Accepted Manuscript

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PII: S0024-4937(18)30168-3
DOI: doi:10.1016/j.lithos.2018.05.011
Reference: LITHOS 4657

To appear in:

Received date: 31 January 2018
Accepted date: 10 May 2018

Please cite this article as: Christoph Beier, Philipp A. Brandl, Selma M. Lima, Karsten M. Haase, Tectonic control on the genesis of magmas in the New Hebrides arc (Vanuatu). The address for the corresponding author was captured as affiliation for all authors. Please check if appropriate. Lithos(2018), doi:10.1016/j.lithos.2018.05.011

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**Tectonic control on the genesis of magmas in the New Hebrides arc (Vanuatu)**

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**ABSTRACT**

We present here new bathymetric, petrological and geochemical whole rock, glass and mineral data from the submarine Epi volcano in the New Hebrides (Vanuatu) island arc. The structure has previously been interpreted to be part of a larger caldera structure but new bathymetric data reveal that the volcanic cones are aligned along shear zones controlled by the local tectonic stress field parallel to the recent direction of subduction. We aim to test if there is an interaction between local tectonics and magmatism and to what extent the compositions of island arc volcanoes may be influenced by their tectonic setting. Primitive submarine Epi lavas and those from the neighbouring Lopevi and Ambrym islands originate from a depleted mantle wedge modified by addition of subduction zone components. Incompatible element ratios sensitive to fluid input (e.g., Th/Nb, Ce/Yb) in the lavas are positively correlated with those more sensitive to mantle wedge depletion (e.g., Nb/Yb, Zr/Nb) amongst the arc volcanoes suggesting that fluids or melts from the subducting sediments have a stronger impact on the more depleted compositions of the mantle wedge. The whole rock, glass and mineral major and trace element compositions and the occurrence of exclusively normally zoned clinopyroxene and plagioclase crystals combined with the
absence of inversely zoned crystals and water-bearing phases in both mafic and evolved lavas suggest that the erupted melt was relatively dry compared to other subduction zone melts and has experienced little disequilibrium modification by melt mixing or assimilation. Our data also imply that differentiation of amphibole is not required to explain the incompatible element patterns but may rather result from extensive clinopyroxene fractionation in agreement with petrographic observations. Thermobarometric calculations indicate that the melts fractionated continuously during ascent, contrasting with fractionation during stagnation in an established crustal magma reservoir. We interpret the occurrence of this fractional crystallisation end-member in a relatively thick island arc crust (~30 km thickness) to result from isolated and relatively rapid ascent of melts, most likely through a complex system of dykes and sills that developed due to the tectonic positioning of Epi in a complex tectonic zone between a compressional environment in the north and an extensional setting in the south. We can show that the alignment of the cones largely depends on the local tectonic stress field at Epi that is especially influenced by a large dextral strike-slip zone, indicating that structural features have a significant impact on the location and composition of volcanic edifices.

**Keywords:** Vanuatu; New Hebrides; silicic melt; island arc; magma evolution; structural control; geodynamics

- Epi melts have experienced no disequilibrium modification by mixing or assimilation
- Melts fractionate continuously while ascending, rather than stagnating
- Magma ascent is through a complex system of dykes and sills
- Epi situated between compressional and extensional regime on thick island arc crust
- Structural features have impact on focusing and composition of island arc magmas
1. **INTRODUCTION**

The origin, ascent, and emplacement of silicic melts at convergent margins leads to a chemical differentiation and stratification of the island arc crust (Arculus, 1999; Bachmann and Huber, 2016; Haase et al., 2011; Taylor et al., 1969) that may eventually act as a precursor for the formation of continental crust (Adam et al., 2012; Polat, 2012). Silicic melts in subduction zone settings form either from partial melting of the hydrothermally altered subducting igneous crust (Pertermann and Hirschmann, 2003; Smith et al., 2006) or from partial melting of the arc crust and assimilation of crustal material as also observed along mid-ocean ridges (Shukuno et al., 2006; Tamura and Tatsumi, 2002; Wanless et al., 2010). A third hypothesis is that they may also form by extensive fractional crystallisation of basaltic magmas (Arculus, 1994; Gill, 1981; Grove et al., 2005; Haase et al., 2006; Haase et al., 2011). Thus, silicic melts provide information on either the deeper subduction zone processes or, alternatively, on those that affect the ascending melts in the island arc at shallower crustal levels.

Silicic melts commonly erupt explosively which appears to be triggered by the injection of mafic melts into a shallow evolved magma chamber (Sparks et al., 1977; Woods and Cowan, 2009), but there is evidence that the injection of silicic melts can also trigger volcanic eruptions (de Silva et al., 2008). Silicic melts are also responsible for the formation of some of the largest volcanic calderas and frequent explosive eruptions observed in island arcs (Carey et al., 2018; Folch et al., 2001; Jellinek and DePaolo, 2003; Kawakami et al., 2007; Walker, 1984; Wright and Gamble, 1999). Thus, understanding the formation, evolution and ascent of melts from the mantle wedge through the island arc crust is important.

The subsequent evolution and eruption from shallow magmatic systems (Kennedy et al., 2018; Self, 2006; Self and Blake, 2008), which may also consist of a complex systems of dykes and sills (c.f. Reid, 2014), is essential for the interpretation of silicic melt compositions and their eruptions. More recently, detailed investigations have questioned the simplicity of
the formation of island arc lavas resulting in explosive, evolved, volatile-rich eruptions. Kemner et al. (2015) showed that the submarine Monowai caldera volcano in the Tonga-Kermadec arc also contain relatively dry, tholeiitic melts that may have formed from mixing melts originating from both adiabatic decompression and fluid-fluxed melting. Hence, melts erupted in island arc calderas may contain crucial information on melt formation and ascent and may be better preserved in systems hosted in relatively thin island arc crust than in the thicker and more complex continental margin arcs (Ruprecht and Bachmann, 2010; Ruprecht and Plank, 2013).

The New Hebrides (Vanuatu) island arc consists of numerous volcanoes that frequently form islands situated in a tectonically complex arc-backarc system (Calmant, 2003; Meffre and Crawford, 2001; Mitchell and Warden, 1971; Ravenne et al., 1977; Taylor et al., 1995; Warden, 1967). Little is known about the composition and evolution of the submarine volcanoes and the impact of the local tectonic stress regime on magma geochemistry (Anderson et al., 2016). Submarine igneous samples from the arc and backarc volcanoes in the New Hebrides are rare (Crawford et al., 1988; Exon and Cronan, 1983; Greene and Exon, 1988; Lima et al., 2017; McConachy et al., 2005; Monjaret et al., 1991). The submarine Epi volcano consists of what has previously been interpreted to be a large submarine caldera supposedly hosting several post-caldera cones. In addition, Epi is situated in the transition zone between a compressional regime in the central New Hebrides arc (Calmant, 2003; Meffre and Crawford, 2001; Taylor et al., 1995) and an extensional regime in the south and thus provides the unique opportunity to investigate the processes and ascent of melts and their interaction with the local tectonic stress field.

Here, we use new major and trace element, and Sr-Nd-Pb isotope data of whole rocks and glasses from five submarine cones east of the island of Epi along with mineral major element chemistry. We show that the addition of sediment-derived fluids/melts released from the subducting slab along with fluids from the igneous slab have a significant impact on the
trace element compositions as expected in an island arc setting. We find no evidence for long-term petrologic disequilibria associated with magma mixing or assimilation within the lithosphere. The melts ascend through the crust in narrow pathways in a spatially limited area in which the impact from the surrounding host rocks and magmatic systems are small. The isolated ascent of the melts through the crust implies that the tectonic regime at Epi may prohibit establishment of a long-lived shallow magma chamber system. Instead the rising melts fractionate along a systems of dykes and sills continuously ascending and fractionating with little to no crustal assimilation.

2. GEOLOGICAL SETTING

The Vanuatu islands are part of the New Hebrides island arc system in the southwestern Pacific Ocean and started forming between the Eocene (Meffre and Crawford, 2001) and Middle Miocene due to the subduction of the Pacific Plate underneath the Australian Plate along the Vitiaz Trench (Schellart et al., 2006; Fig. 1). Collision of the Melanesian Border Plateau with the New Hebrides Ridge in the Middle Miocene caused a NE directed subduction of the South Fiji-North Loyalty Basin. Westward rollback of the subducting slab then led to a clockwise rotation of the New Hebrides arc and backarc extension associated with seafloor spreading in the North Fiji Basin from about 12 Ma (e.g., Schellart et al., 2006). The New Hebrides island arc is divided into three, independently moving blocks where the northern and southern segments are rotating counter-clockwise and clockwise, respectively, whereas the central block is moving eastward (Bergeot et al., 2009; Taylor et al., 1995; Vergniolle and Métrich, 2016). At present, the total rate of westward arc rollback of the southern Vanuatu arc increases from north to south (Schellart et al., 2006; Wallace et al., 2005). The presently active part of the island arc is the consequence of the subduction of the Australian Plate beneath the North Fiji Basin. Between 14°S and 17°S the Vanuatu Trench is unusually shallow due to the subduction of the D’Entrecasteaux Ridges and the West Torres
Massif (Fig. 1a) that also substantially affect the crustal structure of the central New Hebrides arc block (e.g., Bergeot et al., 2009; Meffre and Crawford, 2001). North and south of this central block, the arc started rifting about ~6 and ~3 Ma ago, respectively, opening the Jean Charcot Troughs in the north (Monjaret et al., 1991) and the Coriolis Troughs in the south (Fig. 1a; e.g., Dubois et al. (1977)) and forming the New Hebrides Ridge, a remnant section of the arc. The central block is undergoing compression as a result of collision with the D’Entrecasteaux Ridges and the West Torres Massif and most of the total plate convergence is accommodated by backarc thrusting (Fig. 1a; e.g., Taylor et al. (1995)). From north to south the recently active central New Hebrides island arc consists of the islands from Vanua Lava in the north to Anatom in the south (Monzier et al., 1997; Fig. 1a). All islands consist of several volcanic centres with the youngest ages typically at <1 Ma (Barsdell and Berry, 1990; Gorton, 1977).

The volcanic island of Epi is situated about 260 km south of the present intersection of the D’Entrecasteaux Ridge with the New Hebrides arc at the southern margin of the central block (Bergeot et al., 2009). However, 2-3 Ma ago, the D’Entrecasteaux Ridge may have entered the subduction zone at the recent latitude of Epi (Beaumais et al., 2013). Some 20 km northeast from Epi is the volcanically active island of Lopevi (Beaumais et al., 2013; Handley et al., 2008). The large Kuwae submarine caldera and Karua resurgent dome are situated towards the southeast of Epi and Tongoa (Fig. 1b). Tongoa was previously part of Epi island prior to a caldera-forming eruption at Kuwae in 1452 AD, which is assumed to have been one of the largest eruptions (Volcanic Eruption Index (VEI) 7) in historic times after the 1815 AD Tambora eruption (e.g., Goff and De Freitas, 2016; Monzier et al., 1994). Based on recent fieldwork, Németh et al. (2007) assigned the 1452 AD Kuwae eruption to be mainly submarine in origin and thus of much smaller (<VEI 6) magnitude of explosive eruptions. Today, Epi island hosts two structural provinces and numerous volcanic centres. The western part of Epi is formed by older raised marine and volcaniclastic strata and two Pliocene
volcanic structures (Barsdell and Berry, 1990; Quantin, 1972). The eastern part of Epi is formed almost exclusively from Quaternary volcanic rocks and has distinctive cliffs along its northeast shore, probably representing normal faults. The western wall has previously been interpreted to belong to a large submarine caldera system with a diameter of 9-15 km and water depths ranging from 400 to 500 m that was last active in 2004 (Garae et al., 2004). Our new bathymetric observations, however, question these observations and we will discuss the implication of these observations in more detail below. The area west of Epi island hosts at least seven smaller volcanic cones (each 1-2 km diameter) and three of these cones are aligned from west to east and have been named Epi A, B (recently being re-named Cioan volcano) and C (Greene and Exon, 1988), respectively (Fig. 1c). Historic volcanic activity at Epi has been described in 1920, 1953, 1960, 1984 and 2004 (Smithsonian Global Volcanism Project, (Greene and Exon, 1988). Four additional volcanic cones, named D to G are situated south of Epi B (Fig. 1c). Video imagery during the recent cruise reveals that all cones are covered by volcaniclastic material in various alteration stages, but only Cioan is covered by extremely fresh pumice and lava bombs (Haase et al., 2013). All cones exhibit evidence for previous hydrothermal activity (e.g., bacterial mats and Fe-Mn hydroxide staining) and during the recent cruise we found evidence for diffuse hydrothermal venting at the eastern flank of Cioan.

3. Sampling and Methods

The submarine Epi volcano was sampled during research cruise SO-229 Coriolis Troughs with the German RV Sonne in July 2013 (Table 1; Fig. 1). We sampled the Epi submarine cones during three dives (006-ROV, 013-ROV and 019-ROV) with the Remotely Operated Vehicle (ROV) Kiel 6000 from GEOMAR Helmholtz-Zentrum für Ozeanforschung Kiel. Dive 006-ROV approached the area between Cioan and Epi C from the south-east, dive 019-ROV targeted much of the area between the Cioan and Epi C edifices and dive 013-ROV was
performed from the area between Cioan and Epi C to the flank of Cioan and further south to Epi D. In addition, we performed eight successful TV-guided grabs (TVG) at Cioan and cones Epi C-G (008-TVG, 016-/017-TVG, 020-TVG, 023- to 026-TVG) from which igneous rock samples could be sampled. If fresh volcanic glass fragments (e.g., 006-ROV-11/-13) or matrix glasses (e.g., 006-ROV-01, 025-TVG-02) were available these were separated and analysed. For additional details of sample description see Haase et al. (2013)

For whole rock analyses, fresh cores were cut from the samples, coarsely crushed, washed thoroughly in deionised water to remove possible contamination from seawater, and then powdered in an agate mill at the GeoZentrum Nordbayern, Erlangen. The major element compositions of the matrix glasses and melt inclusions in clinopyroxene, orthopyroxene and plagioclase and the mineral compositions were analysed in thin sections using a JEOL JXA-8200 electron microprobe at the GeoZentrum Nordbayern, Erlangen. For glasses and melt inclusions the microprobe was operated with a 10 µm defocused beam, 15 nA beam current and a 15 kV acceleration voltage calibrated against natural standards (glass VG-A99 (NMNH 113498-1) for SiO₂, TiO₂, Al₂O₃, CaO, MgO, and FeO, glass VG-568 (NMNH 72854) for K₂O, scapolite (NMNH R6600-1) for Na₂O and Cl, chalcopyrite for S, rhodonite for MnO, apatite for P₂O₅). Precision and accuracy relative to A-99 were both better than 5 % (2σ) for all major elements (Beier et al., 2017; Brandl et al., 2012). The mineral analyses were performed using the same instrument operated at 15 nA, 15 kV with a fully focused beam (Freund et al., 2013). Calibration standards for mineral analyses were wollastonite (SiO₂, CaO), TiO₂, Al₂O₃, Fe₂O₃ (as FeO), rhodonite (MnO), MgO, albite (Na₂O), orthoclase (K₂O) and apatite (P₂O₅).

The whole rock sample powders are presented in supplemental Table 1 and their major element compositions were measured on fused glass beads on a Spectro XEPOS plus XRF spectrometer at the GeoZentrum Nordbayern, Erlangen (Freund et al., 2014). The major elements SiO₂, TiO₂, Al₂O₃, Fe₂O₃, MnO, MgO, Na₂O, K₂O, CaO, and P₂O₅ along with a
subset of trace elements (V, Cr, Co, Ni, Cu, Zn, Rb, Sr, Y, Zr, Nb, Ba) were analysed. Loss on ignition (LOI) was determined by weighing 1g of sample before and after heating the samples for 12 hours at 1300°C. Precision and accuracy of the major elements were generally better than 0.8% and 1% (2σ) for most elements and <0.5% for SiO$_2$.

The whole rock trace and glass trace elements were analysed by solution ICP-MS on a Thermo Scientific XSeries 2 quadrupole ICP-MS at the GeoZentrum Nordbayern, Erlangen. Sample powders and standards (BHVO-2, BCR-1) were dissolved following standard techniques using nitric and hydrofluoric acid as reactants (Freund et al., 2013; Lima et al., 2017). The samples were analysed at the same time than those presented in Lima et al. (2017) and standard values also apply to this dataset. Standards for major and trace element analyses are available in supplemental Table 2.

Radiogenic Sr and Nd isotopes were analysed at the GeoZentrum Nordbayern, Erlangen, following techniques described in Haase et al. (2017). Strontium and Nd isotope analyses were performed in static mode using a Thermo Fisher TRITON mass spectrometer with mass fractionation correction factors for $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ and $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. Repeated NBS SRM-987 standard measurements yielded $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.710266 (2σ = 0.000024, n = 3). The reproducibility of the Erlangen Nd isotope standard gave an average of 0.511536 (2σ = 0.0000085, n = 2). The Erlangen Nd isotope standard has a $^{143}\text{Nd}/^{144}\text{Nd} = 0.511541$ relative to the accepted value of 0.511850 for the LaJolla standard. The samples were normalised relative to the Erlangen Nd isotope standard.

Samples for Pb isotope analyses were dissolved using a HF-HNO$_3$ technique. Nitric and HCl acids were double distilled in order to minimise Pb blanks. Analyses were carried out using a $^{204}\text{Pb}/^{207}\text{Pb}$ double spike at the GeoZentrum Nordbayern using a Thermo Neptune Plus MC-ICP-MS. Spiked and unspiked aliquots were measured in static multi-collection mode. Standard runs of the NBS-981 standard (n = 14) gave average values of 16.9422 ±11, 15.4997 ±10, and 36.7239 ±25 for $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$, respectively and
analyses of USGS BCR-2 standard yield 18.8139±7, 15.6334±6 and 38.7206±10 for $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$, respectively. The major element, trace element, and Sr-Nd-Pb isotope data are presented in Table 1, all additional data are presented in supplemental Table 1.

4. PETROGRAPHY AND GEOCHEMISTRY

The Epi caldera samples range from pumiceous aphanitic samples with a glassy, cryptocrystalline vesicular groundmass (e.g., 017-TVG, 024-TVG-01) to relatively phric, hypocrysralline textures (e.g., 026-TVG) to almost entirely phric samples (e.g., 017-TVG-02) with varying amounts of a fine-grained to glassy groundmass (Fig. 2; (Haase et al., 2013)). All samples display evidence for extensive degassing, i.e. vesicularity may reach up to 30 vol.% with the vesicles often stretched implying flow textures. The prevailing mineral assemblage in all samples is plagioclase (up to 70 vol.%), olivine (~10 vol.%), clinopyroxene (~10 vol.%), and minor orthopyroxene, apatite and Fe-Ti-oxides (<10 vol.%). Several samples (e.g., 017-TVG-02) contain aggregates of olivine, clinopyroxene and orthopyroxene — the same mineral assemblage as in the groundmass (Fig. 2). Matrix glasses, glass embayments and glass inclusions occur frequently in the samples. Olivines from eight samples from Cioan, Epi G and Epi C (n=68) display almost constant forsterite (Fo = $100\times\text{Mg}/(\text{Mg+Fe}^{2+})$) contents between cores and rims with a very subtle normal zoning and range from Fo 72 to 83 (average 79±2; Fig. 3). There is no systematic difference between olivine inclusions in plagioclase, matrix olivine, and olivine phenocrysts and antecrysts. Orthopyroxenes analysed in four samples from between cones Cioan and Epi C, at Epi D, and Epi G (n=38) displays almost no compositional zoning and have enstatite (En) contents from 55 to 64 (average 60±2) and ferrosilite (Fs) contents from 33 to 41 (average 36±2) and an average Mg# ($\text{Mg#} = 100\times\text{Mg}/(\text{Mg+Fe}^{2+})$) of 63±2 (Fig. 3). Clinopyroxene phenocrysts in 9 samples from Epi C, Epi G, and Epi D (n= 168) range in wollastonite contents from 30 to 52
(average 42±3), enstatite ranges from 32 to 47 (average 42±2) and ferrosilite contents from 9 to 26 (average 16±4). They have an average Mg# of 72±6. Clinopyroxene crystals are normally zoned with slightly decreasing Mg# and enstatite contents from core to rim. A few crystals display minute changes (enstatite contents change by 2-3) between cores and rims that could be interpreted as inverse zonings. Since these changes are small we consider the overall compositions to be constant or normally zoned from cores to rims. Generally, olivine, orthopyroxene and clinopyroxene occurs in the more primitive lavas (i.e., those with whole rock SiO₂ contents <50 wt.%). However, the overall compositional range of the crystals in the mafic and silicic lavas is comparable (Fig. 3a). Plagioclase crystals from 9 samples from Cioan, Epi C, Epi G, and between Cioan and Epi C, (n= 199) display a much larger range in their compositions ranging in anorthite content (An = 100*(Ca/(Ca+Na+K))) between 31-96 (average 68±21), albite contents of 4-63 (average 31±19) and orthoclase contents up to 6 (average 2±1; Fig. 3). Plagioclase crystals with elevated An contents of >90 occur frequently (n= 50) either as large phenocrysts (>600 µm) or as inclusions (<20 µm) in clinopyroxene and as small crystals in the groundmass. Plagioclase crystals with An >52 frequently occur in lavas with <50 wt.% SiO₂ whereas lavas with SiO₂ >55 wt.% almost exclusively contain plagioclase crystals with An contents <50 (Fig. 3b).

The whole rock samples display a bimodal distribution with the majority of the samples at SiO₂ <50.3 wt.% and MgO contents of >5.9 wt.%; two more evolved whole rock lavas have ~62 wt.% SiO₂ and ~2 wt.% MgO (Fig. 4). The overall range of the whole rock major elements at Epi are similar to slightly larger than those at Lopevi (Beaumais et al., 2013), Ambrym (Handley et al., 2008; Sheehan and Barclay, 2016) or Efate (Raos and Crawford, 2004) at similar to slightly lower K₂O contents. Matrix glasses and glass inclusions have SiO₂ contents of >58.0 wt.% and MgO of <2.7 wt.% (Fig. 4). There is no clear systematic compositional difference between matrix glasses and glass inclusions analysed in clinopyroxene and plagioclase crystals. Hence, we suggest that post-entrapment crystallisation
and re-equilibration did not significantly alter the major element compositions of the glass inclusion for the purpose of this work. Generally, glass and whole rock compositions follow linear trends in most major elements, with the majority of the whole rocks at lower SiO$_2$, Na$_2$O and K$_2$O and higher MgO, FeO and CaO than the glasses.

Sulphur concentrations are generally below 350 ppm for all glasses but cover a large range and Cl/K ratios are generally below 0.1 (Fig. 5) with two samples extending towards 0.18 and Cl/Nb ratios are <700 (not shown, Table 1). Glasses and whole rock samples are enriched in incompatible Large Ion Lithophile Elements (LILE) and Light Rare Earth Elements (LREE) and depleted in Nb and Ta, respectively (Fig. 6 and 7). Europium anomalies expressed as Eu/Eu* ($\text{Eu/Eu*} = \frac{\text{Eu}}{\text{Sm} + \text{Gd}_{\text{N}}}$) are prominent in the more evolved lavas (Eu/Eu* <0.70) from the western cones (Fig. 7b). Lavas erupted at Epi C and G are more mafic (SiO$_2$ ~49 wt.%) and display less pronounced Eu anomalies (Eu/Eu* = 0.92-1.03).

Fluid mobile elements (e.g., Ba, Rb, Pb) are enriched relative to the fluid immobile elements (e.g., Nb, Ta, Hf; Fig. 6). Generally, ratios of Ba/Nb, Th/Nb, and Zr/Nb do not correlate with fractionation indices (e.g., SiO$_2$ and MgO), and published subaerial data from eastern and western Epi cover a similar range but extend to more elevated Ba/Nb ratios (Fig. 7d-f). The Eu/Eu* and Tb/Yb ratios are correlated with SiO$_2$ (Fig. 7a-c). Despite comparable major elements, the Th/Nb ratios of Ambrym and Lopevi are lower than those from Epi, whereas Zr/Nb ratios are comparable to those from Lopevi but higher than those from Epi island and Ambrym (Fig. 7e). Ratios of Nb/Ta range from 14.7-20.7 for the new data from Epi caldera and are within the range of the literature data from Epi island (not shown).

The Sr-Nd-Pb isotope ratios of the Epi lavas display a bimodal distribution with two mafic lavas at slightly lower $^{206}\text{Pb}/^{204}\text{Pb}$ (026-TVG-01 and 026-TVG-05) and all other lavas at $^{206}\text{Pb}/^{204}\text{Pb}$ of ~18.49±0.02 (Fig. 8). Sample 026-TVG-01 also has lower $^{207}\text{Pb}/^{204}\text{Pb}$, $^{208}\text{Pb}/^{204}\text{Pb}$ (Fig. 8), and Ce/Pb but $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.70422, whereas 026-TVG-05 only tends towards lower $^{208}\text{Pb}/^{204}\text{Pb}$ at lower $^{206}\text{Pb}/^{204}\text{Pb}$. $^{87}\text{Sr}/^{86}\text{Sr}$ isotope ratios range from 0.704197 to
0.704232 and the $^{143}$Nd/$^{144}$Nd ratios range from 0.512938 to 0.512954. The trace element and radiogenic isotope ratios at Epi are not correlated, implying a decoupling of the isotope and trace element ratios during the evolution of the lavas.

4.1 THERMOBAROMETRIC CALCULATIONS

We performed a total of 122 thermobarometric calculations using clinopyroxenes, 40 of which were in equilibrium with the co-existing melt (supplemental Table 4 for details), the remaining analyses are clinopyroxene-only calculations. Additionally, 98 plagioclase crystals were used to calculate temperatures and pressures. Calculations were performed using equations 23 (Standard Error of Estimate (SEE) = ±43°C), 25a (SEE = ±0.38 GPa), 31 (SEE = ±0.29 GPa), 32a-32d (SEE = ±0.31 GPa, SEE = ±58°C), and 33 (SEE of calibration data = ±0.16 GPa) from Putirka (2008). We have used the pressures determined from the clinopyroxene barometers as input for the plagioclase calculations of the respective samples, however, it has to be noted that the pressure estimates from plagioclase equilibria have to be treated with caution (Putirka, 2008). In addition, we have calculated H$_2$O contents as estimated from plagioclase equilibria (Putirka, 2008) and compare these with plagioclase hygrometer results using minimum and maximum An contents available for the samples (Lange et al., 2009).

Pressures and temperatures for all Epi lavas straddle ~0.2 to ~0.7 GPa at temperatures ranging from 1,050°C to 1,250 °C (Fig. 9). We do not find evidence for specific depth levels of stagnation in the Epi lavas. Plagioclase barometric estimates display slightly higher pressures than calculated from clinopyroxene equilibria, however, this may be the result of a greater uncertainty of the plagioclase barometric estimates (Putirka, 2008). We note however, that the pressures calculated from clinopyroxene-glass equilibria, clinopyroxene and plagioclase crystals overlap within their intra-sample variability and uncertainty (Putirka, 2008) and range between 8-30 km depth, well within the estimated range of crustal thickness.
for the region around Epi (Monzier et al., 1997). The calculated H$_2$O contents, using plagioclase-liquid equilibria (Putirka, 2008) in lavas with <50.3 wt.% SiO$_2$, range from 2-4 wt.% (Fig. 9c). The more evolved lavas display a bimodal distribution with a group of samples at H$_2$O contents <2 wt.% and one sample deviating to contents of 4-5 wt.%.

5. DISCUSSION

5.1 PRIMITIVE MELTS AND SOURCE COMPOSITIONS AT EPI

The formation and modification of melts in subduction zones has been subject of numerous studies (Arculus, 1994; Gill, 1981; Pearce et al., 1995; Pearce and Parkinson, 1993; Pearce and Peate, 1995; Plank and Langmuir, 1993; Porter and White, 2009; Regelous et al., 1997; Tatsumi and Eggins, 1995; Tatsumi and Kogiso, 2003; Todd et al., 2012) and the general geochemical characteristics of the Epi lavas are similar to those of other subduction zone lavas (c.f. Pearce and Peate, 1995). An enrichment in fluid-mobile elements (e.g., Ba, Rb, Pb) is combined with subtly low immobile High Field Strength Element (HFSE) concentrations and ratios (e.g., Nb/Ta = 14.7-20.7, Nb/Zr = 0.020-0.025; MORB: 16.8±1.2 and 0.062±0.058, respectively (Jenner and O’Neill, 2012); Fig. 6). The HFSE contents of the Epi lavas resemble the lowest concentrations observed in Pacific MORB (Fig. 6) and imply that the mantle underneath the islands has experienced previous melt depletion and that Lopevi and Epi lavas originate from a mantle source in which much of the trace element variability can best be explained by addition of a subducting sedimentary component (Fig. 7; Peate et al., 1997). We note however, that these differences are subtle and may reflect minor heterogeneities of the mantle wedge. The depleted mantle wedge has subsequently been re-enriched by a water-rich component derived from the subducting slab. Slightly elevated water contents are also consistent with the plagioclase-liquid estimates (Fig. 9). We note however, that radiogenic isotopes are not correlated with either fluid- or depletion-sensitive trace element ratios in the Epi lavas, implying that depletion must have been relatively recently (Pearce and Peate, 1995;
Peate et al., 1997) and that a long-term compositional heterogeneity of the mantle wedge may not affect the lava trace element compositions. Instead wedge depletion and addition of fluids is likely to result from the recent subduction process (Fig. 6). The exact origin of chemical depletion of the central New Hebrides island arc mantle wedge remains unclear in the absence of a nearby backarc (Pearce and Peate, 1995; Peate et al., 1997). The elevated Th/Nb ratios in the submarine Epi lavas compared to the subaerial Epi as well as Lopevi and Ambrym lavas indicate that fluids underneath Epi not only originate from the igneous portion of the slab, but may also inherit a trace element signature from the subducting sediments (Pearce and Peate, 1995; Plank, 2005). The fractionation of Ba/Nb and Th/Nb of Lopevi could also be explained by fluid addition from altered oceanic crust (Fig. 7), but Lopevi, Ambrym and Epi lavas display elevated $^{87}\text{Sr}/^{86}\text{Sr}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ similar to those observed in the subducting sediments (Peate et al., 1997; Plank and Langmuir, 1993). Despite the uncertainty in sedimentary compositions entering the New Hebrides island arc the elevated Ba/Nb, Th/Nb, Nb/La and, to a lesser extent, Zr/Nb ratios may best be explained by mobilisation from the subducting sedimentary portion of the slab. The contribution from the sedimentary subduction component is volumetrically small (5-10%, see modelling in Fig. 7), but significantly impacts on the depleted mantle wedge composition as a result of the higher elemental concentrations. The fractionation of Ba and Nb can be explained using a volcaniclastic sedimentary composition (Peate et al., 1997; Plank and Langmuir, 1993) along with partition coefficients at 800°C and 1000°C (Fig. 7; Kessel et al., 2005). The relatively high temperatures imply that the sedimentary component may be added as a melt rather than a fluid.

Slab depths at Epi caldera are on the order of 200 km, similar to those proposed for Ambrym and Lopevi but range from 140 to 160 km underneath Epi island (Fig. 1b; Hayes et al., 2012). At these depths the slab surface temperatures are on the order of 800-900°C (Syracuse et al., 2010), high enough to result in partial melting of sediments (Hermann and
Rubatto, 2009; Hermann and Spandler, 2008; Spandler and Pirard, 2013) in agreement with our best model fit. The occurrence of stronger sedimentary signatures at Epi compared to Ambrym and Lopevi (Fig. 7) may reflect shallower levels of dehydration and partial melting of the subducting sedimentary components at Epi compared to the neighbouring volcanoes (Beaumais et al., 2013; Pearce et al., 2007). The majority of the trace element ratios (e.g., Eu/Eu*, Ce/Yb, Tb/Yb; Fig. 7) are correlated with SiO$_2$ contents, thus suggesting that shallow level crustal processes may have significantly affected the compositions of the melt products erupted at Epi the significance of which we will discuss in detail below.

5.2 Melt Evolution at Crustal Levels

Previous works have stressed the importance of crustal assimilation and fractional crystallisation in the New Hebrides arc lavas (Beaumais et al., 2013; Handley et al., 2008). The Epi glasses and whole rocks cover a wide range in SiO$_2$ and MgO (Fig. 4) and contain numerous normally zoned clinopyroxene and plagioclase crystals (Fig. 3), suggesting that the melts were modified by fractional crystallisation. The fact that the lavas follow linear trends for incompatible elements like K and Zr with increasing SiO$_2$, as well as the constant compositions of Sr isotopes across the entire range of SiO$_2$, suggests that the Epi samples lie on one single liquid line of descent (Fig. 4 and Fig. 7). The difference between the whole rock and glass compositions and the presence of mafic xenocrysts is indicative that much of the major element variability may best be explained by xeno- and antecryst accumulation, and crystal aggregation in the lavas. Even the most primitive glass of the Epi lavas at 58 wt.% SiO$_2$ and 3.4 wt.% MgO (sample 006-ROV-13) has experienced extensive fractionation processes relative to a primitive, mantle-derived melt.

Decreasing Eu/Eu*, and increasing Nb/Yb and (Ce/Yb)$_N$ with increasing SiO$_2$ show that some incompatible element ratios have been significantly affected by fractional
crystallisation (Fig. 7) or, in the case of the whole rock samples by addition of xeno- and antecrysts (see also Monzier et al., 1997). Olivines lack a zoning pattern and have relatively low Fo contents (Fo 72-83 in equilibrium with mantle melts that have Mg# 44-60 (Roeder and Emslie, 1970)) and therefore, they are not in equilibrium with mantle melts or residual mantle (Green et al., 2001; Herzberg and Asimow, 2008; Niu and O'Hara, 2007)(Fig. 3b). Instead, these must have fractionated from a more mafic, parental melt (Mg# of ~72 if in equilibrium with the mantle) that is not directly sampled by one of the Epi submarine cones (Mg# of erupted magmas: 18-40). Similarly, the clinopyroxenes have relatively evolved compositions not in equilibrium with clinopyroxene from the mantle (Mg# ~91; (Putirka, 2008)). It has to be noted that the number of clinopyroxenes with relatively high Mg# decreases towards the more evolved rocks (Fig. 3a) indicative for a continuous crystal fractionation.

The frequent occurrence of plagioclase with An >90 could be related to elevated H$_2$O contents in the magma (Sisson and Grove, 1993). Plagioclase xenocrysts with An contents >70 have Ca/Na ratios ranging from 4-35 and both glasses and melt inclusions ranging from 58.1 to 70 wt.% SiO$_2$ display Ca/Na ratios of 0.5-2.5. These elevated Ca/Na ratios could be indicative of total H$_2$O contents of >2 wt.% in the Epi melts (Sisson and Grove, 1993) and our plagioclase hygrometer results could imply similarly elevated H$_2$O contents (Fig. 9). However, Danyushevsky et al. (1997) showed that high-An plagioclase can also crystallise from dacitic melts saturated in H$_2$O at shallow pressures and does not necessarily require a primary melt with elevated H$_2$O content. We can show that our high An-contents almost exclusively occur in the mafic lavas (Fig. 3) making crystallisation from a silicic melt unlikely. We also note that the majority of high-An plagioclase crystals are normally zoned with significantly lower An contents at their rims (e.g., in sample 026-TVG-07) consistent with an origin from fractional crystallisation. Based on the major element sums of >98 wt.%, high vesicularity of lavas, and our H$_2$O estimations on the most primitive sample using the plagioclase hygrometer (Fig. 9c), we suggest that the melts have minimum H$_2$O contents of 2-
2.5 wt.%. Such H₂O contents are lower than typical mafic arc magmas (cf. Plank et al., 2013) consistent with our observation that the Ba/Nb and Ba/Yb are elevated when being compared to MORB, but are relatively low compared to other island arc volcanoes. Instead the ratios at Epi are more comparable to those commonly found in backarc lavas (Pearce and Stern, 2006).

The general lack of sieve textures, resorption or re-crystallisation rims and the prevalent occurrence of normally zoned clinopyroxene and plagioclase implies that mixing between evolved and basaltic magmas did not influence the melts significantly during their ascent. We observe that sample 024-TVG-01, situated in the hydrothermal area between Cioan and Epi C (Fig. 1c), has two distinct glass compositions that range in SiO₂ between ~60 to ~70 wt.% which implies that the melts did not homogenise. Nevertheless, we do not observe systematic differences between each of the cones sampled. Sulphur concentrations range up to 350 ppm S and overlap with concentrations found in other subduction zone settings (Haase et al., 2006). The values suggest that the melts have experienced significant amounts of degassing (Brounce et al., 2014). The relatively low Cl/K ratios of the glasses provide evidence that little interaction with the surrounding seawater occurred. Combined with our petrographic observations we suggest that the lavas have experienced relatively little crustal assimilation or melt mixing modifications during ascent and eruption but have been subject to significant fractional crystallisation and degassing from a parental mafic magma. These observations, combined with the sparse occurrence of complex disequilibrium textures (e.g., absence of sieve textures, oscillating mineral zonings and recrystallized rims) and our thermobarometric estimates (Fig. 9), suggest that melts have not stagnated in the crust at any specific level and/or for any long period of time.

The trace elements (e.g., low Eu/Eu*, Nb/Yb, high (Ce/Yb)ₙ) and heavy REE (e.g., Dy/Ybₙ) at Epi could reflect amphibole differentiation (supplemental Figs. S1 and S2; Davidson et al., 2007; Davidson et al., 2012). Marcelot et al. (1983) found a few amphiboles in andesitic lavas from Futuna and Sorbadere et al. (2011) conclude that lavas from Aoba may
contain a component derived from amphibole-bearing clinopyroxenite sources. Amphibole-bearing cumulates are possibly stored at lower crustal levels (>800 MPa) similar to those observed at Merapi (Peters et al., 2017). The potential deep crystallization and storage of amphibole in subduction magmas (Davidson et al., 2007) poses the question as to why lavas from Epi do not contain any evidence for amphibole or biotite in their crystallising assemblage, neither in the mafic or evolved lavas and mafic mineral agglomerates. The thermobarometry at Epi suggests that the Epi lavas crystallise over a relatively large range of pressures, from ~0.7 GPa to ~0.2 GPa (Fig. 9). Our rhyolite-MELTS modelling suggests that the large range in SiO₂ contents (46.1-70.2 wt.%) at Epi can be best explained by shallow level crystallisation at relatively high temperatures, likely in excess of 1000°C (Fig. 4 and 9). Crystallisation temperatures are thus too high to facilitate crystallisation of amphibole (Ridolfi et al., 2010). Furthermore, H₂O contents as calculated from plagioclase-liquid equilibria suggest that H₂O contents are too low to result in the fractionation of amphibole (e.g., Krawczynski et al., 2012). Instead, the observed trace element pattern can be explained by differentiation of significant amounts of clinopyroxene, which is in agreement with our petrographic observations, (Ce/Yb)N, (Dy/Yb)N and Dy/Dy* ratios (Figs. S1 and S2) and systematically decreasing CaO/Al₂O₃ ratios with increasing SiO₂ (Fig. S2). In conclusion, the predominant crystallisation of clinopyroxene as opposed to amphibole suggest that crystallization dominantly occurs at shallow levels (<1 GPa; Ridolfi and Renzulli, 2012) and at high temperatures (Ridolfi et al., 2010), where amphibole stability is significantly reduced as the melts approach the surface (Rutherford and Hill, 1993). Thus, the Epi fractionation assemblage consists of olivine, clinopyroxene, orthopyroxene, plagioclase and Fe-Ti oxides with no chemical or petrographical evidence for amphibole or biotite.
5.3 TECTONIC AND MAGMATIC IMPLICATIONS: EVIDENCE FOR A SUBMARINE CALDERA?

The original interpretation that the submarine eastern Epi volcano is a caldera system was based on a geophysical survey in 1984. Greene and Exon (1988) and Crawford et al. (1988) suggested that the three large submarine cones Epi A-C are aligned on the northern rim of a submerged caldera (Fig. 10a). According to their interpretation of seismic profiles, the submarine ridge extending northwards from southeastern Epi island represents the eastern caldera wall. One observation is that the submarine Epi volcanic cones form a stack of edifices of varying age. However, in 2004, the French Institut de Recherche pour le Développement (IRD) undertook a high resolution bathymetric survey of the area and Pelletier et al. (2004) concluded that the Epi A-C cones are not located on the margin of a caldera. Their new bathymetric data along with data collected during cruise SO-229 indicate that the submarine ridge previously interpreted to be the caldera wall may be in fact a fault structure that strikes roughly N-S (Fig. 1c) and has a normal or slight right-lateral trans-tensional shear sense. This structure may represent a major lineament that extends southwards from the rear-arc thrust further north (Fig. 10a). A large submarine canyon breaches this fault-bounded ridge and is the locus of extensive mass transport from around the Epi submarine cones onto the eastern slopes of the island arc. The seismic profile (cf. Fig. 6 of Crawford et al., 1988) previously interpreted as caldera wall and floor could thus also be interpreted as a fault-bounded ridge and accumulation of layered volcaniclastic sediments upslope of the narrow gorge cross-cutting the ridge. On land, we used the publicly available SRTM (Shuttle Radar Topographic Mission) 1 arc-second data to identify possible fault trends. The main trend of fault scarps on the western part of the island of Epi is roughly 150-160°, whereas on the eastern part, fault scarps rotate towards more easterly directions (~100-110°; Fig. 10). From the fault scarps observed on land, there is no evidence for a subaerial continuation of the
submerged caldera structure as previously suggested (Fig. 10). Thus, at present, there is no compelling seismic, structural or bathymetric evidence for the existence of a submerged caldera at Epi. However, a complete bathymetric survey, especially north of Epi will be required to unravel the morphological and structural relationships of the submarine and subaerial Epi volcanic and tectonic structures.

Our new thermobarometric calculations for the submarine Epi lavas preclude the presence of a stable shallow magma reservoir, in which magmas stagnate for prolonged periods of time (Bachmann and Bergantz, 2008; Bachmann and Huber, 2016) emphasizing the possible lack of a crustal magma chamber system underlying a hypothesized Epi caldera. Instead, the polybaric crystallization indicated in lavas from all cones of Epi encompasses almost the entire crustal thickness underneath Epi (Monzier et al., 1997). This suggests that the melts ascend through a complex system of sills and dikes (c.f. Annen et al., 2015; Reid, 2014) in which melts continuously rise from Moho depths (~30 km) to shallower levels of ~8 km or less. Comparable observations were made in the Taupo volcanic zone by Gamble et al. (1999) and in the southern Cascades by Brophy and Dreher (2000). Thus, our petrological evidence for the lack of pro-longed melt stagnation at a given pressure interval (i.e., a stable magma reservoir) is another argument to question the existence of a caldera system at Epi (c.f. Kennedy et al., 2018) as opposed to the neighbouring Kuwae caldera.

5.4 Links between Regional Tectonics and Volcanism

Regional tectonics may control melt ascent paths and storage in volcanic arcs (Chaussard and Amelung, 2014; Hughes and Mahood, 2008) and we consider the New Hebrides island arc to be no exception. The New Hebrides island arc is subdivided into three major segments displaying differences in total convergence and relative motion rate (Taylor et al., 1995). The central segment is moving east relative to the northern and southern segments due to the collision of the D’Entrecasteaux Ridges and the West Torres Massif that is also
responsible for rear-arc thrusting along the central arc segment (Fig. 1). The relative motion between the central and the neighbouring northern and southern arc segments, respectively, is accommodated by strike-slip zones and especially the motion of the southern segment is well constrained by GPS motion vectors and earthquake focal mechanisms (Pelletier et al., 1998; Taylor et al., 1995). This dextral strike-slip zone is of great importance for the volcanic systems of Ambrym, Lopevi, Epi and Kuwae/Karua, that form the volcanically most active part of the arc (Picard et al., 1994). Ambrym, with its 12 km wide caldera and the resurgent domes of Marum and Benbow is located exactly at the transition from regional subsidence in the South Aoba Basin (collapse of the central arc due to collisional uplift in the forearc and thrusting in the rear-arc; (Meffre and Crawford, 2001)) to strike-slip faulting (Picard et al., 1994). Lopevi, Epi and Kuwae/Karua are all located within the strike-slip zone as proposed by Taylor et al. (1995).

The formation of the Epi lavas by continuous fractional crystallisation contrasts with observations from nearby Lopevi volcano (Beaumais et al., 2013; Handley et al., 2008). These differences may provide additional constraints on Epi’s specific tectonic setting. In contrast to the roughly E-W trending strike-slip zone of Taylor et al. (1995), Meffre and Crawford (2001) proposed two distinct strike-slip faults oriented ~45° and cross-cutting the arc. These extend from the trench to Ambrym and Epi, respectively cf. Fig. 12 of Meffre and Crawford (2001)). Here, we suggest that at Ambrym the orientation of the strike-slip zone is roughly E-W in accordance with the major rift intersecting the whole island and the seismic zone south of Ambrym (Fig. 10b). At Epi and on the seafloor SW of Epi, a series of lineaments have been mapped displaying two major strike orientations at ~70° and ~45° (Fig. 10b). This, in combination with the two roughly N-S trending major lineaments (Fig. 10a), may result in a highly fractured and permeable crust. Additionally, Epi is situated not only at the tip point of reversal in lateral but also of vertical crustal motions. According to Meffre and Crawford (2001), the whole area between Ambrym and Efate is affected by uplift, whereas the backarc
(ripping/spreading) and forearc (gravitational collapse following northward migration of collision with the D’Entrecasteaux Ridges) are subsiding south of Epi. North of Ambrym the central section of the arc system (North and South Aoba Basins) is subsiding whereas the fore- and rear-arc crust is being uplifted.

The structural regime of the New Hebrides island arc between Ambrym and the Vate Trough can be interpreted as a dextral strike-slip zone with a complex subset of Riedel shears (Fig. 11a). Changes in stress field from compression to extension (plus rotation) result in a variety of active shear systems (Fig. 11a) with different orientations. In Fig. 11b, we applied this model to the natural system with the result of a consistent structural interpretation of the morphological, seismic and volcanic lineaments mapped at Epi (Fig. 10b).

Even though Epi is located on relatively thick (~30 km) arc crust, the lithosphere is likely to be intensely weakened and fractured and may thus suppress the formation of a shallow, stable magma reservoir. Instead, the system of normal and strike-slip faults and shears supports the formation of dykes and sills (Reid, 2014). The ascent of magmas through a complex system of dykes and sills results in only limited mixing between individual melt batches (Fig. 11c). The small scale major element variability in individual volcanic edifices but similar trace element compositions between different cones may be indicative of a common magma reservoir with melts being extracted at different times. This also explains the alignment of the Epi cones and most other volcanic lineaments between Ambrym and the Vate Trough along synthetic shear bands (Fig. 11b) and stresses the importance of tectonic structures on the melting regime and melt ascent in island arc systems (Anderson et al., 2016; Lima et al., 2017). Interestingly, comparable observations of a tectonic influence in the production and ascent of magma “packages” were recently made at Ruapehu Volcano in the Taupo volcanic zone, New Zealand (Conway et al., 2018).
CONCLUSIONS

Lavas erupted at Epi volcano in the New Hebrides arc display a wide range from mafic whole rock compositions to evolved groundmass glass and glass inclusions. The lavas rise through thick island arc crust but have experienced little interaction with the neighbouring magmatic systems, which have experienced substantial assimilation of crustal material. This could be due to the fact that the wall rocks to these dyke and sill-like conduits may be “armoured” by crystallization from the magma. We conclude that much of the overall petrological character of the Epi lavas is the result of the specific tectonic setting of Epi, situated between two distinct structural domains. The compressional regime in the north and the extensional regime in the south may result in a highly fractured and thus permeable lithosphere that allows the ascent of isolated, distinctive melt batches through a complex system of dykes and sills. These isolated ascent pathways are situated in a broad strike-slip zone that originates from the regional scale tectonic evolution of the New Hebrides arc. The geochemical signatures indicate that the initial mafic melts formed from a slightly depleted mantle wedge by addition of fluids and melts from igneous and sedimentary portions of the slab, respectively. A lack of amphibole and biotite in the mineral assemblage of Epi suggests that crystals formed from a melt not sufficiently saturated in H₂O and/or at temperatures above amphibole stability. Thus, the petrology and geochemistry of silicic, relatively volatile-poor melts in island arcs may be substantially influenced by shallow, crustal processes rather than deep mantle signatures. The continuous range in pressures observed at Epi is inconsistent with a significant and stable magma stagnation in crustal reservoirs but rather indicates a polybaric magmatic system consisting of dykes and sills which also explains the lack of evidence for mixing and assimilation as would be the case in a large, crustal magma chamber. Our observations may also be applicable to other oceanic arcs, especially those that are situated in a complex tectonic regime. Our work stresses the importance of interdisciplinary studies able to connect geological and structural observations with geochemistry.
ACKNOWLEDGMENTS

We acknowledge captain D. Korte and his crew of R/V Sonne and the ROV Kiel 6000 crew for their invaluable help and remarkable efforts during SO-229 and thank the entire scientific party for their immense efforts. PAB thanks N. Augustin and M. Le Saout for their valuable help during multibeam processing and GIS data compilation, we thank S. Meffre for his helpful suggestions and IFREMER for providing additional bathymetric data. We kindly thank M. Regelous and B. Storch for their help in the Erlangen isotope laboratories. We acknowledge support by the Smithsonian Institution in providing calibration and measurement standards for the microprobe analyses. We thank T. Rooney in the Tokyo Haneda ANA Lounge, and J. Gamble for their constructive and helpful reviews that have improved the manuscript considerably and acknowledge constructive comments and editorial handling by A. Kerr. We acknowledge funding of the Bundesministerium für Bildung und Forschung (BMBF) for funding the cruise through project 03G0229A and thank T. Tevi, H. Cook and T. McConachy for their invaluable logistic help from the Australian to the Vanuatu shores and Coriolis Troughs. ChB acknowledges inspirational support from the peaty south coast of Islay. We dedicate this manuscript to Jon Davidson an inspiring, friendly and humorous scientist and colleague.

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**FIGURE CAPTIONS**
Fig. 1. a) Overview Map of the New Hebrides (Vanuatu) island arc. Tectonic structures in a) are from (Meffre and Crawford, 2001) and Patriat et al. (2015). Convergence rates are in cm/year from Calmant (2003). Grid in a) is from GEBCO (IHO-IOC GEBCO Gazetteer of Undersea Feature Names, www.gebco.net). Note that overall slab dip in the New Hebrides (Vanuatu) island arc may be amongst the steepest observed (~70°; (Lallemand et al., 2005) (Hall and Spakman, 2002)) but is varying from <15° over the first 20-40 km to 40° at 40 km depth (Baillard et al., 2015). b) bathymetry of the Epi, Ambrym and Lopevi region using a GMRT grid (Ryan et al., 2009) as background, and high-resolution ship based data from expeditions SO229 (Haase et al., 2013), CALVA (Eissen, 1996) and digitised grids from VATATERM (Ballu and Calmant, 2004). Subaerial elevation model compiled from ASTER Global Digital Elevation Model (GDEM) v2 data. ASTER GDEM is a product of NASA and METI. Depth of the Vanuatu Slab (Hayes et al., 2012) is shown by the coloured contour lines. Other symbols as in a). c) submarine cones east of Epi. Bathymetric grid compiled from GEBCO, high-resolution ship based data from expedition SO229 (Haase et al., 2013), and digitised grids from VATATERM (Ballu and Calmant, 2004). Note that sampling area in c) is on the same length scale than that of Ruapehu volcano in the Taupo volcanic zone discussed in Conway et al. (2018).

Fig. 2. Representative thin section images from submarine Epi caldera samples. a) sample 008-TVG-01g in plane polarised light with large clinopyroxene crystal and groundmass consisting mainly of Fe-Ti oxides, b) sample 017-TVG-01 in plane polarised light, with large clinopyroxene and plagioclase aggregate. Groundmass consists of glass and Fe-Ti oxides. Few small anhedral olivine crystals are also observed in the matrix. c) 017-TVG-02 in crossed polarised light showing phryic sample consisting of large clinopyroxene, plagioclase and olivine crystals, d) sample 017-TVG-07 in plane polarised light showing large aggregate of clinopyroxene and plagioclase in a glassy groundmass, e) vesicular sample 024-TVG-01 in
plane polarised light with large plagioclase crystal in glassy groundmass, and d) sample 026-TVG-01 under crossed polarised light with large clinopyroxene and plagioclase aggregate in fine grained matrix of clinopyroxene, olivine and plagioclase.

Fig. 3. a) Mg# numbers of clinopyroxene, orthopyroxene and olivine crystals. b) Anorthite (An) contents of plagioclase crystals versus SiO$_2$ contents of whole rock samples. Note the occurrence of a limited amount of high An and high Mg# crystals in the silicic lavas providing evidence for the occurrence of mafic xenocrysts in the more evolved lavas.

Fig. 4. Major element variation diagrams of whole rocks, matrix glasses and glass inclusions of the submarine Epi caldera. Panels a-f include non-normalised mineral analyses, panels g-l contain rhyolite-MELTS models (Ghiorso and Gualda, 2015; Gualda et al., 2012) calculated using sample 026-TVG-05 as starting composition. We use sample 026-TVG-05 as the most primitive sample that may best represent a liquid equivalent. This particular sample is sparsely phryic, hypocrystalline with ~10% plagioclase phenocrysts and 1-2% sub-rounded to subangular olivines in a fine-grained matrix (Haase et al., 2013).

Curves shown were calculated with fractionate solids flag using varying water contents and QFM buffer settings (Model A: 0.25 wt.% H$_2$O, 70 MPa, QFM; Model B: 0.25 wt.% H$_2$O, 70 MPa, QFM+1; Model C: 1.5 wt.% H$_2$O, 70 MPa, QFM; Model D: 1.5 wt.% H$_2$O, 70 MPa, QFM+1). A model calculated at 800 MPa fails to reproduce compositions at >55 wt.% SiO$_2$ (not shown). Tick marks in models in panels g-l are rhyolite-MELTS temperature steps of 25°C. Literature data for eastern and central Epi are from Dupuy et al. (1982) and data for western Epi are from Gorton (1977), Dupuy et al. (1982), Barsdell and Berry (1990) and Pearce et al. (2007) for a single sample. For analytical methods see main text.
Fig. 5. Cl/K ratios and S concentrations of the Epi matrix glass and glass inclusions and embayments. Note that two samples extent towards Cl/K ratios of 0.18 (at ~100 ppm S) not show in this figure. Arrows mark potential impact from assimilation of crustal material (increasing Cl/K) and degassing (lowering S), respectively. Limit of detection (l.o.d.) of S inferred from mineral calibration standard (see main text).

Fig. 6. a) Primitive mantle (PRIMA) normalized trace element abundance pattern and b) normal mid-ocean ridge basalt (N-MORB) normalized Rare Earth Element pattern of Epi whole rocks and glasses. PRIMA from Lyubetskaya and Korenaga (2007), N-MORB from McDonough and Sun (1995). Literature data from Gorton (1977), Barsdell and Berry (1990) and Pearce et al. (2007). Note the distinct depletions in immobile source sensitive elements (e.g., Nb, Ta) and enrichments in fluid mobile elements (e.g., Ba, Rb, Pb). Pacific MORB data compiled from Jenner and O'Neil (2012).

Fig. 7. a) Nb/Yb, b) Eu/Eu* and c) (Ce/Yb)N ratios versus SiO$_2$ contents; d) Th/Nb, e) Zr/Nb and f) Nb/La versus Ba/Nb ratios of the Epi glasses and whole rock data. Epi literature data as in figure 3, Ambrym data from Gorton (1977), Peate et al. (1997) and Pearce et al. (2007) Lopevi data from Beaumais et al. (2013) and Handley et al. (2008). Pacific MORB data compiled from Jenner and O'Neil (2012). Melt from depleted MORB mantle calculated using 10% fractional spinel peridotite melting of a depleted MORB mantle source provided by Workman and Hart (2005). Partition coefficients are those compiled and presented by Stracke et al. (2003). Subduction components (SC) sources calculated using partition coefficients by Kessel et al. (2005) for 1000°C (SC$^1$) and 800°C (SC$^2$). Altered oceanic crust from Castillo et al. (2009), fluid calculated at 700°C Kessel et al. (2005). Subducting, dehydrating and melting sediments are compiled in Plank and Langmuir (1993) from Peate et al. (1997). Tick marks are 0%, 5% followed by 10% intervals of adding a sediment fluid/melt to a DMM
Composition. Note that the range in Nb/La ratios in Figure 7f is slightly influenced by fractional crystallisation (i.e. mafic whole rocks ranging from 0.15-0.2, more silicic glasses are at ~0.25).

Fig. 8. a) $^{87}\text{Sr} / ^{86}\text{Sr}$, b) $^{207}\text{Pb} / ^{204}\text{Pb}$ and c) $^{208}\text{Pb} / ^{204}\text{Pb}$ versus $^{206}\text{Pb} / ^{204}\text{Pb}$ ratios of the Epi glasses and whole rocks. Literature data as in Figure 3 and 6. Pacific MORB data are those compiled from Stracke et al. (2003). Epi literature data are from Laporte et al. (1998), Lopevi from Beaumais et al. (2013), Ambrym and sediment data from Peate et al. (1997). Note however, that as a result of scarcity of data the sediment composition may contain some uncertainty.

Fig. 9. a) Temperatures and pressures from clinopyroxene only and clinopyroxene-liquid equilibria and plagioclase-liquid thermobarometers summarised in Putirka (2008), b) temperatures versus whole rock SiO$_2$ contents and c) calculated H$_2$O contents based on plagioclase-liquid equilibria after Putirka (2008). Arrows in c) show degassing and fractional crystallisation trends respectively. Whole rock thermobarometric data are averages from mineral calculations. Note that uncertainties of the calculations are 0.26-0.31 GPa and 46°C for clinopyroxene-liquid and clinopyroxene calculations and ±0.38 GPa and ±43°C for plagioclase-liquid calculations (Putirka, 2008). Note that pressure estimates from plagioclase depend on input pressures previously determined by clinopyroxene barometry. Here, only those pressures are being used where a common agreement between clinopyroxene and plagioclase calculations exists. Note, however, that the pressure estimates from plagioclase-liquid equilibria have to be considered with caution.

Fig. 10. a) Tectonomagmatic map of the area around Epi. Volcanic edifices and fault have been mapped using bathymetric and topographic data also used in Fig. 1b) plus additional bathymetric data from RV Southern Surveyor cruise CoTroVe in 2004 (Arculus et al., 2004)
for the Vate Trough. The proposed outline of Epi caldera from Greene & Exon (1988) is shown by the dashed, dark red line. b) Structural map of the area between Ambrum and Efate. Earthquake data were downloaded from the USGS earthquake catalogue (http://earthquake.usgs.gov) for reviewed earthquakes (1960-2017) with magnitude >4.5 at depths of <30 km. Lineaments mapped based on seismicity are shown in white, those based on morphology in pink and those based on volcanism in yellow.

Fig. 11. a) Model of a dextral shear zone in a compressional and extensional setting. Riedel shears R (synthetic) and R’ (antithetic), and the secondary synthetic P shears. b) Application of the model in a) to the New Hebrides arc area between Ambrum and Efate. Note that the synthetic Riedel shears may be focusing volcanic activity (yellow dashed lines), the antithetic Riedel shears may be accommodating most of the extensional movement (normal faulting) and the secondary synthetic P shears may form a subset of major crustal lineaments visible in morphological changes (pink). c) Synthesis sketch of melting, melt ascent and tectonic focusing underneath Epi. Not to scale.

**TABLE HEADINGS**

Table 1. Representative whole rock and glass major element, trace element and Sr-Nd-Pb isotope data. Full dataset for glasses, glass inclusions and minerals available in supplemental Tables 1 and 3. Methods as described in the main text. Whole rock major element analyses performed by XRF, trace element analyses of glasses and whole rock samples by solution ICP-MS at the GeoZentrum Nordbayern, Friedrich-Alexander Universität Erlangen-Nürnberg following methods also described in Lima et al. (2017). Major elements of the glasses were determined using an electron microprobe. International Geo Sample Numbers (IGSN) registered at System for Earth Sample Registration (SESAR, http://geosamples.org). Errors
for spiked Pb analyses are generally ±0.002, ±0.0019 and ±0.0046 for $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$, respectively.

**SUPPLEMENTAL MATERIAL**

Supplemental Table 1. Whole rock, glass major element, trace element and Sr-Nd-Pb isotope dataset. International Geo Sample Numbers (IGSN) registered at System for Earth Sample Registration (SESAR, http://geosamples.org). Original, shipboard sampling labels refer to ‘Station/Sampling locality (see Fig. 1) - Sampling Method (see main text) - Individual sample number’, i.e. TVG samples were taken by TV grab, ROV samples were taken during dives with Remotely Operated Vehicle Kiel 6000. Whole rock major element analyses performed by XRF, trace element analyses of glasses and whole rock samples by solution ICP-MS at the GeoZentrum Nordbayern, Friedrich-Alexander Universität Erlangen-Nürnberg following methods also described in Lima et al. (2017). Major elements of the glasses were determined using an electron microprobe. Errors for spiked Pb analyses are generally ±0.002, ±0.0019 and ±0.0046 for $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$, respectively.

Supplemental Table 2. Standard data for major and trace element analyses of whole rocks and glasses and electron microprobe standard for mineral analysis.

Supplemental Table 3. Mineral data of the Epi lavas. International Geo Sample Numbers (IGSN) registered at System for Earth Sample Registration (SESAR, http://geosamples.org). Parent IGSN refers to whole rock samples presented in Table 1 and supplemental Table 1.

Supplemental Table 4. Mineral data of the Epi lavas for those samples for which pressures and temperatures were calculated using Putirka (2008). International Geo Sample Numbers
(IGSN) registered at System for Earth Sample Registration (SESAR, http://geosamples.org). Parent IGSN refers to whole rock samples presented in Table 1 and supplemental Table 1.

Supplemental Fig. S1. Fractionation models of clinopyroxene and amphibole for chondrite normalised (Ce/Yb)$_N$, Ce and (Dy/Yb)$_N$ showing that Rare Earth Element Ratios (REE) can best be explained by extensive fractional crystallization of clinopyroxene also observed in the petrography of the Epi lavas. Partition coefficients used are those from Fujimaki et al. (1984).

Supplemental Fig. S2. a) CaO/Al$_2$O$_3$ versus SiO$_2$, b) Dy/Dy* versus SiO$_2$ and c) Dy/Dy* versus chondrite normalised (Dy/Yb)$_N$. Dy/Dy* calculated as in Davidson et al. (2012). Extensive fractionation model similar to that used in Supplemental Fig. S1. Chondrite normalised values from Sun and McDonough (1989).
Tectonic control on the genesis of magmas in the New Hebrides arc (Vanuatu)

Christoph Beier\textsuperscript{1}, Philipp A. Brandl\textsuperscript{2}, Selma M. Lima\textsuperscript{1} and Karsten M. Haase\textsuperscript{1}

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\textsuperscript{2}GEOMAR Helmholtz-Zentrum für Ozeanforschung Kiel, Wischhofstraße 1-3, 24148 Kiel, Germany
Table 1. Representative whole rock and glass major element, trace element and Sr-Nd-Pb isotope data. Full dataset for glasses, glass inclusions and minerals available in supplemental Tables 1 and 3. Methods as described in the main text. Whole rock major element analyses performed by XRF, trace element analyses of glasses and whole rock samples by solution ICP-MS at the GeoZentrum Nordbayern, Friedrich-Alexander Universität Erlangen-Nürnberg following methods also described in Lima et al. (2017). Major elements of the glasses were determined using an electron microprobe. International Geo Sample Numbers (IGSN) registered at System for Earth Sample Registration (SESAR, http://geosamples.org). Errors for spiked Pb analyses are generally ±0.002, ±0.0019 and ±0.0046 for 206Pb/204Pb, 207Pb/204Pb and 208Pb/204Pb, respectively.

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| Sr     | 14±7    | 04±7   | 14±6   | 97±6   | 2±7    | 32±6   | 27±8   | 97±7   | 14±7    | 04±7   | 14±6   | 97±6   | 2±7    | 32±6   | 27±8   | 97±7   |
| 143Nd/144Nd | 0.5129   | 0.5129 | 0.5129 | 0.5129 | 0.5129 | 0.5129 | 0.5129 | 0.5129 | 0.5129 | 0.5129 | 0.5129 | 0.5129 | 0.5129 | 0.5129 | 0.5129 | 0.5129 |
| 207Pb/204Pb | 15.552   | 15.551 | 15.542 | 15.549 | 15.513 | 15.539 | 15.548 | 15.554 |
| 208Pb/204Pb | 38.457   | 38.452 | 38.425 | 38.448 | 38.329 | 38.416 | 38.339 | 38.459 |
**Highlights**

- Epi melts have experienced no disequilibrium modification by mixing or assimilation
- Melts fractionate continuously while ascending, rather than stagnating
- Magma ascent is through a complex system of dykes and sills
- Epi situated between compressional and extensional regime on thick island arc crust
- Structural features have impact on focusing and composition of island arc magmas
Figure 3

(a) Mg# vs. SiO₂ content of whole rock sample [wt.%]
- Olivine rims and matrix crystals
- Olivine cores and phenocrystals
- Orthopyroxene rims and matrix crystals
- Orthopyroxene cores and phenocrystals
- Clinopyroxene rims and matrix crystals
- Clinopyroxene cores and phenocrystals

(b) An content vs. SiO₂ content of whole rock sample [wt.%]
- Plagioclase rims and matrix crystals
- Plagioclase cores and interiors

Key:
- Phenocrysts (008TVG-01)
- Inclusions in pyroxene (17TVG-07)
Figure 4af
Figure 4gl
Figure 5

(a) Cl/K vs. SiO₂ [wt.%]

(b) S [ppm] vs. SiO₂ [wt.%]

- Matrix glasses this work
- Glass inclusions and glass embayments

Assimilation vs. degassing
Figure 9

(a) Graph showing pressure versus temperature with data points for whole rock and glass samples.

(b) Graph showing temperature versus SiO₂ content of whole rock sample with data points for clinopyroxene and plagioclase.

(c) Graph showing H₂O content from P₂O₅ content with data points for plagioclase and arrows indicating fractional crystallisation and degassing.
Figure 10