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NEAR-PRIMARY MANTLE MELTS AND THEIR
IMPLICATIONS FOR THE MECHANISM OF ISLAND ARC

BASALT OXIDATION

BY

ZOE GENTES

A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE

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2015

ABSTRACT

Near-primary melt compositions (i.e., in equilibrium with $>Fo_{88}$ olivine) are rare in arc systems. Yet, such melts provide essential views of mantle-derived melts, without further modification by fractional crystallization or other crustal processes, and reveal the diversity of melt compositions that exist in the arc mantle wedge. This study presents new measurements of naturally glassy, near-primary olivine-hosted melt inclusions from one dredge of Evita seamount (SS07/2008 NLD-02) in the southern Vanuatu arc system. Two distinct basalt types were identified in hand sample upon collection, based on contrasting phenocryst assemblage (Type 1: 1% phenocrysts; Type 2: 15% phenocrysts). We selected melt inclusions from each type and determined major elements and sulfur by EMP, H_2O and CO_2 by FTIR, trace elements by LA-ICP-MS, and $Fe^{3+}/\Sigma Fe$ ratios by XANES. Melt inclusions from both lava types show equilibrium with $\geq Fo_{88}$ olivine, consistent with host olivine compositions, and thus are near-primary melt compositions. Both have high $Mg\#$ (>65), and are basalt to basaltic andesite (49-55 wt% SiO_2). Samples from Type 1 show relatively flat REE patterns, classic high Ba/Th ratios, and positive anomalies in Pb and Sr. In contrast, samples from Type 2 exhibit steeply sloped REE patterns with strong depletions in the HREE that suggest garnet in the source lithology for these magmas. Moreover, the Type 2 samples have low Ba/Th ratios and high La/Yb (29.5-43) and Sr/Y (50-58), which are classically attributed to partial melting of the basaltic slab. The slab surface temperature (SST) was calculated from H_2O/Ce data; Type 1 SST shows temperatures comparable to global arcs ($\sim 767^\circ C$), while Type 2 SST is the hottest yet constrained by this method ($\sim 1041^\circ C$). Volatile analysis reveals that both lava types have had some degassing of H_2O with CO_2 , and give minimum H_2O contents of

each magma: ~3 wt.% for Type 1, ~2.5 wt.% for Type 2. XANES analysis shows that Type 1 samples have $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios similar to global arc basalts (~0.23), while Type 2 samples have $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios that are among the highest measured in natural terrestrial glasses (~0.34), and have much higher concentrations of S. Mixing calculations suggest that Type 2 is not a simple mixture of the Type 1 basalt with an end-member slab melt. Alternate explanations include the possibility that Type 1 and Type 2 are instead the results of a mantle melt component mixing with either slab fluid or slab melt (but not both), or that they are distinct melts from different parts of the wedge that have migrated into the same volcanic system. A global correlation between H_2O and $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratio suggests an oxidized, H_2O -rich component is common to most arcs. The Type 1 magma conforms to this global trend, but Type 2 does not. Despite its highly oxidized condition and high sulfur content, Type 2 is too dry to be the end-member component that appears to be delivering oxidation to most global arc magmas.

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PREFACE

This thesis is an integrated geochemical and volatile analysis of two basalt types from the southern Vanuatu Island Arc system. An abstract pertaining to this research was presented at the American Geophysical Union (AGU) annual conference in 2014 under the title “Near-Primary Mantle Melts and their Implications for the Mechanism of Island Arc Basalt Oxidation.”

This thesis is presented in manuscript format and has been completed with the intention of submission to Earth and Planetary Science Letters. The submission process will begin after this thesis has been approved by the University of Rhode Island.

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Near-Primary Mantle Melts and Their Implications for the
Mechanism of Island Arc Basalt Oxidation

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1. INTRODUCTION

It is important to understand how deep-Earth processes influence the availability of oxygen in solid Earth systems, and vice versa. We know that different materials from the subducted slab give arc magmas distinct geochemical signatures. We also know that the oxidation conditions of arc magmas have been broadly linked to slab signatures (e.g., Kelley and Cottrell, 2009; Brounce et al., 2015). However, we still lack detailed links between specific types of slab-derived materials (e.g., slab fluids, sediment melts, and slab melts) and the redox conditions of arc magmas. We need these links for a complete picture of how subduction modifies the mantle wedge on both local and global scales, and also of the consequences of these deep processes for the construction of island arc crust.

The majority of island arc basalts are presently considered partial melts of mantle-wedge peridotite due to fluxing from water brought down by the subducting slab (Kushiro et al., 1968; Stolper & Newman, 1994; Kelley et al., 2006, 2010; Langmuir et al., 2006). However, some studies support the idea that slab dehydration is not the predominant mechanism for some island arc magma formation (e.g., Kelemen, 2003; Kay, 1978; Yogodzinski and Kelemen, 1998, 2007). Melts of the lithospheric slab (i.e. the basaltic layer) may be more prevalent than previously recognized, and may in fact be an ubiquitous component of arc magmatism (e.g., Moyen, 2009; Sajona 1993). This is particularly evident in some locations where the slab surface is heated enough by the surrounding mantle, such as locations where a torn plate edge is suspended in the mantle (e.g., Defant and Drummond, 1990; Yogodzinski et al., 2001; Plank and Cooper, 2009).

Studies also have shown that oceanic crust ages and its minerals oxidize over time as it moves from spreading centers and is recycled into subduction zones (e.g., Alt et al., 1986). Fluids or melts from the subducting slab contribute a chemical signature to the mantle source for island arc magmas. Resulting island arc lavas have a higher proportion of oxidized iron (Fe^{3+}) relative to reduced iron (Fe^{2+}), expressed as $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratio, than MORBs or OIBs (Carmichael, 1991), and appears to be a result of a more oxidized magma source in the mantle wedge (e.g. Brounce et al., 2015; Kelley and Cottrell, 2009). The composition of arc basalts can be heavily influenced by fluids and sediment or slab melts that infiltrate the mantle wedge in subduction zones (e.g. Plank and Langmuir, 1993) and slab fluids may be capable of driving mantle wedge oxidation (Ballhaus, 1993; Kelley and Cottrell, 2009; Kelley and Cottrell, 2012; Brounce et al., 2015). Magmatic $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios increase toward subduction zones and correlate linearly with H_2O content (e.g., Fig. 1; Kelley & Cottrell, 2009) and element tracers of slab-derived fluids (Brounce, et al., 2014). These observations indicate a direct link between geochemical signatures of subduction and oxidation of the mantle wedge.

Using this direct link, we may be able to see how different subduction signatures influence mantle wedge oxidation, and investigate how the mantle becomes oxidized. The different subduction signatures (slab melting, sediment melting, and slab fluids) may carry different components that may have a role in the oxidation process (e.g. H_2O , S, Fe^{3+} , etc.). To evaluate the role of different slab components in oxidation, we focus on a region where multiple signatures are evident over a small geographical area.

This thesis presents new measurements of major, volatile, and trace elements, in addition to $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios, in naturally glassy melt inclusions hosted by high-Fo olivine

from the Evita seamount in the southern Vanuatu arc system. This study allows for a closer look at the igneous geochemistry of the Evita seamount in order to (1) assess the relative contributions of aqueous fluids, sediment melts, and slab melts to magmas in the southern Vanuatu island arc system, (2) determine the primary oxygen fugacity (fO_2) of the sub-arc mantle, and (3) evaluate how oxidation and slab signatures may be linked.

2. GEOLOGIC SETTING AND SAMPLES

2.1. Geologic Setting of Vanuatu

An ideal place to address the location of and processes related to slab melting and oxidation is the Vanuatu island arc system. It is an intriguing place to study for several reasons. Vanuatu is an intra-oceanic arc, which minimizes the likelihood of magma-crust interactions. The island arc lavas are dominantly basaltic, which makes inferring mantle source easier (Peate et al., 1997). It also has a complex tectonic history that may have allowed different mantle sources and subduction components to be sampled at different locations and times within the arc. The Vitiaz lineament marks a fossil trench where the Pacific Plate once subducted beneath the Indo-Australian Plate to form the Vitiaz arc. Vanuatu was part of this arc before the North Fiji back-arc basin began spreading about 12 Ma (Auzende et al., 1995). At present, a section of the Indo-Australian plate has been subducting eastward at 70° under the Pacific Plate for 7-4Ma (Mitchell and Warden; 1971; Peate et al., 1997), with active volcanism beginning about 6 Ma.

In the southern part of the Vanuatu Arc, an adjacent section of the Indo-Australian plate continues to move horizontally at the surface to the northeast, forming a transform fault (Figure 2a) and the Hunter Fracture Zone (HFZ). This leaves a subducting slab-edge hanging in the mantle, which may set the scene for high temperature on the slab surface and slab melting in a subduction zone (Cooper et al., 2009; Kincaid et al., 2004). Just northeast of the HFZ, at the Hunter Ridge, near-primary arc tholeiites have been reported (Sigurdsson, 1993), along with a wide variety of lithologies including rhyolites, MORBs, and peridotites that reflect the complex tectonic setting of the region.

Vanuatu Arc lavas are dominated by basaltic compositions, with isotopic evidence for sediment involvement that varies systematically, decreasing from north to south (Peate 1997). The most primitive magmas have a wide range of K_2O concentrations, spanning from low-K tholeiites to high-K alkaline lavas. Large-ion lithophile element (LILE) concentrations decrease from the central part of the arc to the southern curve of Vanuatu, arc-front lavas have typical subduction-related trace element signatures (LILE enriched relative to light rare earth elements [LREE], and most samples have depletions in high field strength elements [HFSE]; Peate, 1997). Through most of the arc, melting of subducted oceanic crust is unlikely except under special circumstances, but sedimentary material might be transferred to the mantle wedge by melting; the mantle wedge may thus have been variably modified by addition of this slab-derived material containing a significant sediment component (Peate, 1997). In the southern part of this arc, subducted oceanic crust may find favorable conditions to melt and enter the mantle wedge as well, due to the torn slab edge in this region.

2.2. Preliminary Study and Sampling of Southern Vanuatu Seamounts

In 2008, R/V Southern Surveyor cruise SS07/2008 sampled rifts and spreading centers in the northwestern Lau basin. In transit, the cruise dredged samples from four Vanuatu seamounts, including Eva and Evita (Figure 2b). Seamount Evita produced two different types of basalts identified upon hand sampling, which here are termed Type 1 (NLD 02-01) and Type 2 (NLD 02-02). The Type 1 basalt had approximately 1% phenocrysts; Type 2 had approximately 15% phenocrysts.

Previous work on these samples, including whole rock solution ICPMS analysis and glass LA-ICPMS analysis, confirmed that the two basalt types had different trace element signatures (Gentes, 2012). Type 1 glass analysis showed relatively flat REE patterns, classic negative anomalies in Nb and Ta, and positive anomalies in Pb and Sr, typical of global arc basalts. In contrast, Type 2 lavas exhibit steeply sloped REE patterns, indicated by a high La/Yb ratio, that suggest garnet in the source lithology for these magmas. Moreover, the Type 2 samples have high Sr/Y (50-58), which is classically attributed to partial melting of the subducting slab (Kelemen, 2003). The different lava types suggest the occurrence of different active processes at depth beneath this volcano, both flux melting of the mantle and slab melting. This provides a way to assess the relationships between these two distinct signatures, the processes they represent, and the conditions of their source regions (e.g., temperature, redox, volatile content) at the same location. Doing so at a single volcano removes key variables (e.g., crustal thickness and structure, composition of mantle source) because both magmas presumably have similar ascent paths from source to surface.

At island arc subduction zones, magmas that come to the surface through volcanism may be subject to crustal differentiation processes. Fractional crystallization and degassing can alter the original magma composition, making it difficult to infer the characteristics of the mantle-derived magma at depth (e.g., P-T conditions of the mantle origin, major elements, original volatile content, magmatic redox; e.g. Kelley and Cottrell, 2012; Marsh and Carmichael, 1974; Métrich, 2009; Marsh, 1976; Kelemen, 2003, Kushiro, 1968; Kushiro, 1972; Stolper and Newman, 1994). We thus lose the

ability to link deep processes to these key factors that reflect the characteristics of the magma source.

The best possible way to interrogate primary arc magmas is to sample them directly. Melt inclusions trapped within early-forming phenocrysts may preserve both pre-eruptive volatiles and near-primary melts that have had minimal crystallization relative to their mafic host lavas. Thus, they provide a uniquely pristine view of the mantle. Such melts may reveal the diversity of melt compositions that exist in the arc mantle wedge.

Olivine-hosted melt inclusions can give us a rare glimpse into mantle conditions beneath island arc subduction zones. These glass inclusions, usually 10-200 μm in size, form when basaltic magmas crystallize and solidify, becoming trapped in a crystal. If trapped early enough in a magma's crystallization history, high-forsterite olivine ($>\text{Fo}_{88}$) may preserve near-primary melts, which are effectively unmodified from their origin in the mantle. Olivine crystals are the first to form on the hydrous basalt liquidus (Bowen, 1922) and thus can trap the most primitive of mantle melts, and preserve dissolved magmatic volatiles such as dissolved water, carbon dioxide, and sulfur.

3. METHODS

3.1. Sample Preparation

Four representative pillow basalt samples from Evita seamount, collected at a depth of 1323m on the 2008 cruise SS07/2008, and stored at the Marine Geological Samples Laboratory (MGSL) at the Graduate School of Oceanography at URI, were chosen for melt inclusion analysis (two rocks each from NLD 02-01 and NLD 02-02).

Hand samples were cut and made into thin sections for detailed petrographic analysis. Melt inclusions were prepared as double-polished wafers, exposing the melt inclusion on both sides, to create a clear optical path through the glass with no interference from the host olivine, which is required for some analytical methods described below. Melt inclusions were examined under a petrographic microscope in both plane-polarized light and through crossed polars to identify crystal-free regions and to ensure clean glass was available for analysis (Figure 3). Each melt inclusion was checked to be sure it had paths of clear glass, contained no more than one vapor bubble, was fully enclosed by the host olivine crystal, showed no cracks that could have led to post-entrapment degassing, and contained no visible secondary crystals.

Some samples consisted of clear glass from both a melt inclusion and external, matrix glass still adhered to the olivine crystal. Where possible, this external glass was analyzed, and labelled in figures as either “Type 1 xlg” or “Type 2 xlg,” marked with a circle instead of the diamond shape used for the melt inclusions.

3.2. Detailed Analytical Methods

EMP Analysis

Melt inclusions and host olivines were analyzed for major elements and S concentrations by electron microprobe using the JEOL-8900 5 spectrometer microprobe at the Smithsonian Institution, operating at 10nA, 15kV and with a 10 micron beam diameter (Table 1; Table 2a). Na and K were analyzed with 20 second peak count times to minimize alkali loss in hydrous glasses, followed by Si, Ti, Al, Fe, Mn, Mg, Ca, and P with 30-40 second peak count times (Luhr, 2001). The glasses were analyzed in a second round for S at 80nA and 15kV also using a 10 micron beam. The S concentrations were referenced to the VG-2 standard with 1340 ppm sulfur. Adjacent olivine was analyzed with a point beam, and primary and secondary standards were those used by Luhr (2001).

LA-ICP-MS Analysis

Melt inclusions, host olivines, and some exterior matrix glass were also analyzed for trace element abundances by laser-ablation inductively-coupled plasma mass spectrometry (LA-ICP-MS) at the Graduate School of Oceanography, University of Rhode Island. Analyses were conducted using a Thermo X-Series II quadrupole ICP-MS coupled with a New Wave UP 213 Nd-YAG laser ablation system, using spot sizes ranging from 20-60 μm and 5 Hz repeat rate to maximize ablation time in thin, wafered samples. Beam energy measured at the sample surface for a reference spot of 60 μm and 10 Hz was 0.35-0.41 mJ. Typical ablation duration in the melt inclusions, host olivines, and matrix glass lasted from 20-60 seconds, depending on sample thickness. An example LA-ICP-MS ablation spectrum is shown in Figure 4. Glasses and olivines were analyzed

for 34 trace elements (Table 4); for olivines, we report only 24 minor and trace elements because these were consistently above the detection limit (Table 2b).

Procedures for reducing LA-ICP-MS data follow those outlined by Kelley et al. (2003), using ^{43}Ca as the internal standard for glasses and ^{26}Mg as the internal standard for olivine. Calibration curves were generated from eight natural-composition glasses from the USGS (BIR-1G, BCR-2G, BHVO-2G) and the Max Planck Institute (KL2-G, ML3B-G, StHs6/80-G, GOR132-G, T1-G; Jochum, et al., 2006). A crystal of San Carlos olivine (Fo_{88}) was also analyzed as an in-house reference.

FTIR Analysis

The H_2O and CO_2 contents of the wafered glass samples were determined by Fourier Transform Infrared Spectroscopy (FTIR) at the Graduate School of Oceanography. Infrared spectroscopy has been used for decades to gather transmission spectra of volatiles in glasses (e.g. Stolper, 1982; Dixon, 1995). FTIR spectroscopy has many advantages for measuring volatiles, including excellent analytical sensitivity for H_2O (~ 10 ppm), fine spatial resolution ($\geq 11 \mu\text{m}$), and the ability to determine CO_2 and H_2O species (CO_2 , CO_3^{2-} , H_2O , OH^-) in sample glasses (Devine, 1995).

We followed methods of Kelley & Cottrell (2012) for reduction of FTIR spectra. Spectra were collected from $400\text{-}6000 \text{ cm}^{-1}$ using a Thermo Nicolet iS50 bench FTIR spectrometer coupled with a Continuum microscope equipped with a liquid nitrogen-cooled MCT-A detector, KBr beam splitter, and a tungsten-halogen source. The bench, microscope, and samples were continuously purged with dry air. Thicknesses of the samples were directly measured with a digital piezometric micrometer and ranged

between 20-150 μm (Table 5). Clear, glassy regions of the wafered glass samples were selected optically and the size of the aperture customized to maximize coverage of melt inclusion samples.

Background spectra were collected as a reference before each new spectrum of a sample was collected. Transmission FTIR spectra were read as absorption spectra, and were collected on three different spots in each sample, in addition to a spectral map of the entire inclusion. Dissolved CO_3^{2-} was quantified using the antisymmetric stretching absorptions at 1515 and 1435 cm^{-1} ; the background for fitting the carbonate peaks was done using an empirical least-squares fitting routine developed by Sally Newman. Total peak heights above background for H species were determined by fitting the spectral background with a spline function using OMNIC software. OH^- was quantified using the absorption at 4500 cm^{-1} , and molecular water at the absorption bands of 1630 and 5200 cm^{-1} . Peak heights were analyzed at absorption bands at 3530 cm^{-1} for total H_2O at low H_2O concentrations, whereas 1630, 5200, and 4500 cm^{-1} absorption bands were used for glasses with higher H_2O content (Figure 5; Table 5).

$\text{Fe}^{3+}/\Sigma\text{Fe}$ Analysis

Magmatic $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios are a proxy for system oxygen fugacity ($f\text{O}_2$; Kress & Carmichael, 1991). Wafered inclusions and some external matrix glass were analyzed *in situ* for $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios (i.e., $\text{Fe}^{3+}/[\text{Fe}^{2+} + \text{Fe}^{3+}]$) using micro X-ray Absorption Near Edge Structure ($\mu\text{-XANES}$) spectroscopy, using methods outlined by Cottrell et al. (2009), at beamline X26A, National Synchrotron Light Source, Brookhaven National Lab. Wafered samples were scanned in two dimensions to check that the 9x5 μm XANES beam could

pass through a path of clear glass only, though olivine interference is detectable in the spectra and able to be excluded where necessary (e.g., Kelley and Cottrell, 2009).

Standards used were LW and AII glasses equilibrated at variable oxygen fugacities.

Spectra were recorded from 7020 to 7220 eV with 0.1 eV steps over the pre-edge from 7108 to 7118 eV. Pre-edge spectra were corrected for energy drift and normalized to a value of 7112.3 eV for the LW_0 reference glass (Figure 6). Determination of $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios for basaltic glass is precise within ± 0.005 (Cottrell et al., 2009).

4. RESULTS

4.1 Petrography

In thin section, Type 1 lava shows olivine as the only phenocryst phase (Figure 7a-b). Olivine phenocrysts range in size from large (~0.5-1 mm) to small (~50-100 μm), and all are euhedral, and exhibit skeletal growth textures that enabled the trapping of melt inclusions. The vesicle content of Type 1 lava is estimated at 30-40%, and olivine abundance at ~5-10%. In Type 2 lavas, euhedral to subhedral olivine and clinopyroxene occur as glomerocrysts and isolated phenocrysts 0.25-1 mm in size (Figure 7c-d). The textures of glomerocrysts in Type 2 lavas (e.g., polygonal grain boundaries) suggests at least some olivine and clinopyroxene could be included as xenocrysts, although olivine-hosted melt inclusions and the external glass are similar in trace element composition, suggesting they are genetically related. Vesicularity of Type 2 lavas is ~50%, and total phenocryst abundance is ~5-15%.

4.2. Assessment of Post-Entrapment Crystallization of Olivine

Post-Entrapment corrections are sometimes necessary in order to add back any host crystal that has been lost from the melt inclusion. As melt inclusions cool with the host olivine crystal, it is possible to precipitate a bit of the host mineral from the melt composition onto the inclusion walls, which modifies the glass composition relative to when the inclusion was first trapped. In order to reconstruct the melt inclusion composition at the moment of trapping, we must assess and apply post-entrapment crystallization (PEC) corrections.

After the XANES data were collected, PEC corrections were applied. The equilibrium between melt inclusions and their olivine hosts was assessed by separating FeO* in each melt composition into Fe₂O₃ and FeO using the measured Fe³⁺/ΣFe ratio, and then calculating the olivine in equilibrium with each melt composition. The calculated equilibrium olivine was compared with measurements of the host olivine composition for each sample, and any deviations from equilibrium were attributed to post-entrapment processes (Danyushevsky, 2000).

We use the distribution coefficient $K_{D \text{ ol-liq}}[\text{Fe}^{2+}/\text{Mg}] = [(\text{Fe}^{2+}_{\text{Ol}}/\text{Mg}_{\text{Ol}}) / (\text{Fe}^{2+}_{\text{Liq}}/\text{Mg}_{\text{Liq}})] = 0.3$ (Roeder and Emslie, 1970) to calculate the olivine composition in equilibrium with each melt inclusion, and compare this with the measured composition of its olivine host. If a melt inclusion is in equilibrium with an olivine of lower Fo than its host, then post-entrapment olivine crystallization has occurred. In this case, we add equilibrium olivine back to the melt composition in 0.01% increments until the Fo value of the melt indicates equilibrium with the host. If a melt inclusion appears to be in equilibrium with olivine of higher Fo than its host, then post-entrapment Fe²⁺ diffusion out of the inclusion may have occurred. To correct this, the FeO concentration of the inclusion is increased, and concentrations of the other major elements are proportionally diluted, until the melt indicates equilibrium with its host (Figure 8). Using these corrections, we can account for any post-entrapment processes.

Both Type 1 and Type 2 samples contained olivine hosts that were Fo₈₇ – Fo₉₂. PEC corrections were necessary in 11 out of 12 melt inclusions to equilibrate with the olivine host, and ranged from 2% to 15% olivine added (Table 3). The corrected PEC values for major elements are used for all subsequent analysis and discussion.

4.3. Post-Entrapment Crystallization of Other Phases

Although melt inclusions selected for preparation had no visible daughter crystals, in three cases, daughter crystals were evident only after preparation and analysis. Daughter crystals were found in 3 of the melt inclusions from the Type 2 lava only (Figure 9). Scanning electron microscope [SEM] analysis revealed these crystals to be cpx in all cases. Only the Type 2 samples had daughter crystals. There was no evidence of systematic modification of the major element compositions of these inclusions relative to those with no daughter crystals (Figure 10), and so we choose to show the data. It is important to note that in all subsequent figures, these inclusions with daughter crystals are given different symbols – red stars – and are labeled as “dxtal” for daughter-crystal, where they have been plotted.

4.4. Crystallization of the Evita Magmas

The MgO concentration is a commonly used proxy for extent of crystallization, because mafic minerals crystallizing out of the melt, beginning with olivine (Bowen, 1922), will take MgO out of the melt. Thus, a higher MgO content indicates less crystallization and more primitive magma compositions. Type 1 samples have generally higher MgO concentrations than Type 2 samples (Type 1: 6.6-13.0 wt.% MgO; Type 2: 6.4-7.8 wt.% MgO; Figure 11), and on this basis, one might easily infer that the Type 2 magma has experienced more fractional crystallization than Type 1. Yet, our analyses of both magmas include melt inclusions from $>F_{088}$ olivine (Figure 8), suggesting that inclusions in olivine from both magmas are close to equilibrium with mantle olivine and

have experienced little crystallization since removal from the mantle source despite their differing MgO contents.

We thus also explore molar Mg# ($\text{Mg}/[\text{Mg}+\text{Fe}^{2+}]$) as an alternate proxy for fractional crystallization, since it more closely reflects the extent to which a magma has left Fe-Mg exchange equilibrium with the mantle. Kelemen (2003) defines a primitive magma as $\text{Mg}\# > 60$, and a basalt in equilibrium with Fo₉₀ olivine has a Mg# of 73. Both magmas contain melt inclusions with Mg# 60-77; Type 1 has a maximum Mg# of ~77, Type 2 has a maximum Mg# of ~76. Thus, both Type 1 and Type 2 are near-primary melt compositions that closely reflect unmodified mantle-derived melt compositions in their major elements.

Major element trends among the melt inclusion suites from both magma types are consistent with variable amounts of fractional crystallization from two distinct primary magmas. The FeO* trends for both lavas show that they are tholeiitic to moderately calc-alkaline (Zimmer et al., 2010). In the Hunter Ridge, the island arc tholeiites contained the most magnesian olivines (Fo₉₄), even though those glasses had the highest FeO* at a given MgO content for the suite of magmas studied in the region (Sigurdsson, 1993). In both suites, Na₂O increases as MgO and Mg# decrease, which is expected because it is an incompatible element. Differences in Na₂O content of each magma type at a given Mg# reflect differences in their primary melt compositions, with Type 1 containing higher Na₂O than Type 2. The CaO data suggest Type 2 is clinopyroxene-saturated (CaO decreases with Mg#) and Type 1 is olivine-only (Ca is increasing as Mg decreases), which is consistent with the phase assemblages seen in the thin section petrography. Mid-ocean ridge basalts show decreasing Al₂O₃ with decreasing MgO or Mg#, which is

consistent with co-saturation of olivine and plagioclase. Neither of the Evita magmas suggest plagioclase saturation, because Al_2O_3 is constant or slightly increasing as Mg# decreases, which is also consistent with the petrographic data.

The highest Mg# samples from both magma types show some other differences. From Figure 10, Type 1 has similar Fe, higher Na and Al, and lower Ca than Type 2. Additionally, the data show that Type 2 has higher K_2O than Type 1. Arcs have higher K_2O than MORB due to its incompatible, fluid-mobile nature. Altered oceanic crust has more potassium than MORB to begin with before subduction (Kelley, 2003). Type 1 and Type 2 have similar TiO_2 , which is lower in arcs relative to MORB due to high extents of melting and more depleted mantle sources, a signal seen especially in arcs with back-arc spreading (Kelley, 2010).

4.5. Trace Element Signatures of the Evita Magmas

While major elements can give us a view into the primitive nature and crystallization of the Evita magmas, minor and trace elements provide specific views of the magma source in the mantle wedge. They can reveal incompatible elements, fluid-mobile elements, and melt-mobile elements that have implications for the composition of magma sources. Some studies (e.g., Kay, 1978; Yogodzinski and Kelemen, 1998, 2007) have found enriched, primitive magmas that are heavily influenced by slab melt trace element signatures, i.e., steeply-sloped rare earth element (REE) patterns, depleted in heavy relative to light rare earth elements, with high Sr/Y and La/Yb ratios, and low Y and Yb content. These signatures indicate garnet in the source of the melt, which is hosted in the thermally altered lithospheric slab (Moyen, 2009; Sajona, 1993).

Type 1 samples show relatively flat to slightly LREE enriched REE patterns that are typical of island arc basalts (e.g., the Marianas, Fig. 11). In contrast, Type 2 samples show strong enrichments in the LREE and steeply sloped REE patterns, suggestive of an incompatible-element enriched source and the presence of a residual phase capable of fractionating the REE (e.g., garnet). The Spider Diagram reveals differing anomalies between the magma types (Figure 12). Type 1 samples have a slight positive anomaly in Sr, which is typical of island arc basalts and suggestive of the addition of an aqueous fluid to the mantle source; however, Type 2 samples show negative Pb anomalies and slightly negative Sr anomalies, suggestive of a source that has lost fluid-mobile elements such as Pb and Sr. Both sample types show negative Nb-Ta and Ti anomalies, indicative of a source material with residual rutile or another phase that partitions the HFSE. Type 2, however, has much larger Nb-Ta and Ti anomalies than does Type 1, and Type 2 also displays a negative Hf-Zr anomaly that is absent in Type 1. A negative Hf-Zr anomaly may signify a source material that also contains residual zircon. Type 1 samples also have higher Ba/Th ratios than do Type 2, while Type 2 have higher La/Yb (29.5-43) and Sr/Y (50-58) ratios than Type 1.

4.6. Volatiles and Degassing of the Evita Magmas

We look at volatiles to determine if the melt inclusions have experienced degassing during post-entrapment processes, and to determine the minimum H₂O content for the original magma source of each melt type. Degassing can be assessed by looking at dissolved volatiles of differing vapor/melt solubility. The solubility of CO₂ is highly pressure-dependent, and at crustal pressures, is much lower than H₂O; it will thus be

removed preferentially from a degassing melt. A low CO₂ concentration then means H₂O may have begun to degas.

The model used to calculate the degassing paths was VolatileCalc1.1 (Newman & Lowenstern, 2002). Conditions chosen were closed system, 49% SiO₂ – which was the maximum allowed value for the basalt calculation – the default temperature for basaltic magma (1100°C), the H₂O wt.%, and the CO₂ in ppm. The degassing curves represent the modeled degassing paths of the samples with the maximum H₂O wt.%, and demonstrate that the other samples in the suite follow the path (Figure 13a). Eruption pressure is inferred to be collection pressure, which is calculated to be ~137 bars.

Even these very primitive melts have been modified by the effects of low-pressure degassing during the earliest stages of crystallization. Type 1 samples have low CO₂ concentrations and may have thus degassed some H₂O. Type 2 samples have higher CO₂ and are less likely to be degassed in H₂O. We can, however, say that the data constrain robust minimum values of H₂O contents for both types. The data suggest that Type 1 has at least 3 wt.% H₂O, and Type 2 has at least 2.5 wt.% H₂O. Type 1 also has one outlying sample, NLD-02-01-02_S1, at 4.7% wt.% H₂O, most likely due to its thickness of 16µm, which can cause interference fringes that increase uncertainty. This sample is also the highest Mg# for this type, and is not included in the figure.

We also look at sulfur because it commonly degasses with H₂O (Kelley & Cottrell, 2012). S and H₂O are positively correlated in both magma types, suggesting that they degassed together (Sisson and Layne, 1993; Wade et al., 2006; Kelley and Cottrell, 2012). The highest measured concentrations of H₂O, CO₂, and S are therefore minima for the primary melts. Type 2 samples also have much higher S concentrations than Type 1

samples (Figure 13b). Like iron, sulfur can exist in different valence states, and its solubility is sensitive to redox conditions, increasing with increasing oxidation. Sulfur is more soluble in oxidized magmas because S^{6+} has higher solubility than S^{2-} (Carroll & Rutherford, 1988). Thus, S content gives us some initial indication of the oxidation of each magma type, and we can infer that Type 2 magma may be more oxidized because it has higher S.

4.7. Oxidation of the Evita Magmas

The $Fe^{3+}/\Sigma Fe$ ratio (which is a proxy for fO_2) is useful to assess in these inclusion samples because it gives us a view into the oxidation conditions of their respective source magmas (Figure 14; Table 6). $Fe^{3+}/\Sigma Fe$ ratios were translated into $\log(fO_2)$ and ΔQFM values using the model of Kress & Carmichael (1991) at 1 atm and 1200°C. Type 1 melt inclusions have average $Fe^{3+}/\Sigma Fe$ ratios similar to global arc basalts, at 0.200 ($\Delta QFM+0.8$), with a maximum of 0.233 ($\Delta QFM+1.2$; Figure 15a). Type 2 basalts have high average $Fe^{3+}/\Sigma Fe$ ratios at 0.297 ($\Delta QFM+1.5$), with a maximum of 0.328 ($\Delta QFM+1.9$). The uncorrected PEC ratio for Type 2 is the highest $Fe^{3+}/\Sigma Fe$ ratio yet measured in a natural terrestrial basalt. Type 2 basalts are far more oxidized than Type 1 basalts of similar Mg# (Figure 15b), except for the samples at the highest Mg#, which are similar in terms of oxygen fugacity relative to the QFM buffer.

5. DISCUSSION

5.1. Redox Changes During Magmatic Differentiation

The highest Mg# samples from each magma type are actually fairly similar in oxidation, perhaps suggesting that Type 2 might have become more oxidized by some differentiation process. However, the highest Mg# sample from Type 2 is a daughter-crystallized inclusion; this casts doubt on the fidelity of the $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratio in this melt inclusion, which may have changed during post-entrapment crystallization and cooling. We thus cannot confidently use this data and propose that its composition is possibly not indicative of the unmodified source. Moreover, for all other melt inclusions, we show that at a given Mg#, Type 2 inclusions are uniformly more oxidized than Type 1 inclusions (Figure 15) which suggests a fundamental difference in the oxidation conditions of their respective sources. Thus, we conclude that the similar $f\text{O}_2$ is not real, and that there are no apparent significant changes in magmatic redox that go along with indices of crystallization and differentiation (e.g., MgO or Mg#, Figure 15).

There is some apparent relationship in the Type 1 magma that may relate its oxidation to S degassing. Type 1 magmas show that the lowest S samples have the most reduced $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios (Figure 16), which may relate to the loss of S^{2-} from the melt as SO_2 gas (Kelley & Cottrell, 2012). Although Type 2 magmas have far higher S abundance and range in S concentration than Type 1, S does not correlate with a change in $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios in this magma. This is likely because the magma is too oxidized, and thus all of the sulfur is S^{6+} , which is more soluble in oxidized magmas. The degassing-driven reduction modeled by Kelley & Cottrell (2012) is only possible if S^{2-} is preferentially degassed as SO_2 , leaving electrons behind in the melt that may then reduce

Fe. We thus conclude that sulfur degassing is not responsible for the oxidation conditions of these magmas.

5.2. Conditions of the Mantle Source

If the varying oxidation of these magmas is not changed by differentiation processes, then the differences must come from the mantle source. Both mantle sources have more oxidized conditions than MORB, or average upper mantle conditions. The magma source for Type 1 is consistent with island arc basalts documented in Kelley & Cottrell (2012). Estimates for primary melts in that study were on the order of QFM+1, whereas the Type 1 averaged QFM+0.8. However, Type 2 magmas are not consistent with this previous study, and have the highest oxidation values yet measured for a basalt. The Type 2 samples are far more oxidized than the Type 1 samples, and contain the highest yet recorded $\text{Fe}^{3+}/\Sigma\text{Fe}$ values for terrestrial glasses (Figure 14). Type 2 averages QFM+1.5 (maximum QFM+1.9), which is highly oxidized for a primary magma; one value from Kelley & Cottrell (2012) of QFM+1.8 was a differentiated melt.

5.3. Slab Melt Signatures and Mixing

The mantle sources of each magma type have very different oxidation states; what could be causing this? Trace elements tell us that Type 1 has a classic island arc basalt signature of a mantle melt modified by aqueous fluid addition from the dehydrating slab. The samples have flat REE patterns, with enrichments in fluid-mobile elements over melt-mobile elements (such as a high Ba/Th ratio), a slight positive anomaly in Pb, and a distinct positive anomaly in Sr. Pb and Sr are highly fluid-mobile elements, so their

concentrations will be enriched above elements of similar mantle/melt incompatibility that are not fluid-mobile (e.g., Ce, Nd) if the source has been modified by an aqueous fluid from the slab. These trends indicate that the magma source of Type 1 is mixed with a component of a fluid nature, not a melt.

Type 2 signatures, on the other hand, strongly indicate a slab melt component in the mantle source. The samples show large fractionation across the REE pattern, suggesting garnet in the source lithology for these magmas, and the large LREE enrichment suggests a melt influence, not fluid. The Ba/Th ratio also is low, suggesting more of a melt presence than fluid. HFSE anomalies (Nb-Ta and Hf-Zr) are negative, where they may have been retained by rutile or zircon in a melting slab residue. We can expect to see this in a sediment or slab melt signature, but not in a classic mantle melt. The negative Pb and Sr anomalies also suggest these elements have been removed from the source by previous dehydration. The large Sr/Y ratios suggest deep slab melting; as pressures increase, plagioclase becomes unstable and releases Sr, whereas garnet becomes stable and traps Y, increasing the Sr/Y ratio dramatically with increasing pressure (Moyen, 2009). The La/Yb ratio has a similar trend, which is also high in Type 2.

This slab melt signature can also be identified when we model the slab surface temperature using H₂O/Ce (Table 7; Cooper et al., 2012). Using their equation at 4GPa, and assuming the maximum H₂O content for each type, the calculated slab surface temperature for Type 1 is 781±50°C (Figure 17). This is cooler than the proposed threshold for basaltic slab melting at ~900°C, but is close to the average for global SSTs (~805°C) (Cooper et al., 2012), and is above the H₂O-saturated sediment solidus,

permitting the application of this model for these samples. Type 1 magmas thus likely reflect contributions from slab dehydration and sediment melt components. Type 2, however, has a calculated slab surface temperature of $1041 \pm 50^\circ\text{C}$. This is the hottest slab yet constrained by this method (Cooper et al., 2012), and is well above the threshold for melting of the basaltic slab.

This $\text{H}_2\text{O}/\text{Ce}$ thermometer depends on contributions from the slab fluid only, yet it is important to account for contributions of H_2O and Ce from the mantle wedge. To eliminate this issue, Cooper et al. (2012) developed a method using the Nb/Ce ratio to remove these effects and to un-mix the mantle from the arc composition to reveal the $\text{H}_2\text{O}/\text{Ce}$ ratio of the slab fluid. Nb is minimally added to slab fluids, and is thus normalized to Ce and plotted against $\text{H}_2\text{O}/\text{Ce}$ (Table 7; Figure 18). A highly depleted mantle source with low Nb/Ce will project to a slab fluid with higher $\text{H}_2\text{O}/\text{Ce}$ than a more enriched mantle source with high Nb/Ce. For the Type 1 magma, we choose a depleted, NMORB mantle source, and project through the Type 1 melt composition to a Nb/Ce ratio of 0.02 (assumed to be the maximum Nb/Ce ratio of the slab fluid). This yields a $\text{H}_2\text{O}/\text{Ce}$ ratio for Type 1 of 4650, which translates to a new SST of $767 \pm 50^\circ\text{C}$. Importantly, this new temperature is within the model uncertainty of the temperature determined from the uncorrected $\text{H}_2\text{O}/\text{Ce}$ ratio. We implement the same procedure for the Type 2 magma, but using an EMORB mantle source (as indicated by the trace element patterns; Fig. 12), but because Type 2 has a Nb/Ce ratio of 0.02, we assume that essentially all of the Nb and Ce in this lava originate from the slab component, and Type 2 thus does not require a correction.

This contrast in slab surface temperatures recorded by the two lavas at Evita presents the problem of how the slab surface can be at two different conditions beneath this volcano, and what it may imply for melt and fluid transport in the wedge. At a slab edge, as is present in the location of seamount Evita, relatively hot mantle material may be flowing around the slab edge and heating up the surface of the subducting slab (Kincaid, 2004; Figure 19a). It is possible that flow regimes in the Vanuatu subduction zone are creating two different melting paths near the slab surface beneath the seamount Evita, and thus creating the two different magma types we see in this study. Type 2 could possibly be a mixture of Type 1, which would be the mantle-melt end-member, with a slab-melt end-member (Figure 19b).

We attempted to model how much of a slab melt end-member component may be mixing with the Type 1 magma by using an average major element composition of andesitic slab melts (e.g., Klimm, et al., 2008), and the highest Mg# samples from Type 1 and Type 2. Many of the major elements in Table 8 are explained by ~25-50% mixing of an andesitic slab melt with the Type 1 basalt. The slab melt end-member composition is relatively unknown, so we do not necessarily expect a perfect match. SiO₂ is the most robust major element to look at because its composition is much less sensitive to the phase assemblage of the melting slab. On its own, SiO₂ conservatively suggests a mixture of 23.5% andesitic slab melt end member with 76.5% Type 1 basalt. If we assume these proportions, then the $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratio of the slab melt would be 0.456.

However, we found that the results varied widely (Table 8), and that we cannot reconcile all of the major elements of Type 2 by adding a slab-melt end-member to Type 1. Al₂O₃, CaO, and Na₂O of Type 2 are not mixes of the two proposed end-members,

lying outside their value ranges, and the other elements varied widely. With the inconsistent outcome, we also cannot definitely determine a constrained $\text{Fe}^{3+}/\Sigma\text{Fe}$ value for a slab melt end-member. Thus, although it seems possible that Type 2 is a mixture of Type 1 and a slab melt, the modeling is not conclusive. There are other possibilities to explain what may be happening to produce the differences in these magma types.

One possibility is that Type 1 and Type 2 magmas may not be related strictly to each other (e.g., Type 1 is the wrong thing to mix with to make Type 2), which would mean they need to be separated within the mantle wedge flow regime present at this slab edge. Perhaps Type 1 is the mixing result of a mantle melt with slab fluid added, and Type 2 is a mixture of mantle melt with slab melt added; this would leave the mantle melt, without any additions, as an end-member with which to mix (Figure 19c). We lack sufficient data, however, to reconstruct the mantle melt composition, minus any slab additions, to assess this hypothesis, and note that the differences in HFSE abundances between the two types suggests that the mantle sources are different (Figure 12). It may also be possible, due to complex toroidal flow around the slab edge, that Type 1 and Type 2 magmas are melts of different mantle from different parts of the wedge that are fortuitously delivered to the same ascent path beneath Evita. The magmas could, for example, originate from similar depths, but at different horizontal locations along the slab (Figure 19d). The slab melt signature would originate closer to the hotter slab edge, while the slab fluid signature would be from farther north, away from the slab edge flow.

5.4. Links Between Slab Signatures and Magmatic Oxidation

The final question we address in this study is whether slab melting is truly ubiquitous (e.g. Kelemen, 2003) and responsible for the oxidation seen at global arcs. Type 1 falls on the trend line of previous global data, including MORBS (Figure 20). However, Type 2 does not follow the global trend in $\text{Fe}^{3+}/\Sigma\text{Fe}$ vs. H_2O . This indicates that this slab melt in Vanuatu is not a global component that is wholly responsible for the global trend.

It may be possible that the Type 2 magma had far higher H_2O contents to start with, and simply degassed greatly before the melt inclusions formed. But, judging by the trend in H_2O v. CO_2 (Figure 13), the maximum measured H_2O content is close to the original magmatic value, and this theory does not seem likely. If the magma had greatly degassed H_2O , the magma would have little to no CO_2 . Additionally, the minimal extent of crystallization in the samples suggests that not much H_2O has been lost. When H_2O is degassed, the magma would start to crystallize rapidly as it crosses its liquidus, and plagioclase would saturate because H_2O in the melt suppresses plagioclase crystallization (e.g., Parman et al., 2011). As such, the observed phase assemblage and the primary nature of these magmas do not suggest that significant H_2O has been lost.

Even though it is highly oxidized, the slab melt component cannot be responsible for this trend in oxidation, because it is too dry. It appears from prior work (Kelley & Cottrell, 2009) that there is a component at global arcs that delivers both H_2O and oxidation together. But, Type 2 magma would have needed to have >6 wt.% H_2O in order to be the end-member to which most arc magmas seem to mix in H_2O - $\text{Fe}^{3+}/\Sigma\text{Fe}$ (Figure

19). There must be some other component that delivers both high H₂O and highly oxidized signatures to arc magma sources in the mantle wedge.

Type 2 has much higher sulfur concentrations than Type 1, so we can consider the volatile sulfur for this mechanism by comparing it to the $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios for both types (Figure 16; Kelley & Cottrell, 2012). However, the high sulfur in the oxidized sample may simply be reflecting the S solubility contrast (which, as discussed above, is a strong function of $f\text{O}_2$), and thus we cannot resolve whether slab-derived S is the responsible mechanism for mantle wedge oxidation. We may also speculate about other potential oxidants for the Type 2 magma. If this magma is indeed a mix of slab and mantle melts, it is possible that Fe^{3+} itself has been added from the slab to the wedge, creating the highly oxidized signature. For this to be the case, however, we would expect the more oxidized melts to have higher primary FeO* than the more reduced melts. As seen in Figure 10, Type 1 and Type 2 samples have comparable FeO* values, so this is probably not the case.

6. CONCLUSIONS

Two distinct basalt types were identified at seamount Evita in Vanuatu, based on contrasting phenocryst assemblage (Type 1: 1% phenocrysts; Type 2: 15% phenocrysts), and confirmed by thin section petrography. Major element analysis was conducted to study the crystallization history of the samples, and found that melt inclusions from both lava types show equilibrium with $\geq \text{Fo}_{88}$ olivine, consistent with host olivine compositions, have high Mg# (>65), and have undergone minimal crystallization. The inclusions thus are near-primary melt compositions that have escaped major modification since departing the mantle wedge.

Type 1 melts show flat REE patterns, high Ba/Th ratios, and negative anomalies in Nb and Ta consistent with mantle melting via fluxing from a slab-derived fluid. In contrast, samples from Type 2 strongly indicate a slab-melt signature mixed with a mantle melt. These samples exhibit steeply sloped REE patterns with strong enrichments in LREE over HREE that suggest garnet in the source lithology for these magmas. They also have low Ba/Th ratios, and high La/Yb (29.5-43) and Sr/Y (50-58), which are classically attributed to partial melting of the slab, which is consistent with $\text{H}_2\text{O}/\text{Ce}$ constraints on the slab surface temperature ($>1000^\circ\text{C}$).

Volatile analyses constrain minimum H_2O contents of each primary magma at $\sim 3\%$ for Type 1 and $\sim 2.5\%$ for Type 2. Sulfur is also much more abundant in Type 2 than Type 1, and relationships between S and $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios in Type 1 samples suggest this magma may have experienced reduction by S degassing.

Type 1 samples have $f\text{O}_2$ and $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios similar to global arc basalts, while Type 2 samples have $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios that are among the highest measured in natural

terrestrial glasses. These differences do not appear to be due to differentiation, but instead originate in the magma source. Because the highly oxidized Type 2 samples have trace element signatures that indicate input from a slab-melt source, we infer that the slab melt added to the mantle source of Type 2 magmas was highly oxidized, although not very H₂O-rich, suggesting that slab melts are not a globally-present component that variably mix with arc magmas to deliver both high H₂O and oxidized conditions.

A mixing model also shows that the Type 2 magma is not a simple mixture of a slab melt and a classic, fluid-rich arc basalt (as exemplified by the Type 1 magma). Complex mantle flow, arising from the setting of this region near a suspended plate edge in the mantle, may have allowed for the unusual diversity of magmas, derived from radically different slab conditions, at this volcano.

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Table 1. Major element compositions and Fe³⁺/ΣFe ratios

Sample Name	S1_XLG	S1-a	S1-c	S1	S4	S4_XLG	01_XLG	S2	S3	S4	S1	S2A	S2B	S3	S4
<i>Major Elements</i>								<i>dxtal</i>			<i>dxtal</i>			<i>dxtal</i>	
SiO ₂ WT%	54.79	54.16	50.22	51.67	51.54	52.77	51.40	53.72	49.27	50.18	51.12	51.72	49.90	49.81	51.41
TiO ₂ WT%	0.52	0.48	0.54	0.42	0.62	0.47	0.63	0.56	0.76	0.69	0.65	0.67	0.72	0.62	0.72
Al ₂ O ₃ WT%	15.00	14.78	14.91	14.08	13.87	14.79	15.57	13.87	16.27	13.55	14.10	13.71	13.33	13.87	12.70
Fe ₂ O ₃ WT%	1.55	2.02	-	1.83	1.50	1.70	3.21	2.02	1.66	2.62	1.91	2.55	2.66	2.61	2.68
FeO WT%	6.09	6.02	-	5.44	7.31	6.28	6.08	5.23	3.56	6.18	4.11	5.42	5.22	4.83	5.90
FeO* WT%	7.48	7.76	6.33	7.00	7.99	7.81	8.97	6.49	4.99	8.27	5.69	7.30	7.26	6.96	7.83
MnO WT%	0.12	0.14	0.10	0.15	0.10	0.19	0.13	0.11	0.08	0.10	0.09	0.14	0.13	0.10	0.14
MgO WT%	6.62	7.16	9.32	8.30	4.87	7.31	6.42	3.41	5.29	5.40	5.57	4.28	4.40	4.87	3.87
CaO WT%	11.29	10.92	11.65	10.53	11.65	11.01	10.44	15.73	14.87	14.97	15.35	15.48	15.24	14.71	13.10
Na ₂ O WT%	2.50	2.51	2.15	2.16	1.59	2.54	1.94	1.62	1.58	1.69	1.74	1.65	1.72	1.59	2.22
K ₂ O WT%	0.32	0.32	0.33	0.29	1.49	0.30	1.66	1.09	2.43	1.56	1.57	1.66	1.67	1.49	1.74
P ₂ O ₅ WT%	0.09	0.07	0.09	0.08	0.46	0.08	0.54	0.52	0.90	0.58	0.49	0.62	0.67	0.46	0.62
<i>Total</i>	106.35	106.35	95.65	101.95	103.00	97.28	97.70	104.36	101.65	105.79	102.39	105.20	102.92	101.91	102.95
H ₂ O WT%	1.02	2.43	3.03	5.13	2.10	1.54	-	2.76	2.29	-	1.25	1.97	2.09	1.19	2.11
CO ₂ ppm	75	142	140	142	180	141	-	1275	803	-	842	785	785	122	450
S ppm	26	79	1265	0	147	20	196	6702	2916	2451	2925	2872	3776	2491	5211
Fe ³⁺ /ΣFe	0.186	0.240	-	0.256	0.194	0.196	0.322	0.303	-	-	0.314	0.342	0.356	0.355	0.331
Mg #	60.07	62.67	-	68.82	51.12	61.70	59.35	51.06	-	-	66.37	55.21	56.57	59.99	50.55

Table 2a. Olivine Host compositions and Forsterite value

Sample Name	NLD-02-01-01- S1	NLD-02-01-02- S1	NLD-02-01-02- S4	NLD-02-02-01- S2	NLD-02-02-01- S3	NLD-02-02-01- S4	NLD-02-02-05- S1	NLD-02-02-05- S2	NLD-02-02-05- S3	NLD-02-02-05- S4
<i>Major Elements</i>										
SiO ₂ WT%	39.58	40.36	40.68	38.84	39.79	39.36	39.98	41.27	40.92	39.05
Al ₂ O ₃ WT%	0.02	0.07	0.02	0.18	0.20	0.03	0.11	0.02	0.01	0.12
FeO WT%	10.81	7.14	10.42	11.29	9.02	11.29	8.51	10.07	10.29	12.28
MnO WT%	0.16	0.12	0.18	0.20	0.15	0.21	0.15	0.16	0.18	0.20
MgO WT%	48.80	51.08	48.92	47.61	50.22	47.85	50.44	48.48	49.02	47.32
CaO WT%	0.17	0.17	0.26	0.20	0.22	0.28	0.26	0.27	0.34	0.25
NiO WT%	0.27	0.38	0.20	0.21	0.27	0.20	0.23	0.23	0.22	0.15
(Sum)	99.81	99.32	99.58	98.53	99.86	99.20	99.69	100.50	100.98	99.38
For	0.89	0.93	0.89	0.88	0.91	0.88	0.91	0.90	0.89	0.87

Table 2b. Olivine Host Trace Elements

Sample Name	NLD-02-01-01- S1-a	NLD-02-01-02- S1	NLD-02-01-02- S4	NLD-02-02-01- S2	NLD-02-02-01- S3	NLD-02-02-01- S4	NLD-02-02-05- S1	NLD-02-02-05- S2A	NLD-02-02-05- S3	NLD-02-02-05- S4
<i>Trace Elements</i>										
Li	1.33	-	1.25	2.35	-	-	1.83	-	2.66	2.81
Be	bdl	-	bdl	0.0481	-	-	0.0815	-	bdl	bdl
MgO%	48.8	-	48.9	47.6	-	-	50.4	-	49.0	47.3
P ₂ O ₅ %	0.0059	-	0.0128	0.0197	-	-	0.0466	-	0.0214	0.0392
K ₂ O%	0.00	-	0.00	0.00	-	-	0.00	-	0.00	0.00
CaO%	0.172	-	0.158	0.188	-	-	0.214	-	0.194	0.189
Sc	6.50	-	5.40	5.89	-	-	6.63	-	6.80	6.54
TiO ₂ %	0.00136	-	0.0284	0.00276	-	-	0.00209	-	0.00241	0.00172
V	2.60	-	3.13	2.07	-	-	1.82	-	2.38	2.22
Cr	348	-	331	300	-	-	382	-	233	222
MnO%	0.195	-	0.171	0.213	-	-	0.150	-	0.257	0.241
FeO*%	13.0	-	10.3	13.0	-	-	8.58	-	14.3	13.7
Co	151	-	131	142	-	-	126	-	157	165
Ni	1400	-	1650	1390	-	-	1740	-	1190	1260
Cu	2.28	-	2.65	6.43	-	-	5.76	-	6.90	7.08
Zn	68.2	-	59.5	96.3	-	-	58.3	-	92.6	93.3
Rb	bdl	-	0.0204	bdl	-	-	bdl	-	0.0282	bdl
Sr	bdl	-	0.0230	0.0140	-	-	bdl	-	0.174	bdl
Y	0.0568	-	0.0456	0.0814	-	-	0.0760	-	0.0929	0.0415
Zr	bdl	-	0.218	0.0882	-	-	bdl	-	0.0560	bdl
Nb	bdl	-	0.359	0.00194	-	-	0.00363	-	0.00133	bdl
Cs	bdl	-	0.00266	0.0106	-	-	0.0301	-	bdl	0.000194
Ba	bdl	-	0.0108	0.0101	-	-	bdl	-	0.0497	bdl
La	bdl	-	0.000105	0.00329	-	-	0.00819	-	0.0171	bdl
Ce	bdl	-	0.00191	0.00241	-	-	bdl	-	0.0146	bdl
Pr	bdl	-	0.0000769	0.000643	-	-	bdl	-	0.00351	bdl
Nd	0.0224	-	0.00602	0.00393	-	-	bdl	-	0.00738	bdl
Sm	bdl	-	0.00130	0.00704	-	-	0.0225	-	bdl	bdl
Eu	0.00714	-	bdl	0.00221	-	-	bdl	-	0.000808	bdl
Tb	bdl	-	0.000853	0.00250	-	-	0.00163	-	0.00147	0.000251
Gd(Dy)	bdl	-	bdl	0.00185	-	-	0.0178	-	0.0127	bdl
Dy	0.00861	-	0.0129	0.00212	-	-	0.0244	-	0.00507	bdl
Ho	0.00458	-	0.00265	0.00117	-	-	bdl	-	0.00278	bdl
Er	0.0276	-	0.00853	0.00726	-	-	bdl	-	0.0131	bdl
Tm	bdl	-	0.00155	0.00101	-	-	bdl	-	0.00305	0.000485
Yb	bdl	-	0.0105	0.0190	-	-	0.0267	-	0.0288	0.0161
Lu	0.0103	-	0.00449	0.00582	-	-	0.000800	-	0.00290	0.00329
Hf	bdl	-	0.00151	0.00423	-	-	0.00677	-	0.00386	bdl
Ta	0.00127	-	0.0216	0.00129	-	-	bdl	-	bdl	0.0000467
Pb	bdl	-	0.0342	0.240	-	-	bdl	-	0.0194	bdl
Th	bdl	-	bdl	bdl	-	-	bdl	-	0.00166	0.00286
U	bdl	-	0.000390	bdl	-	-	0.000606	-	0.00950	bdl

Table 3. Post-Entrapment Crystallization Corrected Glass Inclusions

Sample Name	S1-a	S1-c	S1	S4	S2	S3	S4	S1	S2A	S2B	S3	S4	S1	S2A	S2B	S3	S4	S1	S2A	S2B	S3	S4	S1	S2A	S2B	S3	S4	S1	S2A	S2B	S3	S4
Olivine added	2%		0%	9%	15%																											
SiO ₂ WT%	53.85		54.16	50.85	50.08																											
TiO ₂ WT%	0.47		0.48	0.39	0.54																											
Al ₂ O ₃ WT%	14.44		14.78	12.97	12.08																											
Fe ₂ O ₃ WT%	2.02		2.02	1.83	1.50																											
FeO WT%	6.02		6.02	5.44	7.31																											
FeO* WT%	7.84		7.76	7.09	8.66																											
MnO WT%	0.14		0.14	0.14	0.09																											
MgO WT%	8.08		7.16	11.62	10.24																											
CaO WT%	10.67		10.92	9.71	10.15																											
Na ₂ O WT%	2.45		2.51	1.99	2.38																											
K ₂ O WT%	0.31		0.32	0.27	1.30																											
P ₂ O ₅ WT%	0.07		0.07	0.07	0.40																											
Total	98.54		98.58	95.28	96.07																											
H ₂ O WT%	2.38		2.43	4.73	1.83																											
S ppm	77		79	0	128																											
Fe ³⁺ /iFe	0.232		0.240	0.233	0.156																											
Mg #	64.99		62.67	74.72	65.96																											

Table 4. Trace Element Compositions of Olivine-hosted Glass Inclusions

Sample Name	NLD-02-01-01- S1_XLG	NLD-02-01-01- S1-a	NLD-02-01-01- S1-c	NLD-02-01-02- S4	NLD-02-01-02- S4_XLG	NLD-02-02- 01_XLG	NLD-02-02-01- S2	NLD-02-02-05- S1	NLD-02-02-05- S2A	NLD-02-02-05- S3	NLD-02-02-05- S4
<i>Trace Elements</i>											
Li	4.00	3.70	2.85	2.66	5.05	5.86	4.00	3.17	3.44	4.29	6.29
Be	0.338	0.176	0.674	0.147	0.170	1.48	0.754	1.23	1.73	1.96	1.74
MgO%	6.65	8.11	9.35	6.18	32.7	5.64	2.95	4.42	3.47	4.37	3.72
P ₂ O ₅ %	0.0771	0.0895	0.0695	0.0705	0.0985	0.466	0.548	0.403	0.509	0.403	0.632
K ₂ O%	0.293	0.304	0.276	0.219	0.264	1.36	0.883	1.20	1.27	1.18	1.71
CaO%	11.3	12.3	11.4	11.6	11.0	10.3	15.7	14.9	14.9	14.7	13.8
Sc	37.0	37.5	37.7	37.5	40.5	27.5	38.7	46.6	45.2	38.3	37.6
TiO ₂ %	0.505	0.557	0.497	0.486	0.479	0.628	0.546	0.654	0.669	0.593	0.699
V	226	237	224	211	218	263	250	264	274	252	277
Cr	230	273	296	268	464	27.1	283	245	222	215	246
MnO%	0.136	0.107	0.145	0.140	0.260	0.147	0.111	0.079	0.097	0.108	0.137
FeO*	7.19	5.75	7.80	7.41	15.2	8.44	6.24	4.22	5.44	6.40	7.76
Co	25.1	27.4	35.2	26.6	115	27.3	13.4	15.3	13.4	16.3	16.1
Ni	26.1	86.1	107	31.5	562	27.6	10.1	13.8	5.68	12.9	3.38
Cu	83.1	80.2	86.3	93.2	94.6	137	77.2	80.0	90.2	112	145
Zn	59.0	49.3	60.5	58.8	103	66.2	68.0	38.6	57.1	60.7	77.6
Rb	2.71	3.21	2.51	2.49	3.16	26.5	20.5	20.2	24.7	22.1	27.3
Sr	225	196	212	208	209	1220	895	1350	1360	1220	1110
Y	12.7	14.5	12.8	14.8	14.0	20.7	27.7	23.3	27.1	22.4	22.6
Zr	47.3	50.3	48.1	47.2	48.2	124	145	131	165	118	139
Nb	0.618	0.711	0.502	0.627	0.633	2.83	2.48	2.63	3.01	2.66	2.52
Cs	0.0267	0.106	0.0514	0.0146	0.0278	0.167	0.211	0.0527	0.112	0.170	0.165
Ba	47.7	45.4	45.3	42.0	45.1	302	199	315	269	300	239
La	3.63	3.76	3.25	3.37	3.56	57.8	60.3	57.3	77.9	54.9	70.9
Ce	8.29	8.72	7.55	7.52	8.12	126	117	100	146	108	179
Pr	1.27	1.39	1.22	1.16	1.25	16.7	16.7	14.6	21.7	15.2	22.6
Nd	6.24	6.59	6.18	5.52	5.84	64.1	70.5	62.5	92.0	60.5	87.5
Sm	1.77	1.85	1.63	1.68	1.75	11.0	13.6	11.2	16.7	10.6	14.1
Eu	0.606	0.689	0.636	0.566	0.591	2.73	3.28	2.76	3.85	2.65	3.68
Tb	0.337	0.423	0.391	0.358	0.355	0.875	1.17	0.976	1.28	0.899	1.10
Gd(Dy)	2.03	2.34	2.30	2.17	2.51	7.12	9.17	8.03	11.1	7.80	9.03
Dy	2.29	2.73	2.16	2.40	2.21	4.15	5.12	4.52	6.08	4.18	5.00
Ho	0.472	0.549	0.500	0.540	0.510	0.732	0.872	0.884	0.872	0.766	0.827
Er	1.34	1.68	1.37	1.50	1.44	1.80	1.96	2.43	2.31	2.01	2.08
Tm	0.213	0.226	0.200	0.215	0.246	0.248	0.254	0.272	0.274	0.238	0.256
Yb	1.28	1.64	1.59	1.29	1.51	1.48	1.82	1.94	1.82	1.49	1.51
Lu	0.208	0.247	0.235	0.209	0.184	0.234	0.257	0.292	0.296	0.249	0.233
Hf	1.25	1.34	1.43	1.13	1.34	2.68	2.61	2.84	3.13	2.73	2.85
Ta	0.034	0.024	0.044	0.030	0.010	0.118	0.122	0.115	0.155	0.133	0.107
Pb	0.93	1.01	0.93	0.90	0.89	4.58	4.57	4.08	4.28	4.51	5.18
Th	0.38	0.40	0.42	0.47	0.40	5.64	5.63	6.07	6.59	5.68	5.00
U	0.15	0.14	0.14	0.12	0.15	1.67	1.23	1.42	1.77	1.48	1.88

Table 5: FTIR variables and results

Sample Name	NLD-02-01-01- S1_XLG	NLD-02-01-01- S1-a	NLD-02-01-01- S1-c	NLD-02-01-01- S1	NLD-02-01-02- S4	NLD-02-01-02- S4_XLG	NLD-02-01-02- S2	NLD-02-02-01- S3	NLD-02-02-01- S1	NLD-02-02-05- S2A	NLD-02-02-05- S2B	NLD-02-02-05- S3	NLD-02-02-05- S4	NLD-02-02-05- S4
Aperture Size (μm)	50x50	25x35	25x25	30x30	30x30	40x50	30x30	50x30	50x50	30x30	30x30	40x40	35x25	0.072
Thickness (cm)	0.05	0.05	0.05	0.016	0.031	0.031	0.026	0.036	0.132	0.04	0.04	0.04	0.116	0.072
Absorbance (cm ⁻¹)														
1630	0.08	0.32	0.37	0.15	0.03	0.01	0.14	0.18	0.42	0.16	0.16	0.16	0.23	0.32
3530	0.51	1.20	1.40	0.77	0.38	0.42	0.71	0.82	1.64	0.78	0.78	0.83	1.38	1.51
4500	0.01	0.06	0.01	0.03	0.00	0.00	0.02	0.01	0.02	0.02	0.02	0.02	0.01	0.01
5200	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.01	0.02	0.02	0.01	0.01	0.02
H ₂ O WT%	1.02	2.43	3.03	5.13	2.10	1.54	2.76	2.29	1.25	1.97	1.97	2.09	1.19	2.11
CO ₂ ppm	75	142	140	142	180	141	1275	803	842	785	785	122	450	450

Table 6: O-Liq-Thermometry (@ 1atm 1200°C)

Sample Name	S1-a	S1	S4	S4_XLG	S2	S3	S4	S1	S2A	S2B	S3	S4
Putirka 07 + M&G (°C)	1162	1198	1261	1168	1099	1083	1233	1155	1162	1157	1161	1146
log fO ₂	-7.16	-7.09	-8.35	-7.68	-7.19	-6.92	-6.95	-6.89	-6.76	-6.61	-6.45	-6.78
ΔQFM	1.14	1.22	-0.04	0.62	1.11	1.38	1.35	1.41	1.55	1.69	1.86	1.52

Table 7: Slab Surface Temperature (SST) calculation (@4GPa)

Sample Name	NLD-02-01-01- S1_XLG	NLD-02-01-01- S1-a	NLD-02-01-01- S1-c	S4	NLD-02-01-02- S4_XLG	S2	NLD-02-02-01- S1	S2A	S3	S4
H ₂ O (wt%)	1.02	2.43	3.03	2.10	1.54	2.76	1.25	1.97	1.19	2.11
Ce ppm	8.29	8.72	7.55	7.52	8.12	117	99.5	146	108	179
Ce (wt%)	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.01	0.01	0.02
H ₂ O/Ce	1230	2788	4014	2791	1896	236	126	135	110	118
ln(H ₂ O/Ce)	7.11	7.93	8.30	7.93	7.55	5.47	4.83	4.90	4.70	4.77
SST (°C)	889	814	781	814	850	1041	1099	1092	1111	1104
CORRECTION FOR MANTLE MIXING										
Nb	0.62	0.71	0.50	0.63	0.63	2.48	2.63	3.01	2.66	2.52
Nb/Ce	0.07	0.08	0.07	0.08	0.08	0.02	0.03	0.02	0.02	0.01
H ₂ O/Ce	1230	2788	4014	2791	1896	236	126	135	110	118
UNMIXED H ₂ O/Ce	-	-	4650	-	-	236	-	-	-	-
UNMIXED ln(H ₂ O/Ce)	-	-	8.44	-	-	5.47	-	-	-	-
UNMIXED SST (°C)	-	-	767	-	-	1041	-	-	-	-

Table 8: Major element calculations for slab-melt end-member

Major Elements	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Fe ³⁺ /ΣFe
Slab end-member (Klimm, 2008)	58.673	0.793	15.118	2.825	0.052	0.782	4.116	3.297	1.922	0.142	?
Type 1 (Vanuatu)	50.847	0.391	14.078	7.091	0.136	12.975	9.709	1.989	0.266	0.073	0.256
Type 2 (Vanuatu)	52.686	0.513	12.819	7.045	0.098	6.639	14.538	1.499	1.012	0.485	0.303
slab component=	0.235	0.304	-1.210	0.011	0.448	0.520	-0.863	-0.374	0.451	6.026	?
% slab component added to Type 1	23.499	30.421	-120.991	1.071	44.828	51.965	-86.328	-37.417	45.051	602.572	?

Figure 2. (a) Map of Vanuatu illustrating the general volcanic and tectonic setting and the spatial distribution of present-day upper-mantle domains (Heyworth et al., 2010). The Hunter Fracture Zone (HFZ) can be seen to the south of the arc, and the Vitiiaz Lineament is off the diagram, further east. (b) Bathymetry of seamounts Eva and Evita. Collected using an EM300 multibeam system on R/V Southern Surveyor cruise SS07/2008.

Figure 3: Olivine-hosted melt inclusions. (a) NLD-02-02-05_S4. In plane polarized light, we can see that there is a clear brown pool of glass in the melt inclusion inside the host olivine. (b) NLD-02-02-01_S3. In crossed polarized light, glass goes to extinction and appears black when the sample has been double-wafered to expose the glass on both sides.

Figure 4: LA-ICP-MS spectra of NLD 02-01-02_S3_XLG. Background box indicates background collection before the laser ablation of the glass begins. The data used to analyze trace elements exists inside the Peak box.

Figure 5: FTIR spectra for NLD 02-01-02_S4. Aperature size was 30x30um. The peaks at 3530 cm^{-1} and 1630 cm^{-1} are readily apparent. Peaks are also found at 4500 cm^{-1} and 5200 cm^{-1} . Sample had 2.1 wt.% H_2O and ~180ppm CO_2 .

Figure 6: XANES pre-edge spectra for standard glass LW_0.035, a Type 1 glass (NLD 02-01-02_S4), and a Type 2 glass (NLD-02-02-05_S3). The data was collected at Brookhaven National Laboratory, using methods outlined by Cottrell et al. (2009). Pre-edge spectra were corrected for energy drift and normalized to a value of 7112.3 eV for the LW_0 reference glass. Determination of $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios for basaltic glass is precise within ± 0.005 .

Figure 7: Thin section photos of Type 1 and Type 2. (a) NLD-02-01-02 in PPL. (b) NLD-02-01-02 in XPOL. (c) NLD-02-02-05 in PPL. (d) NLD-02-02-05 in XPOL. All photos were taken at magnification of 4.07x (ZEISS Optical Systems).

Figure 8: Host olivine Fo measured directly vs. Fo calculated from inclusion EMP results. Inclusions with daughter crystals are shown with a star, found only in Type 2. These are high-Fo olivines ($>Fo_{88}$). PEC drives the melt-inclusion-calculated Fo to lower values.

Figure 9: Clinopyroxene daughter crystals in the NLD 02-02-05_S1 melt inclusion. Photo taken by SEM during supplemental analysis at the Smithsonian. Photo also shows laser pits from LA-ICPMS analysis.

Figure 10: (a) FeO^* v MgO (b) FeO^* v $\text{Mg\#}$. (c) CaO v MgO . (d) CaO v $\text{Mg\#}$. (e) Na_2O v MgO . (f) Na_2O v $\text{Mg\#}$. (g) Al_2O_3 v MgO . (h) Al_2O_3 v $\text{Mg\#}$. Type 1 and Type 2 Vanuatu samples are plotted against global MORB (Cottrell & Kelley, 2009) and Marianas arc data (Brounce et al., 2015). Marianas samples are averages values from 5 volcanoes: Sarigan, Guguan, Agrigan, Alamagan, and Pagan. Daughter crystal (dxtal) and external glass (xlg) samples are shown.

Figure 11: Chondrite-normalized REEs; normalized to chondrite value by Nakamura (1974). Type 1 and Type 2 samples are compared to a typical arc signal from Agrigan samples in the Marianas (Brounce, 2015). Type 1 samples show trace element signatures comparable to Marianas, while Type 2 do not follow this trend, exhibiting REE patterns that suggest garnet in the source and slab melting.

Figure 12: Spider Diagram of Type 1 and Type 2 samples, compared with Agrigan samples from the Marianas (Brounce, 2015). Normalizing primitive mantle composition taken from Sun & McDonough (1989), elements ordered after Hofmann (1988).

Figure 13: (a) Degassing trends of Type 1 and Type 2 samples, modeled with VolatileCalc1.1 (Newman & Lowenstern, 2002). Conditions: 49 wt.% SiO_2 , closed system, and default temperature of 1100°C for basaltic magma. The red and blue curves represent the degassing path of the sample with the maximum H_2O wt.%. Eruption pressure is inferred to be collection pressure, which is calculated to be ~137 bars. Both samples groups follow fairly consistent degassing trends from a minimum H_2O content for the source magma. (b) Comparing another volatile: H_2O v S.

Figure 14: Histogram showing the distribution of pre-PEC corrected $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratio results. Type 1 basalts (blue) have much lower $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios overall than Type 2 basalts (red). MORBs are reported from Cottrell and Kelley (2011). Marianas are reported from Brounce (2015).

Figure 15: (a) $\text{Mg\#}$ v. $\text{Fe}^{3+}/\Sigma\text{Fe}$. Types 1 and 2 are compared to each other and to global MORB (Kelley & Cottrell, 2012) and Marianas (Brounce, 2015). (b) $\text{Mg\#}$ v. ΔQFM . There is a clear trend in the MORBS of higher QFM with differentiation. This trend is absent in Type 1 and Type 2 from Vanuatu.

Figure 16: $\text{Fe}^{3+}/\Sigma\text{Fe}$ v Sulfur. Type 1 samples are given an inset to show trends more easily. Type 2 samples show significantly higher S contents than Type 1 samples, and additionally have higher $\text{Fe}^{3+}/\Sigma\text{Fe}$ ratios.

Figure 17: Slab surface temperatures of Vanuatu basalts compared to global island arcs. The calculation used to find the SST for Type 1 and Type 2 is taken from Cooper (2012), using H_2O and Ce contents of the magmas, at 4GPa. The global arcs sampled (Cooper, 2012) include the Lesser Antilles, Tonga, Nicaragua, Marianas, Kamchatka, Aleutians, Mexico, Guatemala, Costa Rica, and Cascadia. For global arcs, individual volcanoes are grouped by arc, with a filled green circle representing the arc average SST. Vanuatu SSTs were calculated from samples with the least degassed H_2O content.

Figure 18: Determination of H_2O/Ce in slab fluid components by unmixing the mantle contribution from arc compositions (Cooper, 2012). Vanuatu values are taken from samples with highest H_2O wt%. Figure shows mixing lines between mantle source, arc eruptive, and slab fluid components on H_2O/Ce versus Nb/Ce plots. There are two mantle source compositions (depleted MORB mantle [N-MORB] and enriched MORB mantle [E-MORB]). Slab fluid projections for Type 1 and Type 2 are from their preferred mantle sources. Nb/Ce of N-MORB is 0.311 and E-MORB is 0.553 (Sun and McDonough, 1989). H_2O/Ce of the mantle is 200 (Michael, 1995; Dixon et al., 2002). Nb/Ce of slab fluids is taken at 0.02, where Type 2 falls. For our new Type 1 H_2O/Ce ratio, we get ~4650. For our new Type 2 H_2O/Ce ratio, there is no discernable change. The new ratio is plugged back into the SST equation (Cooper, 2012), to get the unmixed, slab end-member component SST. The new Type 1 SST is $\sim 767 \pm 50^\circ C$, about $14^\circ C$ cooler than previously calculated. Type 2 SST remains the same.

Figure 19: Mantle wedge mixing models. (a) At Evita's slab edge, relatively hot mantle material may be flowing around the slab edge and heating up the surface of the subducting slab (Kincaid, 2004). (b) Type 2 could possibly be a mixture of Type 1, a mantle melt end member, and a slab melt end member, depicted here. Other possibilities include that (c) a mantle melt is mixed with slab fluid to form Type 1 and slab melt to form Type 2, and that (d) complex toroidal flow around the slab edge generates Type 1 and Type 2 magmas from similar depths but apart from each other horizontally on the slab surface, creating two different SSTs.

Figure 20: Type 1 and Type 2 minimum H_2O value plotted against the global trend for $\text{Fe}^{3+}/\Sigma\text{Fe}$ and H_2O (Kelley and Cottrell, 2009). Vanuatu samples shown are the least degassed samples. The trend here shows that while Type 1 oxidation can be explained by the observed global trend for oxidation with H_2O , the Type 2 magma would have needed to have a maximum H_2O content of 6 wt% to fit with the trend. Thus, Type 2 is not part of this global trend, and slab melting does not appear ubiquitous and providing oxidizing conditions at global arcs.