Oxygen isotope variations in Lau Basin lavas

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Abstract

New laser-fluorination oxygen isotope data are presented for volcanic glasses and phenocrysts from Lau and North Fiji Basin lavas. The low oxygen blank of the technique allows accurate analysis of these often scarce igneous phases. $\delta^{18}O$ values of volcanic glass range from 5.53‰ to 6.06‰. Oxygen isotope ratios of ferromagnesian and plagioclase phenocrysts are lower and higher than their host glasses, respectively, and coexisting phases appear to represent oxygen isotope equilibrium at magmatic temperatures. Evolved lavas from the propagating and dying rift tips in the Central Lau Basin display significant $^{18}O$-enrichment with magmatic evolution due to FeTi-oxide crystallisation and require that precursor melts had relatively low $\delta^{18}O$ values immediately prior to oxide saturation. Development of these precursor melts from basaltic parents involved assimilation of hydrothermally altered sheeted dykes or gabbros in recently accreted oceanic crust accompanied by crystallisation of a plagioclase-rich assemblage. The stable isotope geochemistry and petrology of Central Lau Basin lavas constrain magmatic evolution to depths between 2.5 and 5 km beneath the seafloor. A dacite from Valu Fa Ridge is slightly $^{18}O$-enriched relative to Valu Fa basalts, consistent with crystallisation of a mineral assemblage similar to mid-ocean ridge basalts. $^{18}O/^{16}O$ ratios of Lau Basin basaltic glasses decrease with Mg Number; this observation could be reconciled with the assimilation and fractional crystallisation model proposed for initial differentiation of the evolved lavas. However, oxygen isotope data for Central Lau Basin basalts correlate positively with Ba/La while basaltic glasses from throughout the region display a negative correlation between $\delta^{18}O$ and Na suggesting that recycling of subducted oxygen and/or mantle fertility may also exert an influence on $^{18}O/^{16}O$ of primary melts. Presently, the lack of quantitative control over oxygen isotope fractionation in basaltic systems prohibits resolution of these effects. Detailed investigation of less complex suites of lavas, employing the precision offered by laser fluorination, may provide new insights into the behaviour of oxygen at magmatic temperatures. © 1998 Elsevier Science B.V.

Keywords: Lau Basin; Oxygen isotopes; Magmatism; Subduction; Melting; Assimilation

1. Introduction

Unaltered lavas erupted at mid-ocean spreading centres display limited $^{18}O/^{16}O$ variation (5.7 ± 0.3‰; Ito et al., 1987). This is thought to result from melting of an upper mantle reservoir that is uniform with respect to oxygen isotopes, followed by limited degrees of chemical evolution in sub-ridge magma chambers (Muehlenbachs and Clayton, 1972; Taylor...

...and Sheppard, 1986, and references therein). Protracted fractional crystallisation can produce oxygen isotope variations in cogenetic lava suites but this is rare at mid-ocean spreading centres, having been recognised only at the 95°W propagating rift tip of the Galapagos Spreading Centre (Muehlenbachs and Byerly, 1982; Ito et al., 1987). Oxygen isotope ratios are frequently employed in petrogenetic studies of subduction zone magmatism to discriminate between source and crustal contamination of the magmatic system (James, 1981). While the effects of both crystallisation and open system behaviour have been observed in some tholeiitic lava series, the magnitude of variations that may be produced by these processes are small: \( \leq 0.5\%\) (e.g. Taylor, 1968; Anderson et al., 1971; Matsuhisa et al., 1973; Sheppard and Harris, 1985; Woodhead et al., 1987; Ellam and Harmon, 1990). Furthermore, identification of original magmatic \( ^{18}O/^{16}O \) variation in submarine lavas is often obscured by the effects of post-eruptive alteration (Muehlenbachs and Clayton, 1972; Pineau et al., 1976; Taylor and Sheppard, 1986). Phenocrysts or fresh volcanic glass can be analysed to avoid alteration products but these phases are commonly too scarce in seafloor dredge hauls to allow conventional oxygen isotope analysis of entire suites of lavas.

This study presents oxygen isotope data obtained by laser-fluorination (LF) for fresh lavas collected from three sites of active volcanism in the Lau Basin. Laser ablation of bulk samples vastly reduces oxygen blanks compared to the classic ‘bomb’ fluorination method and thus allows routine determination of \( ^{18}O/^{16}O \) ratios in igneous phases (Sharp, 1990; Mattey and Macpherson, 1993). Furthermore, highly refractory phases can be analysed with high precision. Intensive sampling of a propagating rift tip and associated dying rift in the Central Lau Basin has allowed examination of oxygen isotope systematics through a diverse compositional spectrum of lavas generated in an oceanic setting. In an effort to detect subduction-related spatial controls on the \( ^{18}O/^{16}O \) ratios of lavas, oxygen isotope data are also reported for lavas from the northern and southern extremities of the basin. These results are compared with other geochemical properties of the lavas and conventionally obtained \( ^{18}O/^{16}O \) data from nearby localities in an effort to constrain the flux of oxygen from various reservoirs into primary Lau Basin melts.

2. Samples and technique

The Lau Basin is a triangular-shaped back-arc basin situated west of the Tonga Ridge (Fig. 1a), where the Pacific Plate is being subducted beneath the Australian Plate at the Tonga Trench (Karig, 1970). The triangular shape results from rapid, clockwise rotation of the Tonga Ridge relative to the remnant Lau Ridge, and the southward propagation of two phases of ocean spreading (Parson and Hawkins, 1994; Bevis et al., 1995). At around 5.5 Ma the first phase of spreading was initiated, south-east of what is now Peggy Ridge, producing oceanic crust at the Eastern Lau Spreading Centre (ELSC) and the Valu Fa Ridge (Fig. 1a). These spreading centres propagated southwards through extended arc crust between the Tonga and Lau ridges. Then, around 1.5 Ma, the Central Lau Spreading Centre (CLSC) began to propagate through crust formed at the ELSC (Fig. 1a).

Parson et al. (1990) identified three active spreading segments in the Central Lau Basin, representing propagating and dying rifts with an intermediate ‘relay’ section (Fig. 1b). The crests of these ridges were dredged by Cruise 33 of the R.V. Charles Darwin (Pearce et al., 1994). Eruptive products from the ELSC and Intermediate Lau Spreading Centre (ILSC) are dominantly basaltic with Mg No. > 50, although glasses from the northern ELSC site 20 are more evolved (Table 1). Ferrobasalts and andesites erupted at the propagating rift tip of the CLSC display evidence of extreme chemical fractionation and volatile degassing, compared to other Central Lau Basin lavas and mid-ocean ridge basalt (MORB; Pearce et al., 1994; Macpherson and Mattey, 1994; Farley and Newman, 1994). Plagioclase dominated the crystallising assemblage in melts with 4 wt.% < MgO < 7.5 wt.% subduing Sr enrichment and producing negative Eu anomalies during differentiation. Strong FeO, TiO₂ and V depletions in evolved lavas are the result of FeTi-oxide crystallisation (Pearce et al., 1994). Elevated concentrations of incompatible alkali and alkali earth elements and greater vesicularity in ELSC lavas, relative to MORB,
Fig. 1. (a) Map of the Lau Basin and Eastern part of the North Fiji Basin modified from Parson et al. (1994). A heavy box outlines the propagating rift system of the Central Lau Basin shown in (b), and the dashed line marks the approximate extent of oceanic crust formed at the Valu Fa Ridge and East Lau Spreading Centre (ELSC). KK16 was recovered approximately 300 km west of Fiji. SO indicates the location of the Valu Fa samples for which detailed localities are presented by Frenzel et al. (1990). (b) Geometry of the Central Lau Basin neovolcanic zones (stippled) identified by Parson et al. (1990), and dredge sites from Cruise 33 of the R.V. Charles Darwin. CLSC = Central Lau Spreading Centre; ILSC = Intermediate Lau Spreading Centre; ELSC = Eastern Lau Spreading Centre. The CLSC is propagating south at the expense of the ELSC. The ILSC is a ‘relay’ zone (Parson et al., 1990).
may reflect a greater flux of subducted material being recycling close to the arc (Pearce et al., 1994).

The southern end of the Valu Fa Ridge marks the limit of ocean spreading in the Lau Basin (Parson and Hawkins, 1994). Collection sites and petrology of the three glasses analysed from this ridge are described by Frenzel et al. (1990). Major element concentrations of the basaltic andesites are similar to MORB and the dacite (SO35-77KD2) was produced by 'normal basaltic differentiation' of basic melts followed by oxide saturation (Frenzel et al., 1990).

Magma chambers have been seismically imaged, at 1–4 km depth, along the entire length of the spreading segment from which these samples were collected (Collier and Sinha, 1992). Radiogenic isotope and trace element ratios suggest that a small but significant slab flux is present in Valu Fa lavas (Jenner et al., 1987; Sunkel, 1990; Loock et al., 1990); however, Hilton et al. (1993) showed that He isotopic variations are decoupled from Sr, Nd and Pb isotopes as a result of combined degassing and assimilation.

The Northern Lau and North Fiji Basin glasses were collected from disparate sites (Fig. 1a). Sample KK5 was recovered from the western end of Peggy Ridge while KK16-12 was recovered from the central part of the North Fiji Basin (Sinton et al., 1993). Chemically the KK lavas resemble MORB with the exception of slightly elevated \(^{87}\text{Sr}/^{86}\text{Sr}\) ratios that are found in many tholeiitic lavas from the Lau and Fiji basins and Fiji (Sinton et al., 1993). Sample 231KD is a ferrobasalt from the northern edge of the North Fiji Basin (Johnson and Sinton, 1990), where other ferrobasalts have been recovered and magnetic anomalies and oblique magnetic fabric are observed (Sinton et al., 1993; Brocher, 1985). Therefore, 231KD probably evolved from MORB-like magma at a propagating rift tip (Johnson and Sinton, 1990).

### Table 1

<table>
<thead>
<tr>
<th>Sample</th>
<th>Type</th>
<th>Mg No.</th>
<th>(\delta^{18}\text{O (o)})</th>
<th>s.d.</th>
<th>n</th>
<th>Yield %</th>
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</table>

Notes to Table 1:

1. Three numbers of LB samples represent, respectively, dredge site (within site), petrographic type and random fragment number. Full chemical descriptions of each, unique, glass can be found in Pearce et al. (1994) (CLB), Johnson and Sinton (1990), Sinton et al. (1993) (NLB) and Frenzel et al. (1990) (Valu Fa).

2. Lava types based on SiO content: b = basalt; ba = basaltic andesite; Feb = FeTi-basalt; a = andesite; d = dacite.

3. Similarity of Mg No. within each Central Lau Basin site is a function of the samples available and not the overall chemistry of each site (see Pearce et al., 1994).

4. Where more than one determination was made the mean is reported with standard deviation (s.d.) and number of analyses (n) listed.

5. Percent of oxygen expected are based on major element analyses of each glass.


Oxygen isotope ratios of volcanic glass and phenocrysts were analysed using the Royal Holloway University of London laser-fluorination (LF) system (Mattey and Macpherson, 1993). This is similar to conventional oxygen extraction lines (e.g. Clayton and Mayeda, 1963) with the exception that, rather than fluorination of powdered samples in externally heated nickel ‘bombs’, a Nd-YAG laser is used to directly heat uncrushed samples in the presence of a fluorinating agent (Mattey and Macpherson, 1993; Macpherson, 1994). This has the advantage of reducing background contributions, thus permitting routine analysis of samples in the 0.02–2 mg range (Mattey and Macpherson, 1993; Macpherson, 1994). This has the advantage of reducing background contributions, thus permitting routine analysis of samples in the 0.02–2 mg range (Mattey and Macpherson, 1993). Samples were ultrasonically washed in distilled water and again in acetone prior to weighing to ±1 µg, then 0.4 to 2.5 mg of material was selected under a binocular microscope. Care was taken to avoid visible alteration and phenocrysts in glass fragments. Oxygen released from silicates was converted to CO over hot graphite for isotopic analysis on a VG PRISM mass spectrometer. Fractionation of oxygen isotopes during gas handling was avoided through selecting suitable sample sizes and employing graphite that had previously performed many conversions (Mattey and Macpherson, 1993). Isotope ratios are reported as per mil deviation from SMOW in the standard notation.

The difference between δ18O values of any two materials i and j; Δδ18O_{i-j}, is assumed to be proportional to the isotopic fractionation factor α(1000 ln α_{i-j} = δ18O_{i-j}).

The behaviour of ferromagnesian silicates during Nd-YAG laser ablation is described by Mattey and Macpherson (1993). Feldspar is transparent to 1064 nm radiation; however, analyses of infra-red transparent phases can be achieved by mixing with an olivine or clinopyroxene of known isotopic composition. Such mixtures fluorinate readily and 18O/16O ratios calculated for the feldspar are similar to conventional determinations (Macpherson, 1994). Like clinopyroxene, glass reacts vigorously during the early stages of fluorination and care must be taken to prevent sputtering of molten silicate away from the reaction site. Subsequently, glass melts into a pool, at the bottom of the sample hole, and gently reacts until oxygen extraction is complete. Over the period of this study NBS-30 biotite gave δ18O of 5.03‰ (±0.08, n = 80), and all data are normalised to the recommended NBS-30 value of 5.1‰. Reproducibility for various phases can be compared with the Royal Holloway secondary standards San Carlos olivine and GP101 clinopyroxene that gave 4.88‰ (±0.09, n = 26) and 8.25‰ (±0.06, n = 10), respectively (cf. Mattey and Macpherson, 1993). There are no internationally recognised feldspar standards for 18O/16O so analyses of NBS-28 were also conducted by mixing with olivine giving a δ18O of 9.58‰ (±0.17, n = 8, recommended 9.66‰). Therefore greater uncertainty (±0.17‰) is attached to feldspar analyses (Macpherson, 1994). Yields for all standards were 100% (±2%).

3. Results

Results of LF analyses are listed in Tables 1 and 2. The total range of δ18O values in fresh glasses is 5.53‰ to 6.06‰ with replicate analyses generally agreeing to better than ±0.1‰ (1σ, Table 1). The three fresh Northern Lau and Fiji Basin glasses range from 5.64‰ to 5.91‰, similar to bomb fluorination analyses of unaltered whole rock from the Northern Lau Basin (Pineau et al., 1976), and a marginal increase in δ18O values is observed with degree of chemical evolution, as indicated by decreasing Mg Number (Table 1). The Valu Fa glasses also display

<table>
<thead>
<tr>
<th>Sample</th>
<th>Olivine</th>
<th>Yield %</th>
<th>Clinopyroxene</th>
<th>Yield %</th>
<th>Plagioclase</th>
<th>Yield %</th>
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<tr>
<td>LB24-1-1</td>
<td>4.87</td>
<td>98.7</td>
<td>5.28</td>
<td>98.4</td>
<td>6.12</td>
<td>100.8</td>
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* Calculated assuming wt.% oxygen contents of 44.0% in olivine, 45.5% in clinopyroxene, 48.0% in plagioclase.
a minor increase in $\delta^{18}O$ values with decreasing Mg Number, from 5.88%e to 6.06%e. For a given Mg Number, our analyses of Valu Fa glasses fall at the lower end of the range of southern Lau Basin whole rock $\delta^{18}O$ values obtained by bomb fluorination (Vallier et al., 1991; Lichtenstein and Loock, 1993).

$\delta^{18}O$ values of glasses from the Central Lau Basin range from 5.53%e to 6.03%e. The oxygen isotope ratio of LB18-1-2 is significantly lower than other Central Lau Basin basalts (Table 1). This is an olivine–phyric lava that has the highest MgO and Ni contents of the Central Lau Basin suite, features that have been attributed to accumulation or resorption of olivine phenocrysts or xenocrysts (Pearce et al., 1994). Olivine is depleted in $^{18}O$ by approximately 1%e relative to molten basalt (Anderson et al., 1971; Muehlenbachs and Clayton, 1972; LB24-1-1); therefore, olivine resorption should lower $\delta^{18}O$ of a melt, in addition to enriching Mg and Ni. While it may be unreasonable to advocate olivine accumulation alone to explain the entire 0.28%e difference between this glass and the mean value of the remaining basalts, the consistency of oxygen isotope ratio, petrography and chemistry in this sample are noteworthy. In addition to oxygen isotopes, several other geochemical properties of melts may be influenced by phenocryst accumulation, therefore LB18-1-2 is omitted from discussion of magmatic $^{18}O/^{16}O$ systematics. The remaining basic Central Lau Basin lavas (Mg No. > 50) have an average $\delta^{18}O$ value of (5.82%e, ± 0.07%e) that is identical, within error, to both the MORB average and the average of whole rock $^{18}O/^{16}O$ ratios of Northern Lau Basin lavas (Pineau et al., 1976; Ito et al., 1987; Lichtenstein and Loock, 1993; Table 1). The $\delta^{18}O$ values of ferrobasalts and andesites (that have Mg No. < 40; sites 10–13 and 20) encompass the entire $\delta^{18}O$ range of the Central Lau Basin suite and higher values are associated with lower Mg Numbers (Table 1).

Oxygen isotope ratios of phenocrysts are listed in Table 2. With respect to host glasses, $\delta^{18}O$ values of olivine and clinopyroxene phenocrysts are low, while $\delta^{18}O$ values of plagioclase are high. The $\Delta^{18}O_{\text{ol}}$ value of 0.41%e in LB24-1-1 is similar to LF analyses of mantle nodules and to values determined by conventional analysis of basic terrestrial lavas (Anderson et al., 1971; Friedrichsen and Hoerns, 1979; Mattey et al., 1994). The $\delta^{18}O$ values for mineral pairs involving plagioclase fall within the ranges obtained conventionally for basic rocks (Anderson et al., 1971; Friedrichsen and Hoerns, 1979). Values for $\Delta^{18}O_{\text{mineral–glass}}$ obtained for LB24-1-1 are similar to conventionally determined groundmass–phenocrysts pairs, with $\Delta^{18}O_{\text{ol–glass}}$ ($-1.0%e$) and $\Delta^{18}O_{\text{olpx–glass}}$ ($-0.59%e$) values falling in the middle of the previously determined range while $\Delta^{18}O_{\text{plag–glass}}$ lies at the higher end of the conventional data range (Macpherson, 1994, and references therein).

4. Discussion

Exposure of submarine lavas to seawater after eruption and cooling can result in growth of secondary minerals that can significantly enrich the $^{18}O$ content of the rock (e.g. Muehlenbachs and Clayton, 1972; Pineau et al., 1976). Volcanic glass is particularly susceptible to such alteration and caution must be exercised in attributing $\delta^{18}O$ variations to magmatic processes until the extent of post-eruptive alteration can be judged. However, correlations between oxygen isotope ratios and water concentrations, that would be expected if alteration were a major control (Muehlenbachs and Clayton, 1972; Pineau et al., 1976), are not observed in the Lau Basin glasses (Fig. 2). Phenocrysts are less prone to alteration than volcanic glass so extents of alteration can also be estimated by examining $\Delta^{18}O$ values of coexisting glass and phenocrysts (Muehlenbachs and Clayton, 1972; Kyser, 1986). As noted above, isotopic differences between mineral separates and glass appear to be representative of magmatic values and in particular we note the small, positive $\Delta^{18}O_{\text{plag–glass}}$ values obtained for all plagioclase separates. This value would probably be reversed by even small degrees of alteration to glass. These observations suggest that $\delta^{18}O$ values of the glasses have experienced negligible modification since eruption and that $^{18}O/^{16}O$ ratios in coexisting phases of Lau Basin lavas represent equilibrium at magmatic temperatures.

4.1. Oxygen isotope composition of evolved lavas

A strong negative correlation between $\delta^{18}O$ and Mg Number ($R = 0.90$) exists for those Central Lau
Basin lavas with Mg No. < 40 erupted close to the terminations of the propagating and dying rift tips (Fig. 3a). A similar relationship, attributed to magnetite crystallisation, also exists for evolved lavas from the Galapagos Spreading Centre propagating rift tip at 95°W (Muehlenbachs and Byerly, 1982). Oxides possess low δ¹⁸O values, relative to basaltic melt, at magmatic temperatures so oxide crystallisation, even in limited volumes, can produce significant δ¹⁸O enrichment of melts (Friedrichsen and Horns, 1979; Muehlenbachs and Byerly, 1982). The coincidence of δ¹⁸O enrichment with Fe, Ti, and V depletion in the Central Lau Basin suite is strong evidence for operation of a similar process (Fig. 3).

In addition, we note that Δ¹⁸O_{plag−melt} tends toward negative values in evolved melts (Taylor and Sheppard, 1986), also enhancing δ¹⁸O-enrichment during fractional crystallisation.

If fractional crystallisation of a magnetite-bearing assemblage is responsible for δ¹⁸O-enrichment of the evolved Central Lau Basin lavas, the Mg No.−δ¹⁸O relationship suggests that precursor intermediate melts were δ¹⁸O-poor (δ¹⁸O ≤ 5.50‰) relative to basalts erupted nearby (Fig. 3a). Olivine resorption has been mentioned above as a possible mechanism for lowering melt δ¹⁸O values. However, the Central Lau Basin intermediate and evolved lavas approximate a liquid line of descent characteristic of basalt−ferrobasalt−andesite series. Chemical evidence for phenocryst accumulation, which is observed in nearby seamount lavas, is not apparent in the majority of the spreading centre lavas (Pearce et al., 1994). Differences between the inferred δ¹⁸O of intermediate melts and local basalts are not large (i.e., 0.25‰ or more) and are of a similar magnitude to variations observed in individual phases from mantle peridotites (Mattey et al., 1994). While it could therefore be inferred that the intermediate melts were themselves derived from low-δ¹⁸O basaltic parents, there is no evidence for low-δ¹⁸O basaltic glasses of liquidus composition. Chemical evolution has exerted significant control on the major and trace element chemistry of the entire compositional spectrum of Central Lau Basin lavas (Pearce et al., 1994), and therefore we believe that differentiation from basaltic compositions to the point of oxide saturation in propagating rift tip magma chambers involved a decrease in melt δ¹⁸O values. Mechanisms that could lower δ¹⁸O contents during evolution towards lower MgO contents are: (1) a positive value for Δ¹⁸O_{avem−melt}...
Fig. 3. (a) Mg Number versus $\delta^{18}$O for Lau Basin glasses. Error bars are $\pm 1$ s.d. for replicated analyses or $\pm 0.08\%$ otherwise. Symbols and data sources as in Fig. 2. Stippled box represents the average reported for mid-ocean ridge basalts see Table 1. (b) Mg Number versus V (ppm) for Lau Basin lavas Frenzel et al., 1990; Johnson and Sinton, 1990; Sunkel, 1990; Pearce et al., 1994. In the Central Lau Basin suite the onset of V depletion is coincident with Fe and Ti depletion inferring oxide saturation in the melts.

4.1.1. Fractional crystallisation

Previous studies and the data presented here suggest that $\Delta^{18}$O$_{\text{plag-melt}}$, is positive for basaltic melt at magmatic temperatures (Muehlenbachs and Clayton, 1972; Woodhead et al., 1987, and references therein; Tables 1 and 2). Therefore, a positive $\Delta^{18}$O$_{\text{assem-melt}}$ value could be achieved if plagioclase crystallisation sufficiently exceeded the combined crystallisation of olivine and clinopyroxene. Pearce et al. (1994) propose several lines of evidence to suggest that plagioclase was the most abundant phase crystallising from Central Lau Basin basaltic melts including: negligible Sr enrichment, development of negative Eu anomalies, and enrichment of some mildly compatible trace elements, such as V, Co and Sc, as evolution proceeds (Fig. 3b). Therefore, the crystallising assemblage may remove $^{18}$O from basaltic melts in the Central Lau Basin more efficiently than in other suites of basalts.

The relative abundances of plagioclase and ferromagnesian phases in the crystallising assemblage can be estimated since the Sr concentrations in the Central Lau Basin lavas are relatively constant. This suggests that the bulk partition coefficient for Sr was approximately unity. Using reasonable estimates of Sr distribution coefficients between melt and the observed phenocrysts requires crystallisation of olivine : clinopyroxenes : plagioclase in the proportions 1 : 1 : 3. Applying the $\Delta^{18}$O$_{\text{assem-melt}}$ values we have measured in LB24-1-1 (Tables 1 and 2) to these phenocryst proportions produces an estimate for $\Delta^{18}$O$_{\text{assem-melt}}$ of $-0.17\%$. Proportions of 1 : 1 : 6 are required to produce a zero value for $\Delta^{18}$O$_{\text{assem-melt}}$, but would lead to significant Sr depletion in more evolved lavas without depleting $^{18}$O. Therefore, while plagioclase crystallisation was undoubtedly important during evolution of Central Lau Basin melts, and $\Delta^{18}$O$_{\text{assem-melt}}$ was probably more positive than is true of many cogenetic suites of lavas, fractional crystallisation of the observed assemblage cannot produce melts that are $^{18}$O-depleted, relative to their basaltic precursors. Additional processes must be considered to account for the low $\delta^{18}$O values inferred for propagating rift tip lavas immediately prior to oxide saturation.

4.1.2. Assimilation and the depth of magma chambers

Assimilation of $^{18}$O-poor country rock has been proposed to explain low $\delta^{18}$O values in lavas erupted through thick, oceanic (hot-spot) crust in Iceland and Hawaii, and island arc crust in Indonesia (Muehlenbachs et al., 1974; Hattori and Muehlenbachs, 1982; Condomines et al., 1983; Taylor, 1987; Harmon and Gerbe, 1992; Eiler et al., 1996; Thirlwall et al., 1996). Rocks with low $\delta^{18}$O, relative to average MORB, occur at depths of 2–5 km beneath the
(palaeo-)ocean floor in the lower sheeted dykes and layered gabbros of both modern ocean crust and ophiolites (Gregory and Taylor, 1981; Ito and Clayton, 1983; Muehlenbachs, 1986; Stakes et al., 1991; Kempton et al., 1991; Lecoyer and Reynard, 1996). Low $\delta^{18}$O values develop through interaction between crust and seawater at relatively high temperatures ($>300^\circ$C) in the hydrothermal regime associated with ocean ridge magmatism (Gregory and Taylor, 1981). Mineralogical changes associated with this process are the alteration of feldspar and growth of hydrous phases such as amphibole, chlorite and epidote (Ito and Clayton, 1983; Stakes et al., 1991). It is these altered and hydrous, secondary phases at intermediate depths that represent the major $\delta^{18}$O-poor reservoir in oceanic crust. Intrusion of evolved melts into brecciated regions at mid-crustal levels has been reported in the Oman ophiolite and the South West Indian Ocean Ridge (Gregory and Taylor, 1981; Stakes et al., 1991). Such associations could provide an ideal situation for melt–crust interaction and may provide an explanation for low $\delta^{18}$O values in intermediate Central Lau Basin melts.

Evidence for melt stagnation at suitable depths beneath the ocean-floor exists both for propagating rift tip lavas in general, and specifically for the Central Lau Basin suite. Macpherson and Mattey (1994) modelled CO$_2$ degassing of the Central Lau Basin glasses analysed in this work and proposed that volatiles and melt equilibrated at a minimum depth of 2.5 km beneath the seafloor. The absence of amphibole, particularly in evolved, water-rich lavas such as those of the CLSC, suggests a maximum depth for differentiation of 5 km (Spulber and Rutherford, 1983; Farley and Newman, 1994). In addition, experiments on oceanic tholeiite from the Galapagos Spreading Centre suggest that evolution of tholeiitic melt to ferrobasalt and more evolved CLSC compositions is most likely to occur at depths of 3–6 km (Spulber and Rutherford, 1983). Therefore, both the petrology and carbon isotope geochemistry of Central Lau Basin lavas point to stagnation of melts at depths of 2.5–5 km beneath the seafloor.

Simple mass balance calculations can be used to estimate the volume of oceanic crust that must be assimilated to generate the lowest oxygen isotope ratios observed from melts similar to the local basalts. Minimum whole rock $\delta^{18}$O values in ophiolite and modern ocean crust sequences typically range from 3.5‰ to 4.0‰ (Muehlenbachs, 1986, and references therein). Assuming initial CLSC basalts have $\delta^{18}$O values of 5.75‰ (site 15, 7.5 wt.% MgO), then 10% and 15% assimilation of crust with oxygen isotope ratios of 3.5‰ and 4.0‰, respectively, is required to produce the lowest $\delta^{18}$O$_{glass}$ value. Spulber and Rutherford (1983) showed that the point of oxide saturation in propagating rift tholeiites depends on $f_{o_2}$ of the melt but that at 1 kbar the fraction of melt remaining, $F$, would be approximately 0.3. Therefore, if the lowest oxygen isotope ratio occurred immediately prior to oxide saturation, we can estimate that $r$, the ratio of mass assimilated to mass crystallised, is approximately 0.2 (i.e. $\sim 12.5\%/70\%$). These estimates of mass assimilated and $r$ should be regarded as maxima for three reasons. As noted above, the $\Delta^{18}$O$_{assim.-melt}$ value in Central Lau Basin basaltic melts was probably less negative than is regarded typical of most spreading centre melts; therefore, fractional crystallisation would be less efficient in enriching the $^{18}$O content of the melts. Second, the least refractory phases in the wall rock are the hydrous and altered minerals formed by hydrothermal processes. Compared to bulk contamination, selective assimilation of a smaller mass of these $^{18}$O-poor phases could produce similar isotopic shifts in the melts. Finally, a smaller quantity of contaminant would also be required if the melts interacted with very low $^{18}$O/$^{16}$O crust (<3‰) such as found in the Hess Deep (Lecoyer and Reynard, 1996).

Therefore, we suggest that the melts that were precursors to evolved Central Lau Basin lavas assimilated <15% of their own mass of $\delta^{18}$O-depleted oceanic crust during crystallisation of a plagioclase-dominated assemblage. A combination of these processes produced low $\delta^{18}$O intermediate melts. Assimilation of such quantities of wall rock is easily reconciled with the limited size of magma chambers beneath propagating rift tips where waxing magmatism would be accompanied by relatively large temperature contrasts between magma and wall rock (Christie and Sinton, 1981; Perfit et al., 1983). Oxide saturation, at Mg No. $\approx 45$, had a major influence on oxygen distribution due to the large, negative value of $\Delta^{18}$O$_{oxide-melt}$ causing further crystallisation to enrich melts in $^{18}$O (Muehlenbachs and Byerly,
Reversal of $\Delta^{18}O_{\text{plag-melt}}$ in evolved melts may also have contributed to reducing the $\delta^{18}O$ value of the crystallising assemblage with respect to the melts (Taylor and Sheppard, 1986). It is possible that the most evolved melts may have continued to assimilate $^{18}O$-poor crust, but the effects have been masked by oxide crystallisation. Clearly, the interplay of crystallisation and assimilation may generate a wide range of evolutionary vectors in Mg No.$-$ $\delta^{18}O$ space, depending on the mineralogy and isotopic composition of the crust and the value of $\Delta^{18}O_{\text{asem-melt}}$ during differentiation.

Trace element concentrations in CLSC lavas are similar to MORB except for slight enrichment of the large ion lithophile elements (LILE) relative to high field strength elements (HFSE) and light rare earth elements (LREE) with similar compatibilities. Pearce et al. (1994) suggest that these features could result from contribution of either modern or ancient subduction components. However, we note that evolved lavas from the Galapagos propagating rift tips, remote from any subduction zone, also display trace element enrichments that exceed the predictions of closed-system crystallisation models (Mattey and Muir, 1980; Perfit et al., 1983). O’Hara and Matthews (1981) showed that cannibalisation of previous eruptive products by a magma chamber will enrich incompatible element concentrations of melts while the effects on compatible element concentrations would be negligible. Crust beneath the CLSC rift tip is very similar to presently erupting CLSC melts in terms of major and trace elements, but with slightly higher LILE concentration due to their origin closer to the Tonga Arc (e.g. DSDP Site 836, Ewart et al., 1994; Parson et al., 1994). Therefore, assimilation of oceanic crust may be, at least partially, responsible for the small LILE enrichment observed in CLSC lavas, either through the LILE-rich nature of the crust and/or LILE concentration during differentiation (O’Hara and Matthews, 1981).

Crustal assimilation has also been advocated by Hilton et al. (1993) to explain radiogenic He and atmospheric Ar isotopic ratios in evolved spreading centre lavas from Valu Fa Ridge. The Valu Fa spreading centre propagated through attenuated arc crust that is the inferred source for the radiogenic helium (Hilton et al., 1993). However, the CLSC propagated through relatively young back-arc crust generated at the ELSC approximately 1 Ma (Parson et al., 1994). Since there has been little time for in situ growth of $^4$He, assimilation of young crust should be expected to have little effect on He isotope ratios in CLSC melts (Kurz and Jenkins, 1981; Hilton et al., 1993). This is confirmed by a $^{3}$He/$^4$He ratio of 8.6 $R_e$, indistinguishable from the MORB average, measured in a CLSC rift tip glass despite a very low He concentration that would have made the melt highly susceptible to contamination (Poreda and Craig, 1992). Low noble gas concentrations are also observed in glasses from the Valu Fa Ridge, that are thought to result from extensive degassing of noble gases prior to assimilation (Hilton et al., 1993).

However, carbon isotope systematics suggest that $\delta^{13}C$ values of Central Lau Basin glasses were largely controlled by eruption that must occur after assimilation of magma chamber wall rocks (Macpherson and Mattey, 1994). Therefore, the chemical effects of assimilation may provide a useful reference point in resolving pre- and syn-eruptive degassing events affecting ocean ridge magmatism.

4.2. Evolution of Valu Fa, Northern Lau Basin and Fiji Basin lavas

The Valu Fa glasses analysed in this work are representative of a larger suite collected from a small area (< 30 km along the ridge). Chemical compositions of the entire suite approximate a single liquid line of descent involving olivine + plagioclase + clinopyroxene crystallisation, with an oxide joining the assemblage at a late stage (Frenzel et al., 1990). Increasing $\delta^{18}O$ values with decreasing Mg Numbers are consistent with fractional crystallisation of the observed phenocrysts and resemble $^{18}O$-enrichments reported from many cogenetic suites of lavas (Matsuhisa et al., 1973; Muehlenbachs and Byerly, 1982; Sheppard and Harris, 1985; Woodhead et al., 1987). The highest $\delta^{18}O$ value in this study was obtained for the Valu Fa dacite glass and is consistent with oxide saturation during its differentiation. Therefore, $\delta^{18}O$ values of Valu Fa glasses presented here may reflect fractional crystallisation of a melt with $\delta^{18}O$ similar to SO35-124KD1.

Oxygen isotope ratios obtained by Lichtenstein and Loock (1993) for Valu Fa Ridge lavas display
the greatest variation yet observed for fresh Lau Basin lavas and show considerably greater variation than our analyses of glasses from this spreading centre (Fig. 3a). The lower end of the range observed by these workers coincides with our data so may also result from differentiation of a MORB-like melt. The higher $\delta^{18}O$ values, at a similar Mg No., require addition of an $^{18}O$-rich component. While Sr, Nd and Pb isotope ratios of some of the $^{18}O$-rich lavas provide evidence for recycling of subducted material (Loock et al., 1990), coexistence of lavas with variable $\delta^{18}O$ on the same spreading segments suggests either that oxygen isotope heterogeneity exists on a very short length scale in the mantle beneath the Valu Fa Ridge or that melts have interacted with $^{18}O$-rich crust to varying degrees. Establishment of new magmatic plumbing systems associated with ridge propagation provides an ideal environment for crustal assimilation to occur, as has been convincingly demonstrated for radiogenic and rare gas isotopes of some Valu Fa lavas (Hilton et al., 1993).

The arc crust through which the Valu Fa Ridge has propagated may possess considerable $^{18}O/^{16}O$ heterogeneity since it has been constructed by successive cycles of arc magmatism, probably including the development of hydrothermal systems, and sedimentation related to both the Lau and Tonga arcs (Parson and Hawkins, 1994, and references therein). Therefore, rift propagation through rifted arc crust may be even more conducive to crustal assimilation than is the case for propagation through oceanic crust, and may be an important process in the development of the total $\delta^{18}O$ variation observed in Valu Fa Ridge lavas (Fig. 3a). Further investigation of $\delta^{18}O$ variations in Valu Fa lavas and attenuated Southern Lau Basin crust are required to constrain the roles of source and magma chamber processes on what is one of the world’s most petrologically diverse suites of spreading centre lavas.

The Northern Lau and Fiji Basin glasses were

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**Fig. 4.** Relationship between $\delta^{18}O$ values and Ba/La in Central Lau Basin glasses. Error bars as in Fig. 3. Three binary mixing curves for slab-derived hydrous fluid and mantle peridotite are shown: $M$ = contamination of fertile MORB source by fluid derived from altered oceanic crust and sediment; $D$ = contamination of depleted MORB source by fluid derived from altered oceanic crust and sediment; $P$ = contamination of fertile MORB source by fluid derived from sediment. Tick marks indicate the fraction of fluid in the source. Element abundances for mantle sources and fluids from Ewart and Hawkesworth (1987) and Hawkesworth et al. (1991), respectively. An oxygen ratio of 2:1 was assumed for (hydrous) fluid:peridotite. Oxygen isotopic compositions are $\pm 5.75\%$ for melt from uncontaminated peridotite and $\pm 19\%$ for Pacific sediment (Savin and Epstein, 1970).
collected over a wide area and are unlikely to represent a single liquid line of descent (Fig. 1a). While $^{18}$O-enrichment with decreasing Mg No. is observed, and again appears to be concordant with whole rock powder analyses (Fig. 3a), the $\delta^{18}$O variation cannot be unequivocally attributed to fractional crystallisation.

4.3. Source of oxygen in Lau Basin basalts

Oxygen isotope compositions of basaltic glasses (Mg No. > 50) from the Central Lau Basin display a very weak positive correlation between Mg Number and $\delta^{18}$O that may reflect evolution towards $^{18}$O-depleted compositions by the processes outlined in the previous section (Fig. 3a). Assimilation of altered ocean crust has recently been advocated to explain low $\delta^{18}$O values in ocean island basalts from Hawaii and Gran Canaria (Eiler et al., 1996; Thirlwall et al., 1996). We also note that cogenetic suites from Ascension Island and Chichijima Island each display a $\delta^{18}$O minimum at intermediate compositions (Sheppard and Harris, 1985; Dobson and O’Neil, 1989). Therefore, assimilation of hydrothermally altered crust (enhanced plagioclase crystallisation) may play a more important role in the evolution of oceanic basalts than previously realised. However, the major and trace element chemistry of the Central Lau Basin basalts reveals that recycling of subducted water and mantle source fertility influence their chemistry (Pearce et al., 1994). Correlations between oxygen isotope ratios and chemistry of the basaltic glasses that are described in this section suggest that these processes also influence $^{18}$O/$^{16}$O.

Pearce et al. (1994) have demonstrated that water may be recycled into the source of some ELSC and ILSC basalts from the down-going slab. The presence of water is manifested in several ways. First, enrichment of some LILE elements, such as Ba and Rb, in ELSC lavas most probably results from the high mobility of these elements in hydrous fluids released from subducted material (Tatsumi et al., 1986; You et al., 1996). Second, hydrous melting of peridotite close to the Tonga Arc is inferred from high Si$_{8.0}$ values of ELSC lavas that are thought to reflect enhanced orthopyroxene melting at high $P_{H_2O}$ (cf. Jaques and Green, 1980). Finally, a positive correlation between Na$_{8.0}$ and Fe$_{8.0}$ among the basalts may result from variable mantle fertility. Residual sources produce basaltic melts with lower sodium concentrations, at a given MgO content, possibly due to previous extraction of a relatively Na-rich melt (Pearce et al., 1994). The most depleted mantle signatures are found in ELSC lavas close to the arc suggesting that slab-derived fluid may encourage melting of residual peridotite.

Figs. 4 and 5 explore the possibility that recycling and source fertility may influence oxygen isotope ratios of basaltic lavas erupted in the Lau Basin. Fig. 4 displays a positive correlation between $\delta^{18}$O and Ba/La in Central Lau Basin basalts that extends from MORB-like values along a limited trajectory towards possible recycled fluid compositions. Simple models of mantle wedge contamination by fluids derived from altered MORB + sediment (curves M and D) and sediment alone (curve P) have been calculated for fertile and depleted peridotite (see caption to Fig. 4 for details). The best fit to the Central Lau Basin data is obtained for models involving a sediment-derived fluid and/or a depleted composition.
mantle source. Fluid fluxes required to produce the \( \delta^{18}O \) variation among the ELSC and ILSC lavas are \( \leq 0.5\% \) and cannot be resolved on plots involving Sr and Nd isotopes (Hergt et al., 1991).

Our data for Central Lau Basin basaltic glasses display a positive correlation between \( Na_{k6}^{2+} \) and \( \delta^{18}O \), and several bomb fluorination analyses of Northern and Central Lau Basin lavas also fall within error of this trend (Fig. 5). Note that the most significant departures from this trend are due to high \( \delta^{18}O \) values of some Valu Fa lavas that may reflect the presence of a recycled or assimilated \( ^{18}O \)-rich component (see above). Similar relationships can be demonstrated between \( \delta^{18}O \) and other geochemical properties that reflect variations in source fertility, e.g. \( Ti/Zr, Al_{2}O_{3}/TiO_{2} \) and \( CaO/TiO_{2} \) (Ewart and Hawkesworth, 1987; Woodhead et al., 1993). These correlations clearly suggest that a link exists between fertility of the Lau Basin mantle and oxygen isotope ratios of extracted melts. Two origins can be postulated for this relationship: (1) an indirect link as a result of recycling processes, or (2) direct control of \( ^{18}O/^{16}O \) by mineralogy and melting.

An indirect link could arise from a greater fluid flux being required to induce melting in residual mantle (Pearce et al., 1994). In this case, a larger proportion of \( ^{18}O \)-rich, slab-derived fluid would be associated with melting of the least fertile portions of the wedge (Fig. 5). Although both \( ^{18}O \)-rich fluid and depleted (low-\( Na_{k6}^{2+} \)) peridotite would be required to produce the observed melts, the fertility of the source would not directly cause the high \( \delta^{18}O \) values of melts.

Alternatively, mantle fertility may directly produce the oxygen isotope heterogeneity of Lau Basin basalts. Preliminary isotopic fractionation data for the systems enstatite–basaltic melt (Muehlenbachs and Kushiro, 1974) and olivine–tholeiitic melt (Kyser, 1990) suggest that melts may be \( ^{18}O \)-depleted relative to their peridotite residues at very high melting temperatures, i.e. \( \Delta^{18}O_{enst-melt} \) and \( \Delta^{18}O_{ol-melt} \) are positive. Based on then available peridotite data, Kyser et al. (1982) proposed this as a means to enrich the \( ^{18}O \) content of peridotite residues during melting and to explain global oxygen isotope differences between different types of basalt observed by Kyser et al. (1981). Employing these experimental fractionation factors the relationship in Fig. 5 could be interpreted as a result of Lau Basin basalts being extracted from mantle that had previously lost varying fractions of a relatively Na-rich, \( ^{18}O \)-poor melt. However, we believe that this explanation cannot be reconciled with the mass balance of oxygen. Generation of tholeiitic melt from peridotite requires 10–30% partial melting (Jaques and Green, 1980) and, since oxygen abundances are similar in molten and crystalline silicate, oxygen isotope distribution during peridotite melting should be controlled by the residue. Therefore, melting of peridotite with positive \( \Delta^{18}O_{residue-melt} \) values would produce primary tholeiites with \( \delta^{18}O \) values lower than bulk peridotite (i.e. < 5.5\%, Mattey et al., 1994). Fractional crystallisation is unable to enrich melts by more than a few hundredths of a per mil without driving them to SiO\(_2\)-rich compositions, outside the basalt field (Muehlenbachs and Byerly, 1982; Sheppard and Harris, 1985); therefore low \( \delta^{18}O \) basalts would be commonplace on the Earth’s surface. The reverse is true: MORB, the most voluminous molten extract of the mantle, have an average \( ^{18}O \) value greater than the peridotite mean (Table 1; Ito et al., 1987). Thus, we are not convinced that \( \Delta^{18}O_{residue-melt} \) values are positive and do not believe that \( ^{18}O \)-rich melts from residual peridotite are generated due to previous extraction of a low-\( ^{18}O \) melt.

A direct link between \( ^{18}O/^{16}O \) and mantle fertility can be explained if isotopic fractionation factors during melting are similar to values observed and inferred from natural lavas and peridotites (i.e. \( \Delta^{18}O_{enst-melt} > \Delta^{18}O_{ol-melt} > \Delta^{18}O_{cx-melt} \); cf. Mattey et al., 1994). Under equilibrium conditions incongruent melting of orthopyroxene will increase the modal abundance of olivine in peridotite residue producing a larger (more negative) \( \Delta^{18}O_{ol-melt} \) value. Since a greater volume of oxygen is present in the residue than in melts, this increase in \( \Delta^{18}O_{residue-melt} \) will tend to enrich melts in \( ^{18}O \). Orthopyroxene melting is enhanced during melting of residual peridotite (Jaques and Green, 1980); therefore, increasingly residual peridotite will yield increasingly \( ^{18}O \)-rich equilibrium melts. High-\( ^{18}O \) melts could also be generated from residual peridotite by disequilibrium melting. Orthopyroxene consistently has the highest \( \delta^{18}O \) value among the major peridotite phases (Mattey et al., 1994). There-
fore, since orthopyroxene contributes a relatively large fraction to the melt during melting of residual peridotite (Jaques and Green, 1980), melts produced from more harzburgitic sources would be relatively \(^{18}\)O-rich. We note that melts from residual sources tend to have relatively high values of Mg Number (Jaques and Green, 1980) and this could explain the weak Mg No.\(-\delta^{18}\)O correlation in the basalts (Fig. 3a). A larger recycled fluid-flux is still inferred to produce melting of residual peridotite (Pearce et al., 1994), thus producing the Ba/La\(-\delta^{18}\)O correlation of Fig. 4. However, in the scenarios outlined in this paragraph the fluid does not contribute excess \(^{18}\)O to source prior to melting.

At present it is not possible to distinguish whether the correlation between Na\(_{\text{Na}}\) and oxygen isotope ratios in Lau Basin basalts results from addition of an \(^{18}\)O-rich fluid or from melting effects driven by the presence of such a fluid in variably depleted peridotite. Both situations are consistent with the highest fluid flux and most residual peridotite occurring in ELSC lavas close to the Tonga Arc. If mantle fertility does exert a strong control on oxygen isotope ratios then similar correlations may be expected for lavas within and between individual segments of spreading centres worldwide.

A final point worthy of discussion is the lack of correlation between oxygen isotope ratios and water content in the Lau Basin glasses (Fig. 2). In the preceding paragraphs we have outlined three possible origins for \(\delta^{18}\)O variation in Lau Basin basalt, each of which require introduction of a hydrous component to the mantle source. From the arguments outlined in this work and by Pearce et al. (1994), a positive correlation between \(\delta^{18}\)O and H\(_2\)O concentration may be expected through either addition of an \(^{18}\)O-rich hydrous fluid to peridotite or enhanced orthopyroxene melting resulting from high fluid fluxes. On the other hand, water is an incompatible element so its concentration in melts should decrease as the degree of melting increases. At crustal levels, fractional crystallisation will tend to increase both \(\delta^{18}\)O and H\(_2\)O in residual melts while assimilation of \(^{18}\)O-poor oceanic crust would lead to a negative correlation between these properties. Given the possibility of interplay between these processes we are not surprised at the absence of a single, significant correlation between the oxygen isotope ratios and water concentrations in lavas throughout the range of compositions. However, this observation does prompt the following important comment. Water contents are frequently used to correct oxygen isotope analyses of whole-rock powders for the effects of post-eruptive alteration by assuming that all alteration produces simultaneous hydration and \(^{18}\)O enrichment (e.g. Ferrara et al., 1985; Harmon et al., 1987). The Lau Basin data show that evolution of oceanic tholeiites through combined assimilation and fractional crystallisation may elevate water concentrations without modifying \(\delta^{18}\)O and that magmatic H\(_2\)O concentrations can display an order of magnitude variation at near-constant \(\delta^{18}\)O (Fig. 2). This suggests that it may be invalid to correct oxygen isotope ratios of lava suites by assuming a constant initial water concentration.

In summary, the lack of quantitative isotope fractionation data and complementary studies on petrogenetically simpler regions prevents resolution between three models for generation of oxygen isotope variations in Lau Basin basalts. A weak, negative correlation between Mg No. and \(\delta^{18}\)O may reflect differentiation of basaltic melts towards \(^{18}\)O-poor intermediate compositions similar to those postulated as precursors to the most evolved lavas. Alternatively, recycling of a hydrous fluid from the subducted slab may produce \(^{18}\)O-rich lavas either by elevating \(\delta^{18}\)O of the mantle source, or by enhancing orthopyroxene melting in residual peridotite. Further investigation of oxygen isotope variation in lavas from individual segments of spreading centres, remote from subduction zone influences, may help clarify the role of melting in generating \(^{18}\)O/\(^{16}\)O heterogeneity in basaltic lavas worldwide. At that time the relative importance of source fertility and recycling of oxygen in subduction related magmatism may become clearer.

5. Conclusions

(1) Magma chamber processes exert strong influences on oxygen isotope ratios in evolved Lau Basin lavas. Evolution of melts at the Central Lau Basin propagating and dying rift tips initially involved assimilation of a modest amount of hydrothermally altered oceanic crust (< 15%) and crystallisation of
a plagioclase-dominated assemblage. This produced $^{18}$O-depleted intermediate melts and may also have elevated LILE abundances. Enrichment of $^{18}$O in melts commenced at the point of oxide saturation. Oxygen and carbon isotope ratios of propagating and dying rift tip glasses are consistent with differentiation and degassing in magma chambers between 2.5 and 5 km below the seafloor (Macpherson and Mattey, 1994; this study).

(2) The Central Lau Basin data suggest that crustal contamination cannot be discounted merely on the strength of values falling in the range of $\delta^{18}$O values typical of oceanic basalts. The nature of country rocks must be carefully considered before such statements are made. Oceanic basalts may be susceptible to both elevation and reduction of $^{18}$O/$^{16}$O through interaction with hydrothermally altered crust. This conclusion is supported by LF oxygen isotope evidence from ocean islands (Eiler et al., 1996; Thirlwall et al., 1996).

(3) Valu Fa Ridge glasses display a small $^{18}$O-enrichment with chemical evolution consistent with fractional crystallisation of a MORB-like phenocryst assemblage. A significant part of whole rock, bomb fluorination data for lavas from Valu Fa Ridge is coincident with the new data. However, higher $\delta^{18}$O values in some, conventionally analysed, lavas with similar values of Mg No. may reflect assimilation of old arc crust or recycling of subducted sediment (Lichtenstein and Loock, 1993). Sampling sites in the Northern Lau Basin are too disparate to allow firm conclusions to be drawn regarding the role of differentiation in oxygen isotope systematics.

(4) Basaltic glasses from throughout the Lau Basin display correlations between $\delta^{18}$O and geochemical parameters that reflect source fertility. It is unclear whether this is the result of a single process or a more complex melting history requiring higher ($^{18}$O-rich) fluid fluxes to melt increasingly residual peridotite. Insufficient control on oxygen isotope partitioning is presently available from experimental studies or petrogenetically less complex suites of basalts to make qualitative assessments of these relationships. In particular, further investigation of experimental and natural glasses is required to clarify the role of orthopyroxene melting in generating $\delta^{18}$O variations in cogenetic suites of lavas.

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