O}$ ratios in derivative,

evolved lava types, depending on the O-isotope character of the assimilated crust.

Examples of both types of contamination are displayed in Fig. 11. The few samples with the low $\delta^{18}\text{O}$ values of $< + 5.5\text{‰}$ record contamination with crust modified by hydrothermal alteration at high temperature (Taylor 1977). By contrast, OAB and CAB displaying $\delta^{18}\text{O}$ values $\gg + 6\text{‰}$ comprise a majority of the basalts plotted in Fig. 11. In general, this ^{18}O enrichment is accompanied by enhanced $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ ratios and diminished $^{143}\text{Nd}/^{144}\text{Nd}$ ratios, which results in these samples plotting in positions outside the region in Fig. 11 circumscribed by the four mantle components. Both features are readily explained in terms of contamination by isotopically evolved and compositionally heterogeneous ^{18}O -rich crust, either during the ponding and differentiation of primary arc partial melts under open system conditions at the crust mantle boundary, as envisaged by Hildreth and Moorbath (1988) and Davidson et al. (1990) or through AFC and magma mixing processes during subsequent transit or storage in magma reservoirs at high crustal levels (Harmon et al. 1981; James 1982; McBirney et al. 1987; Ellam and Harmon 1990)

V. Continental basalts

Intraplate basalts from continental rift zones (CIB) have a mean $\delta^{18}\text{O}$ value of $+ 6.1\text{‰}$ and exhibit a large range of ^{18}O variation from $+ 4.6$ to $+ 8.3\text{‰}$ (Table 1, Fig. 5g). Whereas alkalic basalts tend to be more ^{18}O rich than tholeiitic basalts in the case of OIB, the converse is true for CIB, ($\delta^{18}\text{O} = + 6.1 \pm 0.6\text{‰}$ for 144 alkalic basalts versus $\delta^{18}\text{O} = + 7.1 \pm 0.7\text{‰}$ for 20 tholeiitic basalts). The $\delta^{18}\text{O}$ range for tholeiitic CIB extends upwards from $+ 5.9\text{‰}$ to over $+ 8\text{‰}$, with $\delta^{18}\text{O}$ values tending to increase with decreasing with Mg#, implying that crustal contamination during fractionation has played an important role in the petrogenesis of these lavas (Harmon et al. 1987). However, it is unlikely that such an explanation can be extended to explain the ^{18}O enrichment observed in some alkali olivine basalts and basanites with Mg# > 67 that have $\delta^{18}\text{O}$ values exceeding $+ 7.0\text{‰}$.

Basaltic lavas from continental flood basalt provinces (CFB) are very poorly represented in our database (Fig. 5 h) because the Columbia River - Snake River Plateau in the northwestern US is the only Neogene example of this type of basaltic volcanism. The 18 entries have a mean $\delta^{18}\text{O}$ value of $+ 5.6 \pm 0.6\text{‰}$ (Table 1), but the population is dominated by samples from the Yellowstone caldera region with variably low $\delta^{18}\text{O}$ values, which Hildreth et al. (1991) argue have been contaminated as a result of assimilation of hydrothermally altered, low- ^{18}O crust. The few tholeiitic basalts of the Columbia River Plateau (Carlson et al. 1981) included in the database have $^{18}\text{O}/^{16}\text{O}$ ratios which extend from the MORB average to $\delta^{18}\text{O}$ values above $+ 6.5\text{‰}$. Together, these features suggest that crustal interaction, as envisaged by Cox (1980) and Arndt and Christensen (1992), may be a

fundamental aspect of the process through which CFB are generated.

Highly potassic, leucite-bearing Si-undersaturated basaltic lavas are found in many arcs worldwide and are a common occurrence throughout the subduction-related Neogene volcanic provinces of the Mediterranean, particularly the central Italian peninsula. Striking features of these potassic lavas are their high water contents and strong enrichment in ^{18}O (Fig. 5i), which are attributes of even the most primitive lavas erupted. For example, the 33 basalts which passed our screening criteria for low water content exhibit a $\delta^{18}\text{O}$ range from $+ 6.4$ to $+ 11.4\text{‰}$ (Table 1). Much has been written about the high- ^{18}O character of the Italian potassic magmatism (e.g. Taylor and Turi 1976; Turi and Taylor 1976; Taylor et al. 1979; Holm and Munksgaard 1982; Ferrara et al. 1985, 1986; Rogers et al. 1986; Varenkamp and Kalamarides 1989; Turi et al. 1991), and the petrogenesis of these unusual lavas continues to be vigorously debated. Ultimately, the high- ^{18}O component in the strongly mafic potassic magmas must be of crustal origin. The critical question centres upon whether the ^{18}O -enriched character of the Italian basalts is a characteristic present within the upper mantle beneath central Italy, and one which subsequently has been inherited during partial melting, or whether it is a feature largely acquired within the continental crust during magma storage and transit. Given the tectonic complexity and unusual geological history of Italy during the Neogene, it is likely that both processes have contributed in substantial measure towards the generation of this unusual signature. Therefore, the model of Turi et al. (1991) – in which parental potassic basalts are envisaged to have been derived from two distinct, ^{18}O -enriched lithospheric domains, with evolved potassic lavas subsequently produced through extensive interaction with a thick, extremely high ^{18}O metasedimentary crust – is quite plausible.

Figure 12 presents plots for continental basalts analogous to those in Figs. 9 and 11 for oceanic and subduction-associated basalts. On these diagrams, the oceanic basalt field from Figs. 7 and 8 is indicated by the shaded field. The $\delta^{18}\text{O}$ - $^{87}\text{Sr}/^{86}\text{Sr}$ and $\delta^{18}\text{O}$ - $^{143}\text{Nd}/^{144}\text{Nd}$ plots for CIB exhibit the same general O-Sr and O-Nd isotope relationships as OIB. The low- ^{18}O trough observed in the O-radiogenic isotope plots for OIB is present in the $\delta^{18}\text{O}$ - $^{87}\text{Sr}/^{86}\text{Sr}$ plot for CIB, but very subdued in the $\delta^{18}\text{O}$ - $^{143}\text{Nd}/^{144}\text{Nd}$ plot because of the fact that far fewer of the samples in the O-isotope dataset have been analysed for Nd-isotopes than for Sr-isotopes. The EM-II branch is clearly identified on the O-Nd and O-Pb plots, whereas the EM-I branch is absent on all plots of Fig. 12. A striking feature of the figure is the strong ^{18}O enrichment at variable $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios that characterizes CIB of Patagonia and the Southwest US; both regions where there is a history of recent subduction. Another distinctive feature of Fig. 12 is the O-Pb plot in which $\delta^{18}\text{O}$ variations for CIB are observed to be inversely correlated with $^{206}\text{Pb}/^{204}\text{Pb}$ ratios. Together, these features imply a major role for the

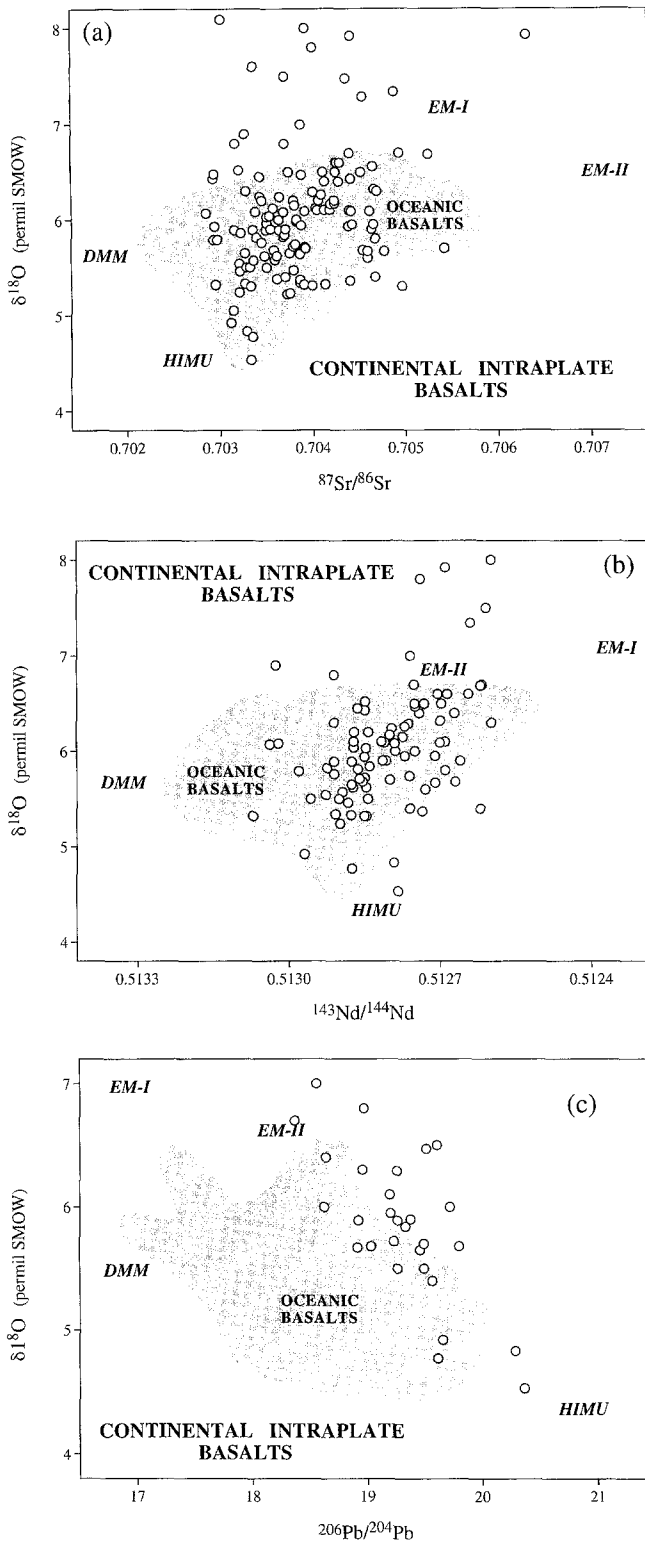


Fig. 12 Plots of $\delta^{18}\text{O}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$ (a), $\delta^{18}\text{O}$ versus $^{143}\text{Nd}/^{144}\text{Nd}$ (b), and $\delta^{18}\text{O}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ (c) analogous to Fig. 9 for continental intraplate basalts. The shaded field in each diagram denotes the oceanic basalt field from Fig. 9. See text for discussion

HIMU reservoir in the mantle source for CIB and also suggest that both asthenospheric and lithospheric mantle contribute to CIB magmagenesis.

Summary and concluding remarks

This paper has presented an overview of available O-isotope data for Neogene basalts (historic lavas, submarine glasses, and lavas with 0.75 wt% H_2O), which comprise 127 analyses of MORB and 616 analyses of oceanic and continental basalts. The major conclusions we draw from our consideration of this data set are that: (1) the mantle source of basaltic magmas is heterogeneous with respect to its O-isotope composition on a significant scale; (2) this isotopic variation has both high- ^{18}O and low- ^{18}O manifestations which are systematic and consistent with radiogenic isotope differences displayed by basalts erupted in different tectonic settings. Up to this point, the focus of this paper has been largely descriptive. In the concluding discussion which follows, we consider the basalt oxygen-radiogenic isotope systematics in light of ideas regarding the origin of mantle compositional heterogeneities and briefly put these observations into a unified, global geodynamic context.

The modern plate tectonic theory introduced the basic idea that the oceanic crust is recycled from Earth's surface back into the mantle. The concept of hot spots in the plate tectonic model led to the notion of fixed mantle plumes and this, in turn, developed into the ideas that subducted oceanic crust, oceanic lithosphere, and/or delaminated subcontinental lithosphere might be stored within the deep mantle for very long (≥ 1 billion year) periods of time before reappearing via the plumes that lead to oceanic, as well as some manifestations of continental, intraplate magmatism.

The most popular model for the mantle is a 2-layer structure (DePaolo and Wasserburg 1976) in which an undepleted lower mantle is overlain by a depleted upper mantle from which the continental crust has been extracted (Hofmann 1989). Portions of this depleted upper mantle have been affected by enrichment processes related to subduction (Hofmann and White 1980; Chase 1981; Ringwood 1982; White and Hofmann 1982; Weaver 1991) and lithospheric delamination (McKenzie and O'Nions 1983; Kay and Kay 1990). Recently, this general model has been expanded upon by Kesson and Ringwood (1989) and Ringwood (1990), who envisage a scenario for slab-mantle interaction in which geochemically enriched domains are generated at two different locations within the upper mantle during subduction. The first occurs at depths between 150–300 km, and involves the partial melting of eclogitic former oceanic crust and the transfer of this melt from the descending slab into the overlying asthenosphere and continental lithosphere. The second occurs in the vicinity of the 650 km seismic discontinuity, where former subducted lithospheric mantle is considered to rest buoyantly upon

the lower mantle and convectively disperse and hybridize within the deepest portions of the depleted asthenosphere. Additionally, the enriched sub-arc mantle wedge may itself be viewed as a potential contributor of enriched components to the depleted upper mantle, either through delamination resulting from tectonic shortening and crustal thickening during subduction (Kay and Kay 1990) or through post-orogenic accretion of an arc onto a continent (Kempton and Harmon 1992). This co-location of both depleted and enriched mantle reservoirs together within the upper mantle can readily explain the high- and low- $^{18}\text{O}/^{16}\text{O}$ ratios observed for basalts when considered in the context of subduction.

As a process that has operated throughout the Phanerozoic, and likely since the Archean, subduction provides the most plausible mechanism for introducing large masses of both high- ^{18}O and low- ^{18}O components into the deep mantle. The descending slab at subduction zones comprises mantle lithosphere, oceanic crust, and sediment. The oceanic crust has been weathered at low temperature in its uppermost parts and subjected to seawater hydrothermal alteration to depths of several kilometres, so that it is variably enriched or depleted in ^{18}O depending on temperature and extent of water-rock interaction. Several O-isotope studies have demonstrated that hydrothermally altered oceanic crust, as exposed in obducted ophiolite complexes (Spooner et al. 1974; Gregory and Taylor 1981; Cocker et al. 1982; Schiffman et al. 1984) and sampled by the DSDP and ODP (Alt et al. 1986; Kempton et al. 1991), varies systematically in its O-isotopic composition from $\delta^{18}\text{O}$ values $> +9\text{‰}$ in the pillow basalt-sheeted dike complex to $\delta^{18}\text{O}$ values $< +4\text{‰}$ in the deep gabbroic section. This difference in O-isotope composition as a function of depth within the oceanic crust is a direct result of increasing temperature of seawater hydrothermal alteration from $< 150^\circ\text{C}$ for the pillow basalt section to $> 300^\circ\text{C}$ for the gabbroic section (Muehlenbachs and Clayton 1972; Ito and Clayton 1983; Alt et al. 1986; Kempton et al. 1991).

The associated complement of authigenic marine sediments and detrital crustal material is in equilibrium with sea water and characterized by very high $^{18}\text{O}/^{16}\text{O}$ ratios (Magaritz and Taylor 1976; Yeh and Savin 1977; McMurtry and Yeh 1981). Although much of the ^{18}O -rich sediment that is subducted, particularly the thick terrigenous sedimentary wedges on some convergent plate margins (von Huene and Schöll 1991), may be subcreted to the continents, some proportion will find its way into the oceanic crust-sediment package that is subducted. During the first 150 km of descent, the most labile components with high $^{18}\text{O}/^{16}\text{O}$ ratios are removed from the sinking slab through dehydration, decarbonation, and possibly partial melting. These high- ^{18}O fluids play a critical role in the inception of subduction-zone magmatism and are responsible for the unique geochemical signature of arc magmas. Ultimately, this ^{18}O -enriched subduction component will be returned to the crustal sphere through arc magmatism or stored within the mantle

wedge later to become fixed in the subcontinental lithosphere.

As a consequence of this geochemical processing of the subducted slab, there may be a bias in terms of material actually recycled into the deeper mantle towards the low- ^{18}O portion of the hydrothermally altered oceanic crustal section. If this is the case, then the widespread presence of low- ^{18}O domains in the upper mantle is to be expected as an inevitable consequence of subduction, particularly if the melts are extracted from the subducted oceanic crust during subduction (Ringwood 1990) and the residual slab accumulates and is stored either at depth in the upper mantle (Hofmann and White 1982; Ringwood 1982) or in the shallow mantle below the depleted domain (Allegre and Turcotte 1985). In fact, eclogite xenoliths in kimberlites, with their extremely wide range in $\delta^{18}\text{O}$ values from ca. $+2$ to $+9\text{‰}$ (Vogel and Garlick 1970; Jagoutz et al. 1984; Ongley et al. 1987), recently have been considered to be ancient, residual fragments of oceanic crust that reflect just such a history (Neal et al. 1990). Thus, it may be more than just coincidence that the eclogite $\delta^{18}\text{O}$ range is essentially that observed for intraplate basalts generated by uprising mantle plumes.

We have seen that ocean floor basalts (MORB) are quite uniform in O-isotope composition, a feature consistent with large degrees of partial melting of a source of relatively homogeneous bulk composition. Subduction-related basalts are derived from the same depleted mantle source after enrichment by hydrophile components contributed from the subducted slab. Oceanic intraplate basalts are produced through variable degrees of partial melting of compositionally heterogeneous mantle generated through the mixing of possibly five isotopically distinct components. Subduction will recycle both high- ^{18}O sedimentary material and sea-water hydrothermally altered oceanic crust variably enriched or depleted in ^{18}O ; lithospheric delamination will return isotopically heterogeneous continental lithosphere and ^{18}O -enriched mafic lower crust into the upper mantle. All of these isotopically distinct domains may contribute to oceanic intraplate magmatism over different spatial and temporal scales. The HIMU sources are one such example of a low- ^{18}O source and portions of the subcontinental lithosphere may be others (Harmon et al. 1986, 1987). We consider that Hawaii provides evidence of an additional low- ^{18}O mantle reservoir of the deeper type envisaged by Ringwood (1990). In the Hawaiian Islands, ^{18}O -enriched alkalic basalts frequently appear during the early and late stages of oceanic island volcanism, when plume interaction with oceanic lithosphere is likely to be most pronounced, whereas the lower- $\delta^{18}\text{O}$ tholeiitic basalts comprise the voluminous main stage of volcanism and likely derive from the centre of an ascending, deep mantle plume. Iceland may be another example, but this remains to be established. Continental intraplate basalts are likely derived from a smaller subset of these components, as well as small-scale enriched domains present locally within the subcontinental lithosphere.

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