Tracing changes in mantle and crustal influences in individual cone-building stages at Mt. Shasta using U-Th and Sr isotopes

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Abstract

²³⁰Th-excess is rare in most arc lavas, but common in the Cascades, yet the origin of such excesses remains unclear. At Mt. Shasta, age-corrected (²³⁰Th/²³²Th) and (²³⁸U/²³²Th) activity ratios range from 1.108–1.290 and 0.987–1.309 (27.3% ²³⁰Th-excess to 6.1% ²³⁸U-excess), respectively. Although small degrees of zircon crystallization (< 0.3%) may yield high (²³⁰Th/²³⁸U)₀ in derivative magmas, high Zr contents, the lack of zircon as a liquidus phase, and low Th/U ratios in Mt. Shasta lavas argue against zircon fractionation. Instead, melting models suggest ²³⁰Th-excesses are imparted on lavas through mixing mantle-derived magmas with partial melts of a mafic amphibolite lower crust where garnet was produced in the residuum through amphibole and plagioclase destabilization. The hot nature of Cascade magmas suggest that high intrusion temperatures promoted dehydration melting in the deep crust.

At Mt. Shasta, the destruction of the ancestral cone (Sand Flat) was followed by four cone-building stages, three of which lie in the age range of U-series geochronology. Lavas within individual eruptive stages have relatively constant (²³⁰Th/²³²Th)₀ that are interpreted to reflect specific mixtures of mantle (*m*) and lower crustal (*lc*) melts that are characteristic of a specific stage (*M_{m:lc}*). High (²³⁰Th/²³²Th)₀ ratios identify higher proportions of lower crust in the Misery Hill stage (*M_{m:lc}* = ~85:15), whereas low (²³⁰Th/²³²Th)₀ ratios reflect the more mantle-like composition of the Shastina lavas (*M_{m:lc}* = ~95:5); in the case of Shastina lavas, very low ⁸⁷Sr/⁸⁶Sr ratios, down to 0.7029, support a substantial

mantle contribution. Changes in $(^{230}\text{Th}/^{232}\text{Th})_0$ ratios correlate with eruptive volume, where the most voluminous stage (Misery Hill) is inferred to have the largest proportion of crustal melt and highest $(^{230}\text{Th}/^{232}\text{Th})_0$ ratios. Variable $(^{230}\text{Th}/^{238}\text{U})_0$ ratios within, and between, eruptive groups likely reflect a combination of residence time in the lower crust and differential assimilation of bulk, non-garnet-bearing crust that had $(^{230}\text{Th}/^{238}\text{U}) = 1$. The volume- $(^{230}\text{Th}/^{232}\text{Th})_0$ relations are accompanied by correlations with $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, where the most radiogenic Sr is associated with the largest eruptive volumes, indicating that the largest magmatic episodes produced the largest amount of lower crustal interaction.

The new U-Th and Sr isotope measurements of this study, along with U-series data for other Cascade centers suggests that interaction with the lower crust exerts greater control on Cascade magma chemistry than previously recognized. Indeed, the relatively dry nature of the Cascades may offer a unique opportunity to better understand the influence of the deep crust in young continental arcs, as larger subduction components in other settings may overprint any lower crustal signature and produce lavas with large ^{238}U -excesses.

Keywords: Mt. Shasta; Cascade Arc; uranium-series isotopes; strontium isotopes; crustal interaction; lower crust

1. Introduction

Magma generation and evolution in continental arc settings is commonly explained in terms of flux-melting in the mantle wedge and differentiation processes in the upper crust. Interactions between mantle-derived melts and deep continental crust may also modify the chemical and volatile compositions of magmas (Hildreth and Moor bath, 1988; Conrey et al., 2001; Annen et al., 2006; Reubi et al., 2011; Ankney et al., 2013). Mineral phases observed in erupted lavas may not reflect deep crustal

differentiation, and “cryptic” fractionation of amphibole and garnet in the lower crust has been invoked to explain the compositions of many arc magmas (e.g., Davidson et al., 2007; Hildago and Rooney, 2010; Lee et al., 2006). Importantly identifying minerals involved in deep crustal magma evolution provide constraints on compositional and thermal characteristics of the lower crust during melting. At upper mantle and lower crustal pressures, amphibole is a near-liquidus phase in mafic arc magmas at $\text{H}_2\text{O} > 5$ wt. % and temperatures $< 1100^\circ\text{C}$. At higher temperatures and lower water contents, plagioclase is the near-liquidus phase. In basaltic to andesitic magmas, plagioclase stability is suppressed and, instead, garnet is a near-liquidus phase at $\text{H}_2\text{O} > 5$ wt. % and temperatures $850\text{--}1100^\circ\text{C}$ (e.g., Mercer and Johnston, 2008; Müntener et al., 2001).

Unambiguous evidence for deep crustal melting and assimilation has been generally limited to settings where chemical and isotopic contrasts between mantle-derived melts and crust are high (e.g., Hildreth and Moorbath, 1988). Where the deep crust is dominated by young mafic igneous rocks, however, crustal interaction is more difficult to trace with commonly used isotope systems such as e.g., Sr, Nd, Pb, Hf, and O (e.g. Bacon et al., 1994; 1997; Hildreth 2007; Schmidt et al., 2008). Moreover, crustal interaction may produce a chemical fingerprint that is difficult to distinguish from that of a subduction component (e.g. Ankney et al., 2013).

Uranium-series isotopes offer a sensitive approach to discriminating between slab dehydration, mantle melting, and upper and lower crustal interaction (e.g. Bourdon et al., 2000; Turner et al., 1997; Zellmer et al., 2000; Berlo et al., 2004; Charlier et al., 2005; Yokoyama et al., 2006; Price et al., 2010). The majority of lavas erupted from active oceanic and continental arcs have ^{238}U -excesses, interpreted to reflect metasomatism of the mantle wedge by U-rich slab fluids (Turner et al., 2003). Significant ^{230}Th -excesses, however, are measured in some continental arc lavas, and are instead ascribed to a variety of

magmatic differentiation processes, including accessory mineral crystallization in the upper crust (e.g., Reagan et al., 2003) and assimilation of lower crustal melts of garnet-bearing or garnet-free protoliths (e.g., Ankney et al., 2013; Jicha et al., 2009).

The Cascades are commonly characterized as a “hot arc” based on the slow subduction of the relatively young Juan de Fuca and Gorda plates (Harry and Green, 1999; Wilson, 1988). High melting temperatures for primitive lavas are expected to promote extensive melting and assimilation of lower crust (e.g., Conrey et al., 2001; Elkins-Tanton et al., 2001). In contrast, previous work at Mt. Shasta argues against crustal interaction and, instead, ascribes chemical and isotopic variations in the more evolved stratocone lavas to a combination of variable slab contributions and magma mixing (Grove et al., 2005). In this study, we integrate the large chemical and isotopic database of the Mt. Shasta volcanic field with new U-Th and Sr isotope measurements to investigate potential mechanisms of ^{230}Th -enrichment, and to better constrain the influence of crustal interaction in magma generation and evolution of the stratocone andesites and dacites at Mt. Shasta. Combined, these data suggest extensive crustal interaction in both lower crustal and upper crustal regions, the magnitude of which correlates with eruptive volume.

2. Geologic Setting

Mt. Shasta is the largest stratovolcano in the Cascades (e.g., Calvert and Christiansen, 2011). The present-day edifice has been constructed by at least four major cone-building events (Figure 1). Basement rocks include the Paleozoic Trinity Ophiolite, as well as the Paleozoic and Mesozoic arc complexes of the Klamath mountains (Brouxel and Lapierre, 1988; Brouxel et al., 1988; Stanely et al., 1990). Following the destruction of the ancestral cone (Sand Flat, ~700–400 ka) in one of the largest known debris avalanches (~45 km³; Crandell, 1989), early volcanism at Mt. Shasta began about 2 km

south of the present-day summit. Active from ~300–125 ka, the Sargents Ridge cone has an estimated eruptive volume of 85 km³. Initiation of the Misery Hill eruptive phase (60 km³) is marked by 114 ka lavas exposed on the eastern flank of the edifice, but is largely identified by units of 60 to 11 ka age erupted from a vent ~300 m south of the summit. A portion of the Misery Hill cone was destroyed by the plinian eruption of the Red Banks pumice at ~11 ka, followed immediately by a westward shift in volcanism to a satellite vent. Based on geochronology and restricted secular variation in paleomagnetic data (D. Champion and R. Christiansen, unpublished data), the large satellite Shastina cone (20–30 km³) is likely to have formed over several hundred years. Volcanism retreated east in the fourth cone-building stage, Hotlum, producing ~49 km³ of lavas and domes of 10.4 ka age and younger. The total eruptive volume at Mt. Shasta is estimated to be 500 km³ over the last ~300 kyr.

Volcanism at Mt. Shasta is characterized by brief periods of intense activity during which lavas and pyroclastic flows were erupted, followed by apparent intervals of repose. Two-pyroxene andesites (60–63 wt. % SiO₂) constitute nearly 90% of the eruptive products. Basalts (< 52 wt. % SiO₂) and basaltic andesites (52–54 wt. % SiO₂) were dominantly erupted from flank (< 2000 m elevation) and regional vents, whereas dacites (63–66 wt. % SiO₂) were emplaced as flows and domes and are more concentrated near the summit. Many andesites and dacites have low FeO/MgO ratios and, therefore, may have unusually high Mg#s (55–65; where $Mg\# = 100 * Mg / (Mg + Fe)$) at high SiO₂ (see Figure A.1B, C), although Fe/Mg ratios may vary in evolved rocks. Each eruptive phase produced a wide range of large-ion-lithophile element contents (e.g. Sr and Ba), and similar variability is measured for other incompatible elements, including U, Th, and Rb (see Figure A.2). Rare earth element abundances (REE) in the stratocone andesites and dacites are characterized by light REE (LREE) enrichments accompanied by heavy REE (HREE) depletions (see Figure A.3).

Interpretations of previously published major- and trace-element contents, as well as volatile and isotopic compositions vary at Mt. Shasta. Some studies argue for a large subduction component and limited crustal interaction (e.g. Grove et al., 2002; 2005), whereas others infer extensive assimilation of crust (e.g. Streck et al., 2007). High pre-eruptive H₂O contents (4–6 wt. %) are inferred for primitive basaltic andesite (BA) and primitive magnesian andesite (PMA) lavas based on phase relations, experimental petrology, and olivine-hosted melt inclusions (Grove et al., 2002; Le Voyer et al., 2010; Ruscitto et al., 2011). Similarly high volatile contents are inferred for intermediate-composition lavas erupted from the main cone of Mt. Shasta (Grove et al., 2005; 2003). Significant upper crustal assimilation in these stratocone lavas has been dismissed based on low FeO/MgO ratios at high SiO₂ (see Figure A.1, Grove et al., 2005). The range of Li, B, Sr, Nd, and Pb isotope compositions has been taken to be consistent with subduction processes, including subducted sediment, slab-fluid addition, or residue from fluid-fluxing in the mantle wedge (Grove et al., 2002; Le Voyer et al., 2010; Magna et al., 2006; Rose et al., 2001; Ruscitto et al., 2011).

In contrast, Streck et al. (2007) assert that the complex phenocryst assemblage of regional PMA lavas near Mt. Shasta indicates contributions from dacitic, basaltic, and ultramafic components, including those obtained from assimilated continental crust. These authors argue that the high Sr contents and high Sr/Y ratios (Figure A.2E, F) cannot be produced by fractional crystallization of primitive mafic magmas that have relatively low Sr, Sr/Y, and abundant plagioclase as a liquidus phase, and instead require a lower crustal contribution that may have involved melting of amphibole-bearing crust and destabilization of plagioclase, releasing Sr into the melt. This scenario has been proposed elsewhere in the Cascades (Ankney et al. 2013; Conrey et al., 2001), and for other arcs (Ducea and Saleeby, 1998; Garrison and Davidson, 2003).

3. Results

We first report below the ^{238}U - ^{230}Th results from 29 stratocone and regional lavas from the Mt. Shasta region, followed by new Sr isotope data, with comparison to results from previous studies. Details on sample selection and analytical methods are given in Appendix B (supplemental material).

3.1 U-Th isotopes

The regional and stratocone lavas from Mt. Shasta have variable U-Th isotope compositions but fall within the range of those reported for other Cascade lavas (Figure 2, Bennett et al., 1982; Trimble et al., 1984; Newman et al., 1986; Volpe and Hammond, 1991; Cooper and Reid, 2003; Reagan et al., 2003; Berlo et al., 2004; Dosseto et al., 2008; Jicha et al., 2009; Ankney et al., 2013.). Using $^{40}\text{Ar}/^{39}\text{Ar}$, ^{14}C , or stratigraphically estimated ages of lavas (A. Calvert and R. Christiansen, unpublished data), activity ratios of ($^{230}\text{Th}/^{232}\text{Th}$) were corrected for ^{230}Th -ingrowth since the time of eruption. We note that ages of older samples (> 60 ka), and therefore the ($^{230}\text{Th}/^{232}\text{Th}$)₀ of these lavas, are comparatively poorly constrained. As noted below, our major conclusions will be drawn from the compositions of younger rocks that have minimal uncertainties in initial Th isotope compositions.

The majority of lavas measured in this study have ($^{238}\text{U}/^{230}\text{Th}$)₀ < 1, indicating ^{230}Th -excess, although small ^{238}U -excesses (i.e. ($^{238}\text{U}/^{230}\text{Th}$)₀ > 1) were measured in nine lavas (Figure 2b). ^{238}U - ^{230}Th isotope compositions are independent of rock type. Two basaltic andesites (09SH1285 and 09SH1286) have relatively high ^{230}Th -excesses (12.3 and 27.3% respectively), whereas a small ^{238}U -excess is measured in the remaining basaltic andesite (09SH1328, 1.7%). Andesites and dacites are similarly variable in ($^{230}\text{Th}/^{238}\text{U}$)₀ composition; the highest ^{230}Th - and ^{238}U -excesses of the stratocone lavas, for example, are measured in andesites (20.0% ^{230}Th -excess in 07SH1255 and 6.1% ^{238}U -excess in 09SHC42).

Variable ^{230}Th - ^{238}U compositions are also found for eruptive units within specific cone-building stages. Most of the Misery Hill and Shastina lavas have $(^{238}\text{U}/^{230}\text{Th})_0 < 1$, ranging from 0.61–17.9% and 7.9–20.0% ^{230}Th -excess, respectively. Small ^{238}U -excesses are measured in two Misery Hill lavas (0.4–2.1%) and one Shastina andesite (0.9%). Hotlum lavas have the lowest overall ^{230}Th -excesses (up to 5.1%) and, instead, the majority of lavas have $(^{238}\text{U}/^{230}\text{Th})_0 > 1$; ^{238}U -excesses up to 6.1 % are measured, although most lavas have $< 3\%$ ^{238}U -excess. Within both the Misery Hill and Shastina cone-building stages, lavas have relatively constant $(^{230}\text{Th}/^{232}\text{Th})_0$ ratios (1.22–1.26 and 1.16–1.18, respectively), with the exception of two lavas, 07SH1249 (Misery Hill, 1.185) and 07SH1255 (Shastina, 1.262). Hotlum lavas are more variable with respect to $(^{230}\text{Th}/^{232}\text{Th})_0$ ratios (1.17–1.23), and increasing $(^{230}\text{Th}/^{232}\text{Th})_0$ is correlated with increasing $(^{238}\text{U}/^{232}\text{Th})$.

3.2 *Sr Isotopes*

Strontium isotope compositions for the Mt. Shasta lavas analyzed in this study vary from 0.7029–0.7038 (Figure 3), spanning a range comparable to that reported for other Shasta area lavas (0.7029–0.7043; Grove et al., 2002; Zeichert, 2004) and the Cascades (0.7028–0.7043; Bullen and Clynne, 1990; Leeman et al., 1990; Bacon et al., 1997; Borg et al., 1997; Jicha et al., 2009b). Measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are equal to the initial ratios within analytical error. Strontium isotope compositions do not vary systematically with SiO_2 content (basaltic andesites, 0.7032–0.7038; andesites, 0.7031–0.7038; dacites, 0.7029–0.7038). Although similarly broad ranges of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are reported for both Mt. Adams and Lassen Peak (Bullen and Clynne 1990; Bacon et al., 1997; Borg et al., 1997; Jicha et al., 2009b), this variability is primarily associated with basalts and basaltic andesites, and evolved lavas are increasingly restricted to more radiogenic compositions at these volcanoes. At Mt. Shasta, however, andesites and dacites extend toward lower $^{87}\text{Sr}/^{86}\text{Sr}$ ratios than those from other Cascade centers.

Distinction in $^{87}\text{Sr}/^{86}\text{Sr}$ ratios is best observed between eruptive groups (Figure 3). Overall, the Shastina samples have the least radiogenic Sr isotope compositions ($^{87}\text{Sr}/^{86}\text{Sr} \approx 0.7029$) and have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios that nearly overlap with Juan de Fuca and Gorda MORB (0.7023–0.7028; Hegner and Tatsumoto, 1987; White et al., 1987). One Shastina lava has a considerably higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratio (07SH1255; Figure 3B). Notably, this andesite, which is relatively mafic for Shastina lavas, also has a high $(^{230}\text{Th}/^{232}\text{Th})_0 = 1.26$. The Misery Hill samples of this study have a larger range in $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7033–0.7038), but may be split into two groups, those with lower $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7033) and those with more radiogenic compositions ($^{87}\text{Sr}/^{86}\text{Sr} \approx 0.7037$ –0.7038). The Hotlum lavas have the largest range of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and vary continuously from 0.7029 to 0.7036.

4. Discussion

We focus our initial discussion on the U-Th isotope compositions of the stratocone lavas to explore potential mechanisms of ^{230}Th -enrichment at Mt. Shasta, including accessory mineral fractionation and crustal melting. Next, we apply non-modal batch melting and mixing models to U-series and trace-element data to distinguish the relative proportions of crust and mantle components between, and within, eruptive units. Finally, Sr isotopes are integrated with U-Th data to constrain the isotopic compositions of mafic precursor magmas and to identify heterogeneities in the crustal column beneath Mt. Shasta.

4.1 Mechanisms for creating ^{230}Th -excess at Mt. Shasta

4.1.1 ^{230}Th -excess as a consequence of zircon crystallization

^{230}Th -excess in continental arc lavas has been attributed to crystallization of accessory phases with high partition coefficients for U relative to Th ($D_U > D_{\text{Th}}$; e.g., Reagan et al., 2003). Zircon has $D_U \gg$

D_{Th} , therefore, small degrees of crystallization will decrease ($^{238}U/^{232}Th$) at constant ($^{230}Th/^{232}Th$)₀ to create large ^{230}Th -excesses. At Mt. Shasta, fractional crystallization models indicate that subtraction of 0.3% by mass of zircon from a parental magma can produce the largest ^{230}Th excess observed at Shasta (Figure 4; Appendix D.1). Although this may seem a small, and therefore reasonable extent of crystallization, subtraction of this amount of zircon would require a parental magma that initially contained 1609 ppm Zr to produce a fractionated lava that had 116 ppm Zr, the average of that measured for Mt. Shasta andesites. There is no evidence in any Cascade lava for parental magmas with such high Zr contents (Bacon et al., 1997; Borg and Clyne, 1998; Bruggman et al., 1999; Conrey et al., 2001; Grove et al., 2002; Zeichert, 2004). Uranium and Th contents of Mt. Shasta lavas further demonstrate that it is unlikely that zircon fractionation can be invoked as a mechanism for generating the observed range of ^{230}Th -excesses. In all eruptive units, magmas were more extensively depleted in Th relative to U (Figure 5), producing low Th/U ratios that are inconsistent with the distribution coefficients of U and Th in zircon ($D_U = 100$, $D_{Th} = 16.7$, Blundy and Wood, 2003).

We next consider the possibility of zircon crystallization in concert with another accessory mineral, such as apatite, to explain the unexpectedly low Th/U ratios. Unlike zircon, apatite favors Th slightly more than U ($D_U = 1.9$, $D_{Th} = 2.82$, Condomines, 1997), and high degrees of apatite crystallization can generate low Th/U ratios. Assuming subtraction of 0.3% by mass of zircon, fractional crystallization models require more than 25% apatite crystallization to replicate the measured Th/U ratios (yellow model line, Figure 5; Appendix D.2). High proportions of apatite crystallization, however, negate the ^{230}Th -excess produced by zircon crystallization, push the lavas into ^{238}U -excess, and would produce very low P_2O_5 contents. Considering all available data, the range of Zr (25–175 ppm) and P_2O_5 (0.10–0.41 wt. %) contents in the Mt. Shasta lavas (Grove et al., 2005; 2002; Zeichert, 2004) permit less 0.035% zircon and 1.47% apatite crystallization, assuming that all Zr and P_2O_5 is partitioned into zircon and apatite,

respectively. At these proportions, zircon crystallization depletes U contents only minimally and produces less than 1% ^{230}Th -excess. An absence of zircon fractionation is further supported by petrographic observations that failed to identify zircon in Shasta lavas (R. Christiansen, person. commun.). In addition, low zircon saturation temperatures calculated for the stratocone andesites and dacites (697–747 °C; Boehnke et al., 2013), in combination with high estimated liquidus temperatures > 800 °C for the majority of Shasta magmas (Grove et al., 2005), indicate that Shasta magmas were under-saturated in Zr. We conclude, therefore, that the large range in Th/U ratios for Mt. Shasta samples cannot be produced by accessory mineral fractionation. Such a conclusion holds for crystal fractionation of accessory minerals at liquidus temperatures, or separation of magmas from a crystal-rich mush.

We instead interpret Th-U variations in the stratocone lavas to, in part, reflect mixing of evolved magmas with a more mafic component, as will be discussed in the following section. Petrographic evidence in support of mixing includes anomalously mafic pyroxene and olivine grains in dacites, as well as pyroxenes that have low Mg# cores (65–72) which are mantled by high Mg# rims (85–92) (Newman et al., 1986; Streck et al., 2007). Fractional crystallization models and phenocryst mineral compositions identify primitive PMA and BA lavas as possible mafic inputs, and suggest that derivative liquids that were compositionally similar to the McKenzie Butte rhyodacite and an evolved PMA represent end-member compositions that may be invoked in mixing models (Grove et al., 2005).

4.1.2 ^{230}Th -excess generated from cryptic garnet fractionation in the lower crust

The failure of accessory phase fractional crystallization models to explain the ^{230}Th -excesses measured in the intermediate-composition stratocone lavas means that a different explanation must be sought. Most arc lavas are highly differentiated and have mineral assemblages that are the product of shallow, relatively late crystallization processes (Davidson et al., 2007). Here we consider minerals other

than those observed in erupted lavas as a possible control on Th/U ratios, including cryptic mineral fractionation in the mid-to lower-crust (e.g., Lee et al., 2006; Davidson et al., 2007; Hildago and Rooney, 2010). Although an uncommon eruptive product, garnet is observed where the lower crustal roots of arcs are exposed. Deep garnet fractionation imparts distinct chemical signatures on magmas, including HREE depletions, unusually high Sr/Y ratios, moderate Fe/Mg ratios, and, as a result of an affinity for U relative to Th ($D_U = 0.31$ $D_{Th} = 0.074$; Rubatto and Hermann, 2007), significant ^{230}Th -excesses may be produced (Ankney et al., 2013; Hildago and Rooney, 2010; Jicha et al., 2009).

Although garnet is a near-liquidus phase in basaltic to andesitic composition magmas at upper mantle and lower crustal pressures (0.8–1.2 GPa) and temperatures (800–1000°C, Alonso-Perez et al., 2009; Müntener et al., 2001), phase equilibria studies suggest that garnet is unlikely to be on the liquidus in parental Cascade magmas (Mercer and Johnston, 2008). Dehydration melting of mafic amphibolite, however, can produce garnet in the residuum at pressures (1–1.25 GPa) and temperatures (850–1000°C) appropriate for the lower crust beneath the Cascades, as demonstrated by vapor-absent melting experimental studies (e.g., Rapp and Watson, 1995; Sen and Dunn, 1994; Wolf and Wyllie, 1994; 1993). Hornblende and plagioclase breakdown to produce garnet according to the reaction (assuming 100 g of material):

(Eqn. 1) 67.47 hbl + 32.53 plag = 20.0 melt + 80.0 residue (residue = 40.59 gt + 40.59 pyx + 7.92 plag + 10.89 hbl) (950 °C, 10 kbar; Wolf and Wyllie, 1993)

Consequently, dehydration melting provides a mechanism for producing a garnet signature where garnet was not initially present in the crust. Moreover, this reaction may influence the trace-element compositions of magmas in equilibrium with the residue, specifically Sr enrichment due to plagioclase breakdown and high LREE/HREE ratios due to residual garnet.

Garnet and lower crustal melting processes have been found to influence magma chemistry at other Cascade centers, including Crater Lake (Ankney et al., 2013), Mt. Adams (Jicha et al., 2009), and Mt. Jefferson (Conrey et al., 2001). To evaluate the role of garnet in producing ^{230}Th -excess at Mt. Shasta, non-modal batch melting models were applied to both lower crustal and mantle sources according to the methods of Ankney et al. (2013). For lower crustal melting, we selected an amphibolite composition that has a modal mineralogy equivalent to the starting materials in equation 1. A spinel lherzolite was chosen as the mantle source, using a starting composition of 51% olivine, 23% clinopyroxene, 21% orthopyroxene, and 5% spinel. Multiple melting mode (P) values for lower crustal and mantle melts are the same as those used by Ankney et al. (2013). The melting reactions for lower crustal P values were determined from modal compositions in equation 1 and mantle P values were determined from the following equation:

$$(Eqn. 2) \ 10.34 \ sp + 40.54 \ en + 49.12 \ di = 67.23 \ melt + 32.77 \ residue \ (residue = 87 \ fo + 13 \ en) \ (Eqn. 3, \text{normalized to 100 g; Ottonello et al., 1984})$$

Melting models were then applied to approximate the U and Th isotope compositions of melts at various mantle (X_{mantle}^{melt}) and lower crustal ($X_{lower \ crust}^{melt}$) melt fractions (Appendix D.3; Table A.2).

Progressive dehydration melting of amphibolite produces increasing amounts of garnet until amphibole is exhausted and the residual garnet and pyroxene begin to destabilize (Eqn. 1; Wolf and Wyllie, 1994; 1993). Accordingly, calculations involving non-modal batch melting models produce high ^{230}Th -excesses at high percent amphibolite melt, up to 34% at $X_{lower \ crust}^{melt} = 0.2$ (Figure 6), where large quantities of garnet are produced. Garnet is not generated by melting spinel lherzolite (Eqn. 2). Batch melting models, however, demonstrate that partial mantle melting may also produce melts that are moderately enriched in ^{230}Th , although the largest excesses are calculated for the smallest melt fractions

(13% ^{230}Th -excess at $X_{\text{mantle}}^{\text{melt}} = 0.01$). It is important to note that the U/Th ratios of modeled mantle melts are not significantly affected by initial mantle composition, given the low partition coefficients of U and Th in ultramafic systems. For example, partial melting of a garnet lherzolite mantle creates ^{230}Th -excesses only slightly higher than that produced by a spinel lherzolite composition (18.2% ^{230}Th -excess at $X_{\text{mantle}}^{\text{melt}} = 0.01$, Ankney et al., 2013).

We combine the effects of mantle and lower crustal melting at various melt fractions through mixing in varying proportions to account for the range of ^{238}U - ^{230}Th compositions measured in the Mt. Shasta stratocone lavas. The extent of ^{230}Th -excess at Mt. Shasta is best approximated by a mixture of mantle and crustal melts at $X_{\text{mantle}}^{\text{melt}} = X_{\text{lower crust}}^{\text{melt}} = 0.15$ (Figure 6), although it is important to note that a variety of melt fractions for the two components can explain the range in ^{230}Th -excess at Mt. Shasta. Broadly, however, relatively high mantle and lower crustal melt fractions are consistent with experimental data for primitive basalts (Grove et al., 2002) and expectations for extensive lower crustal melting in the Cascades (Conrey et al., 2001; Elkins-Tanton et al., 2001).

4.2 ^{230}Th -excess-Ce/Y relations and deep crustal processes

The range of ^{230}Th -excess between eruptive units at Mt. Shasta may be explained by mixing variable crustal melt fractions to a mantle-derived melt (Figure 6), although, in detail, a third component is required. Assimilation of “old” bulk crust that lies along the equiline, in addition to young, garnet-bearing crust, is a more likely explanation for lavas that have moderate ^{230}Th -excesses (lie close to the equiline) than very low percent (< 1%) crustal melting, given the difficulty of extracting small melt fractions (e.g. Bourdon et al., 2000; Price et al., 2007; 2010). We suggest that non-garnet-bearing, mid- to lower-crust was assimilated by lavas that plot near the equiline, but at ($^{238}\text{U}/^{232}\text{Th}$) ratios that were significantly different from those of the mantle. In this scenario, relatively constant mantle and lower

crustal melt fractions mix to produce magmas represented by individual eruptive units that have similar initial ($^{230}\text{Th}/^{232}\text{Th}$) and high ^{230}Th -excesses. These magmas would subsequently undergo different degrees of assimilation of crust that lies along the equiline during ascent, increasing their ($^{238}\text{U}/^{232}\text{Th}$) ratios while maintaining relatively constant ($^{230}\text{Th}/^{232}\text{Th}$) (Figure 6). Mixing relations between mantle melts, garnet-bearing young crust (high ($^{230}\text{Th}/^{238}\text{U}$)₀), and old bulk crust (equiline) with variable U/Th ratios are illustrated in Figure A.4, where inversion of the U/Th ratio has been done to produce linear mixing lines. The majority of Mt. Shasta lavas lie within a three component mixing triangle tied down by a low ($^{232}\text{Th}/^{238}\text{U}$) bulk crustal component on the equiline, although some variability in bulk crust ($^{232}\text{Th}/^{238}\text{U}$) is expected.

Trace elements that are sensitive to garnet fractionation, such as Ce and Y, provide a test of our model that ^{230}Th -excess reflects garnet production during dehydration melting (Figure 7; A.5; Table A.2 Appendix D.4). As garnet production increases with the progressive breakdown of plagioclase and hornblende, lower crustal melts will be simultaneously enriched in ^{230}Th , Ce, but depleted in Y and the HREEs (in garnet $D_{\text{Ce}} = 0.016$, $D_{\text{Y}} = 5.27$, Qian and Hermann et al., 2013). The high Ce/Y ratios that identify a role for garnet are consistent with the high Sr/Y and Dy/Yb ratios of the Mt. Shasta lavas (Figure A.6; A.7), although Sr/Y ratios tend to be more scattered than Ce/Y ratios due to late-stage plagioclase fractionation.

Overall, high Ce/Y ratios between, but especially within, individual eruptive units are correlated with ^{230}Th -excess. Most lavas, however, have a lower ($^{230}\text{Th}/^{238}\text{U}$)₀ than that predicted by melting-mixing models. This may in part reflect magma residence time, as demonstrated by the decay curves in Figure 7. A residence time of 100 kyr is required to reproduce the ($^{230}\text{Th}/^{238}\text{U}$)₀ of the Misery Hill lavas, and ~125 kyr is predicted for the Shastina dacites. This is contrary to the short residence time implied by the

comparatively low phenocryst content of the Shastina lavas (Grove et al., 2005), as well as the rapid eruption rate of the Shastina stage. As an alternative to long residence times, we suggest that assimilation of bulk crust that is old enough to lie on the equiline (i.e. $(^{230}\text{Th}/^{238}\text{U})_0 = 1$) is responsible for the more modest $(^{230}\text{Th}/^{238}\text{U})_0$ of the lavas with respect to Ce/Y. This model implies a greater bulk crustal (equiline) influence in the Misery Hill and Hotlum units, whereas Shastina lavas would contain a smaller bulk crustal (equiline) component. Petrographic evidence in support of assimilation includes granitic, metasedimentary, and peridotite (assumed to be derived from the Trinity Ophiolite) xenoliths included in all eruptive units (Newman et al., 1986; Grove et al. (2005).

It is important to note that, in this combined melting-mixing-assimilation model, the $(^{230}\text{Th}/^{238}\text{U})_0$ ratios calculated for the modeled mixture, and therefore the predicted fractions of bulk crustal assimilation, are highly dependent on the distribution coefficients of U and Th in garnet. Accordingly, both the process (e.g., mixing or assimilation fractional crystallization) and extent of bulk crustal assimilation cannot be robustly quantified without independent constraints (Figures 7, A.5). Instead, this model demonstrates that lower $(^{230}\text{Th}/^{238}\text{U})_0$ ratios relative to those predicted by models of garnet production during lower crustal melting likely result from a combination of variable deep crustal residence times and bulk crustal assimilation.

4.3 Constraints on the proportions of mantle and crustal component

Although there is a wide range of $(^{238}\text{U}/^{232}\text{Th})$ ratios for individual units, the $(^{230}\text{Th}/^{232}\text{Th})_0$ ratios of different cone-building stages are relatively constant (Figure 6). Early U-series studies of the Cascades observed that young (< 10 ka) lavas from the same volcano maintained constant $(^{230}\text{Th}/^{232}\text{Th})_0$ over a range of $(^{238}\text{U}/^{232}\text{Th})$, and consequently hypothesized that the reservoirs feeding each volcano must be homogeneous with respect to Th-isotope compositions (Newman et al., 1986). Our high-precision data

display similar trends at Mt. Shasta over a range of ages, suggesting that Th isotope homogeneity exists on the scale of individual cone-building stages. We interpret the homogeneous $(^{230}\text{Th}/^{232}\text{Th})_0$ of individual cone-building stages, which produce distinct horizontal arrays on an equiline diagram (Figure 6), to correspond to distinct mixing proportions of mantle and lower crustal (high ^{230}Th -excess) melts that are apparently characteristic of specific cone-building stages. This in turn suggests that the nature of magmatic reservoirs that were tapped by successive cone-building stages was fundamentally determined in the lower crust. Below we explore mixing proportions of mantle and lower crustal melts ($M_{m:lc}$) at $X_{mantle}^{melt} = X_{lower\ crust}^{melt} = 0.15$.

Overall, the Misery Hill and Shastina units have the highest and lowest proportion of lower crustal melt, respectively. The higher $(^{230}\text{Th}/^{232}\text{Th})_0$ ratios of the Misery Hill andesites and dacites suggest $M_{m:lc} = \sim 85:15$, whereas the lower $(^{230}\text{Th}/^{232}\text{Th})_0$ ratios measured in the Shastina lavas imply a more mantle-rich mixture of $\sim 95:5$. Notably, some units have exceptions. The high $(^{230}\text{Th}/^{232}\text{Th})_0$ of one Shastina andesite (07SH1255) implies a more dominant lower crustal component, close to but slightly higher than that suggested for the Misery Hill lavas, whereas the youngest Misery Hill dacite measured (07SH1249) has a comparatively low $(^{230}\text{Th}/^{232}\text{Th})_0$, indicating a more dominant mantle component, similar to that interpreted for the Shastina lavas. It is difficult to evaluate if these two lavas record anomalies in mantle and lower crustal melt proportions or unusual $(^{230}\text{Th}/^{238}\text{U})_0$ ratios in the source region.

Hotlum andesites and dacites have a wider range of $(^{230}\text{Th}/^{232}\text{Th})_0$ compositions than those from the Misery Hill or Shastina units. Samples that have high ^{230}Th -excess have low $(^{230}\text{Th}/^{232}\text{Th})_0$, similar to that measured for the Shastina dacites, but become more radiogenic as $(^{238}\text{U}/^{232}\text{Th})$ increases, to ^{238}U -excesses of up to 6%. Variation in the $(^{230}\text{Th}/^{232}\text{Th})_0$ ratios may be interpreted as a consequence of

shifting mantle and lower crustal melt mixing proportions in the Hotlum parent lavas from $M_{m:lc} = \sim 95:5$ to $M_{m:lc} = \sim 85:15$. This explanation, however, fails to account for the apparent correlation of $(^{230}\text{Th}/^{232}\text{Th})_0$ with increasing $(^{238}\text{U}/^{232}\text{Th})$. ^{238}U -excesses are rare in the Cascades, but at other Cascade volcanoes that contain ^{230}Th -excess, ^{238}U -excess is attributed to assimilation of hydrothermally altered upper crust, such as at Crater Lake (Ankney et al., 2013). Although small ^{238}U -excesses are found in two Misery Hill samples ($< 2\%$), the highest ^{238}U -excesses are measured in the Hotlum cone-building lavas, by which time a hydrothermal system at Mt. Shasta might have been well developed. An absence of ^{238}U -excess in the Shastina lavas might therefore be explained by relatively rapid transport from deep crustal levels, minimizing opportunity for upper crustal assimilation.

An important observation is the positive correlation between eruptive volume and $(^{230}\text{Th}/^{232}\text{Th})_0$, and hence proportion of lower crustal melt relative to mantle melt during mixing. The largest cone-building stage, Misery Hill (60 km^3) has the largest proportion of lower crustal melt, and we interpret this to reflect a relatively large lower crustal magmatic footprint that in turn generated relatively large amounts of amphibole dehydration melting. Similarly, the intermediate-volume Hotlum stage (49 km^3) has intermediate $(^{230}\text{Th}/^{232}\text{Th})_0$ ratios, suggesting a comparatively smaller lower crustal melt component. Both Misery Hill and Hotlum lavas have $(^{230}\text{Th}/^{238}\text{U})_0$ that extend to the equiline, indicating additional crustal interaction in the absence of garnet. In contrast, the smallest volume eruptive stage, Shastina ($20\text{--}30 \text{ km}^3$), which erupted very quickly, has the lowest $(^{230}\text{Th}/^{232}\text{Th})_0$ ratios, suggesting the greatest mantle contribution. Moreover, Shastina lavas are dominated by ^{230}Th -excess, indicating minimal interaction with non-garnet-bearing crust, in contrast to those of Hotlum and Misery Hill.

The more primitive regional lavas have the most variable $(^{230}\text{Th}/^{232}\text{Th})_0$ ratios measured at Mt. Shasta, ranging from 1.09 to 1.26. Like Hotlum, $(^{230}\text{Th}/^{232}\text{Th})_0$ ratios appear to increase with decreasing

($^{230}\text{Th}/^{238}\text{U}$)₀. This variability, however, is unlikely to be a consequence of assimilation of hydrothermally altered upper crust. As noted in section 3.1, ages of these older samples (> 60 ka) are comparatively poorly constrained, yielding larger errors for correction for ^{230}Th ingrowth or decay after eruption relative to the younger cone-building stages discussed above. Consequently, we have focused interpretations on the ^{238}U - ^{230}Th compositions of only the more evolved andesites and dacites from the three youngest eruptive stages.

4.4 Sr-Th relations and injection of precursor magmas

Strontium isotopes can be used to further constrain the proportion of mantle and lower crustal melt components, defined in section 4.3 above as $M_{m:lc}$. Although scatter is present, ($^{230}\text{Th}/^{232}\text{Th}$)₀ is positively correlated with $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (Figure 8A). The low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the Shastina andesite and dacites (0.7029), and the more radiogenic compositions measured for Misery Hill lavas (up to 0.7038), are consistent with interpretations that increasing ($^{230}\text{Th}/^{232}\text{Th}$)₀ reflects a larger proportion of lower crustal melt in the Misery Hill unit. The intermediate ($^{230}\text{Th}/^{232}\text{Th}$)₀ ratios in Hotlum lavas, indicating an intermediate amount of lower crustal melt relative to Shastina and Misery Hill, are compatible with the more moderate $^{87}\text{Sr}/^{86}\text{Sr}$ ratios measured for this unit.

Although the Th-Sr relations paint a broadly consistent picture, there is significant Sr isotope variation within individual eruptive units. For example, despite nearly uniform ($^{230}\text{Th}/^{232}\text{Th}$)₀, Misery Hill lavas have a range of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios that is nearly half of that measured for all samples in this study. High $^{87}\text{Sr}/^{86}\text{Sr}$ ratios distinguish Misery Hill lavas that have high ^{230}Th -excess from those on or near the equiline (0.7037 and 0.7033 respectively; Figure 8). Consequently, differential assimilation of bulk crust might be expected to account for variable Sr isotope compositions. Relatively high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, however, have been measured in primitive basalts (up to 0.7045; Figure 3), indicating that Sr isotope

heterogeneity may also reflect variability in the mantle-derived magmas. It is conceivable then that assimilation of deep, non-garnet bearing crust, responsible for decreasing $(^{230}\text{Th}/^{238}\text{U})_0$ within eruptive groups, could also produce a decrease in $^{87}\text{Sr}/^{86}\text{Sr}$ ratios.

The same range of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios is observed in samples with and without ^{230}Th -excess (Figure 8B). As such, Sr isotope variability may signify partial mantle melting processes where primary magmas that dominantly had low, MORB-like $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (Juan de Fuca and Gorda MORB: 0.7023–0.7028, Hegner and Tatsumoto, 1987; White et al., 1987) ascended through and assimilated portions of a crustal column that was heterogeneous with respect to Sr isotopes. Existing data suggest a geographic control on Sr isotope crustal composition (Figure 9). Lavas that have low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were erupted from vents in the northwest quadrant of the volcano, whereas more radiogenic compositions are measured in samples from the south and east. Variable Sr isotope compositions are imparted on eruptive units that sample both crustal compositions (e.g. Sargents Ridge, Misery Hill, and Hotlum). We note that this geographic variability is evident in even the oldest eruptive deposits and is not strongly correlated with U-Th isotope compositions. Accordingly, Sr isotope heterogeneity in the crust likely pre-dated the ancestral Mt. Shasta cone (Sand Flat, 700–400 ka) and may reflect the compositions of the mid- to upper-crust.

5. Model of the Mt. Shasta magmatic system

Figure 10 shows a schematic model of the magmatic processes consistent with the elemental and isotopic variations that may have produced the Mt. Shasta stratocone andesites and dacites. Primitive mantle melts ($X_{\text{mantle}}^{\text{melt}} \approx 0.15$) that contain slight ^{230}Th -excesses ($< 1\%$) and dominantly MORB-like $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are envisioned to have ascended through the mantle wedge and accumulated at the base of the lower crust to create a deep crustal hot zone (e.g., Annen et al., 2006). The high primary

temperatures of mantle-derived magmas inferred to be parental to the Cascades (up to 1300 °C, Elkins-Tanton et al., 2001) should have promoted dehydration melting of an amphibolite lower crust, accompanied by amphibole and plagioclase break down to produce garnet in the residuum (equation 1), generating hydrous melts that had large ^{230}Th -excesses (up to 34%), high Sr/Y and Ce/Y ratios, and HREE depletions (Table A.2). These partial lower crustal melts ($X_{\text{lower crust}}^{\text{melt}} \approx 0.15$) subsequently mixed with mantle-derived magmas in distinct proportions that, on average, created magmas that contained high initial ^{230}Th -excesses (14–30%) and $(^{230}\text{Th}/^{232}\text{Th})_0$ ratios which are characteristic of individual eruptive stages. The magmatic products of this mixing event may have been stored in the deep crust for up to 10^5 years (e.g. Annen et al., 2006), during which assimilation of surrounding deep non-garnet-bearing (equiline) crust may have also occurred, although the U-Th isotope data do not uniquely constrain magma residence time. Both ageing and bulk assimilation processes would act to decrease the initial ^{230}Th -excess of the magmas.

Upon segregation from the lower crust, buoyant mafic magmas ascended a mid- to upper-crustal column that is laterally heterogeneous with respect to Sr isotope compositions. Crust in the northwest quadrant of the Mt. Shasta volcanic field has less radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7029–0.7033) than that in the south and east (0.7034–0.7043). Fractional crystallization, mixing of variably evolved derivative magmas, and crustal assimilation is envisioned to have subsequently occurred over a range of depths to produce stratocone lavas which have variable $(^{230}\text{Th}/^{238}\text{U})$ and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios.

6. Conclusions

Continental arc lavas are often described as the products of flux-melting in the mantle-wedge and differentiation processes in the upper crust. Interactions between mantle-derived melts and deep continental crust, however, may also influence the chemical and volatile compositions of magmas,

although this contribution is difficult to identify where the deep crust is compositionally similar to parental arc magmas. At Mt. Shasta, previous models for magma generation and evolution of the stratocone lavas have dismissed significant crustal interaction, and instead attribute incompatible element enrichments and isotopic variability entirely to a subduction component (Grove et al., 2005). The new U-Th and Sr isotope measurements of this study, combined with the existing chemical and isotopic database from the Mt. Shasta volcanic system indicate extensive interaction in both lower- and upper-crustal regions.

A paucity of ^{238}U -excess and low, MORB-like $^{87}\text{Sr}/^{86}\text{Sr}$ ratios argue against contributions from the subducting slab as the sole source for many trace elements. Instead, we ascribe the large ^{230}Th -excesses (up to 20%) and high Ce/Y and Sr/Y ratios of the stratocone andesites and dacites to assimilation of partial melts from a mafic amphibolite lower crust where garnet was produced in the residuum. Dehydration melting processes in the deep crust may also contribute a portion of the volatiles, as sampled by olivine-hosted melt inclusions, and fluid mobile elements (Grove et al., 2003; 2002; Le Voyer et al., 2010; Ruscitto et al., 2011). A range of ^{230}Th -excess within and between eruptive units reflects variable residence time and variable assimilation of bulk, non-garnet-bearing crust. Distinct $(^{230}\text{Th}/^{232}\text{Th})_0$ ratios are found in individual cone-building stages, indicating that magmas formed from mixing variable proportions of mantle and lower crustal melts. The greatest lower crustal component is found in the largest eruptive stage considered in this study, Misery Hill, presumably reflecting a large magmatic footprint in the lower crust relative to that of the smaller Hotlum and Shastina stages. Although it seems likely that subduction components contributed to the chemical characteristics of primitive Cascade lavas, our results suggest that lower crustal interaction can produce compositions that are similar to those ascribed to subduction. Our focus on the cone-building stages at Shasta indicates a particularly important crustal history for these evolved magmas. Variability in $^{87}\text{Sr}/^{86}\text{Sr}$ - $(^{230}\text{Th}/^{238}\text{U})_0$

relations suggests that variable Sr isotope compositions in the Mt. Shasta lavas are unlikely to be exclusively inherited from magma generation processes in the mantle and may be a consequence of assimilation of crust that is laterally variable in isotopic compositions.

If future work confirms that ^{230}Th -excess is a characteristic feature of Cascade lavas, we suggest that interaction with the lower crust has a greater effect on magma chemistry throughout the Cascades than previously thought. Moreover, the comparatively dry nature of this arc may offer a unique opportunity to understand the influence of the lower crust in continental arc settings. In other young arcs, it is also likely that the deep crust exerts some control on magma composition, but a large subduction component may mask any lower crustal signature and produce lavas with ^{238}U -excess.

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Figure Captions

Figure 1. Simplified geologic map of the Mount Shasta area, showing the four-cone building stages comprising the current edifice. Inset shows the Cascade arc.

Figure 2. Mt. Shasta and Cascade lavas plotted on a traditional U-Th equiline diagram. Mount Shasta data from Newman et al. (1986; diamonds) are plotted with data from Crater Lake (green field) and Mt. Adams (pink field) in (A). Data from this study (squares) are plotted with that of Newman et al. (1986; blue field) in (B). Average error for the Newman et al. (1986) and this study are shown in (A) and (B) respectively. Samples plotting to the left of the equiline have $(^{230}\text{Th}/^{238}\text{U}) > 1$ and are considered to be in ^{230}Th -excess; those plotting to the right of the equiline have $(^{230}\text{Th}/^{238}\text{U}) < 1$ and are in ^{238}U -excess. Cascade lavas are dominated by ^{230}Th -excess. Rock type abbreviations are as follows: basaltic andesite (BA), andesite (A), dacite (D). See Appendix C for complete list of data sources.

Figure 3. $^{87}\text{Sr}/^{86}\text{Sr}$ relative to SiO_2 for Mt. Shasta and Cascades lavas, including Crater Lake (green field), Mt. Adams (pink field), and Mt. Lassen (light grey field). Data from the Juan de Fuca and Gorda MORB are shown as a dark gray field. Mount Shasta data from previous studies are outlined with a dotted line in (B). “Regional lavas” refer to primitive lavas erupted from flank vents and nearby cones. Mount Shasta data from Grove et al. (2002) (circles); Zeichert (2004) (triangles); this study (squares). All errors are smaller than the data symbols. Refer to Appendix C for a complete list of data sources.

Figure 4. Equiline diagram with fractional crystallization models illustrating the effects of zircon and apatite crystallization on ^{230}Th - ^{238}U compositions. Small degrees of zircon crystallization can create large ^{230}Th -excesses (up to 28.5% at 0.3% crystallization) but higher degrees of apatite crystallization are required to generate even moderate ^{238}U -excesses (up to 10.2% ^{238}U -excess at 10% crystallization). Subtraction of this extent of zircon and apatite, however, requires parental magmas with 1609 ppm Zr and 3.0 wt. % P_2O_5 in order to produce lavas with 116 ppm Zr and 0.24 wt. % P_2O_5 , the average of that measured in Mt. Shasta andesites. Individual fractional crystallization models for zircon and apatite are based on Rayleigh fractional crystallization. Initial U and Th contents for fractional crystallization models are based on the average composition of basaltic andesite at Mt. Shasta and are estimated to be 0.77 and 1.94 ppm, respectively. Refer to section 4.1.1 for U and Th partition coefficients in zircon and apatite. We note that the amount of ^{230}Th - and ^{238}U -excess created at a given percent zircon or apatite crystallization is dependent on the choice of U and Th partition coefficients, although reasonable amounts of crystallization are only achieved at very high U and Th partition coefficients, respectively. See Appendix D.1 for a more complete discussion of model parameters.

Figure 5. U and Th contents demonstrate that Mt. Shasta lavas are depleted in Th relative to U (inset), even in the Shastina unit where Zr depletion with increasing SiO₂ suggests possible zircon crystallization. Solid black and grey trendlines represent quantitative linear regressions of the Shastina and Mt. Shasta data respectively. Dotted lines illustrate the effects of variable proportions of zircon and apatite crystallization from a parental magma on U and Th contents: 0.3% zircon and 0% apatite crystallization (blue), 0.3% zircon and 15% apatite crystallization (green), and 0.3% zircon and 25% apatite crystallization (yellow). Fractional crystallization models are based on Rayleigh fractional crystallization, and, given similarly low partition coefficients for U and Th in major phases, crystallization models are simplified to assume that all U and Th is partitioned into zircon and apatite. Distribution coefficients and starting compositions are as in Figure 4. The amount of apatite crystallization required to reproduce Th/U trends of the Mt. Shasta lavas at 0.3% zircon crystallization is dependent on the Th/U ratio of the starting composition and the partition coefficients of U and Th in zircon and apatite. Reasonable amounts of apatite crystallization, however, are only achieved in magmas with very high initial Th/U ratios or by selecting unusually high partition coefficients for Th and U in apatite. Refer to Appendix D.2 for a more complete discussion of model parameters. Mount Shasta data from Grove et al. (2002) (circles); Zeichert (2004) (triangles); this study (squares).

Figure 6. Equiline diagram with non-modal batch melting models for a mafic amphibolite lower crust and a spinel lherzolite mantle. Partial melts of both components are able to generate ²³⁰Th-excess. Melting models and estimated positions of mantle and crustal components are based on Ankney et al. (2013). Initial U and Th contents of the mantle and lower crustal are estimated to be 0.05 and 0.145 ppm and 1 and 2.33 ppm, respectively. Both the mantle and lower crust are assumed to be older than 300 kyr and therefore lie on the equiline. The lower crustal component is constrained by Crater Lake data which converge on (²³⁸U/²³²Th) = 1.30 (Ankney et al., 2013) and is taken to be the composition of the bulk crust. U-series data from Mt. Adams basalts (Jicha et al., 2009) constrain the position of the mantle component at (²³⁸U/²³²Th) = (²³⁰Th/²³²Th)₀ = 1.05. Solid lines denote the U-Th isotope composition of mantle and lower crustal melts at melt fractions (X_{mantle}^{melt} and $X_{lower\ crust}^{melt}$, respectively) ranging from 0.01 to 0.2. The dashed line shows mixing between mantle and lower crustal melts at $X_{mantle}^{melt} = X_{lower\ crust}^{melt} = 0.15$. Provided that garnet is the only phase in our model capable of generating significant ²³⁰Th-excesses (> 10%) at moderate to high melt fractions (0.05 < $X^{melt} \leq 0.20$), the extent of ²³⁰Th-excess produced during melting is largely dependent on the choice of partition coefficient for U and Th in garnet. Refer to Appendix D.3 for a more complete discussion of model parameters. Diamond symbols denote lavas with significant ²³⁰Th-excess, while samples on/near the equiline are shown as squares. The red shaded field represents the ²³⁸U-²³⁰Th composition of melts dominated by interaction with garnet-bearing lower crustal rocks; blue shading shows the ²³⁸U-²³⁰Th composition of melts dominated by interaction with non-garnet-bearing bulk crust.

Figure 7. Variation of (²³⁰Th/²³⁸U)₀ with respect to Ce/Y. A modeled mixture of 15% mantle and 15% lower crustal melts (solid black line; dashed line in Figure 6) predicts high ²³⁰Th-excess at high Ce/Y as a function of garnet production during dehydration melting of the lower crust. A combination of crustal residence times and assimilation of non-garnet-bearing bulk crust may explain the deviation of the Mt. Shasta samples from the modeled mixture (dashed lines). Residence times of 25 kyr (A) and 50 kyr (B) are shown to illustrate how variable residence time affects the amount of bulk crustal assimilation (from

20–80% bulk crustal assimilation or B.C.A.) required to reproduce the $(^{230}\text{Th}/^{238}\text{U})_0$ of the lavas at a given Ce/Y ratio. Notably, one regional lava (09SH1286) has a higher $(^{230}\text{Th}/^{238}\text{U})_0$ than predicted by the modeled mixture. We attribute this, in part, to poor age constraints effecting the eruptive age correction on the $(^{230}\text{Th}/^{238}\text{U})_0$ ratio (Table A.1). Initial U and Th contents and isotopic compositions of the mantle and lower crustal end members are as in Figure 6; initial Ce and Y contents are estimated to be 1.785 and 4.37 and 16 ppm, respectively. A bulk crust with a starting composition of 43 ppm Ce, 19 ppm Y, 1 ppm U, 2.3 ppm Th, and a $(^{230}\text{Th}/^{238}\text{U})$ ratio of 1 was selected for the assimilant. We note that increasing the Ce/Y ratio of the bulk crust decrease predictions of bulk crustal assimilation for the same length of magma storage (25–50 kyr), such that the extent of assimilation cannot be robustly quantified without independent constraints. See appendix D.4 for a more complete discussion of model parameters. Diamond symbols denote lavas with significant ^{230}Th -excess, while samples on/near the equiline are shown as squares. The red shaded field represents the ^{238}U - ^{230}Th of melts dominated by interaction with garnet-bearing lower crustal rocks; blue shading shows the ^{238}U - ^{230}Th composition of melts dominated by interaction with non-garnet bearing bulk crust.

Figure 8. $(^{230}\text{Th}/^{232}\text{Th})_0$ (A, C) and ^{230}Th -excess (B, D) vs. $^{87}\text{Sr}/^{86}\text{Sr}$ at Mt. Shasta. Relatively uniform $(^{230}\text{Th}/^{232}\text{Th})_0$ ratios in individual cone-building stages reflect distinct mixing proportions of mantle and lower crustal melts ($M_{m:lc}$), therefore Sr isotope variability within eruptive groups cannot be attributed to fundamental differences in mantle and lower crustal components. Similarly, the same range of $^{87}\text{Sr}/^{86}\text{Sr}$ is observed in samples with and without ^{230}Th -excess, such that the total amount of non-garnet-bearing crustal assimilation (which decreases $(^{230}\text{Th}/^{238}\text{U})_0$ within eruptive units) cannot be solely responsible for producing variable Sr isotope compositions. Instead, Sr isotope variability may be a function of a mixture of 15% mantle and 15% lower crustal melts (solid line) ascending through and assimilating portions of an old (> 300 kyr) crustal column that is heterogeneous with respect to Sr isotopes. Bulk crustal assimilation (“BCA”; 20–80%; colored dashed lines) is modeled for two crustal end members that have the same $(^{230}\text{Th}/^{232}\text{Th})$ and $(^{238}\text{U}/^{232}\text{Th})$ but different $^{87}\text{Sr}/^{86}\text{Sr}$ ratios: BCA 1 $^{87}\text{Sr}/^{86}\text{Sr} = 0.7035$ (dotted blue lines, based on Klamath Mountain plutons; Barnes et al., 1992) and BCA 2 $^{87}\text{Sr}/^{86}\text{Sr} = 0.7065$ (dotted red lines, estimated from the Trinity Ophiolite; Brouxel and Lapierre, 1988). Residence times of 25 kyr (A, B) and 50 kyr (C, D) are shown to illustrate how variable residence time affects the amount of assimilation required. Initial U and Th contents and isotopic compositions of the mantle and lower crustal end members are as in Figure 6; Initial Sr contents and isotopic compositions are estimated to be 20.3 ppm and 0.7025 and 925 ppm and 0.7028, respectively. Initial U and Th contents of the bulk crustal assimilant are as in Figure 7; initial Sr contents are estimated to be 320 ppm for both BCA 1 and 2. See Appendix D.5 for a more complete discussion of model parameters. Diamond symbols denote lavas with significant ^{230}Th -excesses, while samples on/near the equiline are shown as squares. The red shaded field represents the ^{238}U - ^{230}Th of melts dominated by interaction with garnet-bearing lower crustal rocks; blue shading shows the ^{238}U - ^{230}Th composition of melts dominated by interaction with non-garnet bearing bulk crust.

Figure 9. Simplified geologic map of Mt. Shasta with lava and pyroclastic flow deposits. . Symbol colors show Sr isotope compositions of vents: Blue ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7028$ – 0.7033), red ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7034$ – 0.7043); mixed colors signify vents where at least one outlier was erupted. Symbol shapes indicate age of erupted lavas: West Haight Mountain (10 point star), regional basalts (six point star), Goosenest

(hexagon), Sand Flat (four point star), Sargents Ridge (circle), Misery Hill (diamond), Shastina (triangle), Hotlum (square). Data from Grove et al. (2002), Zeichert (2004), and this study.

Figure 10. Simplified illustration (not to scale) of expected magma generation and evolution processes at Mt. Shasta. (1) Partial melts of spinel lherzolite mantle stalled and accumulated at the base of the lower crust to create a lower crustal hot zone (e.g. Annen et al., 2006) and promote (2) dehydration melting and garnet production in an amphibolite lower crust. Hydrous lower crustal melts subsequently mixed with primitive mantle-derived melts to impart ^{230}Th -excesses and distinct $(^{230}\text{Th}/^{232}\text{Th})_0$ ratios on magmas. The magmatic products of this mixing event may have been stored in the lower crust for as long as 10^5 years (e.g. Annen et al., 2006), during which time fractional crystallization and assimilation of deep non-garnet-bearing crust may also have occurred. (3) Buoyant mafic melts then ascended a mid-to-upper crustal column that was heterogeneous with respect to Sr isotopes (indicated by gradient; tan shading denotes crust with less radiogenic compositions (e.g. BCA 1; $^{87}\text{Sr}/^{86}\text{Sr} = 0.7035$), and light grey shading indicates more crust with more radiogenic compositions (e.g. BCA 2 $^{87}\text{Sr}/^{86}\text{Sr} = 0.7065$). Fractional crystallization, magma mixing, and crustal assimilation occur over a range of depths to produce stratocone andesites and dacites that had variable $^{87}\text{Sr}/^{86}\text{Sr}$ and $(^{230}\text{Th}/^{238}\text{U})_0$ compositions.

Figure 9

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9

Figure 10