Geochemical Investigation of Gabbroic Xenoliths from Hualalai Volcano: Implications for Lower Oceanic Crust Accretion and Hualalai Volcano Magma Storage System

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Word count (excluding abstract): 6508

Figures and Tables: 9 figures and 1 table

References: 60

List of supplements: analytical methods; melt REE concentrations calculation; MELTS modeling; MORB Nd isotopic heterogeneity; 4 figures; 9 tables; 13 supplement references.

16 **Abstract**

17 The patterns of axial hydrothermal circulation at mid-ocean ridges both affect and are
18 influenced by the styles of magma plumbing. Therefore, the intensity and distribution of
19 hydrothermal alteration in the lower oceanic crust (LOC) can provide constraints on LOC
20 accretion models (e.g., “gabbro glacier” vs. “multiple sills”). Gabbroic xenoliths from Hualalai
21 Volcano, Hawaii include rare fragments of *in situ* Pacific lower oceanic crust. Oxygen and
22 strontium isotope compositions of 16 LOC-derived Hualalai gabbros are primarily within
23 the range of fresh MORB, indicating minimal hydrothermal alteration of the *in situ* Pacific
24 LOC, in contrast to pervasive alteration recorded in LOC xenoliths from the Canary Islands.
25 This difference may reflect less hydrothermal alteration of LOC formed at fast ridges than at
26 slow ridges. Mid-ocean ridge magmas from slow ridges also pond on average at greater and
27 more variable depths and undergo less homogenization than those from fast ridges. These
28 features are consistent with LOC accretion resembling the “multiple sills” model at slow
29 ridges. In contrast, shallow magma ponding and limited hydrothermal alteration in LOC at
30 fast ridges are consistent with the presence of a long-lived shallow magma lens, which
31 limits the penetration of hydrothermal circulation into the LOC.

32 Most Hualalai gabbros have geochemical and petrologic characteristics indicating
33 derivation from Hualalai shield-stage and post-shield-stage cumulates. These xenoliths
34 provide information on the evolution of Hawaiian magmas and magma storage systems.
35 MELTS modeling and equilibration temperatures constrain the crystallization pressures of
36 7 Hualalai shield-stage-related gabbros to be ~2.5-5 kbar, generally consistent with inferred
37 local LOC depth. Therefore a deep magma reservoir existed within or at the base of the LOC
38 during the shield stage of Hualalai Volcano. Melt-crust interaction between Hawaiian melts
39 and *in situ* Pacific crust during magma storage partially overprinted clinopyroxene Sr and
40 Nd isotope compositions of LOC-derived gabbros. Although minor assimilation of Pacific
41 crust by Hawaiian melts cannot be excluded, the range of oxygen isotope compositions
42 recorded in Hawaiian lavas and cumulates cannot be generated by assimilation of the *in situ*
43 LOC gabbros, which have relatively uniform and MORB-like $\delta^{18}\text{O}$ values. To first order, the
44 isotopic heterogeneity observed in Hawaiian melts appears to derive from the
45 heterogeneous plume source(s), rather than assimilation of local oceanic crust.

1. Introduction

Hydrothermal circulation at mid-ocean ridges is intimately linked to the processes of oceanic crust accretion. Axial hydrothermal circulation drives cooling and crystallization of axial magmas. In addition, the distribution of magmas in the crust, in particular the presence or absence of a long-live shallow axial magma chamber, influences the depth and degree of axial hydrothermal circulation. Hydrothermal alteration of newly formed oceanic crust dominates the heat and mass exchange between the Earth's lithosphere and hydrosphere. However, the intensity and distribution of hydrothermal circulation in gabbro cumulates, the thickest layer of oceanic crust, is still poorly constrained.

The ca. 1800 Kaupulehu flow of Hualalai Volcano, Hawaii contains abundant dunite, wehrlite and gabbro, and rare websterite and anorthosite xenoliths (Jackson et al., 1981; Clague, 1987; Kauahikaua et al., 2002). A small subset of the gabbros appears to be Pacific lower oceanic crust (LOC) layer-3 gabbros (Clague, 1987; Lassiter & Hauri, 1998). These gabbros can be used to examine the hydrothermal alteration history of LOC, and provide constraints on the mechanism of LOC accretion. The majority of Hualalai gabbros are cumulates derived from Hualalai shield- and post-shield-stage melts (Clague, 1987). These gabbros can be used to investigate the Hualalai magma plumbing system and the role of crustal assimilation in the evolution of Hawaiian magmas.

In this study, we report mineral major and trace element and Sr-Nd-O isotope variations in a suite of 28 Hualalai gabbros, including 16 samples with chemical characteristics indicative of derivation from *in situ* Pacific LOC and 12 samples related to the shield and post-shield stages of Hualalai magmatism. Strontium and oxygen isotope data of LOC-derived gabbros are used to evaluate the extent of hydrothermal alteration of the *in situ* LOC, providing constraints on different models of LOC formation. We examine mineralogical and compositional constraints on the fractionation pressures of Hualalai-related gabbros to

constrain the depth(s) of magma storage, and examine compositional variations in LOC- and Hualalai-derived gabbros to evaluate the effects of melt-crust interaction in the evolution of the *in situ* LOC and Hawaiian magmas.

2. Background

Hawaiian volcanoes commonly evolve through four stages that are characterized by different lava types, magma supply rates and inferred degrees of mantle melting (Clague, 1987). Magma supply rates and degree of melting increase as the volcano moves towards the center of the plume, and then decrease as the Pacific plate moves away from the plume (e.g., Frey et al., 1990). Small volumes of alkalic basalt are primarily erupted during the pre-shield stage. Large volumes of tholeiitic basalt dominate the following shield stage. The post-shield stage is characterized by a return to mostly alkalic basalt and associated differentiated lavas. Finally, some Hawaiian volcanoes erupt small volumes of alkalic basalt, basanite and nephelinite during a rejuvenated stage, usually following a hiatus in activity at the end of the post-shield stage (Clague, 1987).

Hawaiian lavas also display significant temporal and spatial variability in isotopic composition. Post-shield and rejuvenated-stage lavas typically have more depleted isotopic compositions than shield-stage lavas (e.g. Chen & Frey, 1985). In addition, recent Hawaiian volcanoes display systematic spatial-compositional variations that define two parallel linear trends. “Loa trend” volcanoes typically have more “enriched” Sr-Nd isotopic compositions compared to “Kea trend” volcanoes, and are also characterized by higher $^{208}\text{Pb}/^{204}\text{Pb}$ at a given $^{206}\text{Pb}/^{204}\text{Pb}$ (Abouchami et al., 2005). In addition, olivine in “Loa trend” lavas have $\delta^{18}\text{O}$ values that extend as high as +6.1‰, beyond the range of fresh MORB or most mantle peridotites. Olivine in “Kea trend” lavas have $\delta^{18}\text{O}$ values as low as +4.3‰ (Eiler et al., 1996).

The chemical characteristics of “Loa trend” lavas likely reflect ancient recycled oceanic crust and sediments entrained in the Hawaiian plume source (e.g. Eiler et al., 1996; Lassiter & Hauri, 1998). In contrast, several different origins have been proposed for the “Kea trend” lavas, including incorporation of depleted upper mantle asthenosphere (e.g. Chen and Frey, 1985), lower Pacific crust or lithospheric mantle (e.g. Eiler et al., 1996), and recycled lithosphere within the Hawaiian plume itself (e.g. Lassiter & Hauri, 1998). Evaluation of the role of local crust/lithospheric mantle is difficult because the composition of the *in situ* Pacific crust and lithospheric mantle is not well characterized. Furthermore, the extent to which Hawaiian magmas pond and crystallize within the oceanic crust/lithospheric mantle during different stages of Hawaiian volcanism is unclear.

Hualalai Volcano is located on the west coast of the Big Island of Hawaii (Fig. 1). Post-shield alkali basalts cover most of the subaerial portion of Hualalai Volcano. Shield-stage tholeiitic basalts are exposed along the submarine northwest rift zone. The 1800-1801 Huehue and the slightly older Kaupulehu alkali basalt lava flows contain abundant mafic and ultramafic xenoliths that are interpreted as Hualalai cumulates and *in situ* Pacific LOC (Clague, 1987). These xenoliths provide new constraints on the composition and evolution of the local LOC, and on the evolution of Hawaiian volcano magma storage systems.

3. Samples and methods

We examined 28 gabbroic xenoliths from the c.a. 1800-1801 Kaupulehu alkalic basalt flow of Hualalai Volcano. Seventeen samples collected by Dale Jackson and Mel Beeson in 1963 and 1965 are from the Jackson Collection at the Smithsonian National Museum of Natural History. Eleven additional samples were collected by David Clague and Wendy Bohrsen in 1983 and 1984, or by Clague between 1987 and 1996. Previous studies have shown that a small population of Hualalai gabbros (<<1%) has low, MORB-like $^{87}\text{Sr}/^{86}\text{Sr}$

(e.g., Clague and Chen, 1986; Clague, 1987; Lassiter & Hauri, 1998). These samples also have depleted LREE (light rare earth element) and $\text{Sm/Yb} < 1$. We selected samples with low $^{87}\text{Sr}/^{86}\text{Sr}$ and/or low Sm/Yb , thought to derive from the *in situ* LOC beneath the Hawaiian Islands. Samples analyzed also include gabbros with higher $^{87}\text{Sr}/^{86}\text{Sr}$ and/or higher Sm/Yb , which appear to derive from Hualalai shield and post-shield stage magmas.

The gabbroic xenoliths generally have modal mineral abundances ranging from 40-60% plagioclase, 35-55% clinopyroxene, and ~2% spinel. Some gabbros also contain up to 10% orthopyroxene and 20% olivine (point counting from this study and data from Jackson et al., 1981). These gabbros can be classified into three distinct populations based on their mineralogy and chemical characteristics. Group-1 and 2 gabbros contain both clinopyroxene and orthopyroxene, whereas group-3 gabbros lack orthopyroxene. Group-1 and 2 gabbros are distinguished by clinopyroxene Sm/Yb ratios (Fig. S1). Most samples are relatively fresh with only minor alteration (likely iddingsite) on olivine and Fe-oxidation on pyroxene grain boundaries or surfaces. 88Kap-2, 92Kap-3, 63Kap-15, 65-100-110, 65-115-10, 63Kap-7 and 65-60-17 show greater surface alteration. Clinopyroxene in some samples exhibits orthopyroxene exsolution reflecting sub-solidus cooling. In some samples, the presence of highly irregular grain boundaries may indicate incipient partial melting. Planar fluid inclusion arrays in minerals indicative of annealed fractures are also present in some xenoliths.

Clinopyroxene, orthopyroxene, and plagioclase major element compositions were measured by EPMA at the USGS and University of Texas at Austin (UT Austin). Clinopyroxene trace element compositions were measured by LA-ICP-MS (UT Austin). Strontium and neodymium isotopes were measured by TIMS. Neodymium and samarium concentrations were measured by isotope dilution using TIMS. Oxygen isotope ratios were

measured by laser fluorination. All isotopic analyses were performed at UT Austin. Refer to supplementary information for more method details.

4. Results

Major element mineral compositions are presented in Supplemental Tables S1 through S3. Clinopyroxene Mg# [molar $\text{Mg}/(\text{Mg}+\text{Fe}) \times 100$] varies from 73 to 87, and negatively correlates with clinopyroxene TiO_2 content (Fig. S1). Plagioclase anorthite contents (An) range from 57 to 98. Core-rim variation of An is up to 6, but neither cores nor rims are systematically more anorthitic. For some samples, clinopyroxene Mg# variation among grains is up to 2, and plagioclase An variation is up to ~ 10 . However, intra-sample heterogeneity is small compared to inter-sample heterogeneity. Group-1 and 2 gabbros have systematically lower clinopyroxene TiO_2 contents and extend to higher Mg# than group-3 gabbros. Group-2 and 3 gabbros have similar plagioclase K_2O contents, whereas group-1 gabbros have systematically lower K_2O contents than the other two groups (Fig. S2).

Clinopyroxene trace element abundances are presented in Supplemental Table S4. Clinopyroxene from group-2 and 3 gabbros have high LREE concentrations ($[\text{La}] = 0.9 - 2.2$ ppm) and concave-down REE patterns ($\text{La}/\text{Sm} = 0.34-0.81$, $\text{Sm}/\text{Yb} > 1.6$, Fig. 2a). In contrast, clinopyroxene from group-1 gabbros have lower LREE concentrations ($[\text{La}] = 0.01-1$ ppm), depleted LREE/MREE ($\text{La}/\text{Sm} = 0.09-0.56$), and flat to slightly depleted MREE-HREE patterns (e.g., $\text{Sm}/\text{Yb} < 1$; Fig. 2a). Group-1 gabbros are also more depleted in other incompatible trace elements (e.g., Sr, Zr and Hf).

Radiogenic and stable isotope data are presented in Table 1. Both group-2 and 3 gabbros have enriched Sr-Nd isotopic compositions ($^{87}\text{Sr}/^{86}\text{Sr} = 0.70356-0.70377$, $\epsilon_{\text{Nd}} = 4.0-8.3$) similar to Hawaiian lavas (Fig. 3). Group-1 gabbros have depleted isotopic compositions ($^{87}\text{Sr}/^{86}\text{Sr}_{\text{plag}} = 0.70255-0.70287$, $\epsilon_{\text{Nd, cpx}} = 6.6-10.9$), which resemble East Pacific Rise (EPR)

MORB. Several group-1 gabbros have higher $^{87}\text{Sr}/^{86}\text{Sr}$ in clinopyroxene than in plagioclase. In contrast, clinopyroxene and plagioclase from group-2 and 3 gabbros are in isotopic equilibrium.

Clinopyroxene and plagioclase $\delta^{18}\text{O}$ values correlate well with each other with $\Delta_{\text{plag-cpx}} = +0.7 \pm 0.2\text{‰}$ (Fig. 5 inset). Most group-1 gabbros have $\delta^{18}\text{O}$ values in a narrow range ($\delta^{18}\text{O}_{\text{plag}}$ values = +5.7 to +6.2‰, $\delta^{18}\text{O}_{\text{cpx}}$ values = +4.9 to +5.3‰), except one sample with a $\delta^{18}\text{O}_{\text{plag}}$ value of +5.3‰. Oxygen isotope compositions of group-2 gabbros show more variation ($\delta^{18}\text{O}_{\text{plag}}$ values = +5.1 to +6.3‰, $\delta^{18}\text{O}_{\text{cpx}}$ values = +4.3 to +5.5‰). Group-3 gabbros span a narrow range of oxygen isotope compositions ($\delta^{18}\text{O}_{\text{plag}}$ values = +6.1 to +6.2‰, $\delta^{18}\text{O}_{\text{cpx}}$ values = +5.2 to +5.6‰).

5. Discussion

5.1 Origins of Hualalai gabbros

Group-1 gabbros have Sr-Nd isotope compositions that starkly differ from Hualalai and other Hawaiian lavas but overlap with EPR MORB (Fig. 3). Assuming no trapped melt, the REE concentrations of melts in equilibrium with clinopyroxene from group-1 gabbros were calculated using measured clinopyroxene REE concentrations and estimated cpx-melt partition coefficients (see supplement for details). The calculated melts in equilibrium with group-1 gabbros have MORB-like flat to LREE-depleted patterns (Fig. 2b). These characteristics suggest group-1 gabbros derive from layer-3 gabbros of the Pacific oceanic crust beneath the Hawaiian Islands as previously proposed (e.g., Clague, 1987).

The presence of orthopyroxene in group-2 gabbros suggests they derive from tholeiitic melts (Bohrson & Clague, 1988; Fodor & Galar, 1997). Group-2 clinopyroxene REE compositions are in equilibrium with LREE-enriched melts similar to Hualalai lavas (Fig. 2b). The Sr-Nd isotope compositions of group-2 gabbros plot primarily within the field of

Hawaiian lavas (Fig. 3). Group-2 gabbros therefore are most likely related to Hualalai tholeiitic shield stage magmas.

Clinopyroxene from group-3 gabbros are also in equilibrium with LREE-enriched melts resembling Hualalai lavas (Fig. 2b). The Sr-Nd isotope compositions of group-3 gabbros cluster closely within the field of Hualalai alkalic post-shield stage lavas (Fig. 3). In addition, the absence of orthopyroxene in group-3 gabbros suggests they derive from alkalic basalts (Fodor & Galar, 1997). Together, these features suggest that group-3 gabbros likely derive from Hualalai alkalic post-shield stage magmas. In the following discussion, we assign group-1, 2 and 3 gabbros as derived from *in situ* Pacific LOC, Hualalai tholeiitic shield stage magmas and Hualalai alkalic post-shield stage magmas, respectively.

5.2 Hydrothermal circulation in LOC

Near-axis hydrothermal circulation removes a significant portion of the heat from crystallization and cooling of the plutonic LOC (e.g., Phipps Morgan & Chen, 1993; MacLennan, 2005). Interaction with hydrothermal fluids can also substantially modify the chemical and isotopic composition of altered oceanic crust (e.g., Gregory & Taylor, 1981; Alt & Teagle, 2000). The strength and distribution of near-axis hydrothermal circulation remains a matter of debate. Different studies have proposed pervasive hydrothermal circulation through the entire LOC (e.g., Bosch et al., 2004), focused hydrothermal circulation within fracture zones in the LOC (e.g., Coogan et al., 2006), and hydrothermal circulation that only penetrates the upper oceanic crust (e.g., Manning et al., 1996). The strength and distribution of hydrothermal circulation in LOC can be evaluated by examining geochemical signatures of hydrothermal alteration in LOC rocks.

Oxygen and strontium isotopes are sensitive tracers of hydrothermal alteration. Unaltered oceanic crust spans a narrow range of O and Sr isotope compositions [$\delta^{18}\text{O}_{\text{MORB}} \approx +5.4$ to $+5.8\text{‰}$ (Eiler, 2001); $^{87}\text{Sr}/^{86}\text{Sr}_{\text{NMORB}} \approx 0.7022\text{--}0.7032$ (Sun & McDonough, 1989)].

Low temperature hydrothermal alteration results in higher $\delta^{18}\text{O}$ values than NMORB, whereas high temperature alteration results in lower $\delta^{18}\text{O}$ values. Alteration also increases the $^{87}\text{Sr}/^{86}\text{Sr}$ of oceanic crust because seawater has $^{87}\text{Sr}/^{86}\text{Sr}$ significantly higher than fresh MORB [$^{87}\text{Sr}/^{86}\text{Sr}_{\text{seawater}} \approx 0.709$ (e.g., Burke et al., 1982)]. Previous studies of ophiolites and fault-uplifted seafloor oceanic crust sections have documented high $\delta^{18}\text{O}$ values (whole rock $\delta^{18}\text{O}_{\text{wr}} \approx +5$ to $+12.5\text{‰}$) and $^{87}\text{Sr}/^{86}\text{Sr}$ (up to 0.7085) in pillow basalt and upper sheeted dikes, suggesting low-T hydrothermal alteration in the upper oceanic crust (e.g., Gregory & Taylor, 1981; Alt & Teagle, 2000). In contrast, lower sheeted dike and gabbro layers have low average $\delta^{18}\text{O}$ values ($\delta^{18}\text{O}_{\text{wr}} \approx +1.5$ - $+11.5\text{‰}$, $\delta^{18}\text{O}_{\text{plag}} \approx +2.5$ - $+12.5\text{‰}$; e.g., Gregory & Taylor, 1981) and are interpreted to have undergone high-T hydrothermal alteration.

In contrast to typical observations from ophiolitic gabbros, most of the Hualalai LOC gabbros span a narrow range of plagioclase oxygen isotope compositions (Fig. 4a, $\delta^{18}\text{O}_{\text{plag}}$ values = $+5.7$ to $+6.2\text{‰}$). Clinopyroxene from these gabbros have $\delta^{18}\text{O}_{\text{cpx}}$ values ranging from $+4.9\text{‰}$ to $+5.3\text{‰}$. Clinopyroxene $\delta^{18}\text{O}$ values correlate well with plagioclase $\delta^{18}\text{O}$ values ($\Delta_{\text{plag-cpx}} = +0.7 \pm 0.2\text{‰}$, Fig. 5 inset). Plagioclase-clinopyroxene fractionation is $\sim 0.5\text{‰}$ for rapidly quenched gabbroic magmas, and $\Delta_{\text{plag-cpx}}$ increases during cooling of the system (Gregory & Taylor, 1981). The observed $\Delta_{\text{plag-cpx}}$ is consistent with equilibrium fractionation at temperatures $\sim 1150 \pm 60^\circ\text{C}$ (Kyser et al., 1981). If these values are corrected back to $\Delta_{\text{plag-cpx}} = 0.5\text{‰}$, the Hualalai LOC gabbros would have $\delta^{18}\text{O}_{\text{plag}}$ values $\approx +5.6$ to $+6.2\text{‰}$ and $\delta^{18}\text{O}_{\text{cpx}}$ values $\approx +5.1$ to $+5.4\text{‰}$ (assuming $\sim 50:50$ plagioclase/clinopyroxene). Considering analytical uncertainties, these values are within the range of primary magmatic $\delta^{18}\text{O}_{\text{plag}}$ values of $+5.7$ to $+6.2\text{‰}$ and $\delta^{18}\text{O}_{\text{cpx}}$ values of $+5.2$ to $+5.7\text{‰}$ in equilibrium with NMORB (Gregory & Taylor, 1981; Eiler, 2001). Previous studies have observed large variations in $\delta^{18}\text{O}$ values (up to 3‰) over short (mm) length-scales adjacent to micro hydrothermal veins in ophiolitic LOC gabbros (Coogan et al., 2007). Analyses of plagioclase and

clinopyroxene separates could potentially overlook such small-scale alteration. However, hydrothermal micro veins were not observed in Hualalai LOC gabbros. Whole rock $\delta^{18}\text{O}$ values in the Hualalai LOC gabbros (+5.9 to +6.2‰; Lassiter & Hauri, 1998) are also primarily consistent with NMORB (+5.3 to +5.9‰; Eiler, 2001). The oxygen isotope compositions of both the gabbroic whole rocks and mineral separates suggest minimal hydrothermal alteration.

Strontium isotopes in Hualalai LOC gabbros also record minimal interaction with seawater. Both plagioclase and whole rock Sr isotopes in group-1 gabbros [$^{87}\text{Sr}/^{86}\text{Sr}_{\text{plag}} = 0.70254\text{--}0.70284$; $^{87}\text{Sr}/^{86}\text{Sr}_{\text{wr}} = 0.70256\text{--}0.70279$ (Lassiter & Hauri, 1998)] are within the range of fresh NMORB (Fig. 4b). Although clinopyroxene in a few LOC gabbros have higher $^{87}\text{Sr}/^{86}\text{Sr}$ than in plagioclase, there is no correlation between $\delta^{18}\text{O}$ values and $^{87}\text{Sr}/^{86}\text{Sr}$ (Fig. 5), as is typically observed in hydrothermally altered basalts and gabbros (e.g., Hart et al., 1999). Instead, the higher clinopyroxene $^{87}\text{Sr}/^{86}\text{Sr}$ more likely reflect minor interaction between Hawaiian magmas and *in situ* LOC (See section 5.4.2). The combined Sr and O isotope data from the Hualalai LOC gabbros suggest much less hydrothermal alteration than typically observed in ophiolitic or fault-uplifted seafloor gabbros.

The absence of an alteration signature in the Hualalai LOC gabbroic xenoliths does not preclude alteration in other, unsampled portions of the LOC. Some LOC gabbros from ophiolitic sections also appear to have unaltered O and Sr isotope compositions. However, the majority of LOC gabbros in many ophiolites show significant shifts in $\delta^{18}\text{O}$ values and/or $^{87}\text{Sr}/^{86}\text{Sr}$ from fresh oceanic crust values. Gabbros from the Oman ophiolite do not show a significant decrease in alteration signatures with increasing depth (Fig. 4). Although focused fluid flow near fractures has been proposed for portions of the Oman ophiolite, alteration is still observed in most gabbros far from fracture zones (Coogan et al., 2006). These features suggest pervasive hydrothermal alteration from diffuse fluid flow throughout the entire

LOC. The Hualalai LOC gabbros span a range of compositions (e.g., cpx Mg# = 76–86) and 2-pyroxene equilibration temperatures (954–1065 °C, see Table S5.), which suggests they likely represent a random sampling of the *in situ* LOC at various depths. If the *in situ* LOC beneath the Hawaii Islands displayed the same distribution of alteration signatures as is observed in ophiolites, it would be unlikely for alteration signatures to be missing from the LOC gabbroic xenoliths.

The hydrothermal alteration in ophiolitic and fault-uplifted seafloor gabbros may reflect alteration associated with their uplift and exposure, which we refer to as secondary hydrothermal alteration. Isotope tracers (O, Sr) in ophiolitic gabbros exhibit larger shifts from fresh MORB values than in fault-exposed seafloor gabbros (Alt & Teagle, 2000), which likely reflect more intensive secondary hydrothermal alteration associated with the thrusting and emplacement of ophiolites. Recent studies of *in situ* LOC samples (from ODP 1256D) show that the intensity of hydrothermal alteration decreases sharply with increasing depth within the uppermost ~100 m of the gabbro layer (Gao et al., 2012; Höfig et al., 2014). The minimal hydrothermal circulation recorded in the Hualalai LOC gabbros may therefore be more representative of *in situ* LOC than tectonically exhumed gabbros.

5.3 Implications for LOC accretion

Hydrothermal alteration recorded in the Hualalai LOC gabbros may place constraints on the mechanism of LOC accretion. Two endmember models have been proposed. The “gabbro glacier” model proposes that magmas are fed from the underlying mantle to a shallow, long-lived melt lens at the dike-gabbro boundary. Crystallization of gabbroic cumulates occurs in this melt lens, and LOC is formed by downward advection of these cumulates (e.g., Phipps Morgan & Chen, 1993; Quick and Denlinger, 1993). In contrast, the “multiple sills” model proposes multiple magma sills injected into LOC at various depths and times, with LOC produced by crystallization in these sills followed by dominantly

horizontal flow of the solid gabbro away from the ridge axis (e.g. Kelemen et al, 1997). Hybrid models combine these two models, with the top part of the LOC formed in a shallow melt lens and the rest formed in lower magma sills (e.g., MacLennan, 2005).

Heat released from magma crystallization at mid-ocean ridges is primarily removed by near-axis hydrothermal circulation (e.g., Phipps Morgan & Chen, 1993). Different LOC accretion models make different predictions regarding the strength and distribution of this circulation. The “gabbro glacier” model predicts near-axis hydrothermal circulation primarily focused above the shallow melt lens (e.g., Phipps Morgan & Chen, 1993; MacLennan et al., 2004, 2005). In contrast, the “multiple sills” or hybrid models predict deep and intensive near-axis hydrothermal cooling throughout the LOC. Otherwise, the latent heat from sills injected at various depths would trigger remelting of the LOC instead of crystallization (e.g., MacLennan et al., 2004; Thelssen-Krah et al., 2011). Both models permit but do not require focused off-axis hydrothermal circulation penetrating the LOC along fracture zones after melt lenses have solidified. Although the relationship between hydrothermal alteration and LOC accretion is complex, in general, greater deep hydrothermal alteration is expected for the “multiple sills” model (e.g., MacLennan, 2005).

The range of compositions and equilibrium temperatures spanned by Hualalai LOC gabbros (Section 5.2) suggests they do not derive from a single locality. Assuming they represent a random sampling of the *in situ* LOC, we can calculate the probability that hydrothermally altered portions of the lower crust would not be sampled if present. For example, given 16 random samples of LOC material (the number of Hualalai LOC gabbroic xenoliths), if 5% by volume of the *in situ* LOC is altered, the probability that this material would not be sampled is ~44% [$P=(1-x)^N$]. If 20% of the LOC is altered, the probability of not sampling altered regions decreases to ~3%. Therefore, the absence of alteration

signatures in the LOC xenoliths suggests volumetrically minor hydrothermal alteration of the LOC beneath Hualalai, which is difficult to reconcile with the “multiple sills” model.

LOC gabbroic xenoliths have been reported or suggested for several other localities, including Jasper Seamount (Gee et al., 1991), the Canary Islands (e.g., Schmincke et al., 1998), Davidson Seamount (Davis et al., 2007), and Flores Island, Azores (Franca et al., 2008). However, Sr and O isotope compositions are only reported for LOC xenoliths from several Canary Islands (Vance et al., 1989; Hoernle, 1998; Hansteen & Troll, 2003). The majority of these xenoliths have $^{87}\text{Sr}/^{86}\text{Sr}$ shifted to higher values than Atlantic NMORB at a given $^{143}\text{Nd}/^{144}\text{Nd}$ (c.f. Fig. 5 in Hoernle, 1998). Sr-isotope enrichments are most pronounced in Gran Canaria xenoliths. Furthermore, four Gran Canaria xenoliths have lower $\delta^{18}\text{O}_{\text{wtr}}$ values ($\delta^{18}\text{O}_{\text{wtr}} = +3.4$ to $+5.1\text{‰}$; Hansteen & Troll, 2003) than fresh NMORB. These features suggest pervasive high-T hydrothermal alteration of *in situ* LOC beneath the Canary Islands, in contrast to what is observed at Hualalai. The pervasive hydrothermal alteration observed in the Canary Island LOC xenoliths is compatible with the “multiple sills” model.

The oceanic crust beneath the Hawaiian Islands formed at the EPR, a fast spreading ridge, whereas the crust beneath the Canary Islands formed at the Mid-Atlantic Ridge (MAR), a slow spreading ridge. Oceanic crust at slow ridges is generally more fractured than at fast ridges (e.g., Phipps Morgan et al., 1987), which can provide conduits for fluid infiltration and hydrothermal alteration (e.g., Alt et al., 2010). More data from additional localities is needed to determine if LOC from fast ridges is consistently less altered than LOC from slow ridges. However, cooling rates in LOC gabbros from fast ridges decrease significantly with depth in the upper kilometer of the gabbro layer, as estimated from the Ca-in-olivine geospeedometer. In contrast, LOC gabbros at slow ridges display no variation in cooling rates with depth, and cool substantially faster than at fast ridges. The cooling rate profiles

are consistent with more extensive deep hydrothermal circulation and associated cooling at slow ridges (Coogan et al., 2007).

Shallow melt lenses have been consistently detected by seismic imaging beneath fast ridges (e.g., Hooft et al., 1997). In contrast, crustal melt lenses are only occasionally detected beneath slow spreading ridges in the Atlantic Ocean (e.g., Sinha et al., 2006). These observations are consistent with petrologic models that suggest shallower magma ponding at fast ridges. For example, Michael & Cornell (1998) showed that melts at fast ridges fractionate at shallower depths than at slow ridges. Magma fractionation also appears to occur within a narrower range of depths at fast ridges (Michael & Cornell, 1998). MORB glasses from fast ridges also have higher average Cl/K ratios. These features indicate assimilation of shallow, hydrothermally altered materials with elevated Cl/K during shallow magma ponding and fractionation at fast ridges (Michael & Cornell, 1998).

Correlations between MORB isotopic heterogeneity and spreading rate also suggest different magma plumbing systems at fast and slow ridges. Sr isotope heterogeneity at individual ridge segments decreases with decreasing average Mg# (Rubin et al., 2009). In addition, both average Mg# and Sr isotopic heterogeneity correlate with ridge spreading rate, with fast ridges possessing lower average Mg# and less isotopic heterogeneity than slow ridges. Similar correlations are also observed for Nd isotopes. We compiled MORB Nd isotope data from 10 mid-ocean ridge segments as well as samples from ODP Sites 843 and 1256D (see supplement for details). As shown in Fig. S3, ϵ_{Nd} variability is negatively correlated with ridge spreading rate. The lower average Mg# and isotopic heterogeneity at fast ridges is consistent with greater mixing and homogenization of discrete magma batches coupled with protracted fractional crystallization in long-lived magma chambers. Combined with the observation that magmas at fast ridges fractionate on average at shallower depths (Michael & Cornell, 1998), these features are consistent with the presence of a shallow,

long-lived magma chamber, through which most magmas pass prior to eruption. In contrast, the greater isotopic heterogeneity at slow ridges may reflect greater preservation of primary magma heterogeneity due to direct tapping of discrete melt batches from multiple sills intruded at various depths.

Taken collectively, the geochemical and geophysical observations at slow ridges are consistent with the predictions of the “multiple sills” model. Fractional crystallization in a long-lived shallow magma chamber is significant at fast ridges. In principle, hybrid models where magmas pond and fractionate in multiple sills over a range of depths and then feed into a shallow melt lens prior to eruption are consistent with the petrologic observations. However, magma crystallization requires sufficient removal of latent heat, often through hydrothermal cooling (MacLennan et al., 2004, 2005). Therefore, the limited hydrothermal alteration observed in the LOC beneath the Hawaiian Islands, although not conclusive, favors the “gabbro glacier” model. Future studies of LOC xenoliths from additional localities may place further constraints on the distribution of hydrothermal alteration in LOC, and may provide a new avenue to study the processes of LOC accretion.

5.4 Magma storage system of Hualalai Volcano and melt-crust interaction

5.4.1 Evidence for LOC shield-stage magma reservoir of Hualalai Volcano

The depths of magma storage may affect the evolution of Hawaiian magmas in several ways. Pressure affects the liquid line of descent and volatile degassing processes. In addition, the effects of assimilation or melt/wallrock interaction will vary depending on whether melts pond within the volcanic edifice, oceanic crust, or upper mantle. The magma plumbing systems of Hawaiian volcanoes evolve as magma supply rates change (e.g., Clague 1987; Bohrson, 2007 and references therein). The high magma supply during the shield stage provides the thermal impetus to maintain a shallow-level magma reservoir within the volcanic edifice. During the post-shield stage, decreasing magma supply leads to

solidification of this reservoir. The presence of LOC gabbroic xenoliths in Hualalai post-shield stage lavas and the crystallization sequence of Mauna Kea post-shield lavas suggest deep magma reservoirs exist within or beneath the LOC during the post-shield stage, possibly close to the Moho (Clague, 1987; Frey et al., 1990). Several studies propose a deep magma reservoir also coexists with the shallow reservoir during the shield stage (e.g., Clague, 1987; Bohrson & Clague, 1988; Vazquez et al., 2007). Other studies suggest there is only one reservoir during the shield stage, and this reservoir deepens during the shield to post-shield transition (e.g., Frey et al., 1990; Shamberger & Hammer, 2006).

The depths of origin of shield-stage gabbros can be used to infer the depth of the Hualalai shield stage magma reservoir(s), because the gabbros are most likely cumulates from these crystallized reservoirs. We evaluate the crystallization depths of Hualalai shield gabbros using (1) MELTS thermodynamic modeling (Ghiorso & Sack, 1995); (2) comparison of equilibration temperature in shield gabbros and LOC gabbros; and (3) comparison with previous estimates for depth of origin of Hualalai gabbroic and ultramafic xenoliths. Each of these approaches provides an independent, if imprecise, assessment of magma fractionation pressure. Taken together, they provide a consistent picture of the Hualalai volcano shield-stage plumbing system.

Isobaric fractional crystallization of a primitive tholeiite melt was modelled using MELTS over a pressure range of 2-8 kbar. Crystallization depths of Hualalai shield gabbros were then constrained by comparing observed mineralogy and composition with model-predicted crystallization sequences and mineral compositions. We used the primitive Hualalai tholeiite KK-14-5 (Bohrson & Clague, 1988) and an initial water content of 0.4 to 0.7 wt.% as the initial melt composition for the MELTS modeling (Table S8). MELTS models were initiated at the NNO or QFM oxygen buffer and allowed to proceed unbuffered.

Over the range of initial water content and oxygen fugacity investigated, all models were able to produce the major mineral phases present in shield gabbros, and show two consistent results: (1) as pressure increases, the temperature at which plagioclase arrives at the liquidus decreases, and as a result of earlier pyroxene fractionation the An content of the first plagioclase crystallized also decreases; (2) the onset of orthopyroxene fractionation occurs at higher temperature with increasing pressure, resulting in higher Mg# of the first-crystallized orthopyroxene (Fig. 6; more discussion of the MELTS modeling is presented in supplement). The mineral compositions observed in shield gabbros (plagioclase An up to 70, orthopyroxene Mg# up to 87) require fractionation within the pressure range of ~2.5-5 kbar, consistent with pressure constraints (1-5 kbar) of Shamberger & Hammer (2006) based on clinopyroxene geobarometry. This range corresponds to a depth of ~11-18 km, which is within the local oceanic crust (Moore, 1987; Clague, 1987). This depth interval overlaps with but is slightly shallower than crystallization depths previously estimated for Hualalai shield ultramafic xenoliths by Bohrsen & Clague (1988) (~13-28 km).

Comparison of equilibration temperatures between Hualalai shield-stage and LOC gabbros also provides indirect evidence that these two xenolith types derive from similar depths. Hualalai shield gabbros record 2-pyroxene equilibration temperatures [942-1075°C; precision ~±25°C; (Brey & Kohler, 1990)] similar to temperatures recorded in the LOC gabbros (954-1065°C). MELTS modeling suggests crystallization of group-2 gabbros at temperatures higher than 1075°C. Lamellar exsolution textures are observed in pyroxene from some group-2 gabbros. Therefore, the 2-pyroxene temperatures recorded by group-2 gabbros likely reflect subsolidus cooling. In contrast, the Pacific LOC beneath the Hawaiian Islands (Müller et al., 2008) should have cooled to ~400 °C due to conductive cooling over the past ~100 Ma (Pollack et al., 1993), and 2-pyx thermometers in these gabbros should record a pyroxene blocking temperature of ~650 °C (Mitra et al., 1999). However, LOC

group-1 gabbros record much higher equilibrium temperatures , which likely reflects recent re-heating of the LOC by Hawaiian magmas.

Given that the equilibration temperatures recorded by the group-2 gabbros likely reflect post-crystallization sub-solidus cooling, and the temperatures in the group-1 gabbros reflect recent reheating of the LOC, it would seem fortuitous for these xenoliths to record such similar temperatures unless they last equilibrated at similar depths. Bohrsen & Clague (1988) suggested slightly higher sub-solidus temperatures (~1045-1090°C) for Hualalai shield-stage-related ultramafic xenoliths compared to shield gabbros. Combined with the slightly higher pressure estimates for these ultramafic xenoliths (Bohrson & Clague, 1988), this suggests that Hualalai shield ultramafic xenoliths derive from somewhat deeper depths than shield gabbros. In conclusion, although precise pressure estimates for the shield-related gabbroic and ultramafic xenoliths are difficult, it appears most likely that Hualalai shield stage magmas ponded and fractionated within and/or at the base of the LOC. A deep magma reservoir probably existed during shield stage Hualalai volcanism in addition to the shallow intra-edifice reservoir.

5.4.2 Effects of melt-crust interaction on *in situ* LOC

The existence of a magma reservoir within the LOC during shield and post-shield stages of Hawaiian volcanoes provides the opportunity for magmas to interact with the surrounding LOC during magma storage. Several features of group-1 gabbros may record this interaction. These features include Sr-isotope disequilibrium between clinopyroxene and plagioclase in several samples (Fig. 7), and correlations between Sr and Nd isotopes and trace element ratios that trend towards Hualalai magma cumulate compositions (Fig. 8).

Whereas clinopyroxene and plagioclase in group-2 and 3 gabbros have indistinguishable Sr isotopic compositions, clinopyroxene in many group-1 gabbros are out of equilibrium with coexisting plagioclase and extend to more radiogenic compositions (Fig. 7a). The

463 extent of Sr isotope disequilibrium ($\Delta^{87}\text{Sr}/^{86}\text{Sr}_{\text{cpx-plag}}$) correlates with clinopyroxene Sr
464 concentrations and other incompatible trace elements. Clinopyroxene with low
465 incompatible element concentrations (e.g., < 4 ppm Sr) are generally in equilibrium with
466 coexisting plagioclase, whereas incompatible-element-enriched clinopyroxene are not (Fig.
467 7b). Sr isotopic disequilibrium between clinopyroxene and plagioclase has also been
468 observed in gabbros from the Oman ophiolite, which was interpreted to reflect partial
469 exchange with ^{87}Sr -rich seawater-derived hydrothermal fluids (McCulloch et al., 1981).
470 However, the anomalous ^{87}Sr enrichment in clinopyroxene is unlikely to reflect
471 hydrothermal alteration, as there is no oxygen isotope evidence for hydrothermal alteration
472 of the group-1 gabbros (see section 5.2).

473 Both $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ of group-1 gabbros correlate with insoluble trace elements
474 and ratios such as La/Sm, and trend towards Hualalai magma cumulate compositions (Fig.
475 8). Although clinopyroxene $^{143}\text{Nd}/^{144}\text{Nd}$ in group-1 gabbros is also correlated with
476 $^{147}\text{Sm}/^{144}\text{Nd}$, the slope of the correlation suggests an “age” (170 ± 50 Ma) that is significantly
477 older than local oceanic crust (~ 100 Ma; Müller et al., 2008). Therefore, rather than
478 reflecting radiogenic ingrowth, these correlations more likely reflect mixing between LOC
479 cumulates and incompatible-element-enriched Hualalai melts.

480 Interaction between LOC gabbros and Hualalai melts may occur through melt infiltration
481 and reaction along grain boundaries. As an approximation, this interaction can be modeled
482 as simple binary mixing between gabbro clinopyroxene or plagioclase and Hualalai melts.
483 Such mixing affects Sr isotopes in clinopyroxene to a greater extent than coexisting
484 plagioclase, because clinopyroxene has lower Sr concentrations (Fig. 9). The difference in
485 Nd concentrations is smaller between these two mineral phases, so that melt addition has
486 roughly similar effects on plagioclase and clinopyroxene. Mixing models show that the
487 range of both Sr and Nd isotopic variation in LOC gabbros can be generated by addition of

less than 5-10% of either shield- or post-shield-stage melts. This amount of melt addition is unlikely to have significantly affected the major element or oxygen isotope compositions of the LOC xenoliths.

5.4.3 Effects of melt-crust interaction on Hawaiian lavas: the origin of oxygen isotope heterogeneity in Hawaiian lavas

The chemical and isotopic variations in Hawaiian basalts are commonly utilized to infer the origin and length scales of heterogeneities in the Hawaiian plume (e.g., Abouchami et al., 2005). Many studies make the implicit assumption that the compositions of Hawaiian basalts closely track their mantle sources with little modification during transportation and storage in the oceanic crust. However, Hawaiian shield and post-shield stage magmas most likely pond and fractionate in the LOC (see Section 5.4.1). During this ponding, Hawaiian magmas may assimilate *in situ* LOC. Therefore, the role of crustal contamination in the evolution of Hawaiian magmas needs to be assessed.

Hawaiian lavas display greater oxygen isotope variability than fresh NMORB (Eiler et al., 1996). Several studies have proposed that “Loa trend” lavas inherit their high $\delta^{18}\text{O}$ values from ancient recycled oceanic crust or sediment in the Hawaiian plume (e.g., Eiler et al., 1996, Lassiter & Hauri, 1998). However, the origin of the low $\delta^{18}\text{O}$ values of “Kea trend” lavas is still uncertain. Eiler et al. (1996) proposed the low $\delta^{18}\text{O}$ values are generated by assimilation of hydrothermally altered *in situ* LOC gabbros. In contrast, Lassiter & Hauri (1998) argued that both the high and low $\delta^{18}\text{O}$ values in Hawaiian lavas reflect the presence of ancient recycled oceanic lithosphere in the Hawaiian plume. More recently, Wang et al. (2003) and Wang and Eiler (2008) proposed that the low $\delta^{18}\text{O}$ values in Mauna Kea lavas reflect assimilation of volcanic edifice material that has been altered by meteoric water.

Oxygen isotope data from Hualalai gabbroic xenoliths suggest that assimilation of local LOC cannot generate the range of oxygen isotope compositions observed in Hawaiian lavas.

LOC-derived group-1 gabbros have a narrow range of oxygen isotope compositions similar to fresh NMORB (See section 5.2). In contrast, olivine phenocrysts in Hawaiian lavas span a wide range of oxygen isotope compositions ($\delta^{18}\text{O}_{\text{ol}} = +4.3$ to $+6.1\text{‰}$; Eiler et al., 1996). For $\Delta^{18}\text{O}_{\text{melt-ol}} = \sim 0.5\text{‰}$ (Eiler, 2001), Hawaiian magmas therefore have $\delta^{18}\text{O}$ values extending to both higher and lower values than the *in situ* LOC. Assimilation of the *in situ* LOC cannot generate either the anomalously low or high $\delta^{18}\text{O}$ values observed in many Hawaiian melts (Fig. 5).

Most group-2 and group-3 gabbros have oxygen isotope compositions at or beyond the upper limit of the range defined by fresh MORB, consistent with the high $\delta^{18}\text{O}$ values reported for “Loa trend” lavas including the Hualalai post-shield lava analyzed by Eiler et al. (1996). These Hualalai-related gabbros likely crystallized within or at the base of the LOC (See section 5.4.1). Because assimilation of volcanic edifice material cannot have affected cumulates formed within the LOC, the oxygen isotope compositions of these Hualalai cumulates are most likely inherited from the Hawaiian plume source(s). The high $\delta^{18}\text{O}$ values in several shield and post-shield stage gabbros are therefore consistent with previous proposals that high $\delta^{18}\text{O}$ values in Hawaiian lavas reflect recycled high- $\delta^{18}\text{O}$ material in the Hawaiian plume (e.g. Eiler et al., 1996; Lassiter & Hauri, 1998). In addition, although only one Hualalai cumulate with anomalously low $\delta^{18}\text{O}$ is reported here, the presence of low $\delta^{18}\text{O}$ cumulates, if confirmed, would suggest that low $\delta^{18}\text{O}$ values in Hawaiian melts also derive from the Hawaiian plume.

6 Conclusions

Lower oceanic crust (LOC) gabbroic xenoliths from ocean islands and seamounts provide an under-utilized approach to study the compositions and hydrothermal alteration history of LOC. This information can provide further constraints on LOC accretion models, and the

potential influences of melt-crust interaction on the compositions of ocean island basalts. Gabbroic xenoliths from Hualalai Volcano are divided into three groups based on their geochemical and petrologic compositions. Group-1 and 2 gabbros contain clinopyroxene and orthopyroxene, and are consistent with derivation from tholeiitic melts. Group-3 gabbros lack orthopyroxene, and likely derive from alkalic melts. Group-2 and 3 gabbros both have enriched Sr-Nd isotopic compositions similar to Hawaiian lavas, and clinopyroxene REE concentrations in equilibrium with Hualalai-like melts. In contrast, group-1 gabbros have depleted Sr-Nd isotopic compositions that overlap with EPR MORB, and clinopyroxene REE concentrations in equilibrium with MORB-like LREE-depleted melts. The geochemical and petrologic characteristics of group-1, 2 and 3 gabbros suggest they derive from *in situ* Pacific LOC, Hualalai tholeiite shield stage and Hualalai post-shield stage magmas, respectively.

In contrast with observations from many ophiolitic and fault-uplifted seafloor LOC gabbros, the O and Sr isotope compositions of group-1 gabbros indicate minimal hydrothermal alteration of *in situ* LOC beneath Hualalai. The greater hydrothermal alteration in ophiolitic and seafloor gabbros may result from “secondary” hydrothermal alteration during faulting and uplift. The minimal hydrothermal alteration of Hualalai LOC xenoliths also contrasts with pervasive alteration recorded in LOC xenoliths from the Canary Islands. These differences may reflect less hydrothermal alteration of LOC formed at fast ridges than at slow ridges. Combined with other lines of geochemical and geophysical evidence (e.g., greater isotopic heterogeneity, the lack of seismic evidence for shallow melt lenses), the high degree of hydrothermal alteration observed in LOC samples from slow ridges suggests the “multiple sills” model of LOC accretion. In contrast, seismic imaging of shallow melt lenses and reduced isotopic heterogeneity of erupted basalts at fast ridges suggest that magmas in these settings pass through and fractionate within a long-lived

shallow melt lens. This melt lens likely restricts deep penetration of on-axis hydrothermal circulation and cooling of LOC at ridge axes, consistent with the limited hydrothermal alteration of LOC formed at fast ridges.

Group-1 and group-2 gabbros record similar subsolidus 2-pyroxene equilibration temperatures (942 - 1075°C), suggesting derivation from similar depths. MELTS modeling further suggests the crystallization pressures of group-2 gabbros to be ~2.5-5 kbar, roughly consistent with local LOC depth. Therefore, a magma reservoir close to the Moho existed during Hualalai shield stage. Melt-crust interaction between Hawaiian melts and *in situ* Pacific crust during magma storage partially overprinted clinopyroxene Sr and Nd isotopic compositions of group-1 gabbros. Although minor assimilation of Pacific crust by Hawaiian melts cannot be excluded, the range of oxygen isotope compositions recorded in Hawaiian lavas and cumulates cannot be generated by assimilation of the *in situ* LOC gabbros, which primarily have MORB-like $\delta^{18}\text{O}$ values. Therefore, the isotopic variations observed in Hawaiian melts reflect plume source heterogeneity.

Acknowledgements

This work was supported in part by NSF grant EAR-1321937 and by the Jackson School of Geosciences, University of Texas at Austin. We thank T.E. Larson, E.M. Marshall, D.O. Breecker, B.L. Byerly, and R. Chatterjee for laboratory assistance. Some of the xenoliths studied here were collected when DAC and WAB were employed by the U.S Geological Survey, which supported their fieldwork and microprobe mineral analyses. DAC is supported through a grant from the David and Lucile Packard Foundation to the Monterey Bay Aquarium Research Institute. Constructive comments by two anonymous reviewers and the editor Tamsin Mather are appreciated and helped improve this manuscript.

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Figure Captions

Figure 1. Simplified map of Hualalai Volcano. Star marks location where gabbroic xenoliths were collected (modified from Clague, 1987). Inset shows the location of Hualalai Volcano on the Big Island of Hawaii.

Figure 2. (a) Measured clinopyroxene REE concentrations of the Hualalai gabbroic xenoliths normalized to primitive mantle (McDonough & Sun, 1995). (b) Calculated melt REE concentrations in equilibrium with clinopyroxene from the Hualalai gabbroic xenoliths assuming no trapped melt in the xenoliths. REE clinopyroxene-melt K_d values used were calculated using BIGD (see supplement for more details). Fields of Hualalai lavas and EPR basalts are shown for comparison (data from GeoRoc and PetDB).

Figure 3. Clinopyroxene ϵ_{Nd} vs. plagioclase $^{87}Sr/^{86}Sr$. Plagioclase $^{87}Sr/^{86}Sr$ are plotted because Sr-isotopes in some clinopyroxene have been partially reset by melt-crust interaction (see 5.3.2 for more discussion). Fields of Hawaiian lavas and EPR MORB are shown for comparison (data are from GeoRoc and PetDB). Also shown are whole rock Sr-Nd isotope data for basalts from PDP 843 to the west of Hawaii (King et al., 1991), and previously reported whole rock Sr-Nd isotope data from the Hualalai gabbros (e.g., Lassiter & Hauri, 1998).

Figure 4. Oxygen and Strontium isotope compositions of LOC-derived Hualalai group-1 gabbros compared to the Oman ophiolite. Data of the Oman ophiolite is compiled from Gregory & Taylor (1981), McCulloch et al. (1980), Stakes & Taylor (1992), Bosch et al. (2004), Kawahata et al. (2001), Lanphere et al. (1981) and Coogan et al. (2006). Shaded fields of Sr and O isotope compositions of unaltered NMORB are plotted for reference.

Figure 5. $^{87}\text{Sr}/^{86}\text{Sr}$ - $\delta^{18}\text{O}$ in Hualalai gabbroic xenoliths and Hawaiian lavas. For LOC gabbros, plagioclase $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are plotted as they are more resistant to melt-rock interaction (see 5.3.2 for more discussion). Plagioclase $^{87}\text{Sr}/^{86}\text{Sr}$ and clinopyroxene $^{87}\text{Sr}/^{86}\text{Sr}$ are indistinguishable for Hualalai-related group-2 and -3 gabbros. Therefore, we plot clinopyroxene $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for group-2 and -3 samples for which plagioclase $^{87}\text{Sr}/^{86}\text{Sr}$ was not measured. Because most Hawaiian lava $\delta^{18}\text{O}$ data are measured in olivine phenocrysts (data from Eiler et al., 1996), we plotted plagioclase $\delta^{18}\text{O}$ values in equilibrium with measured olivine given the fractionation factor $\Delta_{\text{plag-ol}} = 0.7\text{‰}$ (Eiler, 2001). Inset shows the correlation between plagioclase $\delta^{18}\text{O}$ and clinopyroxene $\delta^{18}\text{O}$ of Hualalai gabbroic xenoliths. Fractionation lines of $\Delta_{\text{plag-cpx}} = 0.7 \pm 0.2\text{‰}$ are plotted for reference.

Figure 6. MELTS modeling results. Lines in the figure shows An of the first plagioclase (open circles) and Mg# of the first orthopyroxene (filled circles) predicted by MELTS modeling at different pressures. Initial composition can be found in Table S8. The effects of various water contents in the initial melt and different initial oxygen fugacity were assessed and presented. Model-predicted crystallization pressure range of Hualalai shield gabbros is constrained by the highest plagioclase An and orthopyroxene Mg# observed.

Figure 7. (a) Comparison of clinopyroxene and plagioclase $^{87}\text{Sr}/^{86}\text{Sr}$. Several LOC xenoliths have higher clinopyroxene $^{87}\text{Sr}/^{86}\text{Sr}$ than for plagioclase from the same sample. Shield- and post-shield related gabbros plot close to the 1:1 line, indicating isotopic equilibrium between plagioclase and clinopyroxene in these samples. (b) Correlation between $\Delta^{87}\text{Sr}/^{86}\text{Sr}_{\text{cpx-plag}}$ in the LOC gabbros and clinopyroxene Sr concentrations. The observed Sr isotope disequilibrium between clinopyroxene and plagioclase most likely reflects melt-crust interaction between Hawaiian melts and *in situ* LOC.

Figure 8. Clinopyroxene $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ versus La/Sm ratios of Hualalai gabbroic xenoliths.

Figure 9. Simple binary mixing model as an approximate for the melt-crust interaction processes between Hawaiian magmas and *in situ* LOC. The filled star represents the average Sr-Nd isotopic composition of Hualalai post-shield (PS) stage magmas, whereas the open star marks the average isotopic composition of Hualalai shield stage magmas. Magma and mineral phases Sr/Nd concentrations used for modeling are $[\text{Sr}]_{\text{magma, shield}} = 250 \text{ ppm}$, $[\text{Sr}]_{\text{magma, PS}} = 450 \text{ ppm}$, $[\text{Sr}]_{\text{cpx}} = 4 \text{ ppm}$, $[\text{Sr}]_{\text{plag}} = 90 \text{ ppm}$; $[\text{Nd}]_{\text{magma, shield}} = 20 \text{ ppm}$, $[\text{Nd}]_{\text{magma, PS}} = 50 \text{ ppm}$, $[\text{Nd}]_{\text{cpx}} = 1 \text{ ppm}$, $[\text{Nd}]_{\text{plag}} = 0.2 \text{ ppm}$. Magma, clinopyroxene and plagioclase Sr and Nd concentrations are chosen based on literature data compiled from GeoRoc, measurement concentrations in this study, and unpublished data from Nicholas Dygert, respectively.

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830 Table 1. Sr- Nd-, and oxygen-isotope compositions of Hualalai gabbroic xenoliths. For $\delta^{18}\text{O}$
831 values, averages of duplicate measurements are presented in this table. Duplicate data are
832 reported in Supplemental Table S6.

<i>Sample</i>	⁸⁷ Sr/ ⁸⁶ Sr	Sm (ppm)	Nd (ppm)	¹⁴³ Nd/ ¹⁴⁴ Nd	ε _{Nd}	δ ¹⁸ O(‰)	⁸⁷ Sr/ ⁸⁶ Sr	¹⁴³ Nd/ ¹⁴⁴ Nd	ε _{Nd}	δ ¹⁸ O(‰)
<i>Clinopyroxene</i>							<i>Plagioclase</i>			
Group-1 MORB										
63Kap-3	0.70265			0.513104	9.1	5.1	0.70262	0.513018	7.4	5.7
65Kap-14d	0.70256			0.513171	10.4	5.3	0.70255			
63Kap-13	0.70268			0.513179	10.6	5.1	0.70266			5.9
65Kap-13	0.70261			0.513071	8.4	5.3	0.70261			6.0
65Kap-14a	0.70256			0.513118	9.4	5.3	0.70255			5.7
65-61-26	0.70295	1.64	4.94	0.512968	6.4	4.9	0.70282	0.512951	6.1	5.8
65-61-26 dup				0.512966	6.4		0.70284			
65-60-32	0.70308	2.07	4.49	0.513134	9.7	5.8	0.70272	0.513138	9.8	6.2
65-7-133	0.70293	1.16	2.59	0.513109	9.2	5.1	0.70257	0.513065	8.3	5.8
65-7-133 dup				0.513102	9.1		0.70255			
65-86-64	0.70348	1.60	4.60	0.512998	7.0	5.1	0.70271	0.512998	7.0	5.7
65-86-64 dup				0.512989	6.8		0.70270			
83Kap-8	0.70295	0.63	1.21	0.513143	9.9		0.70270			5.3
83Kap-9	0.70267	0.50	0.87	0.513192	10.8		0.70266			5.7
87Kap-10	0.70294	0.45	1.57	0.512977	6.6		0.70280			5.7
94Kap-7	0.70260	0.10	0.19	0.513109	9.2		0.70254			5.7
96Kap-7	0.70275			0.513168	10.3		0.70265			5.8
88Kap-2	0.70289			0.513021	7.5					5.8
87Kap-14	0.70265			0.513195	10.9					5.8
Group-2 Shield										
63Kap-15	0.70360			0.512933	5.8		0.70358	0.512974	6.6	5.2
65-115-10	0.70376			0.512867	4.5		0.70377	0.51291	5.3	5.7
65-60-17	0.70359			0.513064	8.3	5.4	0.70359			6.3
84-1801-5	0.70376			0.512873	4.6					
87Kap-11	0.70364			0.512845	4	4.9				
92Kap-1	0.70331			0.512897	5.1					6.1
92Kap-3	0.70359			0.512910	5.3	5.3				
Group-3 Post-shield										
63Kap-7	0.70364			0.512907	5.2	5.6	0.70363	0.512911	5.3	6.1
65-86-92	0.70359			0.512937	5.8	5.4	0.70357			6.0
65-86-92 dup				0.512933	5.8		0.70358			
65-109-146abc	0.70358			0.512917	5.4	5.2	0.70356	0.512917	5.4	6.1
65-109-146abc dup				0.512924	5.6		0.70357			
65-60-221	0.70368			0.512900	5.1		0.70364			
65-100-110	0.70357			0.512927	5.6		0.70356			