

Low $\delta^{18}\text{O}$ in the Icelandic mantle and its origins: Evidence from Reykjanes Ridge and Icelandic lavas

M.F. Thirlwall^{a,*}, M.A.M. Gee^{a,1}, D. Lowry^a, D.P. Mattey^a, B.J. Murton^b, R.N. Taylor^b

^a Department of Geology, Royal Holloway University of London, Egham, Surrey TW20 0EX, UK

^b Southampton Oceanography Centre, Empress Dock, Southampton SO14 3ZH, UK

Received 9 September 2004; accepted in revised form 1 August 2005

Abstract

Oxygen isotope ratios have been determined using laser fluorination techniques on olivine and plagioclase phenocrysts and bulk glasses from the Reykjanes Ridge and Iceland. $\delta^{18}\text{O}$ in Reykjanes Ridge olivines shows hyperbolic correlations with Sr–Nd–Pb isotope ratios that terminate at $\delta^{18}\text{O} = +4.5\text{‰}$ at compositions almost identical to those of moderately enriched lavas on the Reykjanes Peninsula, Iceland. Samples with low $\delta^{18}\text{O}$ show no indication of contamination by oceanic crust such as elevated Cl/K, and are too deep to have been influenced by meteoric water hydrothermal systems. They cannot represent Icelandic melts contaminated in the crust and transferred laterally along the ridge since fissure systems are strongly oblique to the ridge axis. It follows that Icelandic mantle advected along the ridge has low $\delta^{18}\text{O}$. The hyperbolic $^{143}\text{Nd}/^{144}\text{Nd}$ – $\delta^{18}\text{O}$ correlation appears to be more strongly curved than magma mixing trajectories and suggests that melt fractions are $\sim 4.5\times$ greater and source Nd contents $\sim 9\times$ greater in the mantle at 63°N compared with that at 60°N . Primitive lavas from the Reykjanes Peninsula show linear correlations between olivine $\delta^{18}\text{O}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ or $^{206}\text{Pb}/^{204}\text{Pb}$, extending to $\delta^{18}\text{O}$ of $+4.3\text{‰}$ at $^{143}\text{Nd}/^{144}\text{Nd}$ close to the lowest ratios observed in Icelandic magmas. These correlations cannot be produced by melt mixing or crustal contamination because these would yield strongly hyperbolic trajectories. Lower $\delta^{18}\text{O}$ seen in more evolved samples from the Eastern Rift Zone may reflect crustal contamination, though there is some evidence of a mantle source with lower $\delta^{18}\text{O}$ in eastern Iceland. It is very difficult to explain the low $\delta^{18}\text{O}$ of enriched Icelandic mantle sources on current understanding of mantle and crustal oxygen isotopes. There is no obvious reason why such low- $\delta^{18}\text{O}$ sources should not contribute to other ocean islands. No oceanic crustal lithologies exist that could produce the low- $\delta^{18}\text{O}$ enriched sources by recycling into the mantle, and there is no evidence for changes in $\delta^{18}\text{O}$ of ophiolite suites with time, nor of changes during high-P metamorphism. Low $\delta^{18}\text{O}$ appears to be associated with high $^3\text{He}/^4\text{He}$, and we speculate that this signature may be characteristic of the host mantle into which ocean crust was recycled. © 2005 Elsevier Inc. All rights reserved.

1. Introduction

It has long been known that Icelandic lavas have low $\delta^{18}\text{O}$ compared with most volcanic rocks worldwide, with reported $\delta^{18}\text{O}$ in whole rocks ranging from -0.2 to $+6.3\text{‰}$ relative to SMOW and $\delta^{18}\text{O}$ in olivine from $+3.2$ to $+5.8\text{‰}$ (e.g. Muehlenbachs et al., 1974; Condomines et al., 1983; Hémond et al., 1988; Nicholson et al., 1991;

Sigmarrsson et al., 1991, 1992a,b; Hémond et al., 1993; Harmon and Hoefs, 1995; Gee et al., 1998; Eiler et al., 2000a; Skovgaard et al., 2001; Breddam, 2002; MacLennan et al., 2003; Burnard and Harrison, 2005; Macpherson et al., 2005). By contrast, MORB glasses have $\delta^{18}\text{O}$ within the limited range of $+5.21$ to $+5.81\text{‰}$ ($+5.50 =$ mean of a MORB survey, Eiler et al., 2000b; $+5.45 \pm 0.19 =$ mean and 2sd of 62 fractionation-corrected Atlantic MORB from 33 to 40°N , Cooper et al., 2004), while ocean island basalts apart from Iceland have $\delta^{18}\text{O}$ in olivine from $+4.7$ to $+6.1\text{‰}$ (mostly $+5.0$ to $+5.4$, Eiler et al., 1997). This requires the operation of some unusual processes or unusual sources in Iceland, most commonly identified as assimilation of low- $\delta^{18}\text{O}$ Icelandic crust, with several

* Corresponding author. Fax: +44 1784 471780.

E-mail address: matthewt@gl.rhul.ac.uk (M.F. Thirlwall).

¹ Present address: School of Earth and Geographical Sciences, University of Western Australia, 35 Stirling Highway, Crawley, WA 6009, Australia.

studies showing decreasing $\delta^{18}\text{O}$ with decreasing MgO, indicative of assimilation—fractional crystallization (AFC) processes (e.g., Hémond et al., 1988; Nicholson et al., 1991; Eiler et al., 2000a). This argument has considerable force, because Iceland is unique among ocean islands worldwide in both the abundance of low- $\delta^{18}\text{O}$ magmas, and in the presence of thick (up to 40 km, Darbyshire et al., 2000) crust that has had the opportunity to interact in extensive hydrothermal systems with the low- $\delta^{18}\text{O}$ meteoric water characteristic of high latitudes. Consequently, the near-unique low $\delta^{18}\text{O}$ of Icelandic lavas has been regarded as a function of its unique status as the only major high latitude ocean island and subaerial ocean ridge. This is supported by reports of low- $\delta^{18}\text{O}$ magmas in other sub-Arctic volcanic regions (e.g., Kamchatka and some Aleutian arc lavas, Bindeman et al., 2001, 2004) though in these areas magmas with normal $\delta^{18}\text{O}$ are common, and the low- $\delta^{18}\text{O}$ signature is largely restricted to rhyolites and intermediate compositions.

However, a few studies have questioned this consensus. Muehlenbachs et al. (1974) found significant problems with several crustal interaction mechanisms, but also with their alternate hypothesis that Icelandic mantle had low $\delta^{18}\text{O}$ as a consequence of incorporating hydrothermally altered subducted ocean crust. Harmon and Hoefs (1995) noted that $\delta^{18}\text{O}_{\text{WR}}$ (subscript implies whole rock) as low as +3.8‰ was found in relatively primitive ocean island basalts and suggested that this had to reflect mantle $\delta^{18}\text{O}$ heterogeneity. They accepted models in which contamination changes $\delta^{18}\text{O}$ in Iceland magmas, but felt that a plume source for $\delta^{18}\text{O}$ variation in primitive Icelandic magmas was a possibility. Gautason and Muehlenbachs (1998) noted that olivine and clinopyroxene from Tertiary lavas of the Faroes and NW Iceland had low $\delta^{18}\text{O}$ relative to MORB phenocrysts, and questioned whether this might reflect an unusual mantle composition rather than widespread assimilation processes.

With the advent of simple and precise mineral $\delta^{18}\text{O}$ analysis using laser-assisted fluorination (LF, e.g., Sharp, 1990; Mathey and Macpherson, 1993), $\delta^{18}\text{O}$ values determined on whole rock lavas have been shown to be very susceptible to surficial alteration (e.g., Thirlwall et al., 1996; Downes et al., 2001). However, LF studies of Icelandic lava phenocrysts (Gee et al., 1998; Eiler et al., 2000a; Skovgaard et al., 2001), glasses (Breddam, 2002; Burnard and Harrison, 2005; Macpherson et al., 2005), and cumulate nodules (MacLennan et al., 2003) have continued to show low $\delta^{18}\text{O}$, with values as low as +3.2‰ reported in olivine. There is still much disagreement over the origin of the low $\delta^{18}\text{O}$, with Gee et al. (1998) and Eiler et al. (2000a) concluding that it reflects solely crustal contamination, while Skovgaard et al. (2001), Breddam (2002), MacLennan et al. (2003), Burnard and Harrison (2005), and Macpherson et al. (2005) proposed substantial oxygen isotope heterogeneity in the Icelandic mantle. These differences lead to considerable differences in interpretation of other aspects of the magma chemistry, with Eiler et al. (2000a) regarding

correlations between incompatible-element ratios and $\delta^{18}\text{O}$ at Theistareykir (north Iceland) to be the product of crustal assimilation while Stracke et al. (2003) interpreted the isotopic variation in the same sample set to be exclusively the result of mantle processes.

If oxygen isotope heterogeneity in Icelandic mantle sources can really be disentangled from the effect of crustal contamination by low- $\delta^{18}\text{O}$ crust, it would provide a strong constraint on models of Icelandic basalt genesis. Recent models have suggested a major role for recycled oceanic crust in Icelandic magmatism, with, for example, Chauvel and Hémond (2000) and Skovgaard et al. (2001) proposing contributions from gabbroic and basaltic crust to the sources of incompatible-element depleted and enriched lavas, respectively. Foulger et al. (2005) have even proposed that all Icelandic basalts are derived largely from oceanic gabbro. The oxygen isotope heterogeneity observed in ophiolites and modern oceanic crust (e.g., Stakes and Taylor, 1992) would be expected to generate measurable $\delta^{18}\text{O}$ heterogeneity in Icelandic mantle sources for recycled crustal proportions greater than 10–20%.

An obvious test of mantle versus crustal origin for the low $\delta^{18}\text{O}$ is provided by the Reykjanes Ridge, in which Icelandic mantle mixes with a more depleted mantle source along the ridge. Decreases away from Iceland in La/Sm, $^{87}\text{Sr}/^{86}\text{Sr}$, and $^{206}\text{Pb}/^{204}\text{Pb}$ were originally interpreted as the results of mixing between relatively incompatible-element enriched Icelandic mantle and depleted MORB asthenospheric mantle (e.g., Hart et al., 1973; Schilling, 1973), but high precision Pb–Nd–Sr data reported by Thirlwall et al. (2004) indicate that the main gradient between 60 and 63°N is the consequence of mixing enriched and depleted components of Icelandic mantle. Since the Reykjanes Ridge is at depths of 220–2060 m below sea level over this latitude range, it is highly unlikely that Reykjanes Ridge magmas have interacted with crust that has low $\delta^{18}\text{O}$ as a result of interaction with low- $\delta^{18}\text{O}$ meteoric water. Consequently, if low $\delta^{18}\text{O}$ is observed in Reykjanes Ridge magmas, this should unequivocally determine that the low $\delta^{18}\text{O}$ in Icelandic magmas is at least in part of mantle origin.

In this paper, we present results of a detailed oxygen isotope study of samples dredged from the Reykjanes Ridge axis during the cruise CD80 of RRS *Charles Darwin* (Murton, 1994). High-precision Pb–Sr–Nd data were reported for these samples by Thirlwall et al. (2004), major and trace elements by Murton et al. (2002), and He isotope data by Hilton et al. (2000). We show that the incompatible-element-enriched Icelandic mantle that contributes to the Reykjanes Ridge between 60° and 63°N, and at sporadic locations as far south as 57.5°N, has $\delta^{18}\text{O}_{\text{olivine}}$ of +4.5‰ at most, radically lower than observed in other ocean islands. Incompatible-element-depleted sources along the Reykjanes Ridge have normal mantle $\delta^{18}\text{O}_{\text{olivine}}$ of 5.12 ± 0.10 (2sd, $N = 8$ samples, 13 including replicates). We also report the results of a detailed study of oxygen isotopes in Reykjanes Peninsula lavas, including re-analysis of

the samples reported by Skovgaard et al. (2001). $\delta^{18}\text{O}_{\text{olivine}}$ similar to normal mantle values is only present in incompatible-element-depleted lavas, and the higher ratios reported by Skovgaard et al. (2001) are shown to be inaccurate. Correlations are observed with radiogenic isotope ratios, particularly $^{206}\text{Pb}/^{204}\text{Pb}$. The new oxygen data provide us with information about the bulk composition of enriched Icelandic mantle and constrain the contributions of ^{18}O -depleted and ^{18}O -enriched recycled ocean lithosphere to Icelandic magmatism.

2. Analytical techniques

Depending on phenocryst size, rock chips were crushed to 250–1000 μm fragments using a steel plunger and anvil in a 20 mm diameter tube. After sieving and washing in $>18.2\text{ M}\Omega$ water, a hand magnet was used to remove a magnetic fraction, which included most of the lava groundmass in holocrystalline samples. Phenocryst separates were then hand-picked from the less magnetic fraction. For glassy rocks, handpicking alone was used as the glass was not removed by the magnet. $1.7 \pm 0.3\text{ mg}$ of sample was weighed into 16-position Ni sample holders, including four standards (usually positions 1, 2, 8, and 15). The standards were in most cases San Carlos I olivine, which has a preferred $\delta^{18}\text{O} = +4.88\text{‰}$ (calibrated against NBS30 biotite = $+5.10\text{‰}$ relative to V-SMOW, Matthey and Macpherson, 1993), though some were San Carlos II olivine, which has a preferred $\delta^{18}\text{O} = +5.22\text{‰}$, derived by extensive cross-calibration to San Carlos I olivine. Most analyses were performed on the Royal Holloway Mark II laser fluorination system (LaserPrep, Matthey, 1997) attached to a VG Optima mass spectrometer, in which all procedures from fluorination to analysis of the cleaned released O_2 gas were carried out under computer control. A few earlier analyses, primarily for Iceland, used the Mark I manual laser fluorination system (as described by Matthey and Macpherson, 1993) attached to a VG Prism mass spectrometer. These tend to show slightly worse reproducibility than data obtained using the Mark II system, but they have been included in the final sample mean $\delta^{18}\text{O}$ since inclusion does not change the mean by more than 0.08‰ .

Oxygen yields were calculated assuming 46% O in plagioclase (equivalent to $\sim\text{An}_{70-100}$), and oxygen contents of equilibrium olivines and glasses calculated from their major element compositions. Yields mostly ranged from 93 to 102%, and particular care was taken to reproduce $\delta^{18}\text{O}$ in samples with yields at the low end of this range. A few samples with yields of 89–91% are reported in the tables, mostly where there was insufficient sample for replication. In most cases, these lower yields probably reflect loss of occasional grains of finer separates or oxide inclusions. For each day of analysis, data were corrected to the San Carlos olivine results, usually based on the mean of the last three of the four standard runs, which were usually reproducible to $<0.2\text{‰}$. The standard run in the first position on each day is regarded as sacrificial and is used to prime the

sample tray and clean-up line, and contaminants thereby released affected the standard in position two on three occasions. The correction ranged from $+0.21$ to $+0.75\text{‰}$ using a diamond window to the laser cell (14 sessions between 2003 and 2005, and in 1998) and from -0.09 to -0.30 using a BaF_2 window (6 sessions in 1999). The daily correction was usually constant to $\pm 0.1\text{‰}$ over a period of several days. Most samples were analyzed in duplicate and commonly triplicate. Replicate samples were analyzed on more than one day, and using different laser cell windows, so that uncertainty in the correction to San Carlos olivine could be incorporated into estimated sample uncertainty. For example, the four replicate analyses of the isotopically normal olivine 36D5 were performed on four separate days when isotopically abnormal olivines 179D3, 186D2, 175D8, 17D1, and 14D3 were also analyzed.

New Sr–Nd isotope data were obtained using the same techniques as Thirlwall et al. (2004), with Sr being determined on residues from leaching powders at 130 °C for $>1\text{ h}$ in 6 M HCl.

3. Results

3.1. Data replication

Replicate $\delta^{18}\text{O}$ analyses of olivine, plagioclase, and glass in Reykjanes Ridge lavas have total ranges of $<0.20\text{‰}$ (Table 1), apart from 36D5 plagioclase. Several Reykjanes Peninsula samples are rather less reproducible (Table 2). For example, picritic samples from Haleyrbunga include RP80D and RP80E. Of four Optima analyses and two Prism analyses of RP80D olivine four have $\delta^{18}\text{O}$ between $+4.91$ and 5.00‰ , as are both analyses of RP80E olivine, while the other two are around $+4.7\text{‰}$. Similar discrepancies exist for olivine for several other samples, though in some cases early Prism analyses are the most discrepant. For all except RP55C, very coarse phenocryst separates were used, with only 2–4 grains used for each analysis, and it is possible that there may be real heterogeneity in $\delta^{18}\text{O}$ of the phenocryst olivine. Similarly, Eiler et al. (2000a) reported a range of 0.7‰ in olivine from one Theistareykir sample, while Gurenko and Chaussidon (2002) reported SIMS data suggesting a 3‰ variation in a single olivine from Hengill volcano.

Our LF mineral $\delta^{18}\text{O}$ data are in general consistent with mineral and whole rock data obtained by conventional fluorination techniques (e.g., Muehlenbachs et al., 1974), when corrected by appropriate fractionation factors. The main exception is data from Stapafell, where Muehlenbachs et al. (1974) report $\delta^{18}\text{O}_{\text{WR}}$ similar to our olivine data and Condomines et al. (1983) report substantial heterogeneity in $\delta^{18}\text{O}_{\text{WR}}$.

Our new data for the samples studied by Skovgaard et al. (2001) have $\delta^{18}\text{O}$ values up to 0.98‰ lower than the published data (given in parentheses in Table 2). Their data were analyzed on the same Mark II laser fluorination system used here, on two separate days. On the first (24/4/

Table 1
Oxygen isotope data for Reykjanes Ridge lavas

	Latitude (°N)	Depth (m)	By volume			MgO (wt.%)	¹⁴³ Nd/ ¹⁴⁴ Nd	$\delta^{18}\text{O}$ ol	$\delta^{18}\text{O}$ ol	$\delta^{18}\text{O}$ ol	$\delta^{18}\text{O}$ ol	Mean ol	2se	$\delta^{18}\text{O}$ glass	$\delta^{18}\text{O}$ glass	$\delta^{18}\text{O}$ glass	Mean glass	2se	$\delta^{18}\text{O}$ cpx	$\delta^{18}\text{O}$ cpx	$\delta^{18}\text{O}$ pl	$\delta^{18}\text{O}$ pl	Mean pl	2se
			% ol	% pl	% cpx																			
02D6	57.737	1575	0.7	0.3	0.0	8.5	0.513180							5.74			5.74							
14D3	57.520	2060	1	4	0.3	7.7	0.513092	4.93	4.98	4.89		4.93	0.05											
17D1	57.887	1725	10	0.5	0.0	10.7	0.513134	4.72	4.92	4.81	4.73	4.80	0.09											
36D5	58.055	1650	2	17	1.0	6.8	0.513180	5.17	5.07	5.13	5.09	5.12	0.04					5.28	5.39	5.67	5.45	5.56	0.22	
50D3	58.260	1500	3	0.0	0.0	8.6	0.513162							5.74	5.69		5.72	0.05						
80D1	58.968	1150	?	?	?	7.9	0.513155	5.14	5.14			5.14	0.00											
96D1	59.485	1000	2	3	0.3	7.7	0.513152	5.05	5.12	5.00		5.06	0.07											
105D1	59.830	700	3	8	7	8.4	0.513160	5.19				5.19		5.79		5.79		5.49		5.67		5.67		
116D6	60.082	1000	1	3	1	7.7	0.513161	5.04				5.04								5.62		5.62		
121D3	60.148	825	1	6	3	7.8	0.513143							5.49	5.52		5.51	0.03		5.56		5.56		
140D6	61.024	650	5	18	3	7.5	0.513160	5.17				5.17						5.11		5.51		5.51		
143D2	62.263	550	1	1	0.0	7.9	0.513103							5.55		5.55								
144D2	62.292	650	2	8	0.5	7.4	0.513157	5.14				5.14						5.45		5.68	5.69	5.69	0.01	
157D3	61.843	775	3	0.0	0.0	9.1	0.513099	5.13				5.13		5.46	5.51		5.49	0.05						
173D1	62.403	700	5	10	5	7.5	0.513106	5.04	5.05	4.97		5.02	0.05	5.34	5.43	5.48	5.42	0.08	5.30					
175D8	62.489	500	2	0.1	0.0	8.0	0.513073	4.84	4.88	4.91		4.88	0.04											
179D3	62.690	500	2	3	0.0	8.5	0.513062	4.75	4.70			4.73	0.05											
183D2	62.868	550	6	15	4	7.2	0.513078							5.47	5.33		5.40	0.14		5.56	5.50	5.53	0.06	
185D4	62.929	275	4	6	0.1	7.8	0.513056	4.48	4.45			4.47	0.03							5.12	5.27	5.20	0.15	
186D2	63.007	225	3	2	0.0	7.4	0.513039	4.53	4.54			4.54	0.01											

Italics signify measurement using Mark I manual laser fluorination system. $\delta^{18}\text{O}$ is ‰ relative to V-SMOW. MgO and Nd data from Thirlwall et al. (2004). Petrographic information obtained by visual estimation of a thin section.

Table 2
Oxygen isotope data for Icelandic lavas

	Locality	%ol (vol)	%pl (vol)	Max size (mm)		MgO wt.%	¹⁴³ Nd/ ¹⁴⁴ Nd	2se	$\delta^{18}\text{O}$ ol	$\delta^{18}\text{O}$ ol	$\delta^{18}\text{O}$ ol	$\delta^{18}\text{O}$ ol	$\delta^{18}\text{O}$ ol	mean ol	2se	$\delta^{18}\text{O}$ cpx	$\delta^{18}\text{O}$ pl	$\delta^{18}\text{O}$ pl	mean pl	2 se					
				ol	pl																				
TH29	Theistareykir Picrite					15.7*	0.513152		4.29	4.10	4.42	4.28				(4.37)* ¹			4.27	0.13					
408643	Arnahvammur					12.4*			4.38							(5.36)* ²			4.38						
TH40	Borgarhraun					18.6*			4.78							(4.91)* ¹			4.78						
408647	Langaviti					14.5*			4.64							(4.57)* ¹			4.64						
	Eastern Rift Zone and Hekla																								
I646	Krafla 1724					6.0	0.513035												3.18	3.78	3.78				
I647	Laki Grimsvötn					5.7	0.513042		2.56										2.56	3.11	3.24	3.18	0.13		
RP116G	Tungnarhraun					7.7	0.513066												4.01	4.50	4.61	4.56	0.11		
RP116D	Thjorsahraun					6.3	0.513049		3.89	3.71									3.80	0.18	4.50	4.30	4.40	0.20	
I654	Hekla 1970					3.0	0.513017												4.14	<u>5.13</u>		5.13			
RP116H	Hekla 1878					5.6	0.513016												4.31		4.94		4.94		
	Reykjanes Peninsula depleted																								
OS208225	Dimmidalur					11.3*			4.80							(5.19)* ²			4.80						
OS208224	Asar syd (D27)					13.1*			4.90							(5.29)* ²			4.90						
D26	Burfell					11.2*			5.12							(5.75)* ²			5.12						
RP103A	Kleifarvatn	2.2	0.1	0.8	0.5	11.4	0.513032		4.95	5.04	4.92							4.97	0.07						
RP80D	Haleyjabunga	17.4	0.0	3.5	—	16.3	0.513132		4.68	4.91	4.72	4.98	4.99					4.94	4.87	0.11	5.15	5.30	5.23	0.15	
RP80E	Haleyjabunga	12.7	0.0	1.6	—	13.9	0.513130	4	4.92	4.96								4.94	0.04						
OS207901	Haleyjabunga					24.7*			5.03							(5.24)* ²			5.03						
RP107A	Kleifarvatn	5.8	0.0	1.6	—	11.6	0.513124		4.70	4.90	4.89	4.94						4.86	0.11						
RP91E	Fagridalur	8.1	0.0	2.0	—	14.5	0.513134		5.14	5.10	5.12	5.13						5.12	0.02						
RP3	Vatnsheidi	14	3	4	?	18.3	0.513099		4.72	4.79	4.78	4.83	4.78					4.56	4.74	0.08	5.11	5.27	5.19	0.16	
RP82D	Vatnsheidi	0.5	1.6	1.5	1.2	11.2	0.513066		4.52	4.75	4.60	4.61						4.62	0.10		5.02	5.24	5.13	0.22	
RP82C	Vatnsheidi	0.9	0.0	1.5	—	10.0	0.513045		4.47	4.56	4.86	4.80	4.81					4.88	4.73	0.14	5.37		5.37		
OS208222	Vatnsheidi					16.3*			4.57							(4.76)* ²			4.57						
RP95A	Lagafell	37.4	0.0	3.0	—	27.8	0.513145	5	5.07	4.82	4.86							4.92	0.16						
RP95C	Lagafell	5.9	0.0	1.9	—	11.6	0.513148		5.26	5.16	5.12	4.98	4.99					5.10	0.11						
D7	Lagafell					25.6*			5.08							(5.62)* ²			5.08						
RP64O	Hildarvatn	4.2	0.7	0.5	0.3	11.4	0.513111		4.56									4.56							
RP52A	Hamranes	1.0	3.8	1.2	2.2	9.5	0.513094											4.53			5.07	5.07	5.07	0.00	
RP55C	Gunnarsholmi	28.9	0.0	1.9	—	19.6	0.513092		4.69	4.34	4.34	4.50	4.59					4.49	0.14						
RP80C	Skalafell(D1)	0.1	4.3	0.5	5.1	8.0	0.513076	4	4.93	4.73	4.63							4.76	0.18		5.29	5.30	5.30	0.01	
RP105A	NE of Lagafell	1.9	7.5	0.5	2.2	9.9	0.513087	4	5.20	4.90	4.95	4.91						4.99	0.14	5.50	5.54	5.66	5.60	0.12	
	Reykjanes Peninsula enriched																								
RP59F	Helgavik	5.5	0.4	0.7	2.0	9.2	0.513025		3.38									3.38			3.76				
RP65N	Ogmundarhraun	0.1	0.6	—	0.9	7.5	0.513047											4.54			5.16	5.12	5.14	0.04	
RP65H	Drumbur	0.0	6.8	—	1.1	5.5	0.513057											3.80			4.23	4.55	4.51	4.53	0.04
RP42P	Skala-Maelifell	3.8	0.9	0.8	0.8	9.5	0.513005	4	4.58	4.66	4.32	4.26	4.34					4.43	0.16		4.65		4.65		
RP67E	Stapafell	1.8	0.2	0.9	1.4	13.9	0.513010		4.77	4.83	4.80	4.81						4.80	0.03						
RP115J	Sandfell	3.6	1.4	0.8	0.5	9.7	0.513006		4.42	4.23	4.30	4.46						4.35	0.11		4.91		4.91		
RP91L	Fagradalsfjall	4.3	0.0	1.3	—	10.5	0.512979		4.22	4.12	4.33	4.36						4.26	0.11						

Low $\delta^{18}\text{O}$ in the Icelandic mantle

(continued on next page)

Table 2 (continued)

Locality	%ol (vol)	%opl (vol)	Max size (mm)	MgO wt. %	$^{143}\text{Nd}/^{144}\text{Nd}$	2se	ol		opl		cpx	2se	pl		mean	2 se
							$\delta^{18}\text{O}$	$\delta^{18}\text{O}$	$\delta^{18}\text{O}$	$\delta^{18}\text{O}$			$\delta^{18}\text{O}$	$\delta^{18}\text{O}$		
RP100C	7.0	0.0	1.5	11.6	0.512964	4.25	4.32	4.36	4.37	4.33	0.05					
RP98A	0.1	11.6	0.7	7.3	0.513030	4	4.56	4.40	4.65	4.54	0.15		5.26		5.26	
Skogafell																
RP95B	0.8	13.8	0.9	7.9	0.513001	4	4.63	4.50	4.73	4.45	0.10		5.35	5.21	5.28	0.14

Italics signify measurement using Mark I manual laser fluorination system, underlines signify lowish yield (89–91%). Values for mean $\delta^{18}\text{O}_{\text{ol}}$ with no individual measurements estimated from $\delta^{18}\text{O}_{\text{pl}}$ (Section 3.2). $\delta^{18}\text{O}$ is ‰ relative to V-SMOW. $^{143}\text{Nd}/^{144}\text{Nd}$ and MgO data from Thirlwall et al. (2004), except for seven new samples (2se uncertainty given). The original Nd data for RP42P was suspect and it has been re-analysed. MgO data labelled with an asterisk are samples from Skovgaard et al. (2001). Their reported $\delta^{18}\text{O}$ values are shown as, e.g., (5.62)* in the final column for $\delta^{18}\text{O}_{\text{ol}}$, with the superscript 1 or 2 signifying that their analysis was carried out on 24/4/99 or 11/11/99, respectively. ‰ol and ‰opl refer to phenocrysts as volume % of solid rock (i.e., excluding vesicles) based on point counting 1200–1500 points, with maximum size of the phenocrysts given in the next columns. Cpx analyses appear to be from coarse groundmass clinopyroxene.

99), only one standard was used to define the correction to San Carlos olivine I, but the correction was small (-0.08‰) and data for three samples subsequently replicated in this paper agree to within 0.13‰ . On the second day (11/11/99), the three standards analysed required a very large and variable correction to the raw data ($+1.3$ to $+1.8\text{‰}$), and three samples they replicated from 24/4/99 differ by -0.15 to $+0.49\text{‰}$. Their data from 11/11/99 are on average 0.43‰ higher than our new data for the same samples. Several of the Reykjanes Peninsula olivines reported by Skovgaard et al. (2001) are from flows from which we have analysed replicate olivines, and our new data for their samples are much more consistent with our analyses from the same flows (e.g., Lagafell, Table 2). Skovgaard et al. (2001) observed correlations between $\delta^{18}\text{O}$ and several chemical parameters such as the Sr anomaly and Eu/Eu^* . The big change to their reported $\delta^{18}\text{O}$ for 408643 destroys these correlations in the Theistareykir data (e.g., r^2 reduces from ~ 0.85 to ~ 0.1), although they appear still to be present at Reykjanes ($r^2 = 0.54\text{--}0.96$), albeit based on only six samples, from which one aberrant picritic sample was excluded on the grounds of plagioclase accumulation (Fig. 4 of Skovgaard et al., 2001).

Our $\delta^{18}\text{O}$ value of $+4.27 \pm 0.13\text{‰}$ (2se, $N = 4$) determined for olivine from Theistareykir picrite TH29 (supplied by Skovgaard) agrees with the $+4.37$ value reported by Skovgaard et al. (2001) that they analysed on 24/4/99. It is, however, much lower than the value of $+4.70 \pm 0.03\text{‰}$ (1sd, $N = 2$) reported by Eiler et al. (2000a), which is more consistent with the chemistry of this sample. This may reflect an interlaboratory bias, a possibility we have investigated by analysis between 10th and 16th February 2005 of several samples kindly provided by Prof. Eiler, along with some key samples from this study (Table 3). $\delta^{18}\text{O}$ values for our samples in Table 3 conform well to our previous analyses (Tables 1 and 2), and values for Eiler's samples conform well to those reported by Eiler et al. (2000a). On 10/2/05, TH29 olivine yielded $+4.42$ and $+4.63\text{‰}$ for sample supplied by Skovgaard and Eiler, respectively, within error of previous analyses in Table 2 and in Eiler et al. (2000a), respectively, raising the possibility that these may be two different samples. The San Carlos olivine III standard supplied by Eiler yields $\delta^{18}\text{O}$ identical to that of our San Carlos olivine II standard (Table 3).

3.2. Inter-phase ^{18}O fractionation: temperature/composition effects on $\delta^{18}\text{O}$

Where possible two or more mineral phases (olivine, plagioclase, and clinopyroxene) or glass were analyzed for each rock sample. Although phenocryst plagioclase is present in some other key samples (e.g., 186D2), it is frequently small and full of inclusions. For the Reykjanes Ridge, $\Delta^{18}\text{O}_{\text{cpx-ol}}$ averages $+0.28 \pm 0.08\text{‰}$ (2sd, $N = 4$), excepting sample 140D6 ($\Delta^{18}\text{O}_{\text{cpx-ol}} = -0.06\text{‰}$), for which there was insufficient clinopyroxene available to duplicate. $\Delta^{18}\text{O}_{\text{pl-ol}}$ and $\Delta^{18}\text{O}_{\text{gl-ol}}$ average 0.52 ± 0.27 and

Table 3
Oxygen isotope data demonstrating inter-laboratory comparability

10/2/05			14/2/05			16/2/05		
Sample	$\delta^{18}\text{O}$	Source	Sample	$\delta^{18}\text{O}$	Source	Sample	$\delta^{18}\text{O}$	Source
San Carlos ol II (four other samples)	5.26	RH	San Carlos ol II	5.24	RH	San Carlos ol II	5.13	RH
14D3 ol	4.89	RH	San Carlos ol III	5.22	Eiler	San Carlos ol III	5.23	Eiler
San Carlos ol II	5.23	RH	98M2 ol	4.71	Eiler			
17D1 ol	4.73	RH						
36D5 ol	5.09	RH						
TH29 ol	4.63	Eiler						
TH29 ol	4.42	Skovgaard						
San Carlos ol III	5.26	Eiler						
San Carlos ol II	5.09	RH						
98-M2 ol	4.50	Eiler						

All samples run consecutively except where stated. Eiler et al. (2000a) report $+4.70 \pm 0.03$ and $+4.53 \pm 0.04$ for their TH29 and 98-M2 olivines, respectively.

$0.46 \pm 0.26\text{‰}$ (2sd, $N = 6$ and 3 , respectively), comparable to values reported by Eiler et al. (2000c) of 0.65 ± 0.28 and $0.36 \pm 0.22\text{‰}$ for $\Delta^{18}\text{O}_{\text{pl-ol}}$ and $\Delta^{18}\text{O}_{\text{gl-ol}}$, respectively (2sd, $N = 21$ and 9). Similar values are observed for Icelandic mineral pairs (Table 2), with mean $\Delta^{18}\text{O}_{\text{pl-ol}}$ of 0.58 ± 0.24 (2sd, $N = 10$) apart from rather low values of ca. $+0.25\text{‰}$ in RP80D and RP42P. All these values are substantially lower than theoretical values calculated, for example, using factors given by Zhao and Zheng (2003), and we suspect there are problems in extrapolating theoretical factors to high temperature.

We have assessed the effects of fractional crystallization on our data using major element modelling of the lavas and our observed fractionation factors, whilst preserving the temperature variation of Zhao and Zheng (2003). Equilibrium olivine compositions and crystal-liquid equilibration temperatures were estimated using the method of Ford et al. (1983). For the Reykjanes Ridge lavas analyzed here, the MgO range of 9.1–7.2% can be generated by up to 27% fractional crystallization of olivine + plagioclase + clinopyroxene in roughly 1:2.4:1.5 proportions (see Murton et al., 2002), which is expected to lead to increases in $\delta^{18}\text{O}_{\text{ol}}$, $\delta^{18}\text{O}_{\text{cpx}}$, $\delta^{18}\text{O}_{\text{glass}}$, and $\delta^{18}\text{O}_{\text{pl}}$ of 0.03, 0.04, 0.05, and 0.06‰ , respectively, assuming little change in plagioclase An content. A 10% decrease in An would increase $\delta^{18}\text{O}_{\text{pl}}$ by a further 0.03‰ . These changes are insignificant.

Our Reykjanes Peninsula incompatible-element depleted samples all have $>9.5\%$ MgO. Those with $<11.5\%$ MgO have few ($<5\%$), generally small, olivine phenocrysts (Table 2) and probably are close to melt compositions. 10–18% fractional crystallization of olivine and calcic plagioclase is required to reduce MgO from 11.5 to 9.5%, which is expected to cause no significant change to $\delta^{18}\text{O}_{\text{ol}}$ or $\delta^{18}\text{O}_{\text{pl}}$. The more MgO-rich samples contain 5–25% large olivine crystals (Table 2), suggesting they are the results of olivine accumulation. RP3 and RP80D olivines are fairly homogeneous at $\sim\text{Fo}_{89.2}$ and $\sim\text{Fo}_{90.2}$, respectively, with CaO contents indicating an origin through crystallization from basaltic melt rather than as mantle xenocrysts (EA-1, Electronic Annex). The homogeneity of olivine composi-

tions from these and RP82C (EA-1) indicates that the olivines are very unlikely to be accidental xenocrysts from other flows. Calculations using the method of Ford et al. (1983) indicate that for RP3 and RP80D some 21 and 14.5%, respectively, of the observed olivine must be removed to generate a bulk composition in equilibrium with the olivine, and that this would have had 10–12% MgO, similar to magmatic concentrations estimated for Theistareykir by Eiler et al. (2000a). Thus, fractional crystallization or accumulation could have had no significant effect on $\delta^{18}\text{O}_{\text{ol}}$ in these rocks.

The incompatible-element-enriched Reykjanes Peninsula samples reported here include two very magnesian samples, RP67E and RP100C. These only contain small amounts of small olivine phenocrysts, which are at most $\text{Fo}_{87.1}$ and $\text{Fo}_{87.8}$, respectively (EA-1). Although the equilibrium olivine is ca. Fo_{89} , total removal of olivine phenocrysts from RP67E only reduces MgO to 13.1%. Using RP67E as a parent, most of the other enriched Peninsula samples discussed here can be generated by $<20\%$ removal of olivine and plagioclase, with plagioclase accumulation to generate the strongly plagioclase-phyric RP95B and RP98A. This is expected to increase $\delta^{18}\text{O}_{\text{pl}}$ by at most 0.10‰ (RP65N) and have no significant effect on $\delta^{18}\text{O}_{\text{ol}}$. The most evolved sample, RP65H, requires removal of $\sim 45\%$ olivine, plagioclase, clinopyroxene, and magnetite from primitive melts such as RP67E, which is expected to increase $\delta^{18}\text{O}_{\text{pl}}$ by $\sim 0.24\text{‰}$.

There are no parental magmas in this study for the samples from Hekla or the Eastern Rift Zone. Using appropriate parents with $\sim 10\%$ MgO from the Reykjanes Peninsula, we can estimate that $\sim 40\%$ crystallization is needed to produce RP116H and another 42% to produce I654. This is expected to increase $\delta^{18}\text{O}_{\text{pl}}$ by 0.13 and a further 0.36‰ , respectively, though the latter is strongly dependent on the fractionation factor chosen for magnetite. Similar estimates for the change in magmatic $\delta^{18}\text{O}$ were arrived at by Sigmarsson et al. (1992a). Evolved samples I647 and RP116D from the Eastern Rift Zone require 35 and 27% crystallization, respectively, from parents with

~10% MgO, which is expected to result in 0.10‰ increase in $\delta^{18}\text{O}_{\text{pl}}$ and no change in $\delta^{18}\text{O}_{\text{ol}}$. Because $\delta^{18}\text{O}_{\text{ol}}$ is largely invariant with fractional crystallization, most of our discussion centres on this. Where no olivine data are available, we have estimated parental $\delta^{18}\text{O}_{\text{ol}}$ by subtracting 0.5‰ from $\delta^{18}\text{O}_{\text{pl}}$ in more primitive samples, with an additional correction based on the discussion above for evolved samples. These estimated $\delta^{18}\text{O}_{\text{ol}}$ data are given in the mean $\delta^{18}\text{O}_{\text{ol}}$ column of Table 2.

3.3. Reykjanes Ridge oxygen data

Oxygen isotope data from the Reykjanes Ridge are displayed as a function of latitude in Fig. 1. Olivines analyzed from 58.0 to 62.4°N have mean $\delta^{18}\text{O}_{\text{ol}} = +5.11 \pm 0.11\%$ (2sd on 9 samples, 16 total analyses), very close to the lithospheric mantle mean of $+5.2 \pm 0.3\%$ (2sd, Matthey et al., 1994). $\delta^{18}\text{O}_{\text{ol}}$ significantly lower than 58.0–62.4°N samples is observed in a mega-seamount at 57.52°N (14D3), in an unusually primitive composition at 57.89°N (17D1), as well as in all samples north of 62.4°. 17D1 and samples north of 62.5° are significantly lower than the mantle range reported by Matthey et al. (1994). 14D3 and 17D1 have anomalous radiogenic isotope ratios (e.g., low $^{143}\text{Nd}/^{144}\text{Nd}$) and high $^3\text{He}/^4\text{He}$ for their latitude and resemble Icelandic lavas (Taylor et al., 1997; Hilton et al., 2000; Thirlwall et al.,

2004), far south of the main mixing gradient toward Iceland. North of 62.4°, radiogenic isotope ratios are dominated by the contribution from Icelandic mantle. $\delta^{18}\text{O}_{\text{ol}}$ shows hyperbolic correlations with radiogenic isotope ratios, e.g., $^{143}\text{Nd}/^{144}\text{Nd}$ (excluding 17D1, Fig. 2), with low $\delta^{18}\text{O}_{\text{ol}}$ being associated with the lowest $^{143}\text{Nd}/^{144}\text{Nd}$ ratios north of 62.4°N.

Plagioclase and glass data from 57.7 to 62.4°N are very homogeneous, with mean $\delta^{18}\text{O}$ of $+5.60 \pm 0.14\%$ (2sd, $N = 7$) and $+5.61 \pm 0.28\%$ ($N = 7$), respectively, the latter similar to MORB glasses worldwide ($+5.5 \pm 0.3\%$, 2sd, Eiler et al., 2000b). Low $\delta^{18}\text{O}_{\text{pl}}$ is only observed in 185D4, at 62.9°N, but only one other plagioclase, and only one glass, have been analyzed north of 62.4°, from sample 183D2. This has relatively high $^{143}\text{Nd}/^{144}\text{Nd}$ for its latitude and did not yield enough olivine for oxygen analysis. Glass samples from the Reykjanes Ridge appear to show a northward decrease toward Iceland at more southerly latitudes than shown by the olivine data, with $\delta^{18}\text{O}_{\text{gl}} = +5.72$ to

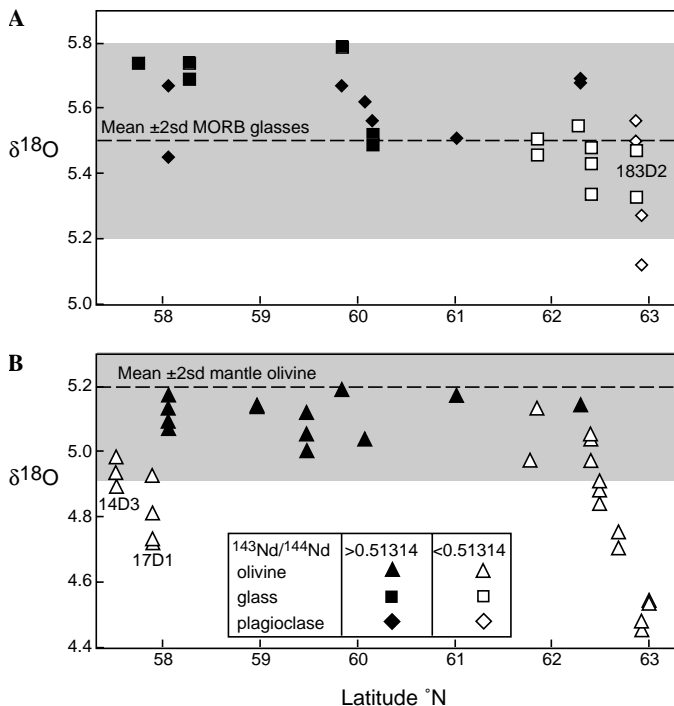


Fig. 1. Variation in $\delta^{18}\text{O}$ of (A) plagioclase phenocrysts and matrix glass, and (B) olivine phenocrysts with latitude along the Reykjanes Ridge. Each individual analysis is shown. Mean mantle olivine $\delta^{18}\text{O}$ from Matthey et al. (1994), mean MORB glasses from Eiler et al. (2000b). 14D3 and 17D1 are samples with anomalously Icelandic Sr–Nd isotope ratios and $^3\text{He}/^4\text{He}$ elevated above background in the southern section of the ridge. 183D2 has anomalously high $^{143}\text{Nd}/^{144}\text{Nd}$ for its latitude and insufficient phenocryst olivine for analysis.

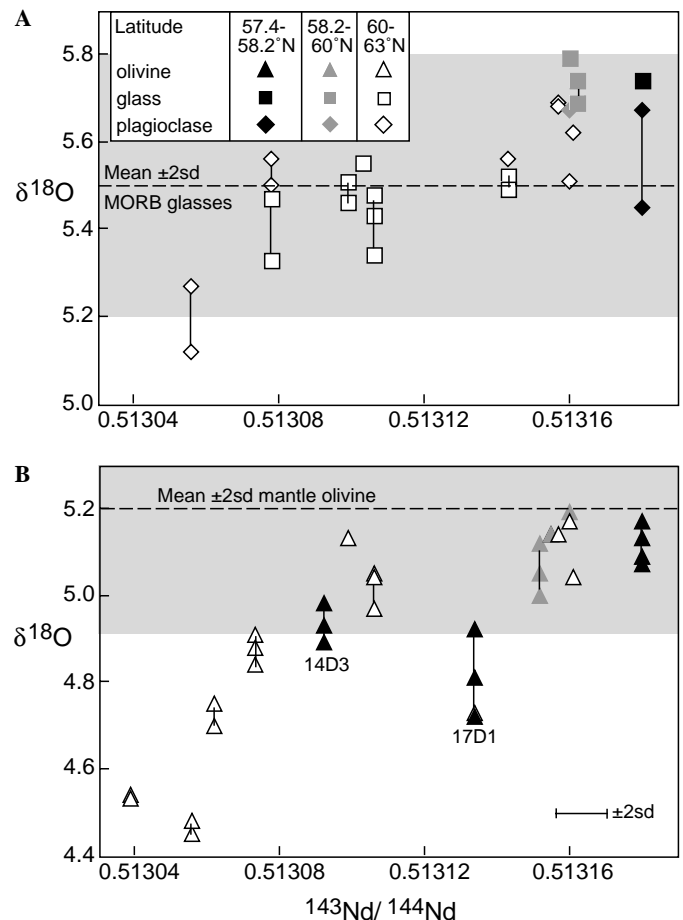


Fig. 2. Variation in $\delta^{18}\text{O}$ of (A) plagioclase phenocrysts and matrix glass and (B) olivine phenocrysts with $^{143}\text{Nd}/^{144}\text{Nd}$ in Reykjanes Ridge lavas. Each individual analysis is shown, with tie-lines between replicate $\delta^{18}\text{O}$ measurements. Mean mantle olivine $\delta^{18}\text{O}$ from Matthey et al. (1994), mean MORB glasses from Eiler et al. (2000b). 14D3 and 17D1 are samples with anomalously Icelandic Sr–Nd isotope ratios and $^3\text{He}/^4\text{He}$ elevated above background in the southern section of the ridge. Nd data from Thirlwall et al. (2004).

5.79‰ south of 60°N, and +5.46 to 5.55‰ from 60 to 62.5°N (Fig. 1). These differences, although small, may be real since one of the more southerly samples, and three of the more northerly, have been replicated.

3.4. Reykjanes Peninsula oxygen data

Preliminary oxygen results for Reykjanes Peninsula lavas were reported by Gee et al. (1998). These data were obtained using the manually operated mark I laser fluorination system and thus not much duplication was carried out. We report the original data in Table 2, together with, for most samples, two to four replicate analyses using the mark II automatic LF system. $\delta^{18}\text{O}_{\text{ol}}$ for Reykjanes Peninsula lavas ranges from +3.38 to +5.12‰. There is no evidence for samples with $\delta^{18}\text{O}_{\text{ol}}$ higher than the lithospheric mantle value of +5.2‰ (Mattey et al., 1994), either in our comprehensive new data or in our re-analyses of samples from Skovgaard et al. (2001). $\delta^{18}\text{O}_{\text{ol}}$ values of +5.3 to +5.76‰ reported by these authors were the result of very large corrections to the raw data interpolated from too few San Carlos olivine standard runs.

$\delta^{18}\text{O}_{\text{ol}}$ values similar to those of normal mantle olivine (>+4.9‰) are restricted to “depleted lavas” on the Reykjanes Peninsula. These have very low incompatible element concentrations (e.g., Nb typically 0.2–1.5 ppm), low ratios of very incompatible to moderately incompatible trace elements (Nb/Zr < 0.09, Fig. 3) and in most cases elevated $^{143}\text{Nd}/^{144}\text{Nd}$ (Fig. 3; Thirlwall et al., 2004). These depleted lavas are commonly accumulative picrites, but range down to 9.5% MgO, and magmatic MgO contents are probably no more than 12% (Fig. 4, Section 3.2). They are very uncommon, occupying less than 5% of the surface area of the Peninsula (Gee et al., 1998), and were probably mostly erupted around 10 ka ago. Several depleted lavas have $\delta^{18}\text{O}_{\text{ol}}$ substantially lower than normal mantle values (ca. +4.5‰, Fig. 3): these tend to have lower $^{143}\text{Nd}/^{144}\text{Nd}$ and higher $^{206}\text{Pb}/^{204}\text{Pb}$ (Fig. 6).

“Enriched lavas” from the Reykjanes Peninsula (defined as having Nb/Zr > 0.09; Fig. 3, Thirlwall et al., 2004) all have $\delta^{18}\text{O}_{\text{ol}}$ less than normal mantle values, with a range of +3.38 to +4.80‰. Only two samples have $\delta^{18}\text{O}_{\text{ol}}$ < +4.25‰: these are RP59F, a large-volume flow that predates the active rift zone on the Peninsula, and RP65H, the most evolved Reykjanes sample analysed, with MgO of 5.5% (Fig. 4). These both show slightly higher $^{87}\text{Sr}/^{86}\text{Sr}$ than other enriched lavas with similar $^{143}\text{Nd}/^{144}\text{Nd}$ on the Reykjanes Peninsula (Table 4, Fig. 5), and RP59F shows Pb–Nd–Sr isotope similarity to Eastern Rift Zone samples (Thirlwall et al., 2004), which also show very low $\delta^{18}\text{O}_{\text{ol}}$ (e.g., Fig. 3).

Apart from the low $\delta^{18}\text{O}$ in evolved sample RP65H, there is no correlation between $\delta^{18}\text{O}_{\text{ol}}$ and MgO in Reykjanes Peninsula samples (Fig. 4), and both enriched and depleted samples with MgO > 10% have $\delta^{18}\text{O}_{\text{ol}}$ around +4.5‰. RP59F and RP65H lie well below apparently linear correlations between $\delta^{18}\text{O}_{\text{ol}}$ and $^{143}\text{Nd}/^{144}\text{Nd}$, Nb/Zr

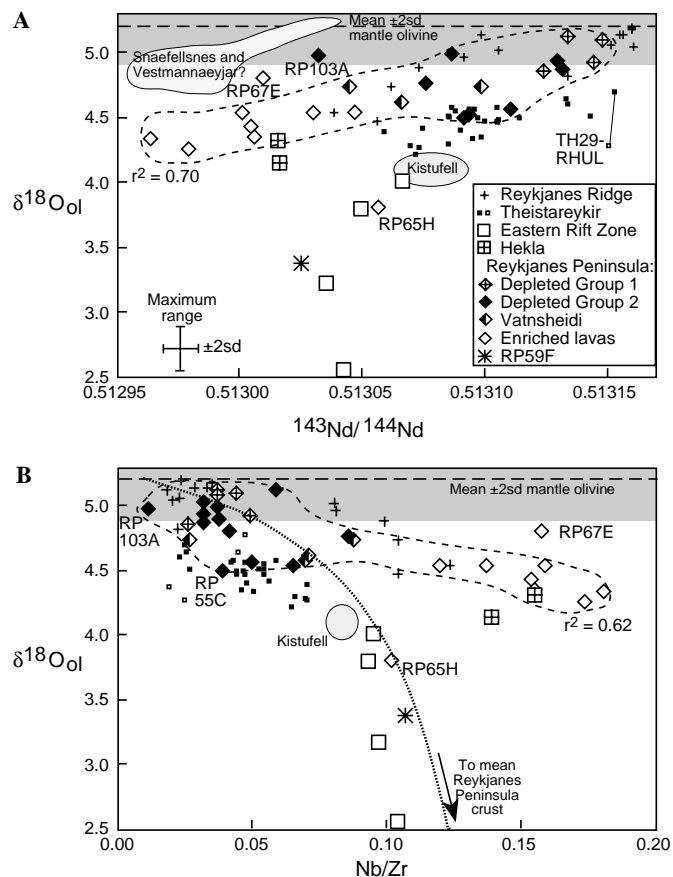


Fig. 3. Relationships between $\delta^{18}\text{O}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ and Nb/Zr in Icelandic lavas. $\delta^{18}\text{O}_{\text{ol}}$ is mean $\delta^{18}\text{O}$ determined on olivine phenocrysts or is parental $\delta^{18}\text{O}_{\text{ol}}$ estimated from plagioclase phenocryst data (seven samples; see Section 3.2). Mean mantle $\delta^{18}\text{O}_{\text{ol}}$ from Mattey et al. (1994). Groups of Reykjanes Peninsula lavas, Nd and Nb/Zr data from Thirlwall et al. (2004) and Table 2. The dashed fields outline correlations for Reykjanes Peninsula lavas with >7% MgO referred to in the text. These exclude RP67E, which has $\delta^{18}\text{O}$ higher than all other enriched lavas, and (in A) only RP103A, which is unique in being a depleted sample (low Nb/Zr) with low $^{143}\text{Nd}/^{144}\text{Nd}$. Data for olivines from Theistareykir primitive lavas from Eiler et al. (2000a) and Stracke et al. (2003), except small open squares: Nb/Zr from Skovgaard et al. (2001) and $\delta^{18}\text{O}$ from Table 2. The Kistufell field is laser fluorination data on primitive glasses ($N = 4$) from Breddam (2002); the Snæfellsnes and Vestmannæyjar field is conventional whole rock data on alkali basalts to trachyte ($N = 6$) from Hémond et al. (1993), all adjusted to olivine values by subtracting 0.5. The mixing line shown is between the most extreme mantle derived incompatible-element depleted lava (Zr = 14 ppm, Nb/Zr = 0.01) and an estimated composition of Reykjanes Peninsula crust ($\delta^{18}\text{O}_{\text{WR}} = +2\text{‰}$, Gautason and Muehlenbachs, 1998; Zr = 80.4 ppm, Nb/Zr = 0.130 based on the average composition of 253 Reykjanes Peninsula lavas, Gee, 1998). Error bar is $\pm 2\text{sd}$ external reproducibility of Nd isotope standards, and maximum total $\delta^{18}\text{O}$ range observed in replicate olivines from Reykjanes Peninsula lavas. r^2 values relate to the correlations outlined by the dashed fields.

and $^{206}\text{Pb}/^{204}\text{Pb}$ (Figs. 3 and 6, correlation indicated by dashed field; $r^2 = 0.62\text{--}0.75$). In both Figs. 3 and 6 RP67E, the most primitive enriched lava, lies above the correlations, while in Fig. 3A RP103A, an unusual low- $^{143}\text{Nd}/^{144}\text{Nd}$ but highly depleted sample (e.g., Nb/Zr, Fig. 3B), also lies above the correlation.

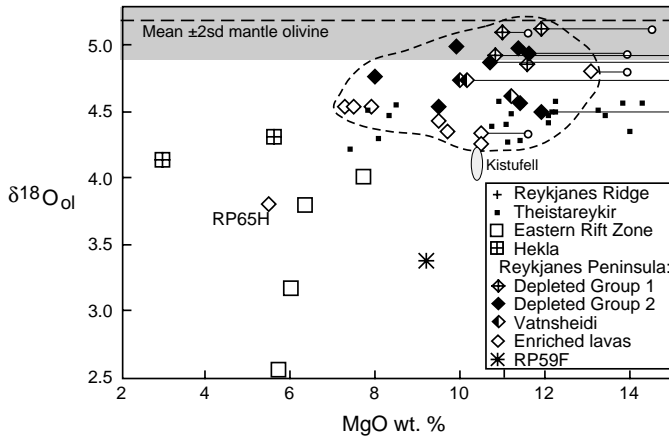


Fig. 4. $\delta^{18}\text{O}$ –MgO relationships in Icelandic lavas. $\delta^{18}\text{O}_{\text{ol}}$ is mean $\delta^{18}\text{O}$ determined on olivine phenocrysts or is parental $\delta^{18}\text{O}_{\text{ol}}$ estimated from plagioclase phenocryst data (seven samples; see Section 3.2). Mean mantle $\delta^{18}\text{O}_{\text{ol}}$ from Matthey et al. (1994). MgO data from Thirlwall et al. (2004), adjusted for olivine accumulation by subtracting F_{090} , or the measured olivine, if available, until either all observed phenocrysts are exhausted or until the calculated melt is in equilibrium with the measured olivine or with F_{090} . Horizontal tie-lines to small circles show the uncorrected MgO contents. Theistareykir data from Eiler et al. (2000a) and Stracke et al. (2003), not corrected for accumulation. Kistufell field is laser fluorination data on primitive glasses ($N = 4$) from Breddam (2002), adjusted to olivine values by subtracting 0.5. Note the wide spread in $\delta^{18}\text{O}_{\text{ol}}$ at 10–12% MgO. The dashed field outlines Reykjanes Peninsula lavas with $>7\%$ MgO that show correlations in Figs. 3 and 6.

3.5. Other Icelandic localities

Table 2 reports LF $\delta^{18}\text{O}$ for olivine and plagioclase from a few other Icelandic localities, namely Hekla, Veidivötn, Laki (Grimsvötn), and Krafla. Pb–Sr–Nd isotope data for these samples were reported by Thirlwall et al. (2004). We have also re-analysed four of the Theistareykir samples

Table 4
Additional Sr isotope data for Reykjanes Peninsula samples

	$^{87}\text{Sr}/^{86}\text{Sr}_{1996}$	$^{87}\text{Sr}/^{86}\text{Sr}_{\text{new}}$	$^{87}\text{Sr}/^{86}\text{Sr}_{\text{new}}$	$\Delta \times 10^6$
RP91E		0.702992 ± 10		–8
RP3		0.703136 ± 09		+37
RP82C		0.703141 ± 10	0.703150 ± 16	–46
RP95A	0.702881 ± 08	0.702861 ± 16		–20
RP95C		0.702924 ± 09	0.702891 ± 14	+36
RP80C	0.703045 ± 09	0.703052 ± 09		
RP105A	0.703049 ± 13	0.703036 ± 14		–13
RP56A		0.703077 ± 13		+11
RP65H		0.703143 ± 11		+15
RP98A	0.703127 ± 11	0.703128 ± 10		+1
RP92C	0.703190 ± 10	0.703195 ± 10		+5
RP42P		0.703178 ± 09	0.703168 ± 10	–35
RP67E		0.703184 ± 09		+7
RP95B	0.703218 ± 09	0.703203 ± 12		–15

All results obtained on powder samples leached for >1 h in hot 6 M HCl. Δ is the difference in $^{87}\text{Sr}/^{86}\text{Sr}$ between the new results reported here and either analyses from ~ 1996 or data reported by Thirlwall et al. (2004). In some cases, this difference is significantly worse than the SRM987 reproducibility of $\sim 18 \times 10^{-6}$.

This may largely reflect poor reproducibility of the leaching technique.

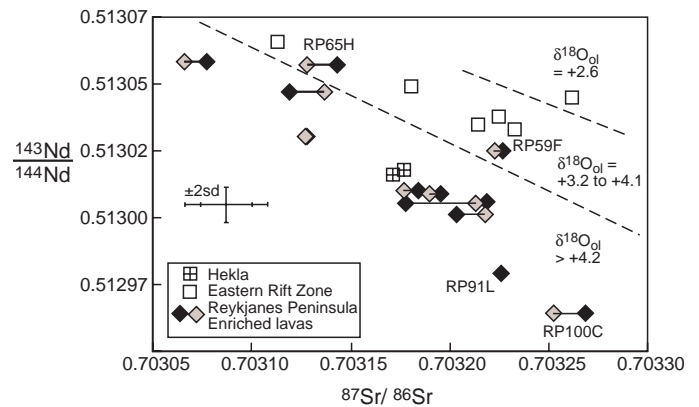


Fig. 5. $^{143}\text{Nd}/^{144}\text{Nd}$ – $^{87}\text{Sr}/^{86}\text{Sr}$ relationships in incompatible-element-enriched Icelandic lavas ($\text{Nb}/\text{Zr} > 0.09$). Note that samples with lower $\delta^{18}\text{O}$ in olivine are displaced toward higher $^{87}\text{Sr}/^{86}\text{Sr}$, possibly suggesting crustal contamination, although there is still a $\delta^{18}\text{O}$ – $^{143}\text{Nd}/^{144}\text{Nd}$ correlation in the suite with $\delta^{18}\text{O}_{\text{ol}} = +3.2$ to $+4.1$. Solid diamonds and squares are post-2000 Sr–Nd analyses, grey diamonds 1994–1996 analyses, connected by tie-lines. Error bar is $\pm 2\text{sd}$ external reproducibility of Sr–Nd isotope standards: the smaller bar for Sr applies to post-2000 analyses. Data from Thirlwall et al. (2004) and Tables 2 and 4.

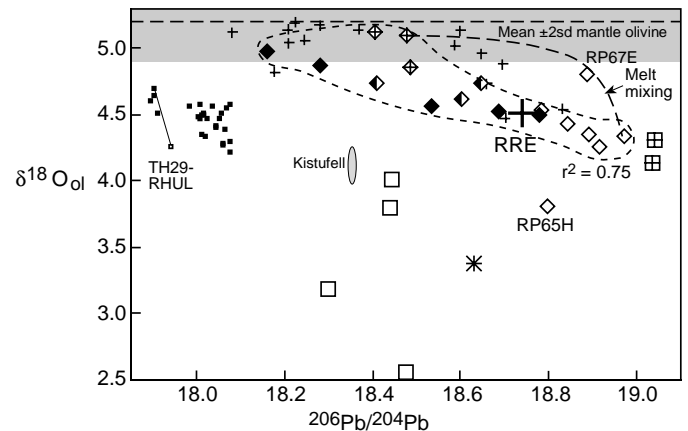


Fig. 6. $\delta^{18}\text{O}$ – $^{206}\text{Pb}/^{204}\text{Pb}$ relationships in Icelandic lavas. $\delta^{18}\text{O}_{\text{ol}}$ is mean $\delta^{18}\text{O}$ determined on olivine phenocrysts or is parental $\delta^{18}\text{O}_{\text{ol}}$ estimated from plagioclase phenocryst data (seven samples; see Section 3.2). Mean mantle $\delta^{18}\text{O}_{\text{ol}}$ from Matthey et al. (1994). Pb data from Thirlwall et al. (2004). Theistareykir data from Eiler et al. (2000a) and Stracke et al. (2003), except TH29-RHUL, which is a Royal Holloway analysis of TH29 olivine with $\delta^{18}\text{O}$ 0.5‰ lower than reported by Eiler et al. (2000a). Kistufell field is laser fluorination data on primitive glasses ($N = 4$) from Breddam (2002), adjusted to olivine values by subtracting 0.5. The dashed field ($r^2 = 0.75$) outlines the correlation for Reykjanes Peninsula lavas with $>7\%$ MgO referred to in the text. The dashed curve is a melt-mixing hyperbola between RP95C and RP100C. Symbols as Figs. 3 and 4. The composition of the enriched end-member on the Reykjanes Ridge (RRE) is shown by a large cross. Reproducibility of $^{206}\text{Pb}/^{204}\text{Pb}$ smaller than symbol size.

reported by Skovgaard et al. (2001), and our analyses agree well with their data obtained on the day when the correction for San Carlos olivine was small. Our Theistareykir re-analyses range from $\delta^{18}\text{O}_{\text{ol}} = +4.10$ to $+4.78\text{‰}$, and are broadly similar to data reported by Eiler et al.

(2000a), apart from TH29 as discussed in Section 3.1. Using these data, and Nd and Nb/Zr data from Stracke et al. (2003), it appears that Theistareykir lavas generally have significantly lower $\delta^{18}\text{O}_{\text{ol}}$ for a given $^{143}\text{Nd}/^{144}\text{Nd}$ and Nb/Zr than lavas from the Reykjanes Peninsula (Fig. 3), though there is overlap. No Theistareykir mean phenocryst olivines (Eiler et al., 2000a) approach the normal mantle value of $\delta^{18}\text{O}_{\text{ol}} = +5.2 \pm 0.3\text{‰}$, but the range of individual phenocryst measurements includes one up to $+4.95\text{‰}$, and $\delta^{18}\text{O}_{\text{ol}} = +5.1\text{‰}$ has been reported from a Theistareykir cumulate xenolith (Maclennan et al., 2003). These authors have shown that the MgO- $\delta^{18}\text{O}_{\text{ol}}$ correlation reported by Eiler et al. (2000a) is not significant when the range in $\delta^{18}\text{O}$ in individual olivine samples is considered, but that there is a significant increase in La/Yb with decreasing $\delta^{18}\text{O}_{\text{ol}}$, similar to the increase with Nb/Zr seen on the Reykjanes Peninsula (Fig. 3B). It is notable that all the Theistareykir samples analysed by Stracke et al. (2003) would have been classified as depleted samples if they had come from the Reykjanes Peninsula.

Our two samples from Hekla were derived from parents with $\delta^{18}\text{O}_{\text{ol}}$ around $+4.1$ to 4.3‰ , based on $\delta^{18}\text{O}_{\text{pl}} = +4.9$ to $+5.1\text{‰}$, but it should be noted that the correction for fractional crystallization is quite large (Section 3.2). They thus lie at the enriched extension of the Reykjanes Peninsula $^{206}\text{Pb}/^{204}\text{Pb}$ - $\delta^{18}\text{O}$ array (Fig. 6), but are offset to slightly higher $^{143}\text{Nd}/^{144}\text{Nd}$ (Fig. 3A). Our Hekla plagioclase data are consistent with conventional basalt to basic andesite $\delta^{18}\text{O}_{\text{WR}}$ of $+4.7$ to 5.1‰ (Sigmarsson et al., 1992a). Our data for the low-MgO, voluminous quartz-tholeiites of Veidivötn, Laki (Grimsvötn), and Krafla have low to very low $\delta^{18}\text{O}_{\text{ol}}$ and $\delta^{18}\text{O}_{\text{pl}}$, consistent with published conventional $\delta^{18}\text{O}_{\text{WR}}$. With the two low- $\delta^{18}\text{O}$ samples from the Reykjanes Peninsula, they lie to substantially lower $\delta^{18}\text{O}_{\text{ol}}$ than the Nd- or Pb- $\delta^{18}\text{O}_{\text{ol}}$ correlations of the Reykjanes Peninsula (Figs. 3A and 6). The limited published $\delta^{18}\text{O}_{\text{WR}}$ data from the off-axis volcanic zones of Snæfellsnes and Vestmannaeyjar lie above the Nd- $\delta^{18}\text{O}$ correlation of Fig. 3, as they have low $^{143}\text{Nd}/^{144}\text{Nd}$ (0.51292–0.51304) but $\delta^{18}\text{O}$ close to MORB values ($\delta^{18}\text{O}_{\text{WR}}$ equivalent to $\delta^{18}\text{O}_{\text{ol}} = +4.7$ to $+5.2\text{‰}$, $N = 6$, if fractionation of 0.5‰ is assumed, Muehlenbachs et al., 1974; Hémond et al., 1993). However, it should be noted that two of these are very evolved (4.0 and 0.7% MgO), and a large correction for fractional crystallization would be necessary to estimate their parental $\delta^{18}\text{O}_{\text{ol}}$. Further, Hémond et al. (1993) report $\delta^{18}\text{O}_{\text{WR}}$ of $+5.75\text{‰}$ for TH29, which is not in equilibrium with the olivine reported by Eiler et al. (2000a).

4. Discussion

4.1. Crustal contamination: Reykjanes Ridge

There are two distinct sources from which some Reykjanes Ridge magmas might derive low $\delta^{18}\text{O}$ by crustal contamination: lower oceanic crust with low $\delta^{18}\text{O}$ resulting from the worldwide process of high-temperature interac-

tion with seawater, or high-latitude low- $\delta^{18}\text{O}$ meteoric water. We consider it highly unlikely that the low $\delta^{18}\text{O}$ could result from interaction with normal oceanic lower crust for the following reasons:

- Although minimum $\delta^{18}\text{O}$ measured in lower oceanic crust samples is $+1.7\text{‰}$, the lowest mean $\delta^{18}\text{O}_{\text{WR}}$ observed in an oceanic gabbro suite is $+4.0\text{‰}$ (actually from gabbroic xenoliths), and the overall mean $\delta^{18}\text{O}_{\text{WR}}$ in oceanic gabbros is $+5.3\text{‰}$ (see compilation of oceanic $\delta^{18}\text{O}$ data in Table 5). If the contaminant had $\delta^{18}\text{O}_{\text{WR}} = +4.0\text{‰}$, a minimum bulk gabbro contribution of 41% would be needed to lavas with $\delta^{18}\text{O}_{\text{ol}} = +4.5\text{‰}$ ($\delta^{18}\text{O}_{\text{magma}} \sim +5\text{‰}$).
- The rest of the oceanic crust has elevated $\delta^{18}\text{O}$, with a mean of $+8.2\text{‰}$ in the pillow basalt section (Table 5). We see little reason why contamination should be restricted to the lower crust, especially given the shallow crystallization pressures inferred for the Reykjanes Ridge by Michael and Cornell (1998).
- There is no objective reason why this process should not occur on other ridges, or in the southern part of the Reykjanes Ridge, unless the increase in crustal thickness towards Iceland (by a factor of <1.6 on our traverse, Weir et al., 2001; and represented by sampling depth in Fig. 7) results in lavas erupted closer to Iceland being more prone to crustal contamination. This mechanism is not expected to yield the abrupt decrease in $\delta^{18}\text{O}_{\text{ol}}$ seen between 550 and 220 m sampling depth, when the change in crustal thickness is gradual (Weir et al., 2001), and there is no change in $\delta^{18}\text{O}_{\text{ol}}$ between 1700 and 550 m depth (Fig. 7).
- This mechanism is unlikely to result in the good correlation observed between $\delta^{18}\text{O}_{\text{ol}}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ or other radiogenic ratios (e.g., Fig. 2). To develop a correlation, the extent of contamination has somehow to be related to the strength of Icelandic radiogenic isotope signal in the gabbros.
- This process would be likely to result in substantial increase in Cl/K ratios, as Cl is introduced to seawater-altered oceanic crust. Michael and Cornell (1998) show that Cl/K in Reykjanes Ridge lavas (0.15) is elevated above normal MORB (<0.07) throughout the ridge, and propose that some of the excess Cl is due to assimilation of oceanic crust. However, in their data there is no sign of a sudden increase in excess Cl around 62.4°N that would correspond to the marked drop we observe in $\delta^{18}\text{O}_{\text{ol}}$, while our specific samples north of 62°N with low $\delta^{18}\text{O}_{\text{ol}}$ have Cl/K = 0.09–0.15 (XRF data from Murton et al., 2002), strongly suggesting minimal interaction with chlorine-rich oceanic crust. 14D3 and 17D1, the samples with lowish $\delta^{18}\text{O}_{\text{ol}}$ at $\sim 57^\circ\text{N}$, appear to have elevated Cl and Cl/K, but so too 02D6 and 36D5, so there is no relationship with $\delta^{18}\text{O}$ (Fig. 7B). These four samples were prepared in an initial sample batch in which sample chips were not water-washed prior to grinding. Chips from samples chemically identical

Table 5
Compilation of published whole-rock $\delta^{18}\text{O}$ for ocean crust lithologies

	Association	Age (Ma)	Location	Ultramafic			Gabbros			Sheeted Dykes			Pillow Basalts		
				Min.	Max.	Mean	Min.	Max.	Mean	Min.	Max.	Mean	Min.	Max.	Mean
Cocker et al. (1982)	Ophiolite	15	Macquarie Island	3.2	4.9	4.1 ^b	3.6	5.7	4.6	4.8	8.8	6.0	7.1	9.7	8.4
Gregory and Taylor (1981)	Ophiolite	97	Samail—Ibra	—	—	—	3.7	6.4	5.0	4.9	11.3	8.2	10.7	12.5	11.7
Stakes and Taylor (1992)	Ophiolite	97	Samail—Wadi Hilti	—	—	—	5.3	6.4	5.6	5.4	10.0	7.9	8.4	12.4	10.2
Stakes and Taylor (1992)	Ophiolite	97	Samail—W. Shafan	—	—	—	3.0	7.0	5.4	3.0	10.5	6.6	—	—	—
Stakes and Taylor (1992)	Ophiolite	97	Samail—W. Rajmi	—	—	—	1.7	6.7	4.9	6.0	8.0	6.6	—	—	7.8
Stakes and Taylor (1992)	Ophiolite	97	Samail weighted mean	—	—	—	—	—	5.3	—	—	7.0	—	—	9.6
Agrinier et al. (1988)	Ophiolite	120	Xigaze, Tibet	2.8	6.4	5.2	4.8	11.5	8.0	7.2	9.5	8.0	9.6	15.6	13.0
Lecuyer and Fourcade (1991)	Ophiolite	420	Trinity, California	—	—	—	4.9	9.0	5.9	4.8	5.8	5.3	8.5	10.1	9.4
Muehlenbachs et al. (2004)	Ophiolite	443	Solund-Stavfjord	—	—	—	2.4	5.9	4.2	3.0	7.1	4.5	5.0	8.8	6.6
Holmden and Muehlenbachs (1993)	Ophiolite	2000	Purtunij	—	—	—	3.9	7.0	5.7	6.1	7.4	6.7	6.3	8.0	7.2
Hansteen and Troll (2003)	Xenoliths in OIB	170	Gran Canaria	—	—	—	3.3	5.1	4.0	4.1	8.6	5.9	4.1	8.6	5.9
Alt et al. (1986, 1995)	Drill core	6	504B, Costa Rica rift	—	—	—	—	—	—	3.5	5.2	4.4	6.1	12.7	7.0
Hart et al. (1999)	Drill core	12	735B, SEIR	—	—	—	2.0	5.8	4.4	—	—	—	—	—	—
Staudigel et al. (1995)	Drill core	120	Bermuda Rise	—	—	—	—	—	—	—	—	—	7.7	19.2	10.0
Lecuyer and Reynard (1996)	Drill core	10	Hess Deep, EPR	—	—	—	2.2	6.5	4.9	—	—	—	—	—	—
Agrinier et al. (1995)	Exposed rift	10	Hess Deep, EPR	3.3	10.0	5.5	2.9	8.0	5.6	4.5	7.2	5.4	3.3	8.2	6.1
mean						4.4 ^a			5.3			6.2			8.4
Putlitz et al. (2000)	eclogite/blueschist	?80	Cyclades	—	—	—	3.4	6.5	5.1	—	—	—	10.7	13.7	12.3
Barnicoat and Cartwright (1997)	eclogite facies		W Alps	—	—	—	—	—	4.8	—	—	—	—	—	—

$\delta^{18}\text{O}$ is ‰ relative to SMOW.

^a Mean excluding value of 10.0.

^b Only two data points.

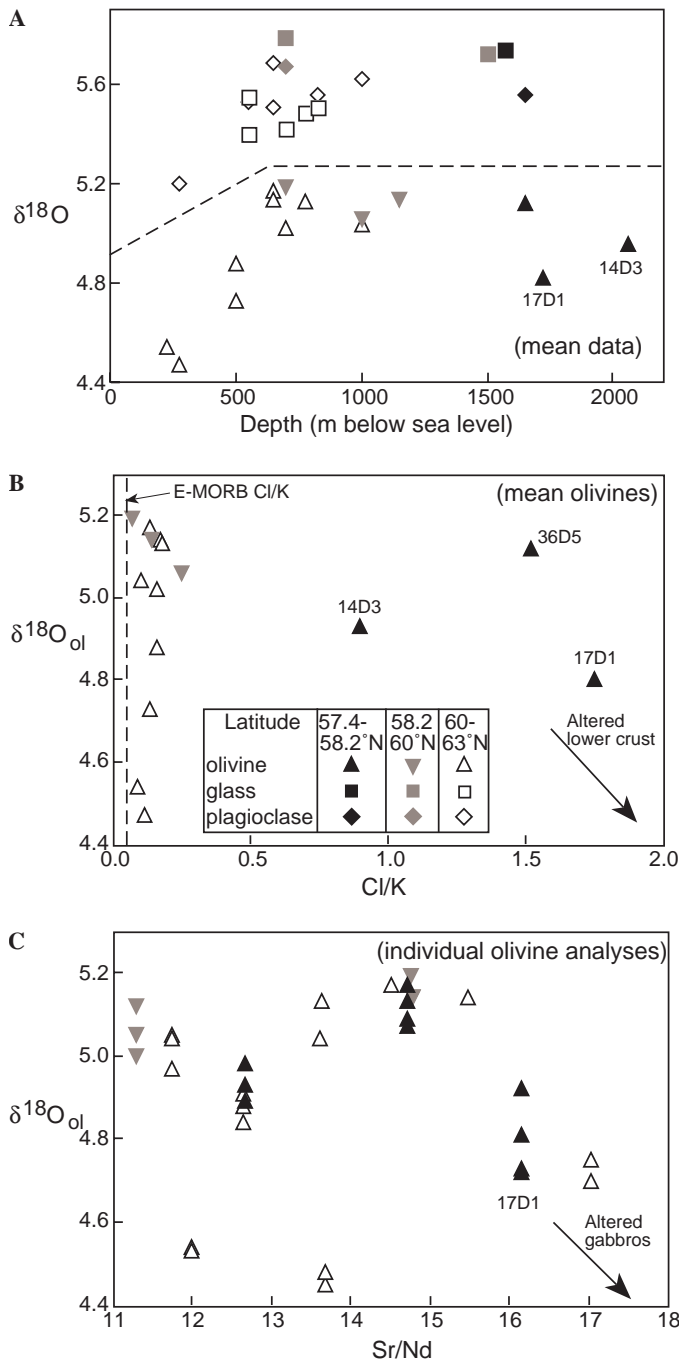


Fig. 7. Relationships between $\delta^{18}\text{O}$ and (A) sampling depth (B) Cl/K and (C) Sr/Nd in Reykjanes Ridge lavas. (A) Mean $\delta^{18}\text{O}$ is shown for olivine phenocrysts below the dashed line, and for plagioclase and glass above the line. (B) Mean olivine data are shown and in (C) individual olivine analyses. 14D3 and 17D1 are samples with anomalously Icelandic Sr–Nd isotope ratios and $^3\text{He}/^4\text{He}$ elevated above background in the southern section of the ridge. Cl/K and Sr/Nd data from Murton et al. (2002), with Cl content increased by 50 ppm to account for a negative intercept in the XRF calibration. E-MORB Cl/K from Michael and Cornell (1998); high Cl/K in samples from $<58.2^\circ\text{N}$ is because they were prepared without the chips being water-washed. Samples from the same localities give normal Cl/K after water washing. Low- $\delta^{18}\text{O}$ samples are not exclusively from shallow depths and do not show elevated Cl/K or Sr/Nd that might suggest contamination by altered gabbroic lower crust.

to 14D3, and from the same seamount, have since been analysed following a water wash, and their Cl/K is much lower at 0.12–0.25.

- There is no indication of higher Sr/Nd, which may be produced through contamination by lower crustal gabbroic cumulates, in samples with low $\delta^{18}\text{O}$ (Fig. 7). Instead, Sr/Nd in all Reykjanes Ridge samples is strongly negatively correlated with moderately incompatible elements such as Y or Zr and positively correlated with Al_2O_3 and MgO, indicating control by plagioclase fractionation (data from Murton et al., 2002).

We now discuss mechanisms by which samples with low $\delta^{18}\text{O}_{\text{ol}}$ could possibly have seen high-latitude meteoric water in their sources or during ascent:

- Is it possible that the parts of the Reykjanes Ridge which show low $\delta^{18}\text{O}$ were at some point emergent and thus developed a meteoric water hydrothermal circulation? Samples with $\delta^{18}\text{O}_{\text{ol}} < +5.0\text{‰}$ north of 62.4° were dredged from depths of 220 to 500 m below sea level, while sea-level was at most 140 m below the present value during the last 500ka (Rohling et al., 1998). The two samples at $\sim 57.6^\circ\text{N}$ with anomalously Icelandic radiogenic isotope ratios, and $\delta^{18}\text{O}_{\text{ol}} < +5.0\text{‰}$, were dredged from depths of 1720 to 2060 m below sea level. Sidescan sonar was used to locate dredging on the ridge axis, resulting in 7 of 9 samples analysed for ^{226}Ra being <8 ka in age (Peate et al., 2001), and most are from identifiable volcanic constructs (e.g., Murton et al., 2002). There is no evidence for tectonic subsidence of the Reykjanes ridge crest.
- Could local seawater in the last ~ 10 ka have had abnormally low $\delta^{18}\text{O}$ as a consequence of low- $\delta^{18}\text{O}$ meltwater from North Atlantic deglaciation? Because this was freshwater, it would have remained close to the surface and not mixed much with denser deeper water. Modern North Atlantic seawater shows limited $\delta^{18}\text{O}$ variation of $+0.1$ to 0.3‰ relative to V-SMOW (Frew et al., 2000) at all depths, except within ~ 400 km of the east Greenland coast, where the upper ~ 50 m shows $\delta^{18}\text{O}$ decreasing from -1 to -2‰ as the coast is approached (Azetsu-Scott and Tan, 1997). These latter workers sampled waters near the head of a 75-km long Greenland fjord with active glaciers and found that, although surface water sometimes had $\delta^{18}\text{O}$ as low as -15‰ , at 100 m it was no more than -1‰ . Even if the meltwater body extended to greater depths after deglaciation, only a short period of time (perhaps ~ 1 ka) would be available for hydrothermal systems to develop with this low- $\delta^{18}\text{O}$ seawater, before it was replaced by normal seawater. We conclude that deglaciation meltwater is very unlikely to have affected the Reykjanes Ridge crust at 220–500 m modern depths, much less at the ~ 1700 m of 17D1. It should be remembered that deglaciation caused sea-level rise. The Cl/K argument discussed above applies equally to this mechanism.

- Could meteoric water from Iceland move laterally under the Reykjanes Ridge to 63°N? This would require a hydraulic head from the Reykjanes Peninsula and an aquifer along which the water could move. The Reykjanes Peninsula has a relatively low topography (mostly <200 m), and it is difficult to see how water could move some 120 km offshore along the volcanically active ridge axis. Further, hydrothermal fluid from the two westernmost geothermal fields on the Reykjanes Peninsula has salinity and $\delta^{18}\text{O}$ similar to those of seawater (Arnórsson, 1978; Sveinbjörnsdóttir et al., 1986), indicating water movement in the opposite direction.
- Could the samples with low $\delta^{18}\text{O}$ represent melts that have moved laterally ~120 km along the ridge after having interacted with Icelandic hydrothermal systems (at <20 km depth)? These samples have radiogenic isotope ratios, Nb/Zr and $\delta^{18}\text{O}_{\text{ol}}$ identical to some of the enriched lavas on the Reykjanes Peninsula (Fig. 3; Thirlwall et al., 2004), and it has been suggested that magmas can be transferred laterally some 70 km along the Icelandic rift zones (e.g., the Laki fissure from Grimsvötn, Sigurdsson and Sparks, 1978). However, the spreading direction and fissure orientation are strongly oblique to the orientation of the Reykjanes Ridge (e.g., Murton, 1994), restricting shallow melt transfer to at most a few km along the ridge. Most Reykjanes Ridge magmas show SiO_2 higher than those from the Peninsula, indicating shallower average melting depths, and apart from 185D4, the samples with low $\delta^{18}\text{O}$ are no exception (data from Gee, 1998; Murton et al., 2002). This indicates that these are normal Reykjanes Ridge magmas.

Although it is impossible to rule contamination out completely as a cause of low $\delta^{18}\text{O}$ in any specific sample, the close relationship observed between the Icelandic Pb–Nd–Sr–He isotope signatures and low $\delta^{18}\text{O}$, both at >62.5°N and at 57.5–58°N (Figs. 2 and 3), provides very strong evidence that low $\delta^{18}\text{O}$ in these sections of the Reykjanes Ridge is a mantle feature and is specifically derived from Iceland.

4.2. Crustal contamination: Iceland

The presence of basaltic crust that has been extensively altered in meteoric hydrothermal systems, resulting in very low $\delta^{18}\text{O}$, means that it is much harder to rule out contamination as a mechanism for producing low- $\delta^{18}\text{O}$ magmas in Iceland. However, the conclusion from Section 4.1, that Iceland is supplying low- $\delta^{18}\text{O}$ mantle to the Reykjanes Ridge, means that we should seek evidence for both the role of low- $\delta^{18}\text{O}$ mantle and low- $\delta^{18}\text{O}$ crust in Icelandic magmas. Most recent models for crustal contamination in Iceland suggest that the contaminant is average basaltic crust or rhyolitic melts therefrom. These would have intermediate $^{143}\text{Nd}/^{144}\text{Nd}$ of ~0.51303 (e.g., Hémond et al., 1993), and Pb isotope ratios that would probably vary regionally (Thirlwall et al., 2004), and thus any crustal contribution

would have little effect on these isotopic systems, except on extreme parent melt compositions. $^{87}\text{Sr}/^{86}\text{Sr}$ is relatively high in geothermal fluids and higher in altered than in unaltered crust (0.7037–0.7042 in hydrothermally altered rocks from the Reykjanes Peninsula, Elderfield and Greaves, 1981), as a result of some incorporation of seawater Sr. This is expected to lead to elevated $^{87}\text{Sr}/^{86}\text{Sr}$ in contaminated samples (Hémond et al., 1993; Gee et al., 1998). Although drillcore samples of altered basalt from the Krafla geothermal system show extreme $\delta^{18}\text{O}_{\text{WR}}$ of –3 to –10‰, altered core material from Plio-Pleistocene basalts near Reykjavik and from the active Reykjanes hydrothermal system shows much less depleted values of +0.8 to +4.8 and +2.8 to +4.4‰, respectively (Hattori and Muehlenbachs, 1982; Sveinbjörnsdóttir et al., 1986), consistent with seawater involvement in Reykjanes hydrothermal systems (Arnórsson, 1978). Gautason and Muehlenbachs (1998) have argued that rift zone rhyolites with $\delta^{18}\text{O}_{\text{WR}} = +2 \pm 1\%$ provide the best estimate of $\delta^{18}\text{O}$ in bulk altered crust. Decrease in the ($^{230}\text{Th}/^{238}\text{U}$) activity ratio has also been used as a tracer of crustal contamination in Iceland (e.g., Nicholson et al., 1991; Sigmarsson et al., 1992b), since it is elevated in MORB but expected to be equilibrium in the altered crust. We consider now each group of Icelandic samples identified by geography in Figs. 3 and 6.

Some Reykjanes Peninsula lavas have nearly identical Pb–Sr–Nd–O isotopic compositions to the most enriched, low- $\delta^{18}\text{O}$ samples from the Reykjanes Ridge (Figs. 3 and 6 and Thirlwall et al., 2004). This implies that the $\delta^{18}\text{O}_{\text{ol}} \approx +4.3\%$ observed in Reykjanes Peninsula samples with low $^{143}\text{Nd}/^{144}\text{Nd}$ (e.g., RP100C, RP91L) is a mantle characteristic. This is supported by the following arguments:

- If the contaminant has $\delta^{18}\text{O}_{\text{WR}} = +2\%$ (from Gautason and Muehlenbachs, 1998; which is lower than $\delta^{18}\text{O}$ observed in altered crust at the Reykjanes hydrothermal system), these lavas require some 24% bulk crustal assimilation. This is only likely to be possible through coupled fractional crystallization, which is not easy to reconcile with the primitive (10.5–11.6% MgO, Fig. 4) and only weakly accumulative character of these samples.
- Although RP100C is from the eastern Reykjanes Peninsula where hydrothermal systems show less seawater input (Arnórsson, 1978), possibly leading to lower $\delta^{18}\text{O}$ in altered crust, RP91L, RP95B, and RP98A have similar Nd–O compositions and are from the west, where hydrothermal systems are seawater-dominated.
- The most extreme samples have $^{143}\text{Nd}/^{144}\text{Nd} = 0.512964\text{--}0.512979$, within error of the lowest values reported from Iceland, with the most radiogenic Pb on the Peninsula (Figs. 3 and 6). Since Icelandic crust has significantly higher $^{143}\text{Nd}/^{144}\text{Nd}$, contamination cannot cause the correlations between $\delta^{18}\text{O}_{\text{ol}}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ or $^{206}\text{Pb}/^{204}\text{Pb}$ observed in the more magnesian lavas of the Reykjanes Peninsula (dashed fields in Figs. 3 and 6).

- The extreme samples (RP91L, RP100C) have low $^{87}\text{Sr}/^{86}\text{Sr}$ for their $^{143}\text{Nd}/^{144}\text{Nd}$ compared with other Icelandic localities and 37–45°N Atlantic MORB (Fig. 5; Dosso and Thirlwall, in preparation; Thirlwall et al., 2004), implying minimal contribution of alteration-affected Sr.

Several primitive incompatible-element depleted samples from the Reykjanes Peninsula also have low $\delta^{18}\text{O}_{\text{oi}}$, around +4.5‰, which would require ~19% bulk assimilation of crust with $\delta^{18}\text{O}_{\text{WR}} = +2\text{‰}$. These possess $^{206}\text{Pb}/^{204}\text{Pb}$ higher than those of depleted samples with mantle-like $\delta^{18}\text{O}_{\text{oi}}$ (Fig. 6), which could suggest assimilation of high- $^{206}\text{Pb}/^{204}\text{Pb}$ crust. However, the very low incompatible element concentrations in these samples (e.g., 0.21 ppm Rb in RP55C) mean that any contaminant must also have very low incompatible element concentrations (e.g., <1.1 ppm Rb), well below the range of the much more abundant incompatible element-enriched Reykjanes Peninsula samples and hence well below average local crust. This is illustrated for Nb/Zr ratios in Fig. 3. Use of a crustal melt as a contaminant would make the situation far worse, as this would be more enriched in incompatible elements at similar $\delta^{18}\text{O}$. An assimilant with much lower $\delta^{18}\text{O}_{\text{WR}}$ could be proposed, but there is no evidence for this in the Reykjanes Peninsula crust. An incompatible-element-depleted low- $\delta^{18}\text{O}$ assimilant could be generated as a residue from partial melting of average crust, but more than 30% melting using reasonable partition coefficients would be needed. Assimilation of ~19% of this residue would be an ad hoc mechanism to explain these low- $\delta^{18}\text{O}$ depleted samples, but is inconsistent with their magnesian character, inconsistent with relatively low $^{87}\text{Sr}/^{86}\text{Sr}$ for their $^{143}\text{Nd}/^{144}\text{Nd}$ (Thirlwall et al., 2004), inconsistent with similar Rb/La ratios to depleted samples with normal mantle $\delta^{18}\text{O}_{\text{oi}}$ (Thirlwall, unpublished data), and also cannot explain the overall Reykjanes Peninsula $\delta^{18}\text{O}$ – $^{143}\text{Nd}/^{144}\text{Nd}$ correlation.

Our Hekla samples lie close to the $^{143}\text{Nd}/^{144}\text{Nd}$ – and $^{206}\text{Pb}/^{204}\text{Pb}$ – $\delta^{18}\text{O}_{\text{oi}}$ correlations of most Reykjanes Peninsula samples (Figs. 3A and 6), suggesting that their inferred parental $\delta^{18}\text{O}_{\text{oi}}$ of +4.3‰ (by calculation from plagioclase) is also a mantle feature. This is consistent with constant conventional $\delta^{18}\text{O}_{\text{WR}}$ of +4.7‰ in basalts from the Hekla area, which increases slightly with differentiation (Sigmarsson et al., 1992a). We cannot rule out the possibility that the constant low $\delta^{18}\text{O}$ of Hekla basalts is a consequence of contamination of their precursors in the lower crust, followed by shallow fractional crystallization, as proposed by Sigmarsson et al. (1992a). However, it seems implausible that this process should yield near-constant $\delta^{18}\text{O}$ values, particularly in Pleistocene basalts which were erupted before the Hekla shallow plumbing system was developed (Sigmarsson et al., 1992a). These authors also used the low ($^{230}\text{Th}/^{238}\text{U}$) values of Hekla basalts as evidence that they had undergone lower crustal assimilation. These are very similar to ($^{230}\text{Th}/^{238}\text{U}$) values reported from other

south central Iceland locations by Kokfelt et al. (2003) and consistent with their model of a regional melting control on ($^{230}\text{Th}/^{238}\text{U}$).

Theistareykir samples (Fig. 3; Eiler et al., 2000a; Stracke et al., 2003; this paper) in general show $\delta^{18}\text{O}_{\text{oi}}$ lower than that of the main Reykjanes Peninsula $^{143}\text{Nd}/^{144}\text{Nd}$ – $\delta^{18}\text{O}_{\text{oi}}$ array which could suggest crustal contamination. However, they show a Nd–O isotope correlation broadly parallel to that of the Reykjanes Peninsula (Fig. 3, excluding our analysis of TH29), and mostly have high MgO contents similar to those of Reykjanes Peninsula lavas. It has been argued that $\delta^{18}\text{O}_{\text{oi}}$ decreases with decreasing MgO, indicating AFC processes (Eiler et al., 2000a), but MacLennan et al. (2003) have shown that there is no significant correlation when the range in $\delta^{18}\text{O}_{\text{oi}}$ within individual samples is considered. It is difficult to envisage that sufficient crustal contamination could have occurred to generate the low $\delta^{18}\text{O}_{\text{oi}}$ of Theistareykir picrites (~+4.6‰; Eiler et al., 2000a) without gross modification of their highly depleted incompatible element chemistry. For example, for crustal $\delta^{18}\text{O} = +2\text{‰}$ (Gautason and Muehlenbachs, 1998), the picrites would require ~16% crustal assimilation, which would require a contaminant with <1 ppm Rb, lower than observed in the more incompatible-element-enriched lavas at Theistareykir (Stracke et al., 2003), which themselves must have Rb lower than that of the incompatible-element-enriched contaminant proposed by Eiler et al. (2000a).

Samples from the large quartz-tholeiite flows of Krafla, Veidivötn, and Laki (Grimsvötn), as well as the old large-volume Reykjanes Peninsula flow RP59F, and the most evolved Reykjanes Peninsula sample analysed (RP65H), have much lower $\delta^{18}\text{O}_{\text{oi}}$ for a given $^{143}\text{Nd}/^{144}\text{Nd}$ or Nb/Zr than the main Reykjanes Peninsula correlation (Fig. 3). Coupled with their lower MgO contents (Fig. 4, apart from RP59F), this could suggest that they have assimilated substantial amounts of low- $\delta^{18}\text{O}$ Icelandic crust, as proposed by Hémond et al. (1988), Nicholson et al. (1991), and Sigmarsson et al. (1991, 1992b). This could be supported by the slightly higher $^{87}\text{Sr}/^{86}\text{Sr}$ observed in these samples for a given $^{143}\text{Nd}/^{144}\text{Nd}$ (Fig. 5), as hydrothermally altered crust will have elevated $^{87}\text{Sr}/^{86}\text{Sr}$. However, primitive basalts (~10.5% MgO) from Kistufell in the Eastern rift zone have Nd–Pb–O isotope ratios similar to our Veidivötn samples (Figs. 3 and 6), and Breddam (2002) argued strongly that their low $\delta^{18}\text{O}$ was a mantle feature. It is also difficult to generate the very low $\delta^{18}\text{O}$ of some of these large-volume flows. Sigmarsson et al. (1991) invoked ~20% bulk assimilation of lower crust with $\delta^{18}\text{O} = -7.2\text{‰}$ to explain the very homogeneous low $\delta^{18}\text{O}$ of the Laki flows, but this is much lower than $\delta^{18}\text{O}$ in any Icelandic rhyolite thought to be generated by crustal melting. Much greater crustal assimilation cannot preserve the observed U–Th disequilibrium. Indeed, Kokfelt et al. (2003) argued that the consistency of ($^{230}\text{Th}/^{238}\text{U}$) in individual Icelandic regions was evidence against significant control of this ratio by crustal assimilation. A similar argu-

ment could be used for oxygen based on the constancy of $\delta^{18}\text{O}_{\text{WR}}$ in samples from Grimsvötn erupted between 1783 and 1998 (Sigmarsson et al., 2000), and we note that MacLennan et al. (2003) have recorded $\delta^{18}\text{O} = +3.2\text{‰}$ in Fo_{88-90} olivines from Theistareykir, which could suggest a mantle source this low.

The heterogeneity in $\delta^{18}\text{O}$ observed in olivine phenocrysts from single Icelandic lava flows (this study; Eiler et al., 2000a) could be regarded as evidence for crustal contamination. However, melt inclusions in primitive olivines from single samples at Theistareykir show variations in incompatible element ratios that span the full range from strongly depleted to enriched (MacLennan et al., 2003), implying that the bulk lavas are mixtures of melts generated from a variety of sources. Since individual phenocrysts typically only contain a limited range of melt inclusion chemistry, it follows that small samples (2–4 grains) of coarse olivine phenocrysts may well not be representative of the average source of the bulk lava.

In summary, although it is impossible to exclude completely crustal interaction as a cause of low $\delta^{18}\text{O}$ for any Icelandic sample, we consider that the Nd–O and Pb–O correlations for relatively primitive Reykjanes Peninsula samples are almost certainly mantle features. In contrast, the very low $\delta^{18}\text{O}$ in some Eastern Rift Zone samples is most likely of crustal origin. We find the close similarity of O–Nd–Sr–Pb isotope ratios in incompatible-element-enriched Reykjanes Peninsula and Ridge lavas particularly compelling evidence for low- $\delta^{18}\text{O}$ mantle. An incompatible-element-enriched Icelandic mantle source with $\delta^{18}\text{O}_{\text{oi}} \approx +4.3\text{‰}$ is supported by comparably low $\delta^{18}\text{O}_{\text{oi}}$ in Fo_{88-90} olivines from Theistareykir cumulate xenoliths, with lower $\delta^{18}\text{O}$ in olivine hosting the most LREE-enriched melt inclusions (MacLennan et al., 2003); by glass $\delta^{18}\text{O} \approx +4.6\text{‰}$ in samples from Hengill with mantle-like $^{40}\text{Ar}/^{36}\text{Ar}$ and elevated $^3\text{He}/^4\text{He}$ (Burnard and Harrison, 2005), and $\delta^{18}\text{O}_{\text{oi}}$ of $+4.2\text{‰}$ in a strongly LREE-enriched primitive basalt with $^3\text{He}/^4\text{He} = 32 \times R_a$ in eastern Iceland (Macpherson et al., 2005). There is also good evidence of a more depleted mantle source with $\delta^{18}\text{O}_{\text{oi}}$ of $+4.0\text{‰}$ at Kistufell in the Eastern Rift Zone (Breddam, 2002).

Our interpretations have changed significantly since the work of Gee et al. (1998) on the Reykjanes Peninsula. The main reasons for this change are:

- Re-analysis of $\delta^{18}\text{O}_{\text{oi}}$ in RP67E, the sample with lowest $^{143}\text{Nd}/^{144}\text{Nd}$ reported by Gee et al. (1998), giving $+4.80 \pm 0.03\text{‰}$ (2se, $N = 4$), instead of the $+4.94\text{‰}$ reported by these authors. There is now no evidence for a source with normal mantle $\delta^{18}\text{O}$ in enriched Reykjanes Peninsula lavas.
- Extension of the dataset to a much lower $^{143}\text{Nd}/^{144}\text{Nd}$ in primitive samples (RP100C) that also have low $\delta^{18}\text{O}_{\text{oi}}$.
- Re-analysis of Sr–Nd isotopes in several Reykjanes Peninsula samples resulting in a much tighter correlation (Fig. 5) that cannot easily be reconciled with addition of hydrothermally altered crust.

- Recognition of low $\delta^{18}\text{O}$ in samples with similar Sr–Nd–Pb isotope ratios on the Reykjanes Ridge.

4.3. $\delta^{18}\text{O}$ values and Nd contents of Reykjanes Ridge mantle components

Thirlwall et al. (2004) used high-precision Pb–Sr–Nd isotope ratios to define mixing between several mantle components in Icelandic and adjacent ridge magmas. From 60 to 63°N along the Reykjanes Ridge, the variation in these isotope ratios can be explained by mixing of two components:

RRD1, a depleted source that resembles North Atlantic MORB apart from very negative $\Delta^{207}\text{Pb}$ and somewhat higher $\Delta^{208}\text{Pb}$ (Fig. 7 of Thirlwall et al., 2004), and

RRE, an enriched source with Sr–Nd–Pb isotope ratios identical to enriched Reykjanes Peninsula lavas with $^{206}\text{Pb}/^{204}\text{Pb} \approx 18.75$.

Three out of four samples with a high Icelandic contribution south of 60°N also lie on these RRD1–RRE radiogenic isotope mixing arrays. The $^{143}\text{Nd}/^{144}\text{Nd}$ – $\delta^{18}\text{O}_{\text{oi}}$ hyperbola seen in 60–63°N Reykjanes Ridge samples (Fig. 2) must also reflect RRD1–RRE mixing and requires that RRD1 have normal mantle $\delta^{18}\text{O}$.

Table 6 and Fig. 8A show results of hyperbolic regressions through the Nd–O data, using the methodology described in Table 6. In principle, these regressions are only valid for those samples that show binary mixing in Pb–Sr–Nd space: this is primarily the 60–63°N samples with the addition of sample 14D3. Sample 17D1 must specifically be excluded from the hyperbolic fits because it lies remote from the RRD1–RRE Pb–Sr–Nd mixing arrays (Thirlwall et al., 2004) and has low $\delta^{18}\text{O}_{\text{oi}}$. Although also not on binary Pb–Sr–Nd arrays with more northerly samples, high- $^{143}\text{Nd}/^{144}\text{Nd}$ samples south of 60°N can be included in the regressions as their Nd–O isotope ratios are identical to those of high- $^{143}\text{Nd}/^{144}\text{Nd}$ samples further north: comparison of regressions 2 and 4 (Table 6) shows that these samples have no influence on the regression results. The best fit hyperbola through all the data (regression 1, Fig. 8A) lies 0.21‰ above 185D4 and systematically below most of the intermediate data points. Hyperbolae 2 and 4 exclude sample 186D2, are more strongly curved, and lie within error (maximum residual 0.1‰) of all samples from 60 to 63°N (Fig. 8A). Hyperbola 3 excludes 185D4 and provides a good fit to all other samples (maximum residual 0.13‰). As noted above, 186D2 does not lie on the binary Pb–Nd–Sr arrays for 60–63°N samples, so hyperbola 2 is the preferred model, but we consider below implications of both hyperbolae 2 and 3.

There is nothing to prevent the composition of RRE from being very close to sample 185D4. Using hyperbola 2, RRE is constrained to have $^{143}\text{Nd}/^{144}\text{Nd} > 0.51304$, but its $\delta^{18}\text{O}_{\text{oi}}$ may be significantly less than $+4.5\text{‰}$. Using hyperbola 3, RRE $^{143}\text{Nd}/^{144}\text{Nd}$ may be somewhat lower. Because oxygen concentrations are likely to be near-con-

Table 6
Summary of hyperbolic regression results for Reykjanes Ridge Nd–O data

Sample set	$^{143}\text{Nd}/^{144}\text{Nd}$ asymptote b	Grad. m	Int. c	r^2	rmsd	Max. resid.	$\delta^{18}\text{O}_D$ 0.51316	$\delta^{18}\text{O}_D$ 0.51318	$\delta^{18}\text{O}_E$	$\delta^{18}\text{O}_E$	$\delta^{18}\text{O}_E$	Nd_E/Nd_D	Nd_E/Nd_D	Nd_E/Nd_D
1. All data	0.512980	-5.92×10^{-5}	5.465	0.856	0.086	0.21	+5.14	+5.17	0.51302	0.51303	0.51304	0.51302	0.51303	0.51304
2. –186D2	0.513043	-9.44×10^{-6}	5.204	0.936	0.050	0.09	+5.12	+5.13	+3.99	+4.28	+4.48	4.5–5.0	3.7–4.0	3.0–3.3
3. –185D4	0.512994	-3.75×10^{-5}	5.355	0.904	0.059	0.13	+5.13	+5.15	0.51305	0.513055	0.51306	0.51305	0.513055	0.51306
4. 60–63°N + 14D3	0.513044	-8.62×10^{-6}	5.191	0.941	0.055	0.10	+5.12	+5.13	+3.86	+4.42	+4.65	16–18	9–11	6.6–7.5
									0.51302	0.51303	0.51304	0.51302	0.51303	0.51304
									+3.91	+4.31	+4.54	6.4–7.0	4.6–5.1	3.6–4.0
									0.51305	0.513055	0.51306	0.51305	0.513055	0.51306
									+3.76	+4.41	+4.65	20–23	11–13	7.6–8.9

Regressions were performed for four subsets of the Reykjanes Ridge $\delta^{18}\text{O}_{\text{ol}}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ data. Regression 1 used all data except for sample 17D1, which is remote from the Pb–Nd–Sr binary mixing arrays of 60–63°N samples, while in regressions 2 and 3 186D2 and 185D4 were, respectively, removed as well from the dataset. 186D2 is somewhat offset from the 60 to 63°N mixing arrays (Thirlwall et al., 2004). Regression 4 uses only data from 60 to 63°N with 14D1, which excludes 186D2 at 63.01°. Linear regressions of $\delta^{18}\text{O}$ against $1/(^{143}\text{Nd}/^{144}\text{Nd} - b)$ were calculated for each subset with the asymptote b determined to maximize the r^2 value of the linear regression. The general equation of the regression is $\delta^{18}\text{O} = c + m/(^{143}\text{Nd}/^{144}\text{Nd} - b)$. rmsd = root mean square deviation of $\delta^{18}\text{O}$ from the regression; max. resid. = maximum $\delta^{18}\text{O}$ residual in the regression, which can be compared with the 0.09‰ maximum 2 σ of replicate analyses (Table 1).

$\delta^{18}\text{O}_D$ is $\delta^{18}\text{O}_{\text{ol}}$ of the depleted mixing component RRD1 calculated from the regression at $^{143}\text{Nd}/^{144}\text{Nd}_{\text{RRD1}} = 0.51316$ and 0.51318 (0.513172 is the maximum value observed between 60 and 63°N, Thirlwall et al., 2004).

$\delta^{18}\text{O}_E$ is $\delta^{18}\text{O}_{\text{ol}}$ of the enriched mixing component RRE calculated from the regression at $^{143}\text{Nd}/^{144}\text{Nd}_{\text{RRE}}$ values shown in the table for each sample set: different values are needed for each sample set since the $^{143}\text{Nd}/^{144}\text{Nd}$ asymptote of the regression excluding 186D2 is higher than measured $^{143}\text{Nd}/^{144}\text{Nd}$ in 186D2.

Nd_E/Nd_D is the Nd concentration ratio in the enriched and depleted mixing components, calculated from the regression at $^{143}\text{Nd}/^{144}\text{Nd}_{\text{RRE}}$ values shown in the table for each sample set. The range corresponds to RRD1 varying from 0.51316 (lower Nd_E/Nd_D) to 0.51318.

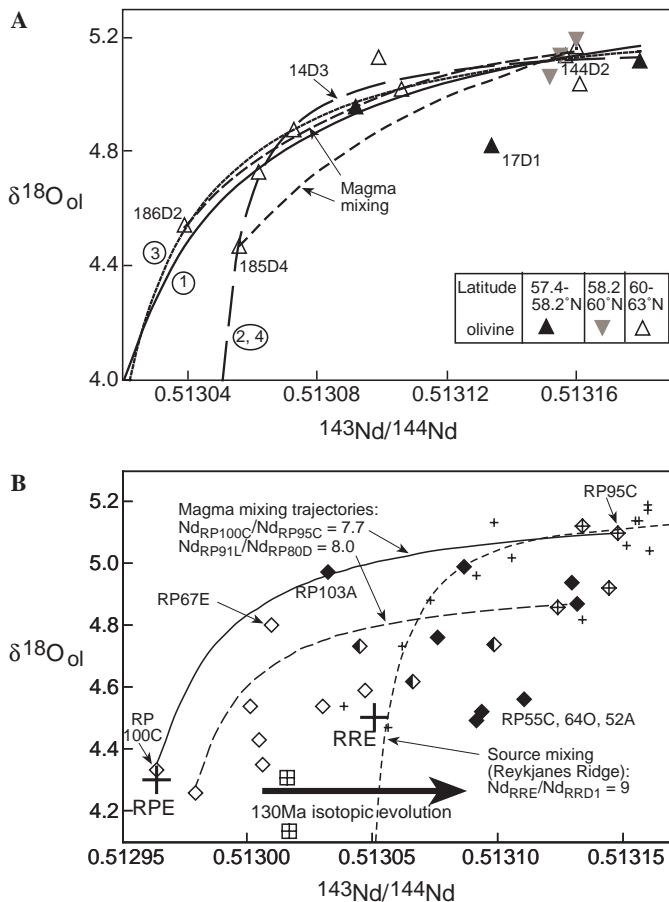


Fig. 8. Modelling of $\delta^{18}\text{O}$ – $^{143}\text{Nd}/^{144}\text{Nd}$ relationships in Reykjanes Ridge lavas (A) and in relatively primitive lavas from the Reykjanes Peninsula (B); symbols as Figs. 3, 4, and 6. All these are thought to be insignificantly affected by crustal contamination (Sections 4.1 and 4.2). $\delta^{18}\text{O}_{\text{ol}}$ is mean $\delta^{18}\text{O}$ determined on olivine phenocrysts or is parental $\delta^{18}\text{O}_{\text{ol}}$ estimated from plagioclase phenocryst data (seven samples; see Section 3.2). Nd isotope data from Thirlwall et al. (2004) and Table 2. (A) Magma mixing trajectories are shown between 144D2 and 185D4, and 186D2; and best-fit hyperbolae for (1) all data excluding 17D1, (2) all data excluding 17D1 and 186D2, (3) all data excluding 17D1 and 185D4 and (4) all 60–63°N data (186D2 is at 63.01°N). Regression parameters for these hyperbolae are given in Table 6. Hyperbolae 1 and 3 are very similar to the 144D2–186D2 melt-mixing trajectory, while hyperbolae 2 and 4 are much more strongly curved than the 185D4–144D2 melt-mixing trajectory. (B) Hyperbola 2 from A is shown for comparison, together with melt-mixing trajectories between two pairs of enriched and depleted Reykjanes Peninsula lavas: the latter are much more curved than the data array, which requires that the observed data be the product of mixing of sources with similar Nd content. The compositions of enriched end-members on the Reykjanes Ridge (RRE) and Reykjanes Peninsula (RPE) are shown (large crosses); RRE can only be generated by mixing RPE with a depleted component (like RP95C) if the latter was much less depleted in Nd than observed now in lavas. The 130Ma isotopic evolution trajectory in B shows the effect on $^{143}\text{Nd}/^{144}\text{Nd}$ of increasing Sm/Nd from 0.258 to 0.420, observed values for RP42P and RP55C.

stant in all possible mixing materials, the curvature of the hyperbolae is governed by the relative concentrations of Nd in the two mixing components, $\text{Nd}_{\text{RRE}}/\text{Nd}_{\text{RRD1}}$. If pairs of values for $\delta^{18}\text{O}_{\text{ol}}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ for RRE and RRD1 are assumed that lie on a given hyperbola, $\text{Nd}_{\text{RRE}}/\text{Nd}_{\text{RRD1}}$ may be calculated (Table 6). This value

becomes larger the further apart RRE and RRD1 are on the hyperbolae. Taking $^{143}\text{Nd}/^{144}\text{Nd}$ of RRD1 to be 0.51316 (observed in high- $\delta^{18}\text{O}$ samples 116D6, 140D6, and 144D2), and RRE to have similar $^{143}\text{Nd}/^{144}\text{Nd}$ to 185D4 or 186D2, $\text{Nd}_{\text{RRE}}/\text{Nd}_{\text{RRD1}} = 9.0$ or 3.6 is obtained for hyperbolae 2 and 3, respectively (Table 6). These values can be used to constrain the nature of the mixing components by comparison with magmatic Nd contents (data from Murton et al., 2002). Fractional crystallization causes some complications, but a rough correction may be made by normalization to constant Y (22 ppm), as Y shows no latitudinal change and correlates well with MgO. The three normal- $\delta^{18}\text{O}$ samples show normalized Nd of 3.8–4.2 ppm, similar to the bulk of samples from 60 to 61°N. 185D4 has normalized Nd of 7.8 ppm, typical of samples from 62.6 to 62.9°N, while 186D2 has normalized Nd of ~ 9.5 ppm, by far the highest in the dataset of Murton et al. (2002). Thus, if the hyperbolae are caused by mixing of melts prior to fractional crystallization, we would expect $\text{Nd}_{\text{RRE}}/\text{Nd}_{\text{RRD1}} = 2.0$ and 2.5 for mixing between 140D6 and 185D4 or 186D2, respectively. Melt-mixing hyperbolae with these values of $\text{Nd}_{\text{RRE}}/\text{Nd}_{\text{RRD1}}$ are also shown in Fig. 8A.

It is quite clear from Fig. 8A that mixing of melts with these Nd concentration ratios cannot explain the curvature of hyperbola 2, but could in principle generate hyperbola 3. Interpretation of this diagram is thus critically dependent on the exclusion of 186D2 from the 60–63°N binary mixing array. If this were valid, it would follow that the mixing components are solid mantle, with Nd contents $\sim 9\times$ higher in the mantle at 63° than at 60°N. Since magmas at 63° have $\sim 2\times$ higher Nd contents than those at 60°N, it would follow that mean melt fractions at 63° are $\sim 4.5\times$ greater than at 60°. Although this inferred change in mean melt fraction is greater than the $<1.6\times$ increase in crustal thickness from 60° to 63° (Weir et al., 2001), crustal thickness does increase by a factor of 5 towards SE Iceland and may not be closely linked to present melt fraction on a ~ 100 km distance scale. This assessment of hyperbola 2 indicates that it may be possible to demonstrate unambiguously both incompatible element enrichment and higher melt fractions in Icelandic mantle. Although these are both commonly accepted in the literature, prior assumptions are needed about the source composition (e.g., that its rare earth element pattern reflects its $^{143}\text{Nd}/^{144}\text{Nd}$). Based on Fig. 8A, the only qualification to this conclusion is the exclusion of 186D2. This could be confirmed by further high-precision analysis of samples from 62.8 to 63.5°N.

4.4. $\delta^{18}\text{O}$ values and Nd contents of Icelandic mantle components

Pb–Nd–Sr isotope ratios in Icelandic lavas can be explained by mixing between at least four mantle components (e.g., Fig. 9, Thirlwall et al., 2004). ID1 is an incompatible-element depleted component with negative $\Delta^{207}\text{Pb}$ and is dominant at Theistareykir and in Reykjanes Peninsula

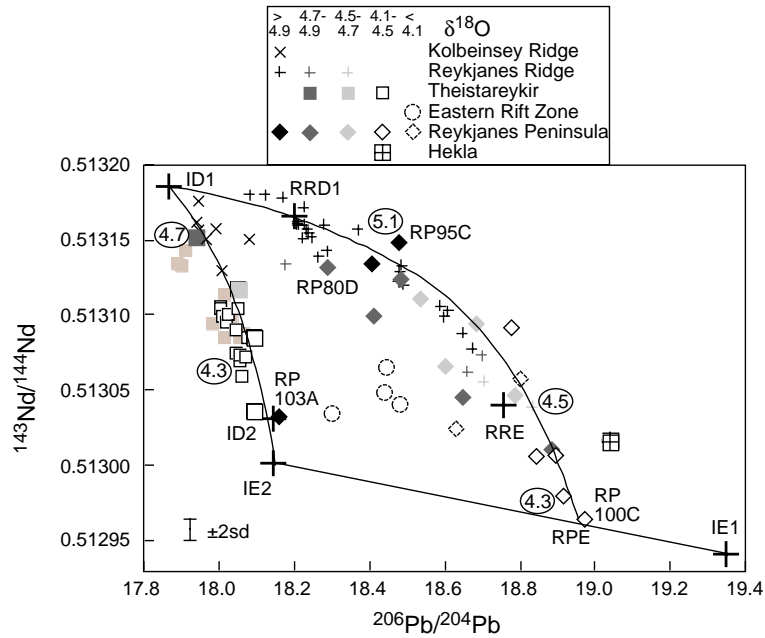


Fig. 9. $^{143}\text{Nd}/^{144}\text{Nd}$ - $^{206}\text{Pb}/^{204}\text{Pb}$ relationships in lavas from Iceland and adjacent ridges, from Thirlwall et al. (2004). Theistareykir data (smaller symbols) from Stracke et al. (2003); these yield slightly lower $^{206}\text{Pb}/^{204}\text{Pb}$ than data from Thirlwall et al. (2004) because of problems with their mass bias correction (see Baker et al., 2005). Symbols for each area are shaded according to their $\delta^{18}\text{O}_{\text{ol}}$ values, as shown in the key, and typical $\delta^{18}\text{O}_{\text{ol}}$ values are shown on the diagram in ellipses. The large crosses indicate mantle components identified by Thirlwall et al. (2004); RPE, the enriched end-member at the Reykjanes Peninsula, is at the intersection between the Reykjanes Peninsula array and the IE1-IE2 mixing line.

samples such as RP95C. It should be noted that RP95C has much higher $^{206}\text{Pb}/^{204}\text{Pb}$ than ID1, and it may be incorrect to extend the Reykjanes Peninsula mixing hyperbola so far. ID1 is distinct from RRD1 in having higher $\Delta^{208}\text{Pb}$ and $^{87}\text{Sr}/^{86}\text{Sr}$. ID2 is a second incompatible-element depleted component with strongly elevated $\Delta^{207}\text{Pb}$ and $\Delta^{208}\text{Pb}$, has some isotopic similarities to EM1, and is dominant in Reykjanes Peninsula sample RP103A. RP103A, RP95C and similar samples, and samples intermediate between these (e.g., RP80D), have $\delta^{18}\text{O}_{\text{ol}} = +4.9$ to $+5.1\text{‰}$, similar to or slightly lower than that of normal mantle. The lower $\delta^{18}\text{O}$ of $+4.5$ to $+4.7\text{‰}$ in olivine from the most depleted Theistareykir samples, and the low $\delta^{18}\text{O}$ in relatively depleted lavas at Kistufell (Fig. 3), could suggest that the depleted component in the Eastern Rift Zone has lower $\delta^{18}\text{O}_{\text{ol}}$, but olivines in Theistareykir nodules with the most depleted melt inclusions have $\delta^{18}\text{O}$ around $+5.0\text{‰}$ (Maclennan et al., 2003).

Two incompatible-element enriched, low- $^{143}\text{Nd}/^{144}\text{Nd}$ mantle components are required by the Pb-Nd isotopic relations in Icelandic lavas shown in Fig. 9, and Öraefajökull represents a third distinct enriched component with elevated $^{207}\text{Pb}/^{204}\text{Pb}$ (Prestvik et al., 2001). IE1 has $^{206}\text{Pb}/^{204}\text{Pb} > 19.3$ (required by Torfajökull, Katla and others) and IE2 has $^{206}\text{Pb}/^{204}\text{Pb} \approx 18.2$ (Fig. 9, Thirlwall et al., 2004). The composition of the local enriched mixing end-members in different Icelandic localities is a mixture between IE1 and IE2. At Theistareykir, the local enriched end-member is IE2-dominated, while at the Reykjanes Peninsula it is an IE1-IE2 mixture with $^{206}\text{Pb}/^{204}\text{Pb} \approx 18.9$

(RPE, Fig. 9). The enriched component on the Reykjanes Ridge, RRE, has $^{206}\text{Pb}/^{204}\text{Pb} \approx 18.75$ so must itself be a mixture between ID1 and RPE (e.g., Fig. 9, Thirlwall et al., 2004). Assuming that RPE provides the low $\delta^{18}\text{O}$ to RRE, it must itself have lower $\delta^{18}\text{O}$, confirmed by $\delta^{18}\text{O}_{\text{ol}} = +4.3\text{‰}$ in primitive lavas such as RP100C (Fig. 9). It follows that IE1 or IE2 or both must have $\delta^{18}\text{O}_{\text{ol}}$ similar to or lower than $+4.3\text{‰}$. The $\delta^{18}\text{O}$ value of RRE can be modelled by a 70–80% bulk contribution to RRE from the Reykjanes Peninsula enriched end-member, RPE (assuming $+5.1$ and $+4.3\text{‰}$ in ID1 and RPE, respectively). This is not at all consistent with contributions of only 10–20% estimated from the Nd isotopic compositions, if the Nd concentration ratio in the mixing materials is assumed similar to the ratio observed in enriched and depleted Reykjanes Peninsula magmas (~ 7.5).

The inference that one or both of the enriched Icelandic mantle components has $\delta^{18}\text{O}_{\text{ol}} \sim +4.3\text{‰}$ can be further investigated by considering data from elsewhere in Iceland. Hekla has a greater proportion of the IE1 component than any Reykjanes Peninsula lava, and based on our limited $\delta^{18}\text{O}_{\text{pl}}$ data and the fairly constant literature $\delta^{18}\text{O}_{\text{WR}}$, it would appear to have a mantle source with $\delta^{18}\text{O}$ similar to RP100C. The lavas with the highest IE2 component in Theistareykir also have $\delta^{18}\text{O}_{\text{ol}}$ similar to RP100C (Eiler et al., 2000a), suggesting that both IE1 and IE2 have $\delta^{18}\text{O}_{\text{ol}} \sim +4.3\text{‰}$. This conclusion is apparently contradicted by the published mantle-like conventional $\delta^{18}\text{O}_{\text{WR}}$ data from the off-axis Icelandic regions of Snæfellsnes and Vestmannaeyjar (Muehlenbachs et al., 1974; Hémond

et al., 1993), which have relatively low $^{143}\text{Nd}/^{144}\text{Nd}$ (means 0.51296 and 0.51300, respectively, Hémond et al., 1993) and radiogenic Pb ($^{206}\text{Pb}/^{204}\text{Pb} = 18.9\text{--}19.2$, Sun and Jahn, 1978, Furman et al., 1991; Chauvel and Hémond, 2000), similar to RPE or IE1. Possible reasons for this contradiction will be explored later.

Fig. 8B shows attempts to model Nd–O isotopic relationships of the more primitive samples on the Reykjanes Peninsula. Melt-mixing trajectories are strongly curved because of very low Nd contents (~ 2 ppm, Gee et al., 1998) in the incompatible-element depleted lavas with high $^{143}\text{Nd}/^{144}\text{Nd}$ and $\delta^{18}\text{O}_{\text{oi}}$. Only three samples lie close to melt mixing lines between lavas with normal mantle $\delta^{18}\text{O}_{\text{oi}}$ (e.g., RP95C) and lavas close to the Reykjanes Peninsula enriched end-member (RP100C). Of these, RP103A clearly cannot be produced by such melt mixing because of its extreme Pb isotope composition (Fig. 6; Thirlwall et al., 2004). RP67E is however close to RP95C–RP100C melt mixing curves in both Nd–O and Pb–O space (Figs. 8B and 6), and this may be an explanation of its offset to substantially higher $\delta^{18}\text{O}_{\text{oi}}$ than both Reykjanes Ridge and Peninsula Nd–O and Pb–O correlations. Identical curves would of course be generated by mixing of mantle sources that produced the depleted and enriched lavas as broadly similar melt fractions.

Some depleted lavas (e.g., RP55C, RP64O, and RP52A) have $\delta^{18}\text{O}_{\text{oi}}$ of +4.4 to +4.6‰, comparable to many of the enriched lavas, elevated $^{206}\text{Pb}/^{204}\text{Pb}$ (some > 18.6), and not much higher $^{143}\text{Nd}/^{144}\text{Nd}$ than some of the enriched lavas (Fig. 8B). RP55C, in particular, is very depleted in incompatible elements (Fig. 3B), but has $^{206}\text{Pb}/^{204}\text{Pb} = 18.78$ (Thirlwall et al., 2004). The extreme incompatible element depletion is not consistent with an origin of their low $\delta^{18}\text{O}_{\text{oi}}$ by either crustal contamination (Section 4.2) or mixing with any enriched melt composition. The oxygen data are consistent with the mechanism proposed by Thirlwall et al. (2004), in which the source of RP55C was generated fairly recently (a few tens of Ma) by melt extraction from a source similar to that of the enriched Reykjanes Peninsula lavas (Fig. 8).

The main sub-linear correlations between $\delta^{18}\text{O}_{\text{oi}}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ or $^{206}\text{Pb}/^{204}\text{Pb}$ in Reykjanes Peninsula lavas (Figs. 3, 6, and 8B) cannot be explained by magma mixing or crustal contamination because these give concave-down trajectories, as the depleted lavas and their sources have very low Nd and Pb concentrations. The correlation either has to represent continuous source heterogeneity, in which case it is very difficult to explain the tight $^{143}\text{Nd}/^{144}\text{Nd}$ – $^{206}\text{Pb}/^{204}\text{Pb}$ hyperbola (Fig. 9), or it has to be a consequence of mixing of mantle sources before the depleted sources developed their present highly incompatible-element-depleted character. Thirlwall et al. (2004) argued that the strong incompatible-element depletion in the sources of the depleted lavas had to be young, much less than 170 Ma and no more than a few tens of Ma. It appears therefore that a continuous range of source compositions were produced by mixing prior to variable source

depletion being introduced by melt extraction. This can resolve the problem with estimating contributions from ID1 and the Reykjanes Peninsula enriched end-member (RPE) to the Reykjanes Ridge enriched end-member, RRE. RRE lies on the main Reykjanes Peninsula Nd–O and Pb–O arrays (Figs. 8B and 6). If the extreme incompatible element depletion in ID1 only takes place *after* it has contributed to RRE, then the bulk contribution of RPE to RRE estimated from Nd becomes closer to that estimated from oxygen. In summary, the proposed sequence of events at Reykjanes is:

- IE1 and IE2 are mixed to produce intermediate enriched end-member RPE with $^{206}\text{Pb}/^{204}\text{Pb} = 18.95$ (Fig. 9).
- This is mixed with ID1, a mildly incompatible-element depleted mantle source, to produce mixed sources that will subsequently act as the sources of most Reykjanes Peninsula lavas (Fig. 9).
- RRE is a homogeneous mixture of these sources (Fig. 9) and is advected southwards along the Reykjanes Ridge to mix with RRD1.
- Extreme incompatible element depletion due to melt extraction then occurs in some of the mixed sources remaining beneath the Reykjanes Peninsula. Depletion seems to occur in all mantle that was originally mildly depleted and in some of the more enriched sources (e.g., RP55C).
- Similar depletion may also occur along the Reykjanes Ridge, but if so the sources are never tapped, or always subordinate to melts of less-depleted sources in magma chambers.

The relationships for Theistareykir (Fig. 3) are less easily interpreted, given the heterogeneity in olivines within single flows. However, young (100–130 Ma) Lu–Hf, Sm–Nd, and ^{238}U – ^{206}Pb pseudoisochrons (data from Stracke et al., 2003) suggest that here too the extreme incompatible-element depletion in some sources is young, allowing these to develop in a few tens of Ma from more enriched low- $\delta^{18}\text{O}$ sources.

4.5. Oxygen isotope constraints on oceanic crust recycling models

Of the five mantle components in North Atlantic magmas identified by Thirlwall et al. (2004), the evidence presented here suggests that both enriched components (IE1 and IE2) possess low $\delta^{18}\text{O}_{\text{oi}}$ of at most +4.3‰, while the three depleted components (ID1, ID2, and RRD1) possess $\delta^{18}\text{O}_{\text{oi}}$ similar to normal MORB or lithospheric mantle, around +5.1‰ (see Fig. 9). Since oxygen isotope fractionation between coexisting phases is much larger at low temperatures, the most obvious explanations of mantle $\delta^{18}\text{O}_{\text{oi}}$ distinct from +5.1‰ involve recycling into the mantle of material that has resided at the earth's surface and developed very different $\delta^{18}\text{O}$. Several studies have proposed recycled oceanic crust in Icelandic mantle sources, both

gabbroic and basaltic crust (Chauvel and Hémond, 2000; Skovgaard et al., 2001), gabbroic crust alone (Foulger et al., 2005), basaltic crust alone (Thirlwall et al., 2004) or ocean island crust (McKenzie et al., 2004). Recycled ocean crust has been used to explain specifically the low $\delta^{18}\text{O}$ in Icelandic mantle by Breddam (2002), Burnard and Harrison (2005), and Macpherson et al. (2005).

Our knowledge of the oxygen isotope composition of oceanic crust (Table 5) comes from studies of ophiolites (Gregory and Taylor, 1981; Cocker et al., 1982; Agrinier et al., 1988; Lecuyer and Fourcade, 1991; Stakes and Taylor, 1992; Holmden and Muehlenbachs, 1993; Muehlenbachs et al., 2004), drill cores (Alt et al., 1986; Alt et al., 1995; Staudigel et al., 1995; Hart et al., 1999), xenoliths (Hansteen and Troll, 2003), and dredged material at fracture zones (Agrinier et al., 1995). Most of these are whole-rock or even composite studies, in order to achieve representative sampling of heterogeneous alteration. Low-temperature alteration of the basaltic lavas in the upper crust results in increased whole rock $\delta^{18}\text{O}$ from +6 to +19‰, with average values from +6 to +13‰. Gabbroic lower crust typically undergoes high-temperature alteration, which decreases whole rock $\delta^{18}\text{O}$ to an observed minimum of +1.7‰, but average values reported from the gabbroic section range from +4.0 to +8.0‰. Only very few samples have $\delta^{18}\text{O}$ below +3.5‰. The range in values in gabbroic crust could reflect incomplete equilibration during alteration, or may be the result of subsequent low-temperature interactions, suggested by high $\delta^{18}\text{O}$ in plagioclase coupled with low $\delta^{18}\text{O}$ in clinopyroxene in some studies (Agrinier et al., 1988). The sheeted dyke complexes of ophiolites appear to have $\delta^{18}\text{O}$ intermediate between the values for basaltic and gabbroic crust, with mean values between +4.4 and +8.2‰.

Chauvel and Hémond (2000) and Skovgaard et al. (2001) used the contrasting trace element signatures (e.g., Sr/Nd, Pb/Ce, Sm/Zr, and Eu/Eu*) of depleted and enriched Icelandic lavas as primary evidence that both recycled gabbroic and basaltic ocean crust were present in the Icelandic mantle source, with gabbroic crust contributing preferentially to the sources of some or all depleted lavas. The relative proportions of recycled crust to host mantle can be constrained if we assume that the host mantle has normal $\delta^{18}\text{O}$. The overlap in $\delta^{18}\text{O}$ values of depleted Icelandic lavas (+4.8 to +5.4‰ in mantle source assuming $\Delta^{18}\text{O}_{\text{mantle-ol}} = 0.3\text{‰}$, Matthey et al., 1994) and gabbroic ocean crust (+4.0 to +8.0‰ averages) could permit, but does not require, substantial gabbroic contributions to their sources. However, the correlations between $\delta^{18}\text{O}$ and gabbro trace element indicators such as Sr/Nd and Eu/Eu* reported by Skovgaard et al. (2001) from Theistareykir are eliminated by our revised oxygen data for these samples, and Thirlwall et al. (2004) have argued that these trace element characteristics may only reflect the recent melting history of the source of the depleted lavas. The $\delta^{18}\text{O}_{\text{ol}} = +4.0\text{‰}$ of the relatively depleted primitive Kistufell lavas would appear too low

to be generated by gabbro recycling, as proposed by Breddam (2002).

The mantle sources of enriched Icelandic lavas are estimated to have $\delta^{18}\text{O}_{\text{mantle}} \leq +4.6\text{‰}$, calculated from $\delta^{18}\text{O}_{\text{ol}} \leq +4.3\text{‰}$, with $\Delta^{18}\text{O}_{\text{melt-ol}} = 0.5$ and $\Delta^{18}\text{O}_{\text{melt-source}} = 0.2$ (Matthey et al., 1994). These clearly cannot be explained by recycling any observed upper oceanic crust (mean $\delta^{18}\text{O} > 6.1$, Table 5) into host mantle with normal $\delta^{18}\text{O}$. To produce these sources by recycling, we can offer four alternatives:

1. ~60% recycling of gabbroic crust with the lowest observed mean $\delta^{18}\text{O}_{\text{WR}}$ of any observed oceanic gabbros (+4.0‰, Table 5) into normal host mantle. Despite suggestions to the contrary by Foulger et al. (2005), this is a completely unrealistic figure, given the lack of gabbro trace element signatures and the normal basaltic major element chemistry of the enriched lavas, which include compositions in equilibrium with Fo₈₇ olivine.
2. Recycling of material from the sheeted dyke zone. This would require >82% of the sheeted dyke suite with the lowest mean $\delta^{18}\text{O}_{\text{WR}}$ (+4.4‰, Table 5). This proportion of recycled basaltic material is completely incompatible with the existing Os isotope data (Skovgaard et al., 2001), and most sheeted dyke complexes have a substantially higher $\delta^{18}\text{O}_{\text{WR}}$.
3. Recycling of upper oceanic crust with average $\delta^{18}\text{O}$ of +8.4‰ (Table 5) into a highly anomalous host mantle with $\delta^{18}\text{O}_{\text{WR}} = +2.8$ to +4.1‰, based on a 32–11% recycled 450 Ma basalt component, respectively (calculated by Thirlwall et al., 2004; from the Os isotope data of Skovgaard et al., 2001).
4. Recycling of highly anomalous upper oceanic crust into normal host mantle. The proportions of recycled basalt calculated from the osmium data would require the oceanic crust to have $\delta^{18}\text{O}$ between +2.7 and –2.7‰ (32–11% crust).

There are two obvious ways to circumvent the problems associated with these alternatives. First, it is possible that the observed oceanic crustal $\delta^{18}\text{O}$ compositions in Table 5 are modified during subduction recycling. Such effects appear to be minimal based on a number of studies of oceanic crustal lithologies exhumed from eclogite/blueschist metamorphism (e.g., Barnicoat and Cartwright, 1997; Putlitz et al., 2000; Table 5). $\delta^{18}\text{O}$ values of garnet in mantle eclogites, which may represent ancient recycled crust, show occasional low values (minimum +2.5‰ in garnet), but >80% of data in a compilation by Lowry et al. (1999) are similar to or higher than normal mantle values (see also Jacob, 2004). A second possibility is that the $\delta^{18}\text{O}$ of modern oceanic crust may not be representative of ancient oceanic crust that has been recycled into the mantle (alternative 4 above). $\delta^{18}\text{O}_{\text{PDB}}$ is progressively lower in older marine carbonates, with minimum Phanerozoic values of around –8‰ in the early Ordovician (e.g., Veizer et al., 1999), which can be interpreted as indicating that ocean water

had $\delta^{18}\text{O}_{\text{SMOW}}$ of around -8‰ at this time. If so, altered Ordovician oceanic crust might be expected to have $\delta^{18}\text{O}_{\text{SMOW}}$ around -8‰ less than modern ocean crust. This might explain the low $\delta^{18}\text{O}$ of Icelandic mantle sources, given the Pb isotopic evidence that they incorporate Lower Palaeozoic ocean crust (Thirlwall et al., 2004). However, there is no evidence whatsoever of unusually low $\delta^{18}\text{O}$ in the two Lower Palaeozoic ophiolites analyzed (Lecuyer and Fourcade, 1991; Muehlenbachs et al., 2004), and these are direct representatives of material that has most probably been recycled into the Icelandic mantle.

The only substantial terrestrial rock units with very low $\delta^{18}\text{O}$ have been produced by large-scale hydrothermal alteration by low temperature meteoric fluids, either at the present day (Icelandic crust, Gautason and Muehlenbachs, 1998; upper crust of Kamchatka, Bindeman et al., 2004) or during past glacial episodes (the Dabie-Sulu high-pressure metamorphic terrain, China, Zheng et al., 2004; the Ruppert Coast terrane, Antarctica, Blattner et al., 1997). Recycling of smaller ocean islands, or ocean islands at lower latitudes (e.g., McKenzie et al., 2004) would not provide sufficiently low $\delta^{18}\text{O}$, because ^{18}O depletion would only develop via the same processes as occur in normal ocean crust. Recycling of an old upper crustal body such as the Dabie-Sulu orogen would have large effects on Sr isotopes, as these have abundant felsic and metasedimentary lithologies with high Rb/Sr. It is difficult to envisage mechanisms to recycle a large thick buoyant body such as Iceland. Sediment derived from high-latitude igneous provinces may be subducted, but sedimentary processes would probably eliminate the low $\delta^{18}\text{O}$ signature.

4.6. A deeper origin for the low- $\delta^{18}\text{O}$ signature?

It thus appears that recycling of crustal materials into the mantle cannot easily generate the low $\delta^{18}\text{O}$ that we infer to be present in enriched Icelandic mantle sources. This does not preclude a role for crustal recycling in generating other geochemical characteristics of Icelandic mantle. This is because its lithophile element signature may be dominated by recycled crust while the host mantle may dominate its $\delta^{18}\text{O}$, as oxygen has similar concentrations in mantle and crust. However, as noted in alternative 3 above (Section 4.5), the host mantle would then have to have significantly lower $\delta^{18}\text{O}$ than observed in lavas ($\delta^{18}\text{O}_{\text{WR}} = +2.8$ to $+4.1\text{‰}$). Since the small amount of He remaining in recycled crust is likely to be radiogenic, the host mantle is also likely to be the source of the elevated $^3\text{He}/^4\text{He}$ ratios common in basalts from Iceland and adjacent ridges (e.g., Condomines et al., 1983; Kurz et al., 1985; Breddam et al., 2000; Hilton et al., 2000;). If both low $\delta^{18}\text{O}$ and elevated $^3\text{He}/^4\text{He}$ were derived from the host mantle, negative correlations between these parameters might be expected for mixing between this and recycled crust. In contrast, positive correlations between these parameters are expected for crustal contamination, as altered Icelandic crust has low $\delta^{18}\text{O}$ and low $^3\text{He}/^4\text{He}$.

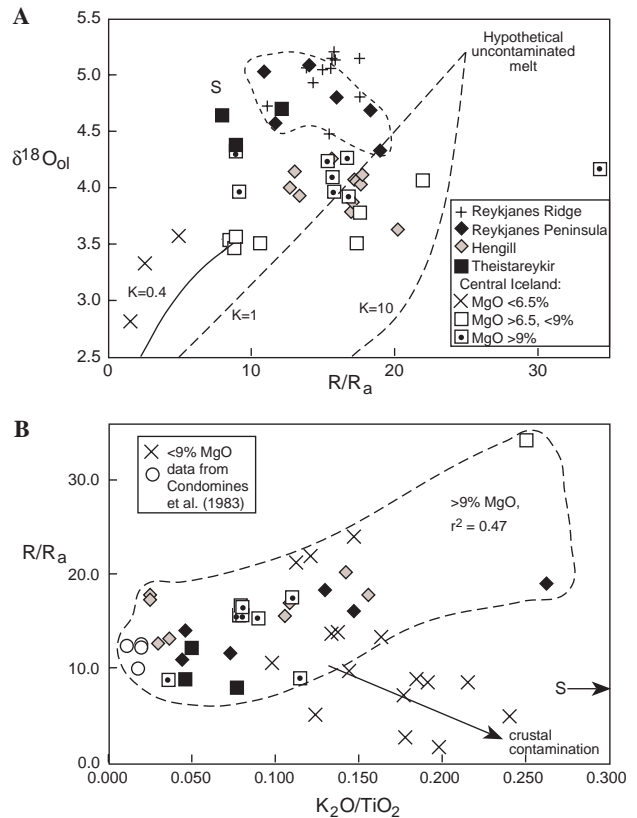


Fig. 10. Relationships between $^3\text{He}/^4\text{He}$ (expressed as R/R_a) and (A) $\delta^{18}\text{O}_{\text{ol}}$ values, (B) $\text{K}_2\text{O}/\text{TiO}_2$ ratios in Icelandic lavas. $\text{K}_2\text{O}/\text{TiO}_2$ is a measure of incompatible element enrichment (equivalent to Nb/Zr or La/Yb) which is available for most samples analysed for He. All oxygen data by laser fluorination except S = Snæfellsnes: conventional whole rock data from Muehlenbachs et al. (1974). Conventional $\delta^{18}\text{O}$ whole rock data of Condomines et al. (1983) are not shown. To calculate equivalent olivine values, 0.5‰ was subtracted from $\delta^{18}\text{O}$ glass data of Burnard and Harrison (2005) and Macpherson et al. (2005), but there would be a significant correction for fractional crystallization for some of the more evolved samples of the latter work. All He and O data are from the same samples except two Reykjanes Peninsula samples, which are from the same flows. Data sources are: Reykjanes Ridge (Hilton et al., 2000; this paper; Murton et al., 2002); Reykjanes Peninsula (He: Breddam et al., 2000; Dixon et al., 2000; O: this paper; $\text{K}_2\text{O}/\text{TiO}_2$: Gee, 1998; Skovgaard et al., 2001); Hengill (Burnard and Harrison, 2005); Theistareykir (He: Breddam et al., 2000; Macpherson et al., 2005; O: this paper; Eiler et al., 2000a; $\text{K}_2\text{O}/\text{TiO}_2$: Skovgaard et al., 2001); Central Iceland (Breddam, 2002; Macpherson et al., 2005). (B) Data from Condomines et al. (1983) have been plotted, and all data from samples $<9\%$ MgO are identified by crosses. Three crustal contamination trajectories are shown in A, using a crustal composition with $\delta^{18}\text{O} = +2\text{‰}$ (Gautason and Muehlenbachs, 1998), $R/R_a = 0.9$, and $0.09 \mu\text{m}^3 \text{He g}^{-1}$ (Condomines et al., 1983). Two trajectories are from a hypothetical uncontaminated magma with $\delta^{18}\text{O}_{\text{ol}} = +5.2\text{‰}$ and $R/R_a = 25$, and illustrate contamination of weakly and moderately degassed magmas ($\text{He}_{\text{magma}}/\text{He}_{\text{crust}} = K = 10$ and $K = 1$, respectively; note that K here is the reciprocal of that used by Macpherson et al., 2005). The third ($K = 0.4$) is from moderately evolved, strongly degassed lavas of Macpherson et al. (2005), and compares well to the behaviour of their most evolved samples. Since most of the primitive ($>9\%$ MgO) glasses are weakly degassed, appropriate contamination trajectories are steep and a wide range of R/R_a in primary magmas is required. R/R_a appears to correlate negatively with $\delta^{18}\text{O}$ as indicated by the dashed field of Reykjanes Peninsula lavas. (B) Primitive lavas show a positive correlation between R/R_a and incompatible element enrichment (dashed field, $r^2 = 0.47$) which itself correlates negatively with $\delta^{18}\text{O}$ (e.g., Fig. 3B).

There is now a substantial body of combined laser fluorination oxygen data and $^3\text{He}/^4\text{He}$ data on the same samples or the same flows (Fig. 10A). Crustal contamination trajectories on Fig. 10A depend primarily on the degassing history of the melt being contaminated. In general, primitive glasses (>9% MgO) show higher ^4He contents than likely in Icelandic crust ($\sim 0.065 \mu\text{cm}^3 \text{g}^{-1}$ in 1 Ma tholeiite, Condomines et al., 1983), and would follow linear or concave-up contamination trajectories, while more evolved glasses show generally lower ^4He contents and would follow concave-down trajectories (e.g., $K=0.4$, Fig. 10A). In principle, it is possible for all the primitive lavas to be the results of contamination of little-degassed magmas with normal mantle $\delta^{18}\text{O}$ and $^3\text{He}/^4\text{He}$ varying from $8\times$ to $35\times$ atmospheric, as illustrated by the trajectories drawn from a hypothetical mantle composition with $R/R_a=25$ (Fig. 10A). However, this conflicts with the evidence presented in this paper that the low $\delta^{18}\text{O}$ in the primitive lavas is mantle-derived. Primitive magmas with MORB-like $^3\text{He}/^4\text{He}$ and $\delta^{18}\text{O}_{\text{OI}}=+5.2\text{‰}$ are not strongly degassed and are thus unable to arrive at these compositions by following concave-down hyperbolae from a high- $^3\text{He}/^4\text{He}$ source. Therefore, there was probably a range in $^3\text{He}/^4\text{He}$ in primary melts, and there is a strong indication that melts with high $^3\text{He}/^4\text{He}$ also have low $\delta^{18}\text{O}$ (e.g., Reykjanes Peninsula array outlined in Fig. 10A, $r^2=0.32$). Burnard and Harrison (2005) interpreted the negative correlation at Hengill ($r^2=0.22$) to be the product of mixing between MORB source mantle and a low- $\delta^{18}\text{O}$ high- $^3\text{He}/^4\text{He}$ plume, but the lavas with highest $\delta^{18}\text{O}_{\text{OI}}$ do not have MORB chemistry or Pb isotope compositions (e.g., Thirlwall et al., 2004). The implication that incompatible-element enriched magmas have high $^3\text{He}/^4\text{He}$ is confirmed by a positive correlation with $\text{K}_2\text{O}/\text{TiO}_2$ for primitive magmas in Fig. 10B, as $\text{K}_2\text{O}/\text{TiO}_2$ in these is a measure of incompatible element enrichment, while more evolved magmas show a decrease in $^3\text{He}/^4\text{He}$ as $\text{K}_2\text{O}/\text{TiO}_2$ increases with fractional crystallization and contamination.

Given the evidence in Fig. 10 that primitive low- $\delta^{18}\text{O}$ magmas have a relatively high $^3\text{He}/^4\text{He}$, it is possible that a common origin should be sought for these features. Anderson (2000), and Foulger and Pearson (2001) suggested that high $^3\text{He}/^4\text{He}$ may be derived from shallow low-U and hence low- ^4He oceanic lithospheric mantle. The maximum $^3\text{He}/^4\text{He}$ on this model corresponds to the initial ratio at the time of isolation from the asthenosphere in a low U/ ^3He environment. The very high $^3\text{He}/^4\text{He}$ ratios in some Icelandic magmas (maximum $37.7\times R_a$, Hilton et al., 1999; compared with a primordial ratio of $\sim 150\times R_a$) are not consistent with the inferred young age of the Icelandic mantle source (Thirlwall et al., 2004). Limited work on the mantle part of ophiolites does show some depletion in $\delta^{18}\text{O}$, but to no greater extent than the gabbros (Table 5), and this depletion is likely to be confined to the shallowest parts. For example, in the Xigaze ophiolite, there is a 1 km thick layer of serpentinized ultramafics with ^{18}O depletion, underlain by

ultramafic rocks with normal mantle $\delta^{18}\text{O}$ (Agrinier et al., 1988). We thus do not consider oceanic mantle lithosphere capable of producing the low $\delta^{18}\text{O}$ or high $^3\text{He}/^4\text{He}$ of Icelandic magmas.

High $^3\text{He}/^4\text{He}$ is commonly regarded as originating from an undegassed reservoir in the lower mantle. Although such reservoirs might have low $\delta^{18}\text{O}$, there is no obvious reason why they should. Hawaiian lavas with high $^3\text{He}/^4\text{He}$ in general have normal mantle $\delta^{18}\text{O}$ (e.g., Eiler et al., 1997). Macpherson et al. (2000) have proposed a mechanism for ^{18}O depletion in the D'' layer at the base of the mantle, involving oxygen isotope fractionation between Mg-perovskite in the lowermost mantle and wüstite in the core. Their calculations suggested that Fe-wüstite would be lower in $\delta^{18}\text{O}$ by $\sim 0.5\text{‰}$ relative to mantle silicates at core–mantle boundary temperatures. They noted that initial core segregation is likely to have taken place at much lower temperatures and pressures in the mantle, where fractionation factors might exceed 1‰ . This is similar to the extent by which enriched Icelandic mantle is inferred to be low relative to normal upper mantle. However, if Icelandic mantle were produced by interaction between lower mantle and core, the core should dominate siderophile element abundances. Although primitive Icelandic lavas have high abundances of platinum group elements compared with those of MORB and Hawaiian lavas, these can be explained by differences in the melting regime (Momme et al., 2003).

4.7. Why is Iceland near-unique in its low $\delta^{18}\text{O}$?

The conclusion that the low- $\delta^{18}\text{O}$ signature of Icelandic magmatism is at least in part of mantle origin leads to the question of why Iceland should be near-unique among ocean islands in its near-ubiquitous low $\delta^{18}\text{O}$ in incompatible-element-enriched lavas. The Sr–Nd–Pb isotope and trace element signature of these is comparable to ocean islands described as young HIMU by Thirlwall (1997), who regarded them as the consequence of recycling Proterozoic and Palaeozoic upper ocean crust into the mantle. As discussed above, this process could not have produced mantle with sufficiently low $\delta^{18}\text{O}$ in Iceland unless the recycled upper ocean crust had mean $\delta^{18}\text{O}$ around $+0\text{‰}$, which would require $\sim 8\text{‰}$ depletion in ^{18}O in the seawater that altered the ocean crust compared with modern seawater. If this were the case, despite the contrary evidence from Palaeozoic ophiolites, it is almost inevitable that other young HIMU ocean islands would also have low $\delta^{18}\text{O}$. Limited studies so far suggest that, although some have olivines with $\delta^{18}\text{O}$ as low as $+4.7\text{‰}$ (low-MgO basalts from Gran Canaria, Thirlwall et al., 1997; low- $^3\text{He}/^4\text{He}$ OIB, Eiler et al., 1997), low $\delta^{18}\text{O}$ is far from common, nor do any OIB suites have consistently low $\delta^{18}\text{O}$. The one exception is the Kea trend lavas of Hawaii (e.g., mean $\delta^{18}\text{O}_{\text{OI}}=+4.8\text{‰}$ at Mauna Kea; minimum of $+4.4\text{‰}$ at Kilauea), where the low $\delta^{18}\text{O}$ has variously been interpreted as a result of assimilation of hydrothermally altered

Pacific crust or island edifice (Eiler et al., 1996; Garcia et al., 1998; Wang et al., 2003), or incorporation of recycled hydrothermally altered oceanic lithospheric mantle (Lassiter and Hauri, 1998; Wang et al., 2003). Their unradiogenic Os is inconsistent with significant contributions from local ~100 Ma Pacific crust (Lassiter and Hauri, 1998), but as noted previously, there is little evidence that alteration significantly affects suboceanic mantle (Table 5): in the Xigaze ophiolite, only the top 1 km of the mantle section is affected (Agrinier et al., 1988).

The only incompatible-element-enriched Icelandic lavas with apparently normal $\delta^{18}\text{O}$ (based on published conventional data) are the alkali basalts from the off-axis regions of Snæfellsnes and Vestmannæyjar. It has previously been argued (e.g., Sigmarsson et al., 1992b) that transitions to lower ($^{230}\text{Th}/^{238}\text{U}$) and lower $\delta^{18}\text{O}$ from these locations towards the rift zones reflect increasing crustal contamination, in tandem with greater melt fractions, toward central Iceland. However, at Snæfellsnes, the underlying crust was formed in an old rift zone and would be a potential low $\delta^{18}\text{O}$ contaminant. There are three possible explanations for these apparently normal $\delta^{18}\text{O}$ values that are consistent with our identification of low- $\delta^{18}\text{O}$ -enriched sources elsewhere in Iceland:

1. That the existing normal conventional $\delta^{18}\text{O}_{\text{WR}}$ data ($N = 5$ Vestmannæyjar including two hawaiites, $N = 8$ Snæfellsnes, including one trachyte; Sigmarsson et al., 1992b; Hémond et al., 1993) are not consistent with our laser fluorination data. This might be suggested by $\delta^{18}\text{O}_{\text{WR}}$ values for TH29 and Haleyjabunga 0.6–1.3‰ higher than $\delta^{18}\text{O}_{\text{ol}}$ from these flows (Hémond et al., 1993; Eiler et al., 2000a; this paper).
2. That another enriched mantle component is present in these areas that has normal (i.e., MORB-like) $\delta^{18}\text{O}$ unlike IE1 or IE2. The high- $^{207}\text{Pb}/^{204}\text{Pb}$ enriched mantle component at Örafajökull appears to be characterized by normal $\delta^{18}\text{O}_{\text{WR}}$ (Prestvik et al., 2001), but unfortunately the existing Pb isotope data from Snæfellsnes and Vestmannæyjar are too imprecise to make a comparison with Örafajökull ($\Delta^{207}\text{Pb} = +4$ to $+8$ from Chauvel and Hémond, 2000; -5 to -1 from Sun and Jahn, 1975; Furman et al., 1991). This hypothesis is supported by low $^3\text{He}/^4\text{He}$ (7 to $8 \times R_a$) in a sample each from Snæfellsnes and Örafajökull (Sigmarsson et al., 1992b) and relatively low $^3\text{He}/^4\text{He}$ at Surtsey ($11 \times R_a$, Kurz et al., 1985), substantially lower than expected from the $^3\text{He}/^4\text{He}$ – $\text{K}_2\text{O}/\text{TiO}_2$ correlation of Fig. 10B.
3. That the normal $\delta^{18}\text{O}$ is in some way related to the smaller melt fractions at Snæfellsnes and Vestmannæyjar, in which respect they more resemble basalts from other young HIMU islands than the rift zone tholeiites. Since the ultimate source of the low $\delta^{18}\text{O}$ in the rift zone magmas cannot easily be the same as the source of the radiogenic Pb and relatively unradiogenic Nd, it may be that the Pb and Nd are hosted in discrete pockets of relatively fertile recycled material in a more refractory low- $\delta^{18}\text{O}$ mantle matrix. If so, then the recycled material could preferentially melt where melt fractions are kept low by a thick lithospheric lid, while larger melt fractions in the rift zones may permit melting of the host low- $\delta^{18}\text{O}$ mantle. This could potentially explain the unique low- $\delta^{18}\text{O}$ of Icelandic magmatism. However, off-axis or other young HIMU magmas would then be expected to have high $\delta^{18}\text{O}$ from recycled upper crust, and there is no evidence for incorporation of a low- $\delta^{18}\text{O}$ residue into the oceanic lithosphere.

5. Conclusions

- (1) Low $\delta^{18}\text{O}$ is present in phenocryst olivine from Reykjanes Ridge lavas at around 63°N from 200 to 500 m sea depths. These samples have radiogenic isotope compositions almost identical to those of incompatible-element-enriched lavas from the Reykjanes Peninsula, Iceland, and samples with anomalously Icelandic isotope ratios from $\sim 57.5^\circ\text{N}$ and >1500 m water depths also have low $\delta^{18}\text{O}$. Reykjanes Ridge lavas show hyperbolic Nd–O isotope correlations which imply that the Icelandic contribution to Reykjanes Ridge magmas has $\delta^{18}\text{O}$ at least 0.6‰ lower than MORB mantle. Samples with low $\delta^{18}\text{O}$ show no indication of crustal contamination with hydrothermally altered oceanic crust or high-latitude meteoric water (e.g., they possess normal Cl/K ratios and their Sr–Nd isotope ratios are appropriate for Iceland–MORB mixtures). The Icelandic contribution cannot be melt that has moved laterally along 120 km of ridge after interacting with low- $\delta^{18}\text{O}$ crust in Iceland because fissure orientations are strongly oblique to the ridge axis.
- (2) Binary mixing hyperbolae between Pb–Sr–Nd and O isotope ratios for 60 – 63°N samples are too curved to be produced by magma mixing and probably require mixing between solid mantle sources. Combined with the Nd contents of primitive magmas at either end, the curvature suggests that basalts around 63°N are the product of $\sim 4.5 \times$ greater melt fractions than those around 60°N , and that the mantle source at 63°N has $\sim 9 \times$ greater Nd concentrations than the mantle source at 60°N .
- (3) There is no evidence in our dataset for mantle sources in Iceland with $\delta^{18}\text{O}$ in olivine greater than that of normal mantle ($+5.2\text{‰}$). The data in Skovgaard et al. (2001) suggesting otherwise are inaccurate analyses.
- (4) Normal mantle $\delta^{18}\text{O}$ in Icelandic rift zones is restricted to lavas with strong incompatible element depletion. Some depleted lavas have $\delta^{18}\text{O}_{\text{ol}}$ as low as $+4.5\text{‰}$; these have $^{206}\text{Pb}/^{204}\text{Pb}$ approaching values in enriched lavas and are probably the result of remelting residues from melting events that produced enriched lavas; contamination cannot preserve the

incompatible element depletion. Depleted sources in eastern Iceland may have lower $\delta^{18}\text{O}$ than at the Reykjanes Peninsula.

- (5) The most incompatible-element-enriched primitive lavas on the Reykjanes Peninsula have $\delta^{18}\text{O}_{\text{ol}} \sim +4.3\text{‰}$, suggesting this is characteristic of enriched Icelandic mantle sources IE1 and IE2. This is lower than found in Reykjanes Ridge magmas at 63°N , consistent with Pb–Nd–Sr evidence that the Icelandic mantle contributing to the Ridge is a mixture of depleted and enriched sources. Lower $\delta^{18}\text{O}_{\text{ol}}$ in more evolved lavas probably reflects crustal contamination, though there is some evidence of mantle sources with lower $\delta^{18}\text{O}$ in eastern Iceland. Linear correlations between $\delta^{18}\text{O}_{\text{ol}}$ and radiogenic isotopes, especially $^{206}\text{Pb}/^{204}\text{Pb}$, in more primitive Reykjanes Peninsula lavas are likely to reflect mixing before the depleted sources developed their extreme incompatible-element depletion.
- (6) The high $\delta^{18}\text{O}$ in depleted sources, and low $\delta^{18}\text{O}$ in enriched sources, are opposite to $\delta^{18}\text{O}$ values expected from models that relate these sources to recycled lower and upper oceanic crust, respectively. A compilation of published ocean crust oxygen data shows that the low $\delta^{18}\text{O}$ source would require $>60\%$ contribution from the oceanic gabbro suite with the lowest observed mean $\delta^{18}\text{O}$, a figure wholly inconsistent with the major and trace element chemistry of enriched basalts. If high- $\delta^{18}\text{O}$ upper oceanic crust is a component in Icelandic mantle, as inferred by many workers, then low $\delta^{18}\text{O}$ in the mixed source must be contributed by the host mantle. A scenario in which the recycled crust was hydrothermally altered by Lower Palaeozoic seawater with $\delta^{18}\text{O}$ of -8‰ relative to SMOW could explain the low $\delta^{18}\text{O}$ source, but is inconsistent with available data from Lower Palaeozoic ophiolites.
- (7) There are strong indications that the low- $\delta^{18}\text{O}$ -enriched source also has high $^3\text{He}/^4\text{He}$, suggesting that it may be of deep origin. However, all models invoking a mantle origin for the low $\delta^{18}\text{O}$, whether by recycling or from a deeper source, raise questions as to why this low- $\delta^{18}\text{O}$ source is not seen in other ocean islands, or indeed in alkaline lavas from areas distal from the rift zones of Iceland. If the low $\delta^{18}\text{O}$ were present in refractory mantle that was host to fertile recycled material, then the larger melt fractions generated by decompression at Icelandic rift zones might allow these to show the low- $\delta^{18}\text{O}$ signature. The fate of the implied low- $\delta^{18}\text{O}$ residue in other islands then becomes an issue of concern.

Acknowledgments

The early part of this research was supported by a NERC studentship to Gee and NERC Grant GR9/8757 to Thirlwall and Murton. This paper was prepared during a Royal Society/Leverhulme Trust Senior Research

Fellowship to Thirlwall. We are grateful to Anna Cecilie Skovgaard for supplying olivine separates that were used at RHUL for oxygen isotope analysis in Skovgaard et al. (2001). Dr. Skovgaard concurs that oxygen isotope data reported therein for samples from the Reykjanes Peninsula were inaccurate. The paper was improved by helpful reviews from John Eiler, John MacLennan, and John Lassiter, and editorial advice from Fred Frey.

Associate editor: Frederick A. Frey

Appendix A. Supplementary data

Supplementary data associated with this article can be found in the online version at doi:10.1016/j.gca.2005.09.008.

References

- Agrinier, P., Hékinian, R., Bideau, D., Javoy, M., 1995. O and H stable isotope compositions of oceanic crust and upper mantle rocks exposed in the Hess Deep near the Galapagos triple junction. *Earth Planet. Sci. Lett.* **136**, 183–196.
- Agrinier, P., Javoy, M., Girardeau, J., 1988. Hydrothermal activity in a peculiar oceanic ridge: oxygen and hydrogen isotope evidence in the Xigaze ophiolite (Tibet, China). *Chem. Geol.* **71**, 313–335.
- Alt, J.C., Muehlenbachs, K., Honnorez, J., 1986. An oxygen isotopic profile through the upper kilometer of the oceanic crust, DSDP Hole 504B. *Earth Planet. Sci. Lett.* **80**, 217–229.
- Alt, J.C., Zuleger, E., Erzinger, J., 1995. Mineralogy and stable isotope compositions of the hydrothermally altered lower sheeted dyke complex, hole 504B, Leg 140. In: Erzinger, J., Becker, K., Dick, H.J.B., Stokking, L.B. (Eds.), *Proceedings of the ODP Science Results 137/140*, pp. 155–166.
- Anderson, D.L., 2000. The statistics of helium isotopes along the global spreading ridge system and the Central Limit Theorem. *Geophys. Res. Lett.* **27**, 2401–2404.
- Arnórsson, S., 1978. Major element chemistry of the geothermal seawater at Reykjanes and Svartsengi, Iceland. *Mineral. Mag.* **42**, 209–220.
- Azetsu-Scott, K., Tan, F., 1997. Oxygen isotope studies from Iceland to an East Greenland fjord: behaviour of glacial meltwater plume. *Marine Chem.* **56**, 239–251.
- Baker, J.A., Peate, D.W., Waight, T.E., Thirlwall, M.F., 2005. Reply to the: comment on “Pb isotopic analysis of standards and samples using a Pb-207-Pb-204 double spike and thallium to correct for mass bias with a double focusing MC-ICP-MS” by Baker et al. *Chem. Geol.* **217**, 175–179.
- Barnicoat, A.C., Cartwright, I., 1997. The gabbro-eclogite transition: an oxygen isotope and petrographic study of west Alpine ophiolites. *J. Met. Geol.* **15**, 93–104.
- Bindeman, I.N., Fournelle, J.H., Valley, J.W., 2001. Low- $\delta^{18}\text{O}$ tephra from a compositionally zoned magma body: Fisher Caldera, Unimak Island, Aleutians. *J. Volcanol. Geotherm. Res.* **111**, 35–53.
- Bindeman, I.N., Ponomareva, V.V., Bailey, J.C., Valley, J.W., 2004. Volcanic arc of Kamchatka: a province with high- $\delta^{18}\text{O}$ magma sources and large-scale $^{18}\text{O}/^{16}\text{O}$ depletion of the upper crust. *Geochim. Cosmochim. Acta* **68**, 841–865.
- Blattner, P., Grindley, G.W., Adams, C.J., 1997. Low- ^{18}O terranes tracking Mesozoic polar climates in the South Pacific. *Geochim. Cosmochim. Acta* **61**, 569–576.
- Breddam, K., 2002. Kistufell: primitive melt from the Iceland mantle plume. *J. Petrol.* **43**, 345–373.
- Breddam, K., Kurz, M.D., Storey, M., 2000. Mapping out the conduit of the Iceland mantle plume with helium isotopes. *Earth Planet. Sci. Lett.* **176**, 45–55.

- Burnard, P., Harrison, D., 2005. Argon isotope constraints on modification of oxygen isotopes in Iceland Basalts by surficial processes. *Chem. Geol.* **216**, 143–156.
- Chauvel, C., Hémond, C., 2000. Melting of a complete section of recycled oceanic crust: Trace element and Pb isotopic evidence from Iceland. *Geochem. Geophys. Geosyst.*, 1, 2000GC000002..
- Cocker, J.D., Griffin, B.J., Muehlenbachs, K., 1982. Oxygen and carbon isotope evidence for seawater-hydrothermal alteration of the Macquarie Island ophiolite. *Earth Planet. Sci. Lett.* **61**, 112–122.
- Condomines, M., Grönvold, K., Hooker, P.J., Muehlenbachs, K., O'Nions, R.K., Óskarsson, N., Oxburgh, E.R., 1983. Helium, oxygen, strontium and neodymium isotopic relationships in Icelandic volcanics. *Earth Planet. Sci. Lett.* **66**, 125–136.
- Cooper, K.M., Eiler, J.M., Asimow, P.D., Langmuir, C.H., 2004. Oxygen isotope evidence for the origin of enriched mantle beneath the mid-Atlantic ridge. *Earth Planet. Sci. Lett.* **220**, 297–316.
- Darbyshire, F.A., White, R.S., Priestley, K.F., 2000. Structure of the crust and uppermost mantle of Iceland from a combined seismic and gravity survey. *Earth Planet. Sci. Lett.* **181**, 409–428.
- Dixon, E.T., Honda, M., McDougall, I., Campbell, I.H., Sigurdsson, I., 2000. Preservation of near-solar neon isotopic ratios in Icelandic basalts. *Earth Planet. Sci. Lett.* **180**, 309–324.
- Downes, H., Thirlwall, M.F., Trayhorn, S.C., 2001. Miocene subduction-related magmatism in southern Sardinia: Sr–Nd- and oxygen isotopic evidence for mantle source enrichment. *J. Volcanol. Geotherm. Res.* **106**, 1–21.
- Eiler, J.M., Farley, K.A., Valley, J.W., Hofmann, A.W., Stolper, E.M., 1996. Oxygen isotope constraints on the sources of Hawaiian volcanism. *Earth Planet. Sci. Lett.* **144**, 453–468.
- Eiler, J.M., Farley, K.A., Valley, J.W., Hauri, E., Craig, H., Hart, S.R., Stolper, E.M., 1997. Oxygen isotope variations in ocean island basalt phenocrysts. *Geochim. Cosmochim. Acta* **61**, 2281–2293.
- Eiler, J.M., Grönvold, K., Kitchen, N., 2000a. Oxygen isotope evidence for the origin of chemical variations in lavas from Theistareykir volcano in Iceland's northern volcanic zone. *Earth Planet. Sci. Lett.* **184**, 269–286.
- Eiler, J.M., Schiano, P., Kitchen, N., Stolper, E.M., 2000b. Oxygen isotope evidence for recycled crust in the sources of mid-ocean-ridge basalts. *Nature* **403**, 530–534.
- Eiler, J.M., Crawford, A., Elliott, T., Farley, K.A., Valley, J.W., Stolper, E.M., 2000c. Oxygen isotope geochemistry of oceanic-arc lavas. *J. Petrol.* **41**, 229–256.
- Elderfield, H., Greaves, M.J., 1981. Strontium isotope geochemistry of Icelandic geothermal systems and implications for sea water chemistry. *Geochim. Cosmochim. Acta* **45**, 2201–2212.
- Ford, C.E., Russell, D.G., Craven, J.A., Fisk, M.R., 1983. Olivine-liquid equilibria: temperature, pressure and composition dependence of the crystal/liquid cation partition coefficients for Mg, Fe²⁺, Ca and Mn. *J. Petrol.* **24**, 256–265.
- Foulger, G.R., Pearson, D.G., 2001. Is Iceland underlain by a plume in the lower mantle. Seismology and helium isotopes. *Geophys. J. Int.* **145**, F1–F5.
- Foulger, G.R., Natland, J.H., Anderson, D.L., 2005. A source for Icelandic magmas in remelted Iapetus crust. *J. Volcanol. Geotherm. Res.* **141**, 23–44.
- Frew, R.D., Dennis, P.F., Heywood, K.J., Meredith, M.P., Boswell, S.M., 2000. The oxygen isotope composition of water masses in the northern North Atlantic. *Deep-Sea Res. I* **47**, 2265–2286.
- Furman, T., Frey, F.A., Park, K.-H., 1991. Chemical constraints on the petrogenesis of mildly alkaline lavas from Vestmannaeyjar, Iceland: the Eldfell (1973) and Surtsey (1963–1967) eruptions. *Contrib. Mineral. Petrol.* **109**, 19–37.
- Garcia, M.O., Ito, E., Eiler, J.M., Pietruszka, A.J., 1998. Crustal contamination of Kilauea volcano magmas revealed by oxygen isotope analysis of glass and olivine from Puu Oo eruption lavas. *J. Petrol.* **39**, 803–817.
- Gautason, B., Muehlenbachs, K., 1998. Oxygen isotopic fluxes associated with high-temperature processes in the rift zones of Iceland. *Chem. Geol.* **145**, 275–286.
- Gee, M.A.M., 1998. Volcanology and geochemistry of Reykjanes Peninsula: plume-mid-ocean ridge interaction. Ph.D. thesis (unpubl.), Royal Holloway University of London.
- Gee, M.A.M., Thirlwall, M.F., Taylor, R.N., Lowry, D., Murton, B.J., 1998. Crustal processes: major controls on Reykjanes Peninsula lava chemistry, SW Iceland. *J. Petrol.* **39**, 819–839.
- Gregory, R.T., Taylor Jr., H.P., 1981. An oxygen isotope profile in a section of Cretaceous oceanic crust, Samail ophiolite, Oman: evidence for δ¹⁸O buffering of the oceans by deep (>5 km) seawater-hydrothermal circulation at mid-ocean ridges. *J. Geophys. Res.* **86**, 2737–2755.
- Gurenko, A.A., Chaussidon, M., 2002. Oxygen isotope variations in primitive tholeiites from Iceland: evidence from a SIMS study of glass inclusions, olivine phenocrysts and pillow rim glasses. *Earth Planet. Sci. Lett.* **205**, 63–79.
- Hansteen, T.H., Troll, V.R., 2003. Oxygen isotope composition of xenoliths from the oceanic crust and volcanic edifice beneath Gran Canaria (Canary Islands): consequences for crustal contamination of ascending magmas. *Chem. Geol.* **193**, 181–193.
- Harmon, R.S., Hoefs, J., 1995. Oxygen-isotope heterogeneity of the mantle deduced from global O-18 systematics of basalts from different tectonic settings. *Contrib. Mineral. Petrol.* **120**, 95–114.
- Hart, S.R., Schilling, J.-G., Powell, J.L., 1973. Basalts from Iceland and along the Reykjanes Ridge: Sr isotope geochemistry. *Nat. Phys. Sci.* **246**, 104–107.
- Hart, S.R., Blusztajn, J., Dick, H.J.B., Meyer, P.S., Muehlenbachs, K., 1999. The fingerprint of seawater circulation in a 500-meter section of ocean crust gabbros. *Geochim. Cosmochim. Acta* **63**, 4059–4080.
- Hattori, K., Muehlenbachs, K., 1982. Oxygen isotope ratios of the Icelandic crust. *J. Geophys. Res.* **87**, 6559–6565.
- Hémond, C.H., Condomines, M., Fourcade, S., Allègre, C.J., Óskarsson, N., Javoy, M., 1988. Thorium, strontium an oxygen isotopic geochemistry in recent tholeiites from Iceland: crustal influence on mantle-derived magmas. *Earth Planet. Sci. Lett.* **87**, 273–285.
- Hémond, C.H., Arndt, N.T., Lichtenstein, U., Hofmann, A.W., Óskarsson, N., Steinthorsson, S., 1993. The heterogeneous Iceland plume—Nd–Sr–O isotopes and trace-element constraints. *J. Geophys. Res.* **98**, 15833–15850.
- Hilton, D.R., Grönvold, K., Macpherson, C.G., Castillo, P.R., 1999. Extreme ³He/⁴He ratios in northwest Iceland: constraining the common component in mantle plumes. *Earth Planet. Sci. Lett.* **173**, 53–60.
- Hilton, D.R., Thirlwall, M.F., Taylor, R.N., Murton, B.J., Nichols, A., 2000. Controls on magmatic degassing along the Reykjanes Ridge with implications for the helium paradox. *Earth Planet. Sci. Lett.* **183**, 43–50.
- Holmden, C., Muehlenbachs, K., 1993. The ¹⁸O/¹⁶O ratio of 2-billion-year-old seawater inferred from ancient oceanic crust. *Science* **259**, 1733–1736.
- Jacob, D.E., 2004. Nature and origin of eclogite xenoliths from kimberlites. *Lithos* **77**, 295–316.
- Kokfelt, T.F., Hoernle, K., Hauff, F., 2003. Upwelling and melting of the Iceland plume from radial variation of ²³⁸U–²³⁰Th disequilibria in postglacial volcanic rocks. *Earth Planet. Sci. Lett.* **214**, 167–186.
- Kurz, M.D., Meyer, H., Sigurdsson, H., 1985. Helium isotopic systematics within the neovolcanic zones of Iceland. *Earth Planet. Sci. Lett.* **74**, 291–305.
- Lassiter, J.C., Hauri, E.H., 1998. Osmium-isotope variations in Hawaiian lavas: evidence for recycled oceanic lithosphere in the Hawaiian plume. *Earth Planet. Sci. Lett.* **164**, 483–496.
- Lecuyer, C., Fourcade, S., 1991. Oxygen isotope evidence for multi-stage hydrothermal alteration at a fossil slow-spreading center: the Silurian Trinity ophiolite (California, USA). *Chem. Geol.* **87**, 231–246.
- Lecuyer, C., Reynard, B., 1996. High-temperature alteration of oceanic gabbros by seawater (Hess Deep, Ocean Drilling Program Leg 147): evidence from oxygen isotopes and elemental fluxes. *J. Geophys. Res.* **101**, 15883–15897.
- Lowry, D., Matthey, D.P., Harris, J.W., 1999. Oxygen isotope composition of syngenetic inclusions in diamond from the Finsch Mine, RSA. *Geochim. Cosmochim. Acta* **63**, 1825–1836.

- MacLennan, J., McKenzie, D., Grönvold, K., Shimizu, N., Eiler, J.M., Kitchen, N., 2003. Melt mixing and crystallization under Theistareykir, northeast Iceland. *Geochem. Geophys. Geosyst.* 4, art. No. 8624.
- Macpherson, C.G., Hilton, D.R., Matthey, D.P., Sinton, J.M., 2000. Evidence for an ^{18}O -depleted mantle plume from contrasting $^{18}\text{O}/^{16}\text{O}$ ratios of back-arc lavas from the Manus Basin and the Mariana Trough. *Earth Planet. Sci. Lett.* 176, 171–183.
- Macpherson, C.G., Hilton, D.R., Day, J.M.D., Lowry, D., Grönvold, K., 2005. High $^3\text{He}/^4\text{He}$, depleted mantle and low- $\delta^{18}\text{O}$, recycled oceanic lithosphere in the source of central Iceland magmatism. *Earth Planet. Sci. Lett.* 233, 411–427.
- Matthey, D.P., 1997. LaserPrep: an automatic laser-fluorination system for micromass 'Optima' or 'Prism' mass spectrometers. *Micromass Application Note* 107, 8.
- Matthey, D.P., Macpherson, C.G., 1993. High-precision oxygen isotope microanalysis of ferromagnesian minerals by laser-fluorination. *Chem. Geol. (Isot. Geosci. Sect.)* 105, 305–318.
- Matthey, D.P., Lowry, D., Macpherson, C.G., 1994. Oxygen isotope composition of mantle peridotite. *Earth Planet. Sci. Lett.* 128, 231–241.
- McKenzie, D.P., Stracke, A., Blichert-Toft, J., Albarède, F., Grönvold, K., O'Nions, R.K., 2004. Source enrichment processes responsible for isotopic anomalies in oceanic island basalts. *Geochim. Cosmochim. Acta* 68, 2699–2724.
- Michael, P.J., Cornell, W.C., 1998. Influence of spreading rate and magma supply on crystallization and assimilation beneath mid-ocean ridges: evidence from chlorine and major element chemistry of mid-ocean ridge basalts. *J. Geophys. Res.* 103, 18325–18356.
- Momme, P., Óskarsson, N., Keays, R.R., 2003. Platinum-group elements in the Icelandic rift system: melting processes and mantle sources beneath Iceland. *Chem. Geol.* 196, 209–234.
- Muehlenbachs, K., Anderson, A.T., Sigvaldason, G.E., 1974. Low- O^{18} basalts from Iceland. *Geochim. Cosmochim. Acta* 38, 577–588.
- Muehlenbachs, K., Furnes, H., Fonneland, H.C., Hellevang, B., 2004. Ophiolites as faithful records of ancient seawater: the Solund-Stavfjord Ophiolite Complex as a Late Ordovician example. In: Dilek, Y., Robinson, P.T. (Eds.), *Ophiolites in Earth History. Spec. Publ. Geol. Soc. London*, vol. 218, pp. 1–14.
- Murton, B.J., 1994. RRS Charles Darwin Cruise 80, 01 September to 01 October 1993. The PETROS Programme: geological sampling and bathymetric surveying of the Reykjanes Ridge between 57°N and 63°N , southwest of Iceland. IOSDL Cruise Report, pp. 32.
- Murton, B.J., Taylor, R.N., Thirlwall, M.F., 2002. Plume-ridge interaction: a geochemical perspective from the Reykjanes Ridge. *J. Petrol.* 43, 1987–2012.
- Nicholson, H., Condomines, M., Fitton, J.G., Fallick, A.E., Grönvold, K., Rogers, G., 1991. Geochemical and isotopic evidence for crustal assimilation beneath Krafla, Iceland. *J. Petrol.* 32, 1005–1020.
- Peate, D.W., Hawkesworth, C.J., van Calsteren, P., Taylor, R.N., Murton, B.J., 2001. ^{238}U - ^{230}Th constraints on mantle upwelling and plume-ridge interaction along the Reykjanes Ridge. *Earth Planet. Sci. Lett.* 187, 257–272.
- Prestvik, T., Goldberg, S., Karlsson, H., Grönvold, K., 2001. Anomalous strontium and lead isotope signatures in the off-rift Öraefajökull central volcano in south-east Iceland. Evidence for enriched endmember(s) of the Iceland mantle plume? *Earth Planet. Sci. Lett.* 190, 211–220.
- Putlitz, B., Matthews, A., Valley, J.W., 2000. Oxygen and hydrogen isotope study of high-pressure metagabbros and metabasalts (Cyclades, Greece): implications for the subduction of oceanic crust. *Contrib. Mineral. Petrol.* 138, 114–126.
- Rohling, E.J., Fenton, M., Jorissen, F.J., Bertrand, P., Ganssen, G., Caulet, J.P., 1998. Magnitudes of sea-level lowstands of the past 500,000 years. *Nature* 394, 162–165.
- Schilling, J.-G., 1973. The Icelandic mantle plume, geochemical evidence along the Reykjanes Ridge. *Nature* 242, 565–578.
- Sharp, Z.D., 1990. A laser-based microanalytical method for the in situ determination of oxygen isotope ratios of silicates and oxides. *Geochim. Cosmochim. Acta* 54, 1353–1357.
- Sigmarsson, O., Hémond, Ch., Condomines, M., Fourcade, S., Óskarsson, N., 1991. Origin of silicic magma in Iceland revealed by Th isotopes. *Geology* 19, 621–624.
- Sigmarsson, O., Condomines, M., Fourcade, S., 1992a. A detailed Th, Sr and O isotope study of Hekla: differentiation processes in an Icelandic volcano. *Contrib. Mineral. Petrol.* 112, 20–34.
- Sigmarsson, O., Condomines, M., Fourcade, S., 1992b. Mantle and crustal contribution in the genesis of Recent basalts from off-rift zones in Iceland: constraints from Th, Sr and O isotopes. *Earth Planet. Sci. Lett.* 110, 149–162.
- Sigmarsson, O., Karlsson, H.R., Larsen, G., 2000. The 1996 and 1998 subglacial eruptions beneath the Vatnajökull ice sheet in Iceland: contrasting geochemical and geophysical inferences on magma migration. *Bull. Volcanol.* 61, 468–476.
- Sigurdsson, H., Sparks, R.S.J., 1978. Lateral magma flow within rifted Icelandic crust. *Nature* 274, 126–130.
- Skovgaard, A.C., Storey, M., Baker, J., Blusztajn, J., Hart, S.R., 2001. Osmium-oxygen isotopic evidence for a recycled and strongly depleted component in the Iceland mantle plume. *Earth Planet. Sci. Lett.* 194, 259–275.
- Stakes, D.S., Taylor Jr., H.P., 1992. The Northern Samail ophiolite: an oxygen isotope, microprobe and field study. *J. Geophys. Res.* 97, 7043–7080.
- Staudigel, H., Davies, G.R., Hart, S.R., Marchant, K.M., Smith, B.M., 1995. Large scale isotopic Sr, Nd and O isotopic anatomy of altered oceanic crust: DSDP/ODP sites 417/418. *Earth Planet. Sci. Lett.* 130, 169–185.
- Stracke, A., Zindler, A., Salters, V.J.M., McKenzie, D., Blichert-Toft, J., Albarède, F., Grönvold, K., 2003. Theistareykir revisited. *Geochem. Geophys. Geosyst.* 4, GC000201.
- Sun, S.-S., Jahn, B.-M., 1975. Lead and strontium isotopes in post-glacial basalts from Iceland. *Nature* 255, 527–530.
- Sveinbjörnsdóttir, A.E., Coleman, M.L., Yardley, B.W.D., 1986. Origin and history of hydrothermal fluids of the Reykjanes and Krafla geothermal fields, Iceland. *Contrib. Mineral. Petrol.* 94, 99–109.
- Taylor, R.N., Thirlwall, M.F., Murton, B.J., Hilton, D.R., Gee, M.A.M., 1997. Isotopic constraints on the influence of the Icelandic plume. *Earth Planet. Sci. Lett.* 148, E1–E8.
- Thirlwall, M.F., 1997. Pb isotopic and concentration constraints on the evolution of HIMU mantle. *Chem. Geol.* 139, 51–74.
- Thirlwall, M.F., Graham, A.M., Arculus, R.J., Harmon, R.S., Macpherson, C.G., 1996. Resolution of the effects of crustal assimilation, sediment subduction and fluid transport in island arc magmas: Pb–Sr–Nd–O isotope geochemistry of Grenada, Lesser Antilles. *Geochim. Cosmochim. Acta* 60, 4785–4810.
- Thirlwall, M.F., Jenkins, C., Vroon, P.Z., Matthey, D.P., 1997. Crustal interaction during construction of ocean islands: Pb–Sr–Nd–O isotope stratigraphy of the shield basalts of Gran Canaria. *Chem. Geol.* 135, 233–262.
- Thirlwall, M.F., Gee, M.A.M., Taylor, R.N., Murton, B.J., 2004. Mantle components in Iceland and adjacent ridges investigated using double-spike Pb isotope ratios. *Geochim. Cosmochim. Acta* 68, 361–386.
- Veizer, J., Ala, D., Azmy, K., Bruckschen, P., Buhl, D., Bruhn, F., Carden, G.A.F., Diener, A., Ebner, S., Godderis, Y., Jasper, T., Korte, C., Pawellek, F., Podlaha, O.G., Strauss, H., 1999. $^{87}\text{Sr}/^{86}\text{Sr}$, $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ evolution of Phanerozoic seawater. *Chem. Geol.* 161, 59–88.
- Wang, Z.G., Kitchen, N.E., Eiler, J.M., 2003. Oxygen isotope geochemistry of the second HSDP core. *Geochem. Geophys. Geosyst.* 4, art. No. 8712.
- Weir, N.R.W., White, R.S., Brandsdóttir, B., Einarsson, P., Shimamura, H., Shiobara, H., et al., 2001. Crustal structure of the northern Reykjanes Ridge and Reykjanes Peninsula, SW Iceland. *J. Geophys. Res.* 106, 6347–6368.
- Zhao, Z-F., Zheng, Y-F., 2003. Calculation of oxygen isotope fractionation in magmatic rocks. *Chem. Geol.* 193, 59–80.
- Zheng, Y-F., Wu, Y-B., Chen, F-K., Gong, B., Li, L., Zhao, Z-F., 2004. Zircon U–Pb and oxygen isotope evidence for a large-scale ^{18}O depletion event in igneous rocks during the Neoproterozoic. *Geochim. Cosmochim. Acta* 68, 4145–4165.