Petrogenesis and structure of oceanic crust in the Lau back-arc basin

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Abstract

Oceanic crust formed along spreading centers in the Lau back-arc basin exhibits a dramatic change in structure and composition with proximity to the nearby Tofua Arc. Results from recent seismic studies in the basin indicate that crust formed near the Tofua Arc is abnormally thick (8-9 km) and compositionally stratified, with a thick low-velocity (3.4-4.5 km/s) upper crust and an abnormally high-velocity (7.2-7.4+ km/s) lower crust (Arai and Dunn, 2014). Lava samples from this area show arc-like compositional enrichments and tend to be more vesicular and differentiated than typical mid-ocean ridge basalts, with an average MgO of ~3.8 wt. %. We propose that slab-derived water entrained in the near-arc ridge system not only

enhances mantle melting, as commonly proposed to explain high crustal production in back-arc environments, but also affects magmatic differentiation and crustal accretion processes. We present a petrologic model of Lau back-arc crustal formation that successfully predicts the unusual crustal stratification imaged in the near-arc regions of the Lau basin, as well as the highly fractionated basaltic andesites and andesites that erupt there. Results from phase equilibria modeling using MELTS indicate that the high water contents found in near-arc parental melts can lead to crystallization of an unusually mafic, high velocity cumulate layer. Best-fit model runs contain initial water contents of ~0.5-1.0 wt. % H₂O in the parental melts, and successfully reproduce geochemical trends of the erupted lavas while crystallizing a cumulate assemblage with calculated seismic velocities consistent with those observed in the near-arc lower crust. Modeled residual melts are also lower density than their dry equivalents, which aids in melt segregation from the cumulate layer. Low-density, water-rich residual melts can lead to the eruption of vesicular lavas that are unusually evolved for an oceanic spreading center.

1.0 Introduction

Back-arc basins are productive centers of crustal formation that generate extensive regions of oceanic crust through rifting and organized spreading. Because their mantle melting regions are influenced by mantle flow, thermal structure, and chemistry above a subducting lithospheric slab, back-arc spreading centers host a diversity of crustal structure, composition, and processes not typically found at mid-ocean ridges. Numerous petrological and geophysical studies along back-arc spreading centers have shown changes in melt production and lava chemistry that correlate with proximity to the active arc (e.g., Stolper and Newman, 1994;

Martinez and Taylor, 2002; Langmuir et al., 2006). Current models and observations suggest that back-arc spreading centers located close to an active arc entrain water (and other related components) released from the dehydrating, subducting slab, which enhances mantle melting beneath the spreading centers (e.g., Kelley et al., 2006). This leads to a higher melt flux to the ridge and erupted lavas that are more arc-like in composition than typical mid-ocean ridge basalts (MORB).

While increases in melt production and crustal thickness in back-arc settings are often ascribed to high water contents, recent evidence in the Lau back-arc basin points to associated differences in crustal structure as well (Peirce et al., 2001). The L-SCAN active source seismic tomography experiment along the Eastern Lau Spreading Center identified large regions of an anomalously thick crust with an unusual vertical stratification formed at back-arc spreading centers located close to the active Tofua arc (Dunn and Martinez, 2011; Arai and Dunn, 2014). This near-arc crust has a thick upper layer of unusually low seismic velocities and a lower layer of very high seismic velocities (7.2-7.4+ km/s) just above the base of the crust. The lavas that form the volcanic layer of this crust are both more vesicular (e.g., Pearce et al., 1994) and more differentiated than is typical of mid-ocean ridge basalts (e.g., Pearce et al., 2005; Escrig et al., 2009), with an average of ~3.8 wt. % MgO, and carry trace element and isotope signatures derived from the nearby subducting slab (e.g., Escrig et al., 2009; Bézous et al., 2009). In contrast, spreading centers located farther from the arc form crust with a more normal geophysical structure for their spreading rate, and primarily erupt basalts. The transition between these two types of crust occurs over just a few kilometers, and closely correlates with major changes in bathymetric depth and morphology, Bouguer gravity, and the depth to the top of the sub-ridge magmatic system (e.g., Dunn and Martinez, 2011; Dunn et al., 2013).

The process that generates this unusual crustal stratification is currently unknown. Similar high velocity lower crusts have been reported in other back-arc regions, including the Yamato Basin in the Japan Sea (Sato et al., 2014; Hirahara et al., 2015), the Izu-Bonin-Mariana Arc (Kodaira et al., 2007; Takahashi et al., 2008, 2009), and the Aleutian arc (e.g., Shillington et al., 2004; Behn and Kelemen, 2006), as well as rifted continental margins (Korenaga et al., 2000). Sato et al. (2014) propose that the anomalous back-arc crust in the Japan Sea is the result of higher mantle potential temperature (e.g., Kelemen and Holbrook, 1995; Korenaga et al., 2002). However, this is likely not the underlying mechanism in the Lau basin, where mantle potential temperature is thought to be hottest in the northern areas of the basin, decreasing to the south where the anomalous crust is actually observed (Kelley et al., 2006; Wiens et al., 2006; Wei et al., 2015).

To explain the stratified, anomalous crust, Arai and Dunn (2014) instead proposed that entrainment of slab-derived water in mantle melts alters crystallizing phase proportions by suppressing plagioclase relative to olivine and clinopyroxene, leading to crystallization of a dense layer of mafic-to-ultramafic cumulates at the bottom of the crust and a thick, more silica-rich upper crust (Fig. 1). While this process was not modeled in their paper, water has long been understood to play a major role in the generation, evolution and eruption of magma in subduction zone settings. Slab-derived water lowers the mantle solidus (e.g., Kushiro et al., 1968), driving melt production in the overriding mantle wedge (e.g., Davies and Bickle, 1991; Stolper and Newman, 1994). Water also suppresses mineral liquidus temperatures (e.g., Nicholls and Ringwood, 1973), leading to changes in magma crystallization sequences (e.g., Grove and Baker, 1984; Sisson and Grove, 1993), and is associated with production of silica-rich lavas (e.g., Gaetani et al., 1994). While much attention has been given to water's effects on mantle melting

and eruption processes, some of its effects on crustal differentiation and formation are difficult to constrain due to the complicated magmatic processes typical of arc systems (e.g., crustal assimilation and contamination). Back-arc spreading centers are simpler settings to study the effects of water enrichment on the crustal magmatic system, as they entrain water-rich material from the subducted slab while lacking some of the additional complications found at arcs; crustal assimilation, for example, is likely less than at arcs with their thick sequences of heterogeneous crust.

In this study, we apply thermodynamic models of phase equilibria to the extensive geochemical datasets available along Lau back-arc spreading centers to test the crustal formation hypothesis of Arai and Dunn (2014). We compare predicted magma evolution trends and estimations of crustal seismic properties with available geochemical and geophysical observations to constrain likely formation conditions for this unusual oceanic crust.

1.1 Setting

Opening due to rollback of the Tonga trench (e.g., Hawkins, 1994), the Lau back-arc basin is a wedge-shaped extensional basin located west of the currently active Tofua arc (Fig. 2). Unlike some back-arc extensional settings where broad crustal rifting and subsequent partial melting and modification of pre-existing arc crust has been proposed (e.g., Takahashi et al., 2008), most new crustal formation in the eastern Lau back-arc occurs primarily at well-organized spreading centers. Seismic imaging along these spreading centers indicates a single narrow axial magmatic system (e.g., Jacobs et al., 2007; Dunn et al., 2013), and sonar and magnetic data indicate a narrow and persistent axis of crustal formation for >2 Myr (e.g., Sleeper, 2011; Austin, 2012; Sleeper and Martinez, 2014). The current spreading configuration in the central Lau basin

initiated in the north and propagated southward along the Central Lau Spreading Center (CLSC), Eastern Lau Spreading Center (ELSC), and Valu Fa Ridge (VFR), resulting in the basin's current wedge shape (Fig. 2) (e.g., Hawkins, 1995; Taylor et al., 1996).

The ELSC and VFR, which are the focus of this study, consist of smaller ridge segments divided by second- and third-order discontinuities (ELSC1-4 and VFR1-2; Fig. 3a). These ridge segments range from ~100 to 40 km from the active arc, and show variations in crustal thickness and lava compositions consistent with higher contributions from subduction-related melts toward the south (e.g., Langmuir et al., 2006; Martinez et al., 2006; Pearce and Stern, 2006; Escrig et al., 2009). Despite decreasing spreading rates toward the south (from ~96 to ~40 mm/yr), the shallow, axial high morphology and thick crust in the south indicate higher magma supply close to the active volcanic arc (Figs. 2, 3) (e.g., Martinez and Taylor, 2002; Martinez et al., 2006). The CLSC and northern ELSC segments (ELSC1-2) are currently located far enough from the arc that the arc's influence is greatly reduced, and their magma supply and resulting crustal thickness are comparable to MOR of similar intermediate spreading rates. ELSC3 is characterized by a transition in ridge morphology and axial depth, where the deep (~2700 m) ridge axis in the north shoals abruptly to ~1800-2200 m in the south (Fig. 3b), roughly coincident with the appearance of the seismically-imaged axial magma chamber (AMC) reflector (Harding et al., 2000; Jacobs et al., 2007).

The variation in melt supply and crustal accretion observed along the spreading centers is also recorded in older, off-axis crust. The anomalous near-arc crust and more typical oceanic crust form a broad crustal zoning in the basin, with a remarkably sharp transition between the two (Fig. 2). These regions have therefore been identified as two distinct crustal zones, called Domain II (arc-proximal) and Domain III (arc-distal) respectively (Martinez and Taylor, 2002;

Dunn and Martinez, 2011). Domain II crust forms at spreading centers <50 km from the active Tofua arc (Dunn and Martinez, 2011; Arai and Dunn, 2014) and is currently being produced along the VFR. ELSC1 and ELSC2 are currently forming Domain III crust, with the southernmost ELSC (segments ELSC3 and ELSC4) producing a narrow zone of crust that is transitional between Domains II and III (Dunn and Martinez, 2011).

At 8-9 km thick, the near-arc Domain II crust is considerably thicker than typical oceanic crust formed at MOR, which averages ~6-7 km globally (e.g., White et al., 1992). Models of mantle melting indicates the thick crust in this region is best explained by high melt productivity due to the presence of water in the mantle wedge (Kelley et al., 2006; Harmon and Blackman, 2010). Domain II crust also exhibits anomalously high seismic velocities in the lower crust, averaging 7.2-7.4 km/s with some local regions exceeding 7.5 km/s (Arai and Dunn, 2014). These values are higher than that expected for a gabbroic lower crust, which typically has wave speeds of ~6.8-7.2 km/s (e.g., Dunn and Forsyth, 2007). Meanwhile, the low velocity layer that forms the upper crust (seismic layer 2) is abnormally thick (2.9-3.8 km) with unusually low velocities (3.4-4.5 km/s) that extend deeper than expected for a typical volcanic (porous) layer alone (Jacobs et al., 2007; Arai and Dunn, 2014). These observations can be explained by the presence of more evolved rock compositions and higher porosities in the upper crust than typical for crust produced at mid-ocean ridges (Jacobs et al., 2007; Arai and Dunn, 2014).

Lava samples from the present Domain II ridge axis (VFR) and the transitional zone are more evolved than typical MORB (Fig. 3c), with abundant basaltic andesites, andesites and a few dacites sampled on the VFR close to the arc (e.g., Pearce et al., 1994; Langmuir et al., 2006). Domain II lavas also show strong enrichments in subduction-related components relative to normal MORB (or the more MORB-like Domain III lavas erupted further to the north), including

enrichments in fluid mobile elements and related ratios (e.g., Ba/Th, Fig. 3d) as well as enrichments associated with a sedimentary contribution (e.g., Th/La, Fig. 3e, and La/Sm, $^{206}\text{Pb}/^{204}\text{Pb}$, not shown) (Escrig et al., 2009). Samples tend to be highly vesicular (Pearce et al., 1994), and available analyses of water contents, and/or proxy estimates, from the VFR and near-axis seamounts have values as high as 1-5 wt. % H_2O (e.g., Michael et al., 2011).

Oceanic crust formed farther north along the ELSC and away from the arc (>70 km; Domain III) is similar in structure and composition to global averages for its spreading rate. Its thickness and seismic velocities are consistent with intermediate-rate oceanic crust (Dunn and Martinez, 2011; Dunn et al., 2013). Lavas are primarily back-arc basin basalts (BABB) that have undergone low-to-moderate amounts of crystallization (~6-8+ wt.% MgO), with low enrichments in components typically ascribed to derivation from the slab (Fig. 3), consistent with the interpretation that the ridge has moved too far away from the subducting slab to entrain very much subduction-related material into the melting region (e.g., Hawkins, 1995; Pearce, 1994; Peate et al., 2001; Kent et al., 2002).

2.0 Data and Methods

The along-axis variations of the ELSC and VFR and the window they provide into arc and back-arc melting processes has made them the subject of numerous sampling expeditions and geochemical studies, resulting in an extensive compositional dataset from which to examine crustal formation processes. The major element compositional data used in this study come from the compilation of Gale et al. (2013), which released new sample analyses in addition to providing a cleaned compilation of all the published data available in PetDB (www.earthchem.org/petdb; Lehnert et al., 2000) for the ELSC and VFR (complete references in

Supplementary Material). We primarily use major element compositions for the thermodynamic modeling, but rely on additional trace element compositional data to evaluate consistency and check for indicators of other processes (e.g., crustal assimilation, contamination, or magma mixing) that might affect the samples' geochemical compositions.

We have chosen a representative subset from these available data for forward modeling. To limit the effects of other process variations, we have excluded off-axis samples, which may be significantly older or have erupted off-axis. We also exclude samples from overlapping spreading centers (OSCs) and segment ends, which show clear excursions from the general along-axis trends in the Lau data (Fig. 3) (e.g., Escrig et al., 2009). Because they show clear deviations from ridge segment centers, we model these samples separately (see the Discussion section for further analysis of these samples, with full results available in the Supplementary Material). The exclusion of OSC samples also removes the outlier E-MORB sample DR41-1, identified by Escrig et al. (2009) as having isotopic and trace element enrichments similar to Samoan lavas. For the purposes of modeling crystallization trends, we also exclude the few samples with MgO < 3.0 wt. %, where data are sparse and forward models are known to be less reliable.

Forward modeling of fractional crystallization was done using alphaMELTS (Antoshechkina and Asimow, 2010), the latest release of the MELTS algorithms for thermodynamic modeling of phase equilibria in magmatic systems (Ghiorso and Sack, 1995; Asimow and Ghiorso, 1998; Smith and Asimow, 2005). Major element composition modeling include SiO₂, TiO₂, Al₂O₃, FeO*, MgO, CaO, Na₂O, K₂O, P₂O₅, and H₂O. Model runs are isobaric and fractional, with an oxygen fugacity of FMQ-2, which commonly shows good matches to mid-ocean ridge basalt data. We also tested more oxidizing conditions, as might be

expected in more hydrous environments, but found these model runs consistently produced worse fits to the observed trends. We simulate pressures (P) ranging from 0.05 GPa-0.22 GPa to approximate conditions from the shallow crustal melt lens down to the Domain II lower crust.

We tested multiple starting compositions (C_0) (see Supplementary Material), fixing relative abundances of all oxides except initial water composition (H_2O_i), which is allowed to vary. Model results end up being relatively insensitive to reasonable variations in starting composition, so we present results for a single starting composition (MOD3) for the remainder of this paper. We tested variable water contents ranging from 0 to 3.0 wt.% H_2O_i , renormalizing oxide concentrations accordingly so they still total 100%.

Model outputs from alphaMELTS (referred to simply as MELTS below) crystallization runs include predictions for the abundance and composition of liquid and solid phases as a function of temperature, and we compare modeled liquid compositional trends against analyzed glass compositions at the scale of individual ridge segments. We also use the calculated mineral compositions and their proportions to estimate the mode and composition of the earliest crystallizing phases. Assuming the lower crust is composed primarily of the early products of fractional crystallization, we use modeled solid assemblages to calculate estimated seismic wave speeds of the lower crust for comparison with results from seismic tomography (Hacker et al., 2003; Hacker and Abers, 2004). We calculate the proportion of each mineral phase (olivine, clinopyroxene, plagioclase) predicted to have fractionated by MELTS for a given solid fraction. Since the physical properties of a mineral are also dependent on its composition, we use the predicted mineral compositions of these fractionated phases to cast each mineral into proportions of appropriate end-members for which physical property data are available: forsterite and fayalite (olivine); anorthite and albite (plagioclase); diopside, hedenbergite and jadeite (clinopyroxene).

We then calculate wave speeds (Hacker and Abers, 2004) for the bulk crystal assemblages at conditions appropriate to off-axis lower crust (0.22 GPa, 800°C). Although differences in lower crustal pressure accompany differences in crustal thickness across the study area, the predicted difference in seismic velocities due to this effect are < 0.1 km/s and smaller than the resolution of the seismic tomography, and are therefore neglected from further consideration.

3.0 Results

Predicted liquid compositions for example model runs are shown in Figure 4 with the modeled dataset shown for comparison (see Supplemental Material Fig. S1 for results from additional oxides). Data from ELSC1 and ELSC2 (red and orange symbols in Fig. 4) lie along a visibly distinct trend from ridge segments to the south (ELSC3-VFR) in Al_2O_3 , FeO , TiO_2 , Na_2O , and to a lesser degree CaO and K_2O vs. MgO . For Al_2O_3 , FeO^* , TiO_2 , and Na_2O , this offset is consistent with a delay in plagioclase crystallization. Plagioclase crystallization produces a clear kink in these oxide trends, so its appearance in the crystallizing assemblage is easy to detect in lava compositional trends. MELTS runs with higher water contents show a delay in the onset of plagioclase crystallization (Figs. 4, 5), preferentially crystallizing olivine and clinopyroxene initially (Fig. 5), leading to higher Al_2O_3 contents and lower FeO^* , TiO_2 and Na_2O contents for a given MgO (e.g., Sinton and Fryer, 1987). A few oxides exhibit relatively poor fits to the data for all model runs: SiO_2 , CaO , and K_2O show steeper slopes in the data than predicted by MELTS for all P and H_2O_i . Model fits are generally worse at low MgO contents, as systematic differences between the model and the data accumulate with each temperature step. We also note that inflections in FeO^* , TiO_2 , and SiO_2 vs. MgO trends at ~ 4.0 wt. % MgO indicate the onset of

crystallization of an Fe-Ti oxide phase, which does not appear in MELTS runs until below 3.0 wt. % MgO due to the low fO_2 used in the modeling.

Major oxide trends from the northern ridge segments ELSC1 and ELSC2, currently producing Domain III type crust at the ridge axis, are best matched by MELTS runs with low water contents (<0.5 wt. %) (Fig. 4). Ridge segments in both Domain II (VFR) and the transitional region (ELSC3 and ELSC4) are best matched by MELTS runs with higher water contents (~0.5-1.0 wt. %) at low to moderate pressures (Fig. 4). All data are better matched for model runs with low to moderate pressures (~1.0-1.5 kb), but overall the data fits are not very sensitive to pressure. The best-fit run varies slightly depending on the oxide-MgO pairing, but the most water-sensitive oxide trends (Al_2O_3 , FeO^* , TiO_2 , P_2O_5) are consistently fit best by a relatively narrow range of model conditions.

In order to compare model results across these multiple components, we calculate an overall misfit for each model run and each oxide-MgO pairing, and then combine these misfits into a total misfit for all oxides of interest (calculation details given in Supplementary Materials). For each model run, we calculate a mean scaled absolute deviation (MSAD) for each oxide-MgO pairing, normalized by the standard error of the data from an idealized fractionation trend for that oxide-MgO pair – see Supplementary Materials for complete details. The resulting mean scaled absolute deviation (MSAD) for a given model run can be thought of as the number of standard errors from the sample trend for any given oxide. This allows us to directly compare model fits across different oxides with different average concentrations and different amounts of scatter, and combine them into a total average MSAD for all oxides for a given set of model conditions. Independent of the starting composition, SiO_2 and K_2O consistently show poor model fits for all

conditions and contribute little to the understanding of this system. We therefore omit them from the total average MSAD.

Final calculated total misfits for individual ridge segments are shown in Figure 5. Major oxide trends from the northernmost ridge segment ELSC1, currently producing Domain III type crust at the ridge axis, is best matched by MELTS runs with low water contents (~ 0.2 wt. % H_2O_i) and low pressures (up to ~ 1 kb) (Fig. 5a). ELSC2, which has recently produced a narrow strip of crust classified as Domain III (Fig. 1), shows more variability and a wider range of model fits, with good fits at ~ 0.4 wt. % H_2O_i at low-to-moderate P, ranging to slightly higher H_2O_i at higher P (Fig. 5b). Ridge segments in both Domain II (VFR) and the transitional region (ELSC3 and ELSC4) require MELTS runs with higher water contents at low-to-moderate pressures to match their compositional trends. ELSC3-4 major oxide trends are best matched by MELTS runs at ~ 0.4 - 0.7 wt. % H_2O_i (Fig. 5c), and VFR is best matched by ~ 0.5 - 1.0 wt. % H_2O_i (Fig. 5d). All data are better matched for model runs with low to moderate pressures (~ 0.10 - 0.15 GPa).

3.1 Effect of water on seismic properties

Water affects the MELTS-predicted bulk fractionated solids (Fig. 6), which can lead to detectable differences in geophysical properties. To illustrate the effects of water on crustal seismic velocities, we calculate seismic velocities (Fig. 7) for the MELTS-predicted bulk fractionated solids (e.g., Fig. 6) at various crystal extents and initial water contents. In all runs, early crystallizing phases are very mafic with relatively high seismic velocities, which decrease with continuing crystallization. At a given crystal fraction, higher H_2O_i contents produce compositions with higher estimated V_P values, with an increase of ~ 0.2 - 0.3 km/s for the first ~ 1

wt. % H_2O_i . Further addition of water leads to smaller increases in V_P (an additional ~ 0.1 km/s for the next ~ 1 wt. % H_2O_i). Lower-crustal velocities determined by Arai and Dunn (2014), which are averages of the lowermost 2 km of crust just above the Moho taken every ~ 8 km along their seismic profiles, show mean differences in V_P of a similar magnitude, from ~ 7.0 - 7.2 km/s in Domain III to 7.2 - 7.4 + km/s in Domain II.

Despite uncertainties in both the thermodynamic modeling and the determination of V_P , there is excellent agreement between the estimated and observed seismic velocities at water contents that best match lava evolution trends. Best-fit model runs for Domain III and Domain II lava compositions, crystallized to ~ 20 - 25% to approximate the extent of the lower crust over which the seismic averages were taken, predict seismic velocities ~ 7.1 km/s and ~ 7.2 - 7.4 + km/s, respectively.

4.0 Discussion

Our results from MELTS modeling suggest that even relatively low to moderate water enrichments (~ 0.5 - 1.0 wt. % H_2O) can produce important changes in differentiation processes and seismically-detectable changes in crustal structure. MELTS model runs are sensitive to the effects of water even at very low (~ 0.2 - 0.5 wt. %) water contents. A number of studies have emphasized the importance of even small amounts of water (< 0.5 wt. % H_2O) in the crystallization process of MORB (e.g., Danyushevsky 2001; Asimow et al., 2004), although more experiments are needed to better quantify these effects, resolve discrepancies, and improve calibration of thermodynamic models (e.g., Médard and Grove, 2008; Almeev et al., 2012).

The modeling results broadly agree with previous studies indicating higher water contents in the south of the Lau Basin (e.g., Kelley et al., 2006; Michael et al., 2011), though the

best-fit model runs for the VFR tend to have lower parental H₂O on average than previous estimates. Because VFR lavas are highly differentiated, their primary magma water contents are difficult to reconstruct and such efforts are complicated by degassing, diffusion or other modification in melt inclusions, and contamination by seawater; direct measurements of water data from these samples is limited and mostly come from ELSC samples. Hahm et al. (2012) find dissolved H₂O concentrations of 0.23-0.24 wt. % in lavas from the northern ELSC segments. Bézous et al. (2009) identify a “high H₂O” group within ELSC1 with water contents up to ~1.5 wt. % (at ~8.5 wt. % MgO), however we note that this group comes from the southern end of the ridge segment near the OSC — samples from the OSC limbs show clear deviations from the main ridge segments and were removed from our main dataset and modeled separately (see below). Melt inclusions from near-axis seamounts at the southern end of the VFR have measured water concentrations up to 2.4 wt. % H₂O in lavas with ~0.55-0.65 wt. % H₂O (Kamenetsky et al., 1997). There are few other published analyses of water contents on the VFR, and most reported estimates for Domain II and the transitional crust are calculated from K (Michael et al., 2011) or Ti (Kelley et al., 2006) values and assuming constant ratios to H₂O. Measured and calculated H₂O contents from ELSC and VFR glass samples range from 0.2-2.5 wt. % on the ELSC to 2-5 wt. % on the VFR (Michael et al., 2011). However, these high values may in part reflect differentiation, particularly for the evolved lavas in Domain II. In addition, the Michael et al. (2011) study focused largely on seamounts and may represent more extreme compositions than typically found on the ridge (e.g., Kamenetsky et al., 1997; Haase et al., 2009). We also note that there is an inherent difference in the scale of sampling between such petrological studies, which represent spot-analyses of glass or rocks from a few locations, and the seismic study results, which are averaged over 2 km x 8 km.

Estimates of mantle source H₂O contents in the Lau Basin based on mantle melting models also show a strong gradient between the northern and southern spreading segments, ranging from a mean source H₂O of ~0.04 for the ELSC (all segments) to ~0.22 wt. % in the VFR (Kelley et al., 2006). Harmon and Blackman (2010) also explore source water contents and mantle potential temperature in the Lau back-arc region with 2-D numerical models of regional mantle flow, thermal structure and melting, and require a source H₂O > 0.22 wt.% to match observed crustal thickness and lava water contents for the VFR, with substantially less water for the northern ELSC. While it is difficult to draw direct comparisons between these different data types and proxies, results are generally consistent in requiring higher water contents along the VFR, with even more extreme compositions measured in off-axis seamounts.

The modeling is less sensitive to pressure than composition. Where AMC reflectors are observed in seismic reflection surveys (ELSC3-4 and VFR), their depth ranges from ~2-3 km (Jacobs et al., 2007), in good agreement with the low-to-moderate best-fit fractionation pressures modeled. This does not preclude some crystallization also occurring in the lower crust – eruptions are most likely to sample material from the shallow parts of the magmatic system and lava samples are therefore biased to the final (shallowest) crystallization pressures, particularly in the presence of a steady-state melt lens. Numerous studies of MORB (e.g., Grove et al., 1992; Herzberg, 2004; Eason and Sinton, 2006) and their phenocryst-hosted melt inclusions (e.g., Wanless and Shaw, 2012; Wanless et al., 2015) suggest that crystallization occurs over a range of depths. ELSC2, which lacks an AMC reflector, is the only segment to show poor fits at low pressure (~0.05 GPa and less). We note that samples from ELSC2 have lower CaO/Al₂O₃ for their MgO than the other segments, indicating a slightly earlier onset of clinopyroxene fractionation, which can be indicative of high pressure crystallization. Most but not all samples

from the OSC also have slightly lower $\text{CaO}/\text{Al}_2\text{O}_3$ on average for their MgO , perhaps indicating higher crystallization pressures.

Implicit in the MELTS modeling is the assumption that crustal assimilation and contamination does not play an important role in generating the observed major element compositional trends. Although crustal assimilation and/or magma mixing are frequently called upon as mechanisms for generating evolved lava compositions in arc settings, we do not see evidence for assimilation or magma mixing playing a significant role in the major element oxide trends. Recent studies addressing the role of crustal contamination processes find stronger Cl enrichments relative to K_2O , H_2O and TiO_2 and other slab-derived components in basaltic samples from the ELSC than in VFR samples, consistent with crustal assimilation of Cl-rich altered oceanic crust material in Domain III (Kent et al., 2002; Hahm et al., 2012). We also note the lack of correlation between MgO and $^{87}\text{Sr}/^{86}\text{Sr}$ (Escrig et al., 2009), which one might expect if assimilation of altered oceanic crust were playing an important role in generating high magma SiO_2 contents. A similar investigation into the generation of silicic crust at back-arc spreading centers in the Manus Basin also found a lack of evidence for assimilation of altered oceanic crust, and concluded that it must play a minor role, if any, in generating the broad compositional spectrum (50-75 wt. % SiO_2) found there (Beier et al., 2015). While we cannot rule out some assimilation and contamination of altered crust or gabbro, we find no evidence that such a process contributes significantly to the composition of the eruptive products.

By excluding samples near OSCs and other ridge offsets, we have by coincidence removed samples in the otherwise fairly ‘dry’ Domain III that show enrichments of a wet slab-derived component (Bézos et al., 2009). While this allows us to better characterize the more ‘typical’ compositions and processes operating along the northernmost ridge segments, these

deviations are striking. While the MELTS modeling is not very pressure-sensitive, the group of samples from the OSC limbs of Domain III are best fit by model runs that are both higher pressure and higher water content than the rest of Domain III (Supplementary Material, Fig. S4). Some of these samples exhibit the high-Al, low-Si signature characteristic of samples near the ends of ridge segments from mid-ocean ridges worldwide (e.g., Eason and Sinton, 2006), and may reflect higher average pressures of crystallization than at ridge segment centers. Dry versions of these high-Al, low-Si basalts are strongly associated with regions of low magma supply, such as near ridge terminations, fracture zones, and at slow spreading centers, and are thought to be related to enhanced conductive cooling in these regions (e.g., Herzberg, 2004; Eason and Sinton, 2006).

While higher crystallization pressures may play a role in generating these OSC samples, their elevated H₂O (Bézos et al., 2009), Ba/Th (Fig. 3) and similar enrichments of fluid-mobile elements also require higher contributions from a subduction-related source. This does not appear to be along-axis mixing from the wetter ELSC2; seismic imaging of the crustal magmatic system at this OSC (20°10' S) shows that the low velocity regions associated with the two ridge limbs are discontinuous down to at least ~5 km depth, indicating the crustal magmatic system and the bulk of the associated thermal anomalies are distinct throughout the mid-to-upper crust (Dunn et al., 2013). Further south in Domain II, Sleeper and Martinez (2014) also noted a correlation between OSCs and geochemical tracers of subduction input, and suggested that the OSCs lie along slab trajectories from arc volcanoes and represent “hot fingers” of more hydrous melting in the mantle wedge (e.g., Tamura et al., 2002).

We suspect that the geochemical processes and signatures seen at segment ends are actually present to some extent all along the segment, but get swamped by the more robust

magma systems and greater degree of magma mixing at the segment center. The less abundant, enriched subduction-related mantle components may be more easily sampled in areas that lack a persistent, melt-dominated system. We also note that this region sits on the edge of the crustal transition and appears to have switched from generating Domain III-like crust to wetter, transitional crust and back several times in recent history (Fig. 1). Regional mantle imaging indicates this is approximately the latitude at which the low velocity anomalies in the mantle under the volcanic arc and the spreading centers transition from being connected (at ~20-35 km depth) to distinct (within the resolution of the imaging) (Zha et al., 2014).

4.1 Effects of water on physical processes

In addition to modifying the crystallizing phase proportions of an evolving magma, the presence of water also affects magma density, with consequences for melt separation and eruption. Water lowers the density of magma relative to a dry magma of an otherwise equivalent composition (e.g., Grove and Baker, 1983). This effect increases as the magma evolves due to the incompatible behavior of water, which leads to increasing concentration in the residual melts and further suppression of plagioclase crystallization relative to denser phases. Density calculations from the MELTS model runs illustrate this effect: Figure 8 shows the density of the residual melt as a function of MgO for various water contents as well as the difference between the bulk solid and liquid densities, plotted as % of bulk solid. Higher water contents lead to lower liquid densities (Fig. 8a) and significantly larger density differences between the fractionated solid phases and the residual melt — up to almost 20% for the best-fit VFR model runs (Fig. 8b). This has important implications for melt segregation, since larger density differences may lead to more efficient melt separation and buoyant rise. This may help

contribute to the unusually strong chemical stratification inferred from the seismic structure of Domain II. In combination with the generally higher magma supply found in Domain II and the transitional region, enhanced melt segregation may assist in the development of the shallow AMC seismically imaged all along the ridge axis in these regions, where melt accumulates in sufficient quantities to produce strong reflectors (Harding et al., 2000; Jacobs et al., 2007). Across-axis seismic tomography profiles (Dunn et al., 2013) show what appears to be a very narrow magmatic system beneath ELSC4 that extends to shallow depths (~2 km) with the seismically-imaged melt lens near the top, while the low-velocity zones underlying Domain III are broader and deeper (up to ~3-4 km depth). Results from mantle Rayleigh wave tomography also suggests mantle melt extraction is more efficient for the ridge segments near the arc, as the low seismic velocity anomalies are weaker in the south indicating lower melt retention (porosity) despite an increase in melt supply to the ridge (Wei et al., 2015).

Density effects may contribute to the distinct distribution of lavas found along the water-rich segments versus the dry Domain III, as well (Fig. 8c). The coincidence of low MgO lavas on segments with a crustal melt lens primarily reflects low temperature differentiation made possible by maintaining long-lived magma bodies in the shallow crust. However, Domain II lavas are even more differentiated on average than typical MORB from the Eastern Pacific Rise or other fast-spreading ridges that exhibit long-lived AMCs, indicating these magmas often remain eruptible at lower MgO. Typical dry MORB magmas reach a density minimum at the point of plagioclase saturation (~8 wt. % MgO), which is close to a typical MORB composition (e.g., Stolper and Walker, 1980). Once plagioclase starts to fractionate, the melt density increases (Fig. 1a), quickly approaching densities for typical basaltic crust and making eruption more difficult. While more evolved lavas do sometimes erupt along mid-ocean ridges, they are

much more rare, and tend to be associated with unusual settings (e.g., ridge terminations, plume-influenced ridge segments). However, lavas erupted in the near-arc regions of the ELSC and VFR are much more evolved than typical MORB, spanning from ~1 to >7 wt. % MgO with an average of ~4 wt. % (Fig. 8c). As first noted by Asimow et al. (2005), MELTS model runs with $H_2O_i > \sim 0.4$ show calculated liquid densities continue to decrease with continued crystallization, even after plagioclase saturation (Fig. 8a). While other factors such as magma supply (overpressure), volatile exsolution, and viscosity will also play significant roles in controlling eruption, their decreasing density may help keep these water-rich magmas buoyant and eruptible even to very evolved compositions.

4.2 Crustal formation at back-arcs

Arai and Dunn (2014) proposed that the high-velocity lower crust observed in seismic experiments in the Lau back-arc basin could be explained by the presence of water and its predicted effects on magmatic differentiation, and we have now shown the feasibility of that model. We believe this mechanism provides the best current explanation for the high-velocity lower crust and thick, low velocity upper crust. Additionally, changes in physical properties of the magma due to the presence of water (i.e., density and viscosity) may further contribute to generating strongly fractionated crust.

Oceanic crust with a high velocity lower crust has been imaged in the Yamato Basin in the Japan Sea, with a crustal thickness of ~10-15 km and up to 7.4 km/s wave speeds in the lowermost crust (Sato et al., 2014). There it is proposed to form via underplating of ultramafic material due to an anomalously hot asthenosphere during the opening of the basin (Sato et al., 2014; Hirahara et al., 2015), similar to high velocity lower crust adjacent to continental margins

(e.g., Kodaira et al., 1995). While high mantle potential temperature is thought to be a mechanism for generating ultramafic lower crust (e.g., Kelemen and Holbrook, 1995; Korenaga et al., 2000, 2002), this is unlikely in the Lau basin, since estimated mantle potential temperatures are lower in the south where the anomalous Domain II crust is currently forming (Kelley et al., 2006; Wiens et al., 2006; Wei et al., 2015). An analysis by Kelley et al. (2006) suggests that, while mantle potential temperature seems to be responsible for most excess melt generation (and correspondingly thick crust) beneath many back-arc basins, the Lau back-arc appears to be an exception, showing a steep linear increase in extent of melting with mantle source H_2O_0 . This is further supported by Harmon and Blackman (2010), who modeled mantle flow, thermal structure and melting at Lau back-arc spreading centers and found that differences in source water content (0.15 wt. % and >0.22 wt. % H_2O_0 for the ELSC and VFR, respectively), not potential temperature (1300°C for both), best explain the observed differences in crustal thickness between the VFR and ELSC. Furthermore, we note that Domain II lavas have relatively lower fractionation-corrected FeO contents than Domain III (Fig. 4), which is inconsistent with predictions for higher potential temperatures; melting runs using pMELTS (Ghiorso et al., 2002) indicate that the temperature increase required to produce a 2-3 km increase in crustal thickness also produces a 1-2% increase in fractionation-corrected FeO (e.g., Asimow and Langmuir, 2003) due to Fe's pressure sensitivity (e.g., Langmuir and Hanson, 1980; Klein and Langmuir, 1987). Modest increases in source water content, on the other hand, can produce the same increase in crustal thickness while predicting slightly lower FeO contents (Asimow and Langmuir, 2003), consistent with Domain II lava compositions. This also distinguishes Lau from the Yamato Basin, where thick back-arc crust is associated with significantly higher FeO contents (~8-11+ wt. % FeO vs. 5-7.8 wt. % FeO at similar MgO)

(Hirahara et al., 2015), and suggests that processes might vary significantly between back-arc basins.

The Lau back-arc is thought to be unusually water-rich compared to other back-arcs globally (e.g., Kelley et al., 2006), perhaps in part due to fast convergence rates. However, our MELTS calculations suggest that even moderate amounts of water (> 0.5 wt. %) in the parental magma may have an observable effect on the seismic structure. It is possible this process plays a role in other water-rich back-arcs (e.g., Havre Trough, Mariana Trough, Manus Basin), as well as contributes to the crustal structure of the arcs themselves (e.g., Aleutian arc, Izu-Bonin-Mariana arc) where deep high velocity lower crusts have been observed (e.g., Shillington et al., 2004; Behn and Kelemen, 2006; Kodaira et al., 2007; Takahashi et al., 2008, 2009).

5.0 Conclusions

Slab-derived water leads to increases in melt supply and crustal thickness at back-arc spreading center segments close to the active arc (< 50 km). In this study, we demonstrate that water can also produce significant changes in the seismic structure of the oceanic crust produced along the ELSC and VFR in the eastern Lau Basin. We model magma differentiation paths using thermodynamic models of phase equilibria, comparing modeled liquid compositional trends against analyzed sample compositions from individual ridge segments. Predicted crystallizing assemblages and mineral compositions are also used to calculate seismic wave speeds of the lower crust for comparison with seismic tomography results. Compositional trends from lavas erupted along the northern ELSC > 70 km from the active arc are best matched by MELTS runs with low initial water contents (~ 0.2 wt. % H_2O_i), while higher water contents (~ 0.5 - 1.0 wt. % H_2O_i) are required to match lava compositions along the VFR (< 50 km from the arc). Best-fit

model runs for the VFR also have more mafic early crystallizing phases due to the presence of water, with calculated seismic velocities of $\sim 7.2\text{--}7.4$ km/s at solid fractions that approximate the volume of the lower crust, consistent with the anomalously high seismic velocities imaged there. We present a petrological model of crustal formation along back-arc spreading centers in the Eastern Lau Basin, in which slab-derived water close to the arc suppresses plagioclase crystallization and leads to the formation of an ultramafic lower crust with high seismic velocities, while successfully predicting major element compositional trends of the erupted lavas.

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Figure 1. A schematic cartoon illustrating hypothesized crustal formation processes at arc-distal (Domain III) and arc-proximal (Domain II) regions of the ELSC. Figure modified from Arai and Dunn (2014). In this model, slab-derived water is entrained in the ridge melting zone near the active arc (right column), leading to an increase in melt supply to the ridge and producing a thicker crust. As these melts cool and evolve in the crustal magmatic system (gray fields, upper right), their moderately high magmatic water contents lead to suppression of plagioclase crystallization relative to olivine and clinopyroxene, leading to more mafic cumulate phases during the early stages of crystal fractionation. The buoyant, water-rich residual melts segregate and rise into the upper crust where they form an axial melt lens (dark gray) and continue to crystallize, eventually forming a thick, unusually evolved volcanic layer. Where the spreading center is further than ~70 km from the active arc (left column), the water content of melts is low and water-induced differentiation is minimal.

Figure 2. Shaded relief map of seafloor topography in the southern Lau Basin. ELSC (Eastern Lau Spreading Center) and VFR (Valu Fa Ridge) axes are shown in red. Crustal domains (Martinez and Taylor, 2002; Dunn and Martinez, 2011) are outlined by dashed black lines, separated by a narrow transitional zone (striped). As illustrated in Figure 1, Domain II is the region of anomalous oceanic crust formed near the active arc; Domain III is the region of crust formed further from the arc. (Inset) Regional map of the Lau basin and surrounding areas. Grey shading indicates areas where the seafloor is shallower than 2000 m below sea level. Open triangles indicate locations of active volcanoes of the Tofua arc. The dotted line denotes the axis of the Tonga trench, which is surrounded by the 7000-m contour. Black lines show the current

configuration of spreading centers in the Lau basin: VFR; ELSC; CLSC (Central Lau Spreading Center); FR (Fonualei rift). The red box indicates area of main panel.

Figure 3. (a) Multibeam bathymetry of the ELSC and VFR showing sample locations (black dots). (b) Axial depth profile from Martinez et al. (2006). Gray bar along the left edge denotes area where magma lens reflectors are observed in along-axis multichannel seismic data (Harding et al., 2000; Jacobs et al., 2007). The change in crustal type correlates with the disappearance of the axial magma lens reflector and significant increase in axial depth. (c)-(e) Along-axis compositional variation of MgO and representative trace element ratios, color-coded by ridge segment. Samples from OSC limbs or near the ends of ridge segments are shown in different symbols. Domain III crust includes ELSC1 (red symbols) and ELSC2 (orange); the VFR (blue) currently forms Domain II crust, while ELSC3 (green) and ELSC4 (teal) make up the transitional zone. (c) Along-axis variation of lava MgO content. Samples from Domain II and transitional crust exhibit lower and more variable MgO contents than Domain III samples on average. (d) Along-axis variation in Ba/Th ratio, which reflects enrichment in fluid mobile elements, showing high Ba/Th in transitional and Domain II crust, with the strongest enrichments in ELSC3. Samples from the limbs of ELSC1 that forms the OSC with ELSC2 at 20°10' S also show enrichments in Ba/Th relative to the main ridge segment. (e) Along-axis variation in Th/La ratio, which reflects sedimentary contribution to the arc magmas (e.g., Plank, 2005), showing an increasing enrichment closer to the arc. Escrig et al. (2009) argue that the decoupling between Ba/Th, which peaks in the transitional zone (Fig. 3d), and the sediment-derived component, which shows a steady, gradual increase towards the south (Fig. 3e), reflects a compositional change of the subduction input.

819

820 **Figure 4.** ELSC and VFR major element oxide compositions, plotted as a function of MgO
821 (symbols as in Figure 3). Samples from OSC limbs or near the ends of ridge segments have been
822 removed from the dataset (see text for discussion). Lines denote crystallization model runs with
823 different initial water compositions ranging from 0.0 to 1.0 wt. % H_2O_i . Model runs shown were
824 calculated using MELTS at 0.01 GPa pressure, FMQ-2, with fractionating solid phases.

825

826 **Figure 5.** MELTS model misfits as a function of pressure and H_2O_i content for (a) ELSC1, (b)
827 ELSC2, (c) ELSC3-4, and (d) VFR. Lower values (darker blues) denote better fits to data.
828 Misfit calculated using a total Mean Scaled Absolute Deviation normalized to spread for each
829 major element oxide-MgO pair (see text for description). ELSC1 is best matched by runs with
830 low parental water contents (~ 0.2 wt. %). While fairly insensitive to pressure, model runs with
831 higher water contents are required to match near-arc magma evolution trends, with runs with
832 ~ 0.5 -1.0 wt. % H_2O_i showing the best fit to VFR (Domain II) samples.

833

834 **Figure 6.** Crystallizing phase proportions of olivine (ol), clinopyroxene (cpx) and plagioclase
835 (plag) predicted by MELTS (0.02 GPa, MOD3 starting composition, FMQ-2) as a function of
836 temperature for initial water contents of (a) 0.2 wt. % and (b) 1.0 wt. %. Water delays
837 crystallization of all mineral phases to slightly lower T, but has the strongest effect on
838 plagioclase. Selected runs are at lower crustal pressures, but similar mineral relations hold at
839 shallow crustal pressures as well.

840

Figure 7. Calculated lower crustal seismic velocities vs. parental magma water content for selected extents of crystallization at 0.22 GPa and 800°C, consistent with cool, off-axis lower crust. Lines denote 0.10, 0.15, 0.20, and 0.25 solid fraction. V_P is calculated from the MELTS-predicted mineral mode using the method of Hacker et al. (2003) (see text for description). Colored bands denote range of lower crustal velocity averages (lowermost 2 km, comparable to a solid fraction of ~0.20-0.25) for Domain II (blue) and Domain III (red) from Arai and Dunn (2014).

Figure 8. (a) Calculated residual liquid densities and (b) density difference between bulk solid and liquid (as percent of bulk solid density) vs. MgO for MELTS model runs with different starting water compositions and a pressure of ~0.22 GPa. Low pressure MELTS runs (~0.01 GPa) are nearly identical. (c) Histogram plot of lava sample MgO content for Domain III (red), transitional (green) and Domain II (blue) crust. Samples from Domain II and Transitional-type crust tend to have lower MgO contents with a distribution centered on ~4 wt. % MgO, while Domain III samples are restricted to >6 wt. % MgO. The liquid density of relatively dry starting compositions (<0.4 wt. % H_2O_i) increases during crystal fractionation, quickly approaching typical densities of oceanic crust and inhibiting melt ascent and eruption. The presence of water in the starting liquid composition leads to a larger density contrast between the melt and crystallizing phases, which may aid in the segregation, extraction and eruption of more evolved lavas.

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8