Petrology of "Mt. Shasta" high-magnesian andesite (HMA): A product of multi-stage crustal assembly

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ABSTRACT

Occurrences of high-Mg andesite (HMA) in modern volcanic arcs raise the possibility that significant volumes of continental crust could be directly derived from Earth's mantle. Such rocks are commonly associated with subduction of young, warm oceanic lithosphere or occur in areas heated by mantle convection. A relatively rare occurrence near Mt. Shasta in the Cascades volcanic arc has been considered to represent one such primary mantle-derived magma type, from which more evolved andesitic and dacitic magmas are derived. Recognition that the Shasta area HMA is actually a hybrid mixed magma, calls into question this notion as well as the criteria upon which it is based. We report new chemical and mineralogical data for samples of the Shasta HMA that bear on the components and processes involved in its formation. Several generations of pyroxenes and olivines are present along with different generations of oxide minerals and melt inclusions. The most magnesian olivines (Fo₉₃) exhibit disequilibria textures, exotic melt inclusions, and reaction rims of Fo₈₇ composition; these crystals along with spongy, ~Mg# 87 orthopyroxene crystals are interpreted to be xenocrystic and do not signify a primitive mantle derivation. The groundmass is andesitic with moderate MgO content, and melt inclusions of similar compositions are hosted by equilibrium olivine ($\sim Fo_{87}$). The bulk magma (whole rock) is more magnesian, but primarily due to incorporation of mafic minerals and ultramafic xenolith debris. We propose that the exotic crystal and lithic debris in these rocks is derived from (1) dacitic magmas of possible crustal derivation, (2) prograded ultramafic rocks in the underlying crust, and (3) random lithic debris and crystals derived from conduit wall rocks and earlier intruded magmas within the feeder plexus beneath Shasta. The HMA is inferred to represent a mixture between evolved dacitic and primitive basaltic magmas as well as incorporation of xenolithic crystal cargo. There is no compelling evidence that HMA is present in large volumes, and it is not considered to be an important parental liquid to more evolved magmas at Shasta.

Keywords: High-Mg andesite, Mt. Shasta, Cascades arc, mineral zoning and populations; New Advances in Subduction Zone Magma Genesis

INTRODUCTION

This paper presents a detailed description of high-Mg andesites (HMA) associated with Mt. Shasta, a prominent quaternary composite volcano in the northern California Cascade Range. The interpretation of HMAs is a significant fulcrum in evaluating subduction-related magmatic processes and carries important implications for understanding the formation of continental crust in general. Most estimates of average crustal composition are essentially andesitic, so a fundamental issue regards the extent to which crust-forming materials are directly derived from the mantle. The occurrence of adakitic magmas in many modern volcanic arcs may be considered to suggest that this is possible. However, we present compelling evidence that the Shasta HMA is a product of complex mixing, and question its interpretation as a primitive mantle-derived

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magma. Proponents that it is a primitive mantle-derived magma argue that (1) the effects of hybridization are insufficient to shift its composition significantly from direct melts of a subduction-modified mantle source, (2) such magmas are reasonable parents to more evolved magmas at Shasta, and (3) they therefore must comprise a significant volume. Conversely, we believe the available data summarized in this paper indicate that (1) the HMA can reasonably be considered to be a mixture between the more volumetrically dominant basaltic and dacitic magmas produced at Shasta, (2) certain features (e.g., presence of Fo₉₃ olivine) assumed to be diagnostic of primitive mantle derivation are artifacts of interaction with ultramafic rocks in the underlying crust, and (3) HMA is not likely to be parental to the more evolved andesites and dacites at Mt. Shasta. A critical question concerns the actual volume of such magma and, based on surface exposures (essentially one local vent occurrence), HMA appears to be a relatively minor product. We would argue that this is the case for most volcanic arcs. Certainly for the Cascades, we favor the view that basaltic magmas are the dominant contribution from the sub-arc mantle, and their

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injection into the crust promotes melting to produce dacitic or more evolved magmas. Although fractional crystallization is inevitable, many intermediate composition arc magmas likely are products of mixing between local mafic and silicic magmas.

Here, we present detailed petrographic, mineralogical, and geochemical data for HMA lavas from a satellite vent on the Whaleback shield volcano. These rocks were previously studied by Anderson (1974), who first described their mineralogy and investigated melt inclusions in the phenocrysts. A series of important papers (Stone et al. 1989; Baker et al. 1994; Müntener et al. 2001; Grove et al. 2002; Bindeman et al. 2005; Le Voyer et al. 2010; Martin et al. 2011; Ruscitto et al. 2011) address aspects of their petrology and geochemistry and/or present results of experimental work bearing on their formation and evolution. Assuming that HMAs are little-modified mantle-derived magmas, Grove and coworkers proposed that they were produced by melting of a mantle wedge previously modified by influx of slab-derived hydrous fluids or melts. In contrast, our work (Streck et al. 2007a, 2007b, 2007c) clearly documents the hybrid nature of the Shasta area HMAs, which brings into question their interpretation as primitive magmas and demands reconsideration of previous models constructed on this premise.

Figure 1 presents representative compositional variations in Shasta lavas and serves to illustrate the controversy regarding their petrogenesis. When plotted vs. SiO₂, all major element oxides define approximately linear arrays that collectively differ from trajectories defined by fractional crystallization experiments (Müntener et al. 2001; Grove et al. 2002, 2005). Similar near-linear arrays are seen for other Cascades volcanoes (e.g., Mt. St. Helens: Smith and Leeman 1987, 1993; Mt. Lassen: Clynne 1999; Mt. Baker: Baggerman and DeBari 2011; Mt. Rainier: Sisson et al. 2014), and all could be produced by mixing between local basaltic and dacitic magmas. Cognate melt inclusions (HMA-type in Fo₈₇ equilibrium olivines; "HMA MI" in Fig. 1) and groundmass in the Shasta HMA ("HMA GM" in Fig. 1) plot essentially on the linear arrays, whereas the bulk HMA lava deviates from the array for MgO in a manner consistent with addition of exotic Mg-rich components (Streck et al. 2007c). One such component comprises forsteritic (Fo₉₃) olivines that contain "primitive basaltic andesite" (PBA-type) melt inclusions that are distinct from all known Shasta lavas and attest to the exotic nature of the host olivines (Ruscitto et al. 2011; cf. our Supplement II in the Appendix¹ for more details). Beginning with either local calcalkalic basalt or bulk



FIGURE 1. Harker diagrams for Shasta lavas showing SiO₂ vs. (**a**) MgO, (**b**) CaO, (**c**) Al₂O₃, and (**d**) Na₂O (all in wt%). Plots for K₂O, FeO*, and TiO₂ (not shown) also define near-linear arrays (labeled "mixing" in panel **a**) for the lavas (data from Grove et al. 2002). Samples are distinguished as basaltic, and high-Mg andesite (HMA) lavas; also plotted are groundmass ("GM", this study) and average "HMA-" and "PBA-type" melt inclusions [annotated "R" for average of Ruscitto et al. (2011) and "PBA" for our average; cf. Table 1], and average melt inclusions ("LV") from several Shasta basaltic lavas (Le Voyer et al. 2010). Yellow field shows loci of experimental liquids produced from primitive mafic lavas (Sisson and Grove 1993; Müntener et al. 2001; Pichavant et al. 2002; Grove et al. 2002, 2003; Sisson et al. 2005; Whitaker et al. 2008). Heavy arrowed curves show fractional crystallization trends for water-bearing liquids produced from a Shasta basaltic andesite (85-44; blue, FCb) and the Shasta HMA (85-41b; black, FCa) at pressures between 0.8 and 1.2 GPa; at lower pressures, plagioclase occurs as a near-liquidus phase, causing depletion of Al₂O₃ (effect shown in panel **c** for experiments (data taken from review by Gao et al. 2016). These are screened to exclude mica-bearing starting compositions (that tend to produce granitic melts). At pressures below ~1.5 GPa (gray field), plagioclase is a residual phase, whereas above 1.5 GPa (red-outlined open field) garnet is present, and these minerals strongly influence the Al₂O₃ and Na₂O contents of partial melts; in **a** the two sets of experiments overlap and are not distinguished.

HMA, the experimentally determined fractionation pathways cannot replicate observed compositions of more evolved lavas. However, experimental partial melts of mafic amphibolites (cf. Gao et al. 2016) indicate that the most evolved (dacitic) magmas could potentially be produced by partial melting of such rocks at deep crustal levels. This raises the possibility that intermediate lava compositions could be produced by mixing between mantle-derived basaltic magmas and melts of preexisting crustal rocks. Here, we further characterize the magmatic and other mixing components identifiable in HMA, and speculate on how these magmas formed in the context of petrogenetic scenarios proposed for the Cascades. Ultimately, this work bears on models for arc magmatism in the Cascades and worldwide.

SAMPLES AND ANALYTICAL METHODS Field relations

HMA occurrences at Mt. Shasta and surroundings are largely restricted to a scoria cone on the NE flank of the Whaleback shield volcano, located 5 km northeast of Mt. Shasta summit (Figs. 2 and 3). All samples of "Mt. Shasta" HMA investigated by Anderson, Grove and coworkers, and by us were obtained from this location. No other occurrences of HMA have been reported, albeit Barr et al. (2008) mention a lava flow of HMA composition in the vicinity. Although not directly dated, the age of the HMA scoria cone is estimated to be 160 ± 16 ka (A. Calvert, personal communication). Quarry exposures reveal two tephra sequences separated by a strong angular unconformity (Fig. 3; cf. Anderson 1974). Layers of the lower sequence are on average thinner (<30 cm) than those of the upper sequence (0.5 to 2 m). The relative proportions of ash:lapilli:blocks vary, and individual strata range from thin layers comprising mainly coarse ash/fine lapilli, to well-sorted layers consisting almost entirely of lapilli, to thicker layers containing abundant blocks and bombs. Samples investigated in this study consist of volcanic ejecta blocks (bombs) and tephra samples, as well as a crosscutting dike (MS-HMA15) that represents the youngest eruptive event discernible in the quarry exposure. Melt inclusions from the two tephra samples (MS-13-04 and MS-14-04) were previously studied by Ruscitto et al. (2011).

Analytical methods

Whole-rock samples were analyzed for major and trace elements by XRF (Washington State University, Geoanalytical Laboratory) and, for selected samples, for ICP-MS trace elements (Rice University and Washington State University), ICP-AES B concentrations (Rice University), and Sr and Pb isotopic compositions (University of Arizona). These data are presented in Supplemental¹ Table A1. For comparison, we also include in this Table our re-analysis of sample 85-41b along with analyses of the same sample as reported by Grove et al. (2002) and by Magna et al. (2006). LA-ICP-MS analyses were also conducted on representative minerals and groundmass using facilities at ETH (Zürich) and Rice University. All mineral major element analyses were performed using a Cameca SX100 electron microprobe housed at Oregon State University and Accessed remotely from Portland State University. Analyses were made using a 30 nA beam current,



FIGURE 2. Oblique Google Earth view of Mt. Shasta and the adjacent Whaleback shield volcano from a northeast perspective. Red cross indicates locality of samples described in this paper. Inset map shows general location of this area.



FIGURE 3. (a) West side of Whaleback Quarry, showing locations (dots or vertical bar) of analyzed samples in the context of the exposed stratigraphy. (b) Hand sample of MS-16-04 showing olivine-rich xenolithic fragments. (c) MgO vs. SiO_2 for all Whaleback Quarry samples. (d) Cr vs. MgO, showing the up-section increase in these components. All samples of the pyroclastic layers are volcanic bombs except for MS-13-04 and MS-14-04 that are tephra samples used by Ruscitto et al. (2011); these two samples are not plotted as there are no bulk compositional data for tephra samples.

15 kV accelerating voltage, and a 1 μ m diameter focused beam. Counting times ranged from 10 to 30 s on peak positions and half these times on background positions. Na was always analyzed first to maximize count rates. Standardization utilized a set of natural materials. Representative mineral analyses are reported for pyroxenes, olivines, oxide minerals, and plagioclase, (Supplemental¹ Tables A2, A3, A4, and A5, respectively).

LA-ICP-MS data were acquired at ETH using a 193 nm ArF Excimer laser coupled to an ELAN 6100 ICP quadrupole mass spectrometer, and at Rice University using a 213 nm YAG laser coupled to an Element 2 magnetic sector mass spectrometer. Similar laser settings (10 Hz repetition rate) were used in both labs yielding a fluance of about 12 J/cm² on the ablation site, with typical beam size of $30-60 \ \mu m$ on minerals and 70 $\ \mu m$ on groundmass. Trace element contents are reported for selected minerals (Supplemental¹ Table A6) and for HMA groundmass (Table 1).

Sr and Pb isotopic compositions (Supplemental¹ Table A1) were determined by MC-TIMS and MC-ICPMS, respectively, using methods described by Thibodeau et al. (2007, 2015). ^{ar}Sr/⁸⁶Sr ratios are reported with precision better than 0.002% relative to a value of 0.71025 for the NBS-987 reference material. Pb isotopic ratios are reported with external precision better than 0.03% on each ratio based on replicate analyses of NBS-981 reference material with sample ratios normalized to the isotopic composition reported by Galer and Abouchami (1998).

BULK-ROCK COMPOSITIONS

Major and trace element compositions are reported for samples that bracket the stratigraphy exposed in the Whaleback quarry. Our analyses compare closely with published data for equivalent samples (Baker et al. 1994; Grove et al. 2002; Magna et al. 2006) as seen by direct comparison for sample 85-41b (Supplemental¹ Table A1). Overall, there is little variation, especially with regard to incompatible elements, and the averaged concentrations for our samples have relative standard deviations smaller than 5-7% for most elements. It is notable that Cr (489-603 ppm) and MgO (7.9-9.1 wt%) contents systematically increase up section (Fig. 3d). Cr and Ni contents are highest in the dike, which represents the latest eruptive phase that was sampled. There is essentially negligible variation in Pb isotopic composition, whereas the dike has slightly lower 87Sr/86Sr than the three earlier ejecta blocks that were analyzed. These subtle changes in composition suggest that the magma reservoir may have been zoned (i.e., toward slightly more mafic bulk composition with depth), or that the eruptions were accompanied by influx of a similar but slightly more mafic magma as eruptions proceeded. To complement this study we also present new analytical data for representative samples of local basement rocks (e.g., Trinity Ophiolite) and eight other lavas from the Shasta region (cf. Supplemental¹ Table A7).

PETROGRAPHY AND MINERAL COMPOSITIONS

Overview

HMA blocks and lapilli have vesicular textures with 15-20% phenocryst-sized (i.e., >80 µm) crystals and/or aggregates of olivine, clinopyroxene, and orthopyroxene as well as smaller groundmass/matrix crystals of these same minerals set in a finegrained, largely crystalline groundmass. Among phenocryst-sized minerals, orthopyroxene (Opx) is most abundant followed by subequal amounts of clinopyroxene (Cpx) and olivine, in approximate proportions of 45:30:25, respectively. In the lowermost tephra layers, Cpx is clearly more abundant than olivine, but the proportion of olivine increases slightly toward the top of the sequence (as reflected by increasing MgO in bulk rocks; see above). Pyroxene compositions show a large composition range (Fig. 4), but the most remarkable feature of these rocks is their diversity in terms of mineral texture and compositional zoning of individual crystals. Notably, many Cpx and Opx grains are characterized by conspicuous reverse zoning with Fe-rich cores and Mg-rich overgrowth rims. These minerals are described in detail in the following section, but clearly they signify a complex assembly history for the HMA magma. It should also be noted that xenolithic fragments (up to 0.5 cm) of disaggregated ultramafic material and/or aggregates of olivine ± pyroxene occur in numerous blocks examined (cf. Fig. 3b). Plagioclase is largely confined to the groundmass, and rare phenocryst-sized grains appear to be in disequilibrium with the host magma. Other minor minerals include Cr-spinel or ilmenite inclusions in ferromagnesian silicates, rare Fe-Cu-Ni sulfide inclusions found predominantly as inclusions in olivines (mainly in micro-xenoliths; cf. Stone et al. 1989), and sparse oxide minerals in the groundmass.

Mineral compositions, zoning, and populations

The textures and mineral compositions of Shasta HMA provide critical constraints on the formation of these rocks. Here we provide a detailed description of this remarkable assemblage.

Pyroxenes. Phenocryst-sized pyroxenes appear to have formed in several different growth environments. Based on distinctive textures, zoning characteristics, and modes of occurrence, we have defined three groups or "types." In some cases, these are further subdivided based on subtle variations in these features.

Type 1 pyroxenes include both Opx and Cpx crystals that exhibit conspicuous step-like reverse zoning with variably thick (<10 to ~50 μ m) high-Mg overgrowth rims on lower-Mg interior cores (Fig. 5; cf. Streck 2008). Overgrowth Mg# values exhibit similar ranges for both Cpx (90–86) and Opx (93–81), with values predominantly near 87. BSE images of the overgrowths (Fig. 5) reveal no discernible internal discontinuities (e.g., resorption

TABLE 1. HMA groundmass, whole-rock, and melt inclusion compositions

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This	Groundmass	S.D.	HMA WR	S.D.	HMA MI	S.D.	PBA MI	S.D.	Enrichment	Opx MI ^f	S.D.	Ruscitto	HMA MI	PBA MI
paper	Average ^a		Average ^b		Average ^c		Average ^c		Factor ^d	Average		et al.	Average ^c	Average
	(n = 7)		(n = 7/9)		(n = 5)		(n = 14)			(n = 5)		(2011)	(n = 5)	(n = 14)
SiO ₂	61.63	1.20	58.14	0.25	58.48	1.17	55.62	1.43	1.06	67.75	1.70		58.46	53.54
TiO ₂	0.68	0.09	0.59	0.01	0.70	0.03	0.68	0.03	1.15	1.10	0.30		0.70	0.58
AI_2O_3	16.28	1.40	14.79	0.24	16.86	1.12	15.08	0.73	1.10	17.14	0.90		16.82	12.93
FeO*	4.49	0.33	5.65	0.11	5.68	0.39	4.37	0.33	0.80	3.76	0.05		5.70	7.02
MnO	0.08	0.01	0.10	0.001	0.10	0.01	0.09	0.01	0.76	0.07	0.05		0.10	0.07
MgO	3.60	0.50	8.56	0.40	5.45	0.39	8.88	0.99	0.42	0.59	0.37		5.54	12.92
CaO	7.99	0.60	8.18	0.11	8.74	0.60	12.35	0.96	0.98	4.41	0.73		8.70	10.56
Na₂O	4.10	0.30	3.12	0.10	2.86	0.56	2.25	0.43	1.31	3.21	0.37		2.86	1.79
K₂O	1.00	0.15	0.74	0.03	0.92	0.11	0.37	0.07	1.35	1.68	0.06		0.91	0.31
P_2O_5	0.15	0.02	0.130	0.003	0.21	0.08	0.32	0.07	1.15	0.29	0.08		0.22	0.27
Mg# ^g	64.6 ^e	-	76.1	-	66.8	-	81.0	-	-	24.8	-		67.1	79.4
Fog	87.0 ^e	-	91.5	-	87.7	-	93.5	-	-	56.0	-		87.8	92.4
							Trace elem	ents ^h						
U	0.96	0.11	0.69	0.04	0.92	0.08	0.37	0.08	1.39				0.92	0.37
Rb	19.5	1.9	14.1	1.4	20.5	2.0	4.8	2.1	1.38				20.5	4.8
Pb	5.2	0.6	3.83	0.31	6.4	0.8	3.1	0.7	1.36				6.4	3.1
Nb	3.16	0.47	2.48	0.20	3.10	0.07	1.5	0.2	1.27				3.10	1.5
Ce	28.6	3.1	22.8	0.8	33.3	2.2	28.2	3.6	1.25				33.3	28.2
Ba	221	21	187	11	256	16	80	18	1.18				256	80
La	12.1	1.3	10.6	0.3	11.4	1.0	9.8	0.5	1.14				11.4	9.8
Та	0.20	0.01	0.18	0.01	0.18	0.05	0.08	0.02	1.11				0.18	0.08
Th	2.2	0.3	1.98	0.08	2.0	0.3	1.04	0.21	1.11				2.0	1.04
Pr	3.4	0.2	3.08	0.10	3.6	0.3	3.3	0.3	1.10				3.6	3.3
Zr	101	11	91.5	8.3	95	8	68	7	1.10				95	68
Nd	14.1	1.4	12.8	0.5	13.8	1.9	12.6	0.9	1.10				13.8	12.6
Sm	2.8	0.3	2.64	0.11	2.7	0.4	2.3	0.5	1.06				2.7	2.3
Eu	0.90	0.04	0.85	0.04	0.96	0.09	0.8	0.1	1.06				0.96	0.8
Yb	1.1	0.1	1.04	0.05	1.2	0.3	0.8	0.2	1.06				1.2	0.8
Sr	801	61	765	20	867	54	1047	73	1.05				867	1047
Hf	2.4	0.3	2.36	0.09	2.8	0.4	1.8	0.3	1.02				2.8	1.8
Er	1.1	0.1	1.14	0.04	1.0	0.3	0.9	0.3	0.96				1.0	0.9
Υ	10.5	1.1	11.6	1.0	11.1	1.1	7.9	1.3	0.91				11.1	7.9
Dy	1.9	0.2	2.15	0.08	2.0	0.3	1.4	0.2	0.88				2.0	1.4
Gd	2.2	0.2	2.52	0.18	2.1	0.6	1.7	0.5	0.87				2.1	1.7
Sc	15.4	2.0	22.7	0.4	28	7	31	6	0.68				28	31
Cr	59	22	548	40	361	246	1019	1395	0.11				361	1019

^a Groundmass is from sample MS-04-16 (this study); analyses by EMP (major elements) and LA-ICPMS (trace elements); X-fe3 = 0.22. ^b Average HMA whole-rock (from Supplemental Table A1, this study); based on 7 major element and 9 trace element analyses. ^c Compositions of olivine-hosted HMA- and PBA-type melt inclusions are from Ruscitto et al. (2011); cf. Supplement II. ^d Enrichment Factor = Groundmass concentration/Average WR concentration. ^e Mg# and equilibrium olivine Fo are calculated assuming Xfe3 = 0.22. ^f Melt inclusions in Type 1 antecrystic low Mg# (71–73) cores. ^g For all analyses except first column, calculated Mg# and olivine Fo content assume X-fe3 = 0.15. ^h Trace element data are not corrected for olivine addition.



FIGURE 4. (a) Al and (b) Cr vs. Mg# in analyzed HMA pyroxenes. Lowest Mg# (59) of one orthopyroxene (Opx) plots offscale. Low Mg# compositions (<80) are found in cores of Type 1 pyroxene phenocrysts while highest Mg# (>88) are typically found in rims of phenocrysts. Range of typical Type 1 pyroxene compositions shown by solid bar for rims and open bar for crystal cores. Resorbed Opx crystals have compositions clustering around Mg# 87 (cf. Figs. 5, 6, and 7; see text).

surfaces) and appear consistent with continuous rim growth. The overgrowths typically show a subtle outwardly progressive decrease in Mg# (Fig. 6); thicker growth bands may display step-like decreases (Figs. 5d and 6b). Although Opx overgrowth tends to be initially higher in Mg# than Cpx overgrowth, data on Type 3 pyroxenes (see below) clearly attests that both minerals co-crystallized. Overgrowths on both Cpx and Opx exhibit similar ranges in Cr₂O₃ content (from >1 to ~0.2 wt%; Fig. 5), and Mg# and Cr contents are generally positively correlated. However, sometimes there is decoupling of Mg# and Cr as exemplified by some grains where maximum Cr content occurs in the middle of the overgrowth band (Fig. 6a; also Fig. 7e of Streck 2008).

Interiors (cores) of Type 1 pyroxenes have distinctly higher Fe contents than the rims with Mg# < 80, except as noted below. Where analyzed, zoning in the cores is always normal with Mg# decreasing outward from the interior (Fig. 5b). However, based on differences in composition and zoning characteristics within the cores, we define three sub-types that could signify distinct sources or growth environments for these pyroxenes.

Type 1A pyroxene cores are Fe-rich (Mg# between 59–75 overall) with little zoning (Mg# range <2%) in any given grain (Figs. 5a, 5c, and 5d). Type 1B crystals have relatively homogeneous but higher Mg# (76–80) cores, but they exhibit a thin (~10–50 µm) outer selvage of slightly more Fe-rich composition

(Mg# <76). Included in this group are some larger (>500 μ m) crystals with more extremely zoned outer cores (Mg# down to ~70; Fig. 5b). Cores of some Type 1B crystals exhibit a growth discordance or hiatus as evidenced by zoning unconformities or even patchy zoning (cf. Streck 2008). Finally, Type 1*C* crystals are also characterized by more magnesian cores (Mg# always >76), but differ from Type B grains by the absence of any more Fe-rich compositional domains.

Interfaces between low-Mg# cores and high-Mg# overgrowths in Type 1 pyroxenes are mostly anhedral, albeit subhedral to euhedral forms are sometimes preserved (Fig. 5). On the other hand, the external morphology of Type 1 crystals is nearly always euhedral due to precipitation of the late overgrowth rims. Some Type 1 pyroxenes occur as glomerocrystic clusters or aggregates (± inclusions of other minerals); these are more common in the lowermost tephra sequence but are observed throughout the deposit.

Other notable features are as follows. Some Type 1 grains clearly are fragments of larger crystals that must have been fractured or comminuted shortly before eruption of the HMA magma. This is most evident where low-Mg# portions terminate against the groundmass with straight and angular shapes but with no signs of overgrowth or resorption, whereas other portions of the same crystal do exhibit the high-Mg# overgrowths characteristic of intact Type 1 pyroxenes (Fig. 5). Also, some Type 1 pyroxenes contain relicts of distinctive "patchy zoned" grains (defined below as Type 2 pyroxenes) that appear to have served as nucleation seeds for Type 1 overgrowths. Rarely, some Type 1 pyroxenes are strongly resorbed. And finally, the Fe-rich cores of some Type 1 Opx and Cpx crystals contain melt inclusions of dacitic bulk composition that are now variably crystallized (Streck et al. 2007a; see below and Table 1).

Two-pyroxene geothermometry (Putirka 2008) was applied to subsets of Type 1 pyroxenes that we interpret to have been essentially coeval. Pyroxene pairs representing high-Mg# (~87) overgrowths give temperatures near 1100 °C, that plausibly reflect conditions shortly before the HMA eruption. In contrast, pyroxene pairs representing low-Mg# (~70) cores yield temperatures around 1000 °C. The latter are similar to estimates for 1980 Mount St. Helens dacites (~970 °C; cf. Smith and Leeman 1987) and are considered realistic. For comparison, Ruscitto et al. (2011) estimate entrapment temperatures near 1050 ± 35 °C for melt inclusions in HMA Fo₈₇ phenocrystic olivines; this is consistent with slightly higher temperature for the host magma.

Type 2 pyroxenes (Opx only) are distinguished by the presence of distinctive resorption channels (i.e., crystals have pronounced coarse spongy cellular textures) and/or patchy zoning (Fig. 7). The resorption channels have a vermicular appearance under the microscope or in backscattered images (Fig. 7). These channels are either filled with partially crystalline material (indistinguishable from adjacent groundmass) or appear to be empty (Figs. 7a, 7b, and 7c). Patchy zoning in the remaining orthopyroxene shows great variability in spatial distribution and compositional gradation. These areas appear lightly speckled (at a scale of \sim 5–20 µm) within a dark matrix (Fig. 7b), although larger patches with complex compositional gradation also occur (Fig. 7c). The morphology and compositional gradients of patchy zoned domains seem inconsistent with simple crystal growth from a single cooling magma, and more likely reflect diffusional or melt reaction processes by



FIGURE 5. BSE images of representative Type 1 (antecrystic) pyroxenes; note that lowermost grain in panel **d** is a Type 3 pyroxene. Each phenocryst is labeled to indicate crystal type (Opx vs. Cpx). White dots show loci of individual analyses; adjacent numbers are Mg#. White lines within crystals in panels **b** and **d** indicate loci of the analytical traverses shown in inset. Black empty areas in Opx crystal of **a** are artifacts of sample preparation. Scale bar is 200 μ m in all panels.

which high-Mg Opx was replaced by lower-Mg Opx—possibly in response to interaction with lower-Mg# melt preceding entrainment of these crystals in the HMA magma.

A notable subtype of Type 2 pyroxene is characterized by distinct overgrowths of very low Mg# (<70) Opx that appear to infill cracks and channels (Figs. 7b and 7d; also Fig. 2c of Streck et al. 2007a) and seemingly postdate formation of spongy resorption in the higher-Mg# Opx host. For most Type 2 pyroxenes, the last discernible crystallization "event" is recorded by formation of euhedral, and usually higher-Mg# overgrowths similar to those observed in Type 1 Opx (Figs. 7b and 7c; Fig. 2c of Streck et al. 2007a; Fig. 3b of Streck 2008). Nevertheless, some Type 2 crystals do have an anhedral habit and, rarely, they occur as glomerocrystic clusters (Fig. 7d). Finally, small-scale (< ~100 μ m) comminuted Type 2 crystal debris is common in the groundmass.

Type 3 pyroxenes (both Cpx and Opx) are distinguished by the absence of significant concentric zoning (except for the very outermost few micrometers of crystals; Figs. 5, 7b, and 7c). Mg# ranges for Cpx (90–83) and Opx (93–86) largely overlap. Some Type 3 pyroxenes display hour-glass zoning that has been attributed to rapid growth, particularly in pyroxene from sub-alkaline magmas (Brophy et al. 1999). They occur predominantly as microphenocrysts (<80 μ m) or small phenocrysts (<200 μ m), and very seldom as crystals larger than a few hundred micrometers. The textural habits of Type 3 pyroxenes (e.g., intergrowths sharing crystal faces or mutual overgrowths; Fig. 7c) unequivocally suggest simultaneous growth of these pyroxenes coincident with the crystallization of high Mg# overgrowths on Type 1 and 2 pyroxenes. Some small Type 3 crystals have skeletal forms that are consistent with rapid growth. Rapid growth and mostly small size of Type 3 grains suggest crystallization close to eruption and this in turn suggests that variation of Mg# of Type 3 pyroxene is a function of how the melt composition changed during groundmass crystallization.

In summary, Type 3 Opx and Cpx appear to have formed directly from the HMA liquid, whereas Types 1 and 2 clearly formed from different liquids and represent antecrystic and/or xenocrystic material.

Olivine. Three distinct olivine populations are recognized based on compositions, textures, and crystal sizes. Type 1 olivines (Fig. 8) appear to be xenocrystic and are rare (\pm 3 grains per thin section). They generally comprise large (typically >500 µm, but up to 3 mm) crystals with highly forsteritic compositions (Fo_{92–94}, with most near Fo_{93–93.6}). They are mostly anhedral to



FIGURE 6. Analytical traverses across two Type 1 clinopyroxene crystals at point spacing of ~1.5 μ m. From right to left, traverses start within low Mg# core portion and end at the rim. Importantly, step profiles across low to high Mg# domains are similar for both faster (Fe,Mg) and slower (Cr) diffusing cations. Also, approaching the rims, Cr sharply decreases (albeit not as low as the core compoitions), whereas Mg# drops less dramatically (only to 82–84). These two profiles were used to calculate relaxation times for diffusion; optimal fits suggest times on the order of a few months (see text).

subhedral in form but exhibit several distinct terminations against the groundmass. In some cases, the anhedral margins are clearly resorbed and show compositional zonation to lower Fo content (Fig. 8d) similar to that of true phenocrysts (\sim Fo₈₇, see below) in the HMA magma. In other cases, the olivine is overgrown by a reaction selvage of polycrystalline Opx that in places also penetrates into the olivine (Fig. 8a). The anhedral margin is sometimes fringed by rapidly crystallized skeletal (\sim Fo₈₇) olivine (Fig. 8b; cf. Fig. 1 of Streck et al. 2007a) with seemingly little reequilibration of the original olivine. In places, the skeletal fringe grades into a continuous euhedral overgrowth, albeit the anhedral form of the original grain is still evident (Fig. 8c). Last, the anhedral margins of some Fo₉₃ olivines terminate against groundmass without any preserved selvage (Fig. 8b, top and lower right-hand side), consistent with comminution of the xenocrysts near the time of eruption.

Type 2 olivines (Fig. 9) appear to be true phenocrysts of the

final HMA magma with maximum size of 400–500 μ m in the longest dimension. They commonly have skeletal forms with spiny to straight edges (Fig. 9a) and feathery shapes. Some are almost completely closed, euhedral grains with triangular inclusions of trapped melt (Fig. 9b). Occasionally, skeletal olivines partially enclose vesicles that predate or were contemporaneous with olivine growth (Fig. 9a). These textures are consistent with rapid growth (cf. Lofgren 1980; Schiano et al. 2010). The compositions of skeletal olivines are narrowly defined (between Fo₈₅₋₈₉ with most near Fo₈₇, Fig. 10) and equivalent to olivine microphenocrysts and overgrowths on Type 1 olivines, and also similar to Mg# of late-formed pyroxenes.

Type 3 olivines occur primarily as multi-grain glomerocrysts with somewhat distinctive compositions (Fo₈₈₋₉₁). Varied extinction angles among grains confirm that clusters consist of multiple, interlocking crystals and that these are in fact derived from disaggregated xenoliths (Figs. 9c and 9d). This interpretation is supported by rare observation of undulose extinction within single crystals of olivine. Such glomerocrysts appear to make up the majority of xenolithic olivine debris observed in hand samples (cf. Fig. 3b). Contacts between crystals are often filled by veinlets consisting dominantly of Opx, but with abundant spinel and rare plagioclase (Figs. 9c, 9d, and 11b). Rare individual olivine crystals having anhedral form and similar compositions are assumed to have the same derivation. NiO contents in the olivines vary from ca. 0.15–0.4 wt%, and correlate loosely with Fo content (Fig. 10). Type 2 phenocrystic olivines form a distinct group from the others, whereas there is considerable overlap between Types 1 and 3.

Oxides. Oxide minerals are quite rare in samples of HMA. They occur as sparse inclusions in pyroxene and olivine, and are most abundant in Opx-dominated veinlets within Type 3 olivine glomerocrysts (Figs. 9 and 11). In contrast, oxide inclusions are exceedingly rare in Type 1 olivines. Oxide minerals are not observed as phenocrysts in HMA but do occur in the groundmass as accessory chromian spinel. Based on detailed modal counts under reflected light, the stratigraphically lowest samples (MS-HMA2 and MS-HMA4) contain traces (50–100 grains/thin section) of very small (~10 μ m) groundmass grains, whereas the stratigraphically highest samples (MS-HMA14 and MS-HMA15) contain slightly higher abundances (ca. 200–400 grains/thin section). Compared with crystallized groundmass in typical basaltic to andesitic rocks (>1000 Fe-oxide grains/thin section) these are surprisingly low abundances.

Compositions of oxide inclusions vary with the specific enclosing mineral host. Within high-Mg pyroxene and olivine, oxide inclusions invariably are chromites or chromian spinels (Cr# =80–45, Mg# = 75–45), whereas in Fe-rich pyroxene cores (Mg# < 75) ilmenite is the only oxide inclusion found (Supplemental¹ Table A4). The highest-Mg# spinels occur in Type 1 olivines, whereas spinel inclusions in Type 3 glomerocrystic olivines and within their Opx-dominated veinlets have intermediate Mg# (68–55) and Cr# (80-50). Spinel from a rare anhedral xenocrystic olivine-Opx association (inset of Fig. 11a) has Mg# and Cr# at the low end of these ranges (55 and 50, respectively); these, and spinels from glomerocrystic Type 3 olivine clusters, compositionally overlap with spinel from the nearby Trinity Ophiolite (Quick 1981; Ceuleneer and Le Sueur 2008). Finally, the lowest Mg# spinels (Mg# = 55–15, Cr# = 83–60) occur as inclusions in late-formed pyroxenes



FIGURE 7. Transmitted light (**a**, inset of **d**) and BSE images (**b**, **c**, **d**) of Type 2 pyroxenes. High Mg# core of Type 2 pyroxene is characterized by strong resorption channels leading to coarsely spongy texture and by patchy zoning with occasionally distinct overgrowth of low Mg# pyroxene (**b** and **d**). In many cases, complex and mostly xenocrystic core is overgrown by high Mg# euhedral rim (**b**, **c**, and partly **d**). Type 3 clinopyroxenes are present in upper center of panel **b** and upper left of panel **c**; the latter has an Opx core overgrown by Cpx and is adjacent to an Opx microlite (dark gray shade) with Mg# ~90. Inset in **d** is a transmitted light image of same grain. Scale bar is 200 μ m in each panel; dots show analysis loci with Mg#.

and olivines as well as in the groundmass (Fig. 11).

Plagioclase. Plagioclase occurs dominantly as groundmass crystals (or microphenocrysts). Only seldom does it occur as large phenocryst-sized crystals (≤1 crystal/thin section), except in the lowermost tephra deposits where both isolated grains and mono-mineralic glomerocrysts are slightly more common (a few crystals/thin section). Because all such grains are variably resorbed (i.e., mildly rounded to strongly sieve-textured) and somewhat variable in composition (An₇₈₋₄₈; Fig. 12), they are interpreted to be xenocrysts. Groundmass plagioclase is more uniform in composition (~An₆₀) and similar to rare inclusions in late-formed Type 3 pyroxene (Fig. 12). Only a melt that has evolved during late stage crystallization can explain the groundmass plagioclase composition because the bulk composition of HMA has a Ca/Na ratio too high to crystallize An₆₀ plagioclase. Groundmass composition (Table 1) would be in equilibrium with An_{68} plag (at K_D = 2; cf. Sisson and Grove 1993); this suggests that groundmass plagioclase appeared even after some more Cpx groundmass crystallization reduced the Ca/Na ratio sufficiently. Furthermore, only the highest observed An content of large crystals can be potentially explained by equilibrium crystallization using a $K_{\rm D}$

value of 2. This is consistent with textural observations indicating its xenocrystic nature.

Proportion and distribution of pyroxene and olivine populations

Volume proportions of essential minerals in the HMA magma, and distributions of all distinctive phenocryst-sized pyroxene and olivine sub-types were determined by detailed inventory for one representative thin section (sample MS-16-04). For simplicity, two mineral populations were distinguished: grains (1) with maximum dimension greater than 200 μ m, and (2) smaller than 200 μ m (including some microphenocrysts). BSE images and electron microprobe data were collected for all phenocryst-size pyroxenes except for 15 Type 2 Opx grains, for which identification was obvious from textural evidence. All olivines were surveyed by textural criteria and compositional data were acquired on all large olivines and a subset of smaller grains. On the other hand, only representative pyroxene and olivine microphenocrysts were imaged by BSE and analyzed by electron microprobe because they are so numerous. All surveyed pyroxene grains were classified as discussed above and each type is distinguished in Figure 13. Results are also



FIGURE 8. Transmitted light (**a** and **c**) and BSE images (**b** and **d**) of Type 1 (xenocrystic) olivines. (**a**) Shows micro-crystalline (mostly Opx) reaction selvage inside and around olivine. (**b**) Shows lighter reaction fringe of Fo_{87} olivine on more magnesian crystal; no overgrowth is observed along the top and lower right edges of the crystal, juxtaposing Fo_{93} olivine against groundmass that crystallized ~ Fo_{88-85} olivine. (**c**) Anhedral Fo_{93} relic olivine core jacketed by euhedral Fo_{88} olivine. (**d**) Remnant of requilibrated Fo_{93} olivine overgrown by Fo_{86} olivine. Scale bar is 200 µm in each panel. Dots indicate analysis loci with Fo content; in **b** are shown compositions along analysis traverse.

presented in an interactive digital map that can be accessed online (https://www.pdx.edu/geology/streck-Shasta-HMA) where BSE images along with electron microprobe analyses are displayed. A high-resolution BSE image of the entire thin section surveyed is found in Supplemental¹ Figure A7.

Based on our detailed grain inventory, "phenocryst-like" components in this sample make up 18% by volume (after correction for 35% vesicularity); actually, these are nearly all "antecrystic" or "xenocrystic". Type 1 and 2 pyroxenes dominate the population of larger crystals. Estimated modal proportions (corrected for vesicularity) of phenocryst-size minerals is: 20% Type 1 Cpx and Opx (Mg#75), 50% Type 2 Opx (Mg#89), and 30% Type 1 olivine (Fo₉₃). Among smaller crystals (<200 μ m), Type 3 ("equilibrium") and Type 1 ("antecrystic") pyroxenes occur in subequal proportions, whereas Type 2 ("xenocrystic") crystals are less abundant. Notably, all pyroxene types appear to be randomly distributed and well mixed at the thin section scale. Data on olivines indicate that all distinctly large olivines are Type 1 ("xenocrystic"), whereas the smaller grain population is dominated by Type 2 ("equilibrium") olivines. Implications of these findings are discussed below.

Trace element compositions of minerals and groundmass

Type 1 clinopyroxene crystals were analyzed for trace element content to further delineate the contrast between low Mg# core and high Mg# overgrowth and to estimate compositions of equilibrium melts that likely produced them. Representative analytical results are presented in Supplemental¹ Table A6. Primary observations are that concentrations of incompatible trace elements are much higher in low Mg# cores than in high Mg# rims and overgrowths. Furthermore, REE patterns of low Mg# interiors indicate a pronounced negative Eu anomaly typically observed in evolved silicic igneous rocks (cf. Grove et al. 2005). And finally, melt compositions calculated with REE partition coefficients (Lee et al. 2007; Severs et al. 2009) yielded REE concentrations similar to or slightly higher than those of naturally occurring dacite lavas of Mount Shasta. Use of slightly higher partition coefficients calculated (cf. Brophy 2008) for magmatic SiO₂ contents of Shasta dacites produces even closer agreement. Thus, the pyroxene cores are inferred to have crystallized from such magmas. We note that this is also the case for high-Mg basaltic andesite and andesite from the North Cascades (cf. Sas et al. 2017). In contrast, allowing for

uncertainties in partition coefficients, estimated melts equilibrated with the pyroxene rims are similar in composition to HMA and consistent with in situ formation of these overgrowths.

Groundmass was analyzed at multiple areas of sample MS-16-04 and averaged results are listed in Table 1, where they can be compared to average analyses for melt inclusions from Ruscitto et al. (2011). Melt inclusions hosted by \sim Fo₈₇ olivines are similar (mostly within analytical uncertainty) to the average groundmass, and are herein designated as "HMA-type." In contrast, melt inclusions hosted in \sim Fo₉₃ olivines are of a distinctive "primitive basaltic andesite" (or "PBA") type (Ruscitto et al. 2011). The compositions of groundmass and HMA-type melt inclusions lie essentially on a hypothetical mixing line between Shasta basalts and dacites (cf. Fig. 1). Moreover, subtraction of phenocryst-size minerals in observed modal proportions (see above) from the bulk composition of sample MS-16-04 indicates that this estimated "liquid" is similar in composition to the analyzed groundmass and HMA-type melt inclusions.

The bulk composition of MS-16-04 (and all other HMA samples) differs from the groundmass composition with respect to elements that are concentrated in ferromagnesian minerals or

diluted by their addition. That is, the groundmass composition is more silicic (61 wt% SiO₂) and poorer in MgO, Sc, Cr, and Ni. In detail, relatively incompatible elements (e.g., U, Th, Rb, Pb, Nb, Ta, Zr, and LREE) are variably enriched by factors of 1.05 or greater, whereas less incompatible elements (including HREE and Y) are slightly depleted (cf. Table 1, where trace elements are listed in decreasing order of enrichment in the groundmass) relative to the HMA whole rock. These patterns are consistent with selective accumulation in the bulk magma of clinopyroxene, which has partition coefficients greater than unity for the latter elements.

Sr, Nd, and Pb isotopic compositions

New Sr and Pb isotopic data for four samples are in close agreement with previously published data (cf. Supplemental¹ Table A1). Average isotopic values for HMA are as follows: ⁸⁷Sr/⁸⁶Sr: 0.70301 \pm 6; ²⁰⁶Pb/²⁰⁴Pb: 18.873 \pm 17; ²⁰⁷Pb/²⁰⁴Pb: 15.588 \pm 7; ²⁰⁸Pb/²⁰⁴Pb: 38.463 \pm 24. A single measurement of ¹⁴³Nd/¹⁴⁴Nd (0.51297) is available from Grove et al. (2002). These data lie within ranges for both basaltic and more evolved (andesite-dacite) lavas from Mt. Shasta (⁸⁷Sr/⁸⁶Sr: 0.7027–0.7038; ¹⁴³Nd/¹⁴⁴Nd: 0.5127–0.5130; ²⁰⁶Pb/²⁰⁴Pb: 18.75–19.18; ²⁰⁷Pb/²⁰⁴Pb: 15.56–15.70; ²⁰⁸Pb/²⁰⁴Pb:



FIGURE 9. Images of other olivine variants. (**a**) BSE image of skeletal Type 2 olivine; (**b**) transmitted light image of skeletal but nearly euhedral Type 2 olivine; this texture is consistent with rapid crystal growth or quenching. (**c** and **d**) Transmitted light images of two Type 3 olivine clusters with cumulate-like textures; these are both single lapilli extracted from tephra samples MS-13-04 and MS-14-04. Intergrain and surrounding material are Opx veins described in text and seen in Figure 11b. Dots indicate analysis loci with Fo content.



FIGURE 10. Fo vs. NiO compositions of olivine crystals (xls) of different types. Type 2 (phenocrystic) and Type 3 (micro-xenocrystic) olivines are shown in Figure 9. Type 1 olivines are part of "single crystals" and compositional spread of crystal interiors is indicated by bracket. Compositions of Type 1 olivine fall into fields of mantle peridotites except those with less than 0.3 wt% NiO and those higher than Fo₉₃ that fall outside of the compilation by Sobolev et al. (2005) as shown by grayed field.

38.32–39.05; cf. Supplemental¹ Table A7 and Grove et al. 2002). Although these data may be considered representative of domains in the underlying mantle wedge, there is partial or near overlap with present day isotopic compositions of mafic and ultramafic rocks of the Trinity Ophiolite (cf. Supplement I; 87Sr/86Sr: 0.7033-0.7160; ¹⁴³Nd/¹⁴⁴Nd: 0.5123–0.5136; Gruau et al.1998; Jacobsen et al. 1984). Also, gabbro and ultramafic rocks from the nearby Castle Lake area (Schwindinger and Anderson 1987) have 87Sr/86Sr (0.7044-0.7050) and Pb (206Pb/204Pb: 18.6-18.8; 207Pb/204Pb: ~15.6; and ²⁰⁸Pb/²⁰⁴Pb: 38.3-38.5) compositions close to those of Shasta area lavas (cf. Supplemental¹ Table A7). And, finally present-day Sr isotopic values for many Mesozoic age plutonic rocks of the nearby Klamath Mtns. are in the range 0.703-0.704 (Allen and Barnes 2006). Detailed comparison of Sr-Nd-Pb isotopic compositions of all these rocks is shown in Supplemental¹ Figure A1. Thus, on the basis of these data, it is difficult to exclude involvement of such crustal rocks in Shasta area magmas, either via magma-wallrock interaction or by partial melting. In particular, magma interaction with ultramafic wallrocks would have small leverage on magmatic isotopic compositions considering the low concentrations of Sr, Nd, and Pb in such rocks.

DISCUSSION

The petrogenesis of the Mount Shasta HMA is at the forefront of the discussion of the origin and significance of HMA magmas in general (Kelemen et al. 2003; Grove et al. 2002; Streck et al. 2007a; Barr et al. 2007; Kelemen and Yogodzinski 2007; Martin et al. 2011; Ruscitto et al. 2011; Mitchell and Grove 2015; Walowski et al. 2016) and our data presented above bear strongly on this issue. We examine our data and those of other recent studies with the goal of evaluating scenarios to explain the observed chemical and mineralogical features of the Mount Shasta HMA. Specific petrogenetic scenarios that have been proposed include: (1) HMA represents a magma type that was derived more or less directly from the mantle with only minor modification by interaction with other magmas or with crustal rocks; (2) HMA is derived from a more mafic precursor magma via fractional crystallization \pm open system processes like wallrock assimilation; (3) HMA is a product of magma mixing with some of the crystal cargo derived from the respective mafic and silicic mixing end-members \pm other contributions; and (4) HMA is a product of partial melting of crustal lithologies with the addition of some crystal cargo from other sources. The data presented here and in previous papers is used to assess the likelihood of these scenarios.

Origin of crystal and melt components in HMA

The diverse compositions and the complex textural relations of phenocryst-sized minerals in the Mt. Shasta HMA preclude the notion that the observed mineral assemblage formed by equilibrium crystallization of a single melt similar to HMA in bulk composition. Therefore, to understand the complex processes by which HMA was assembled, we first need to consider the origins of the various mineral and lithic components and the processes that can be inferred from the petrographic and geochemical evidence. We begin with discussion of the latest stage of crystallization, immediately prior to eruption.

Late-stage crystallization: Products from hybridized melt

Textural and compositional criteria clearly indicate that the latest-formed near-liquidus minerals in HMA are the Type 3 pyroxene microphenocrysts and equivalent overgrowths (all with Mg# near 87 or slightly higher) on Type 1 and 2 pyroxenes and the skeletal (or triangular inclusion-bearing) Type 2 olivines and equivalent overgrowths (all near Fo₈₇) on Type 1 and 3 olivines. These minerals clearly crystallized contemporaneously and shortly before eruption of the HMA magmas. This interpretation is supported by compositional overlap between rare spinel inclusions in these ferromagnesian minerals and in the groundmass (Fig. 11). Variation in Mg# and other elements (e.g., Cr) observed in Type 3 pyroxene and in the zonation pattern of overgrowth of Type 1 and 2 pyroxenes likely records melt evolution as olivine and pyroxene progressively crystallized during this late stage of crystallization. We infer that the ferromagnesian minerals crystallized from a common silicate melt having Mg# near 66 and evolving to a liquid near Mg# 60 (assuming bulk $K_D = 0.3$). Furthermore, this silicate liquid was saturated with regard to plagioclase only during the terminal crystallization period when groundmass crystals formed, and minimal oxide saturation is indicated by the sparse occurrence of Cr-rich spinels. The nature of this liquid is further constrained by compositions of melt inclusions in Fo₈₇ olivines (Ruscitto et al. 2011), and by our analyses of interstitial groundmass (Table 1). Together, these constraints indicate that the late-stage melt component in the HMA magma is andesitic in composition, with specific characteristics as follows: ~62 wt% SiO₂, ~5 wt% FeO*, \sim 20 ppm Rb, and \sim 60 ppm Cr, with Ba/Nb = \sim 70, La/Yb = \sim 11, and $Sr/Y = \sim 76$, (Table 1). We note that the groundmass composition would equilibrate with olivine of ~Fo₈₇ assuming molar ferric iron fraction $(X-fe_3) = 0.22$. Using the olivine-liquid thermometer of Lee et al. (2009), estimated magmatic temperature is ~1083 °C

(b)

grd-HMA2,4



FIGURE 11. Compositions of (a) spinels, and (b) olivine-spinel pairs. BSE image in a shows spinel-bearing olivine inclusion within a reacting disequilibrium orthopyroxene grain. Reflected light image in b shows oxide-rich Opx vein in olivine micro-xenoliths of Figures 9c and 9d. Spinel variants are distinguished by symbols as follows: spinel in veins (mostly Opx) of olivine micro-xenoliths (open orange triangles); spinel in olivine of micro-xenoliths (inverted open blue triangles); spinel in olivine included within the pictured Mg#84 orthopyroxene (open red cross); groundmass spinel (small open green circles) grouped by sample in which they are found. Other symbols: rare spinel in skeletal Fo_{87} olivine (filled green diamonds); rare spinel in late grown Type 3 Cpx and in rim of Type 1 Cpx (filled green circles); rare spinel in Fo_{93} Type 1 olivine (half-filled squares). Also shown are compositional fields for spinel of the Trinity Ophiolite (Quick 1981) and experimental spinels (black crosses) (Grove et al. 2003).

on an anhydrous basis and close to 1000 °C if magmatic water content approached ~3 wt% (cf. Ruscitto et al. 2011). How this magma likely formed is discussed below.

Textural and compositional observations support the notion that many disequilibrium components observed in the HMA hybrid magma were assembled shortly prior to generating the liquid for this late-stage crystallization and thus shortly prior to eruption. For example, crystallization of late (micro)phenocrysts and formation of outermost overgrowths is inferred to be rapid. The sharpness of interfaces separating core from overgrowth in Type 1 and 2 pyroxene and the similarities in compositional profiles for fast (Fe, Mg) and slow (Cr) diffusing elements across this transition imply that the overgrowths developed rapidly (Figs. 5 and 6). Diffusional relaxation times were estimated (cf. Costa et al. 2008) to simulate observed compositional profiles across the coreovergrowth transition for selected pyroxenes depicted in Figure 6. For these calculations, a temperature of 1100 °C was assumed (based on our two-pyroxene geothermometry); this is consistent with estimated entrapment temperatures for melt inclusions in Fo₈₇ olivine (Ruscitto et al. 2011). Calculations were made using the Costa spreadsheet modified with diffusivities appropriate for clinopyroxene (Costa et al. 2008; Costa, personal communication). Optimal fits were obtained for relaxation times on the order of a few months (~1 month for pyroxene bP6, Fig. 6a; ~3 months for pyroxene bP10, Fig. 6b). A short timescale for magma interaction is also supported by (1) narrowness (~10 µm; cf. Fig. 5a) of euhedral pyroxene and olivine selvages on earlier formed ferromagnesian minerals. These are consistent with growth in ~100 days at realistic crystallization rates (ca. 10⁻¹⁰ cm/s; Cashman 1993), and (2) development of hourglass zoning in pyroxene microphenocrysts



90

80

70

X

FIGURE 12. FeO* vs. mol% An compositions of plagioclases. Key: large crystals that are phenocryst-sized or glomerocrystic (L); HMA groundmass crystal (G); rare plagioclase inclusions in Mg#77 pyroxene (I); and plagioclase composition from experiments at 0.1 and 200 MPa (E; Grove et al. 2003). Composition range for plagioclase from Mt. Shasta dacites is indicated by horizontal arrow. Vertical arrows indicate predicted An contents for plagioclases hypothetically in equilibrium with indicated K_D values that increase with magmatic water content (Sisson and Grove 1993); solid arrows with HMA composition magma and open arrows for dacitic melt inclusion (left) and of HMA groundmass (right) of Table 1. The observed plagioclase compositions preclude their formation from HMA magma under water-rich conditions.

90

80

70

grd-HMA15

ard-HMA1



FIGURE 13. Map of crystals and vesicles in thin section of sample MS-16-04. All pyroxene crystals $>70 \ \mu\text{m}$ in size were classified and color coded according to the textural criteria discussed in the text. Histograms show distribution of pyroxene types according to size. An interactive version of this map provides access to BSE images and compositional data for analyzed crystals can be accessed at: https://www.pdx.edu/geology/streck-Shasta-HMA. See also Supplemental¹ Figure A8 for high-resolution BSE image of entire thin section.

consistent with rapid crystallization of sub-alkaline magmas (cf. Brophy et al. 1999; Skulski et al. 1994).

Restriction of plagioclase to the groundmass, and its absence as a phenocryst indicates that the late liquid was undersaturated with respect to feldspar until groundmass began to crystallize. This could be a consequence of elevated water content in the late-stage magma, which is estimated to be about 3% (Ruscitto et al. 2011). Also, the absence of oxide phenocrysts (and paucity of oxide minerals in the groundmass) suggests that magmatic redox conditions were too reducing to stabilize these minerals earlier. This observation is notable considering that redox conditions are near the fayalite-magnetite-quartz buffer (i.e., already relatively reduced) for typical Cascades mafic lavas (Smith and Leeman 2005; Rowe et al. 2009).

In contrast, Ruscitto et al. (2011) estimate that selected melt inclusions record relatively oxidizing redox conditions (oxygen fugacity near NNO+1.5 log units) based on measurements of sulfur speciation. Under such conditions, we would expect HMA-like liquids to be well saturated in oxide phases. This is not evident from the observed mineral assemblage and raises a question as to origin of the melt inclusions. We note that five of six analyses of S speciation were made on melt inclusions from ca. Fo₉₃ olivines, all of which, according to our evaluation, are considered to be xenocrystic (cf. Ruscitto et al. 2011).

Evidence for silicic (dacite) contributions: Glimpses into the murky magmatic past

Inference of precursor events and magmatic history must be gleaned from sparse clues based on the population of "nonequilibrium" crystals in HMA. Type 1 pyroxenes clearly represent a penultimate magmatic stage as, texturally, they post-date Type 2 pyroxenes. Fe-rich cores of Type 1 pyroxenes provide direct

evidence for involvement of felsic melt components in HMA. However, variations in zoning patterns between different subtypes indicate multiple contributions and a complex assembly history. Type 1A low-Mg#(59–75) pyroxene cores are consistent with their precipitation from melts having Mg# between 30-47 (assuming bulk K_D near 0.3; Sisson and Grove 1993). More directly, compositions of rare melt inclusions in Type 1A pyroxene cores implicate a parental liquid of roughly dacitic composition (Table 1). These melt inclusions provide evidence that all low Mg# pyroxene (i.e., Type 1A cores) crystallized from silicic melts with low FeO* and MgO contents (e.g., ~4 and ~1-2 wt%, respectively), albeit our estimated Mg#s are lower than those of typical Mt. Shasta dacites (cf. Grove et al. 2005). Low Cr contents in Type 1A pyroxene cores (150-700 ppm in Cpx and ~200 ppm in Opx; Supplemental¹ Table A6) are also consistent with their formation from low-Cr liquids not unlike Shasta dacites (that typically contain ca. 40-70 ppm Cr; Grove et al. 2005). Furthermore, relatively low Al₂O₃ contents in these pyroxene cores imply that they formed at low pressures (Fig. 16, see later discussion) and effectively preclude sub-crustal storage and crystallization of the silicic host melt. Calculated REE patterns for melts in equilibrium with low-Mg# Cpx cores (Fig. 14) display large negative (ca. 0.8-0.4) to negligible Eu anomalies. The negative Eu anomalies strongly implicate involvement of feldspar in the generation of the melts from which Type 1A pyroxene cores precipitated. Interestingly, such inferred silicic melts are more evolved (lower Mg# and more fractionated REE) than observed for most Mt. Shasta dacites (Grove et al. 2005). Nevertheless, some Type 1A cores as well as some Type 1B and 1C pyroxene cores have compositions (e.g., Mg# ~75; REE systematics, Fig. 14) consistent with their formation from melts similar to erupted Shasta dacites.

Another question concerns the progressive normal zonation observed in Type 1B and 1C Opx and Cpx cores (Mg# varies between ~82-75 from interior to outermost part, except in Type 1B that includes some portion with Mg# 75). This type of zonation is seemingly consistent with progressive fractional crystallization of the host melt, but could also be explained by mixing of the host melt with a more evolved (lower Mg#, essentially dacitic) magma. The possibility that the Fe-rich cores were derived from disaggregated cumulate or plutonic wall rocks rather than from a magma is considered unlikely based on the presence of melt inclusions and the preservation of relict euhedral outlines in some Type 1 cores, both of which are consistent with their original formation as phenocrysts in a silicic melt. These features, in combination with short time estimates for the crystallization of high Mg# overgrowths, are consistent with crystal growth from dacitic melt shortly before this magma mingled with other components to produce the final HMA end product. We conclude that mixing to form the hybrid HMA involved two or more distinctive and contemporaneous magmas in addition to entrainment of crystalline wall rocks (cf. Ruscitto et al. 2011), to yield antecrysts of dacitic origin.

Finally, we note that similar petrographic and mineralogical features are not uncommon, and have been interpreted in terms of injection of relatively hot mafic magmas into crustal-level melt reservoirs or crystal mush zones beneath other arc volcanoes (cf. Baggerman and DeBari 2011; Price et al. 2012; Humphreys et al. 2013). For example, iron-rich cores and complex reverse zoning in pyroxenes from Ruapehu are consistent with mixing between

mafic and more silicic magmas and/or reaction between entrained "dacitic" pyroxenes and invasive mafic magma (Nakagawa et al. 2002).

Evidence for incorporation of crustal-derived crystal cargo

As far as entrainment of crystalline wall rocks is concerned, we do recognize several distinct crystal populations that we consider unlikely to have been derived directly from magmatic sources. These are interpreted to be disaggregated xenocrystic debris of likely crustal derivation. In particular, it has been speculated that a possible source for such debris in Shasta lavas could be the Trinity Ophiolite that is inferred to underlie the Mt. Shasta area (cf. Baker et al. 1994).



FIGURE 14. (a) REE profiles for Type 1 clinopyroxene cores and rims, (b) calculated melts in equilibrium with Type 1 core compositions using D_{REE} in clinopyroxene (see text) and HMA inclusions and groundmass (Table 1, Fig. 1), and (c) representative Shasta dacite lavas (Grove et al. 2003) and whole-rock HMA averages of our data and of sample 85-41b whole rock (cf. Supplemental¹ Table A1). Stippled field in **b** shows range for Shasta dacites using data of Grove et al. (2003). Note that REE profile for HMA whole rock is similar to that of Shasta dacites.

Exotic pyroxenes. Distinctive disequilibrium features (e.g., resorption channels and patchy zoning) in all Type 2 Opx crystals are taken as evidence for their derivation from coherent country rocks that were infiltrated by melt along cracks and grain boundaries. The pervasive presence of coarse spongy textures instead of a progressive increase in width and intensity of the resorption channels toward crystal rims are features consistent with melt infiltration. Compositions of the overgrowths and infillings (~En₆₅) implicate involvement of a significantly less magnesian melt than the late HMA host, but instead more like the inferred dacitic melt postulated to generate Cpx with Eu anomalies. We suggest that features of Type 2 pyroxene formed in a nearly solid (or crystal mush-like) environment prior to incorporation of Type 2 pyroxenes into the same magma where Type 1 pyroxenes resided. Subsequent crystallization of high Mg# overgrowth on both Type 1 and 2 pyroxenes ties these crystal populations together and suggests that the enveloping melt was initially slightly more magnesian than melt inclusions within Fo₈₇ olivines (Ruscitto et al. 2011) or the HMA groundmass to explain beginning rim overgrowth with Mg# >87 (Table 1; Figs. 5 and 6). We note that REE patterns and incompatible element concentrations of HMA and those of Mt. Shasta dacites (Fig. 14; Supplemental¹ Table A1 and Table 1) are nearly indistinguishable and overlap with those inferred from Type 1 pyroxenes. These observations allow the interpretation that the bulk magma was essentially dacitic prior to the final hybridization/mixing assembly that produced the HMA (i.e., increased MgO, Cr, and Ni and subtly decreased Si, and other incompatible elements).

Exotic olivines. A country rock origin is also inferred for Type 3 olivine clusters. Irregular crystal interfaces and diverse optical orientations of the aggregate grains suggests that these inclusions are micro-xenoliths akin to those observed macroscopically in hand sample. Disaggregation of the xenoliths accounts for isolated grains having similar composition and morphology. The abundance and distinctive (for most) compositions of spinel grains in cross-cutting Opx-rich intergranular seams (Figs. 9 and 11b) is notable considering the paucity of spinels in HMA, and these features support an exotic origin for the micro-xenoliths (Fig. 10). Compositions of Type 3 olivines are intermediate between Types 2 and 1 olivines, but similar to those of many primitive Cascades basalt phenocrysts (Conrey et al. 1997; Smith and Leeman 2005; Rowe et al. 2009). Thus, they could be derived from olivine-rich cumulate sequences beneath the Cascades arc-either in the underlying crust or upper mantle-or from underlying accreted mafic-ultramafic bodies such as the Trinity Ophiolite (e.g., Quick 1981). Considering that the modern Cascade volcanic belt rests on a stack of accreted convergent margin terranes of Paleozoic to Mesozoic age, there could be many sources for the lithic debris found in HMA and other Mt. Shasta lavas (cf. Grove et al. 2005; Snoke and Barnes 2006).

Type 1 olivines are also considered to be exotic based on their conspicuously large size, their anhedral to subhedral (and sometimes fragmental) morphology prior to development of late euhedral overgrowths of Fo_{87} olivine, and their highly refractory composition (~Fo₉₃). Such Fo₉₃ olivines largely plot outside the Sobolev et al. (2005) field for olivines in mantle-derived rocks (cf. Fig. 10). These features distinguish them from all other olivines in HMA (and in all other Shasta lavas as far as we are aware), and rule out an equilibrium relationship to the late hybrid melt. Recent discovery of rare melt inclusions with compositions distinct from the late-stage host melt (Ruscitto et al. 2011) and lower δ^{18} O values compared to other minerals in HMA (Martin et al. 2011) led these authors to support our xenocrystic interpretation (cf. Streck et al. 2007a). However, the origin of these olivines (and their melt inclusions) remains enigmatic.

We doubt that these olivines are true phenocrysts precipitated from any known mafic magma in the Shasta area. If magmatic, their forsteritic composition would imply formation from a liquid with very high Mg# (near 84, assuming $K_D = 0.3$ and normal redox conditions), and such melts could only be derived from a highly refractory source not commonly recorded in mantle-derived xenoliths. Highly forsteritic olivines have been found elsewhere, and attributed to formation from highly oxidized magmas wherein much of the iron occurs as Fe3+ (e.g., Cortes et al. 2006; Blondes et al. 2012). However, highly oxidizing magmatic conditions would favor precipitation of magnetite-rich spinels and, as noted earlier, spinel inclusions are rare in Type 1 olivines and would not be consistent with the precipitation of the observed sulfides (Stone et al. 1989). Alternatively, highly forsteritic olivines (up to ~Fo₉₆) can form by prograde heating of serpentinite (e.g., Vance and Dungan 1977; Evans 2008) in which much of the protolith iron is oxidized and sequestered in secondary magnetite segregations. So, another possible source is from entrainment of (or melt reaction with) serpentinite bodies in the underlying crust. This scenario is supported by the redox state of melt inclusions in Type 1 olivines (as discussed above).

Relict plagioclase: Further evidence for evolved components

The sparse occurrence of strongly resorbed grains and aggregates of phenocryst-sized plagioclase within tephras of the Whaleback Quarry further implicates involvement of evolved magmas and/or plutonic contaminants in HMA magma. We note that observed plagioclase compositions overlap those in Shasta dacites, whereas they are generally too An-poor to have recrystallized from a melt with Ca/Na similar to that in HMA (Fig. 12).

Comparison of observational and experimental petrology of HMA

Despite overwhelming evidence in support of a hybrid or mixing origin for HMA as presented here and proposed in earlier studies (Anderson 1974; Streck et al. 2007a, 2007b, 2007c), it is instructive to review and contrast the observed mineral assemblage with phases produced experimentally from HMA melt composition (Müntener et al. 2001; Grove et al. 2003, 2005). Experiments have been conducted over a range of conditions spanning uppermost mantle (i.e., 1.5 and 1.2 GPa; Müntener et al. 2001; Grove et al. 2003) to crustal conditions (800–0.1 MPa; Grove et al. 2003). The starting material used in most experiments is sample 85-41c, which is nearly identical in composition to our sample MS-HMA11 (Fig. 3c, Supplemental¹ Table A1). However, experiments at 1.5 GPa used a modified HMA composition to which an excess of 2.5% FeO was added (Grove et al. 2003).

If a natural partial melt of HMA composition crystallized during ascent to the surface, it is reasonable to expect that observed phenocrystic mineral phases and compositions should correspond with those produced at appropriate experimental conditions. Figure 15 (also see Fig. 1) illustrates the experimental liquid line of descent with which we compare our mineral database.

As a general rule, near-liquidus mineral assemblages for all high-pressure (1.2 and 1.5 GPa) experiments comprise only pyroxenes (Opx±Cpx), whereas low-pressure experiments produced mostly olivine, Opx or Cpx, plus spinel or plagioclase or no Alrich phase. Both Opx and Cpx (with Mg $\# \le 86$) crystallized in low pressure experiments along with plagioclase after residual liquid SiO₂ content exceeded ~59 wt%. In most cases the experimental mineral assemblage does not correspond with the natural assemblage (e.g., olivine or a pyroxene is missing, or plagioclase is present). Only one 200 MPa run, with 75% liquid remaining, closely replicated the observed liquidus assemblage olivine + Opx + Cpx + spinel (Fig. 15a). However, in this example, Mg#'s (82-84) for the experimental pyroxenes and olivine are lower than observed in the natural minerals (Fig. 15b). No experimental olivine comes close to the Fo₉₃ composition of Type 1 olivines. Natural pyroxenes also have significantly higher maximum Cr2O3 contents than any experimental ones (Fig. 16). Al₂O₃ contents for high-Mg# (>82) pyroxenes are comparable except for those in the 1.5 GPa experiment. We note that only low-pressure experiments reproduced naturally observed Al contents for low-Mg# pyroxene cores (Fig. 16). Although we argue here that the cores cannot have crystallized from HMA magma, this observation is consistent with low-pressure conditions for their formation.



FIGURE 15. Summary of phase assemblages and compositions of experiments using Mt. Shasta HMA (85-41c) as starting composition. Data are from Müntener et al. (2001) for water undersaturated experiments at P = 1.2 GPa and temperatures ranging from 1230 °C to 1030 °C, and from Grove et al. (2003) for experiments under dry conditions at P = 0.01, water-saturated conditions at 0.2 GPa, and water-undersaturated conditions at 1.5 GPa over temperatures ranging from 1300 to 940 °C.

It is particularly noteworthy that oxide inclusions in Type 1 pyroxene cores are ilmenites whereas oxide phases that crystallized in HMA experimental liquids (e.g., 61.7 wt% SiO₂ melt of HMA #85-41c; Grove et al. 2003) are chromian spinels; such spinels also occur in the actual HMA groundmass (Fig. 11). This evidence indicates that Type 1 pyroxene cores are not genetically linked to a HMA parent or its derivative liquids. They are most plausibly derived from dacitic liquids of unrelated origin. Spinels appear to be generally sparse in the experiments, as is also the case for the natural samples we have examined. Spinels in one 200 MPa experiment are compositionally similar to spinel inclusions that occur in late crystallizing (i.e., phenocrystic, Mg# ca. 87) olivine and pyroxene as well as groundmass spinel. On the other hand, no experimental spinels are similar to those included in Fo₉₃ olivines (Fig. 11). Finally, plagioclase is only common in relatively evolved experimental melts or at low (near-surface) pressure. This jives with the observation that "equilibrium" plagioclase in natural HMA is restricted to groundmass crystals.

Overall, most of the experiments fail to replicate first-order



FIGURE 16. Comparative compositions of pyroxenes from experiments using Mt. Shasta HMA (#85-41c) as starting composition and naturally occurring pyroxenes of this study. Experimental data are from Müntener et al. (2001) and Grove et al. (2003); see Figure 15 for more details.

mineral compositional or assemblage features documented in the natural samples. Specifically, they do not support the premise that the observed mineral assemblage could be in equilibrium with a melt of HMA composition. This is not surprising considering the documented presence of clearly xenocrystic minerals in HMA.

Evidence from melt inclusions

Two recent studies investigated melt inclusions in olivines from the HMA (Ruscitto et al. 2011), and from high-alumina olivine tholeiites (HAOT), calcalkalic basalts, and basaltic andesites (BA) from the Mt. Shasta area (Le Voyer et al. 2010). Ruscitto et al. (2011) discerned two melt components in HMA minerals. A low-Ca (or HMA-like) component is trapped in Type 2 (~Fo₈₇) olivines and thus is considered representative of the melt from which all late-stage phenocrystic or overgrowth minerals crystallized. The average composition of such melt inclusions is slightly more magnesian than HMA groundmass, but has incompatible trace element abundances similar to those in our HMA groundmass average (Table 1). After correction for post-entrapment crystallization (PEC) to the host olivine composition (ca. Fo_{87}), olivine equilibration (trapping) Ts are calculated to be near 1150 °C for these inclusions (cf. Supplement¹ B, Fig. B3). A second high-CaO melt component (primitive basaltic andesite, or PBA) is restricted to inclusions in Type 1 olivines (ca. Fo₉₃), and differs from all other bulk lava or melt inclusion compositions so far recognized at Mt. Shasta (Supplement¹ Fig. A6). After PEC-correction to the host olivine composition, compositions of these inclusions correspond to higher trapping Ts (\sim 1190–1260 °C, assuming X-fe3 = 0.15; cf. Supplement II). This PBA melt component is also distinguished by relatively high concentrations of H₂O (maximum ca. 3.5 wt%), Cl (>0.2 wt%) and Sr (most >1000 ppm), but low contents of Ba (most <100 ppm), K (ca. 0.3 wt%), and other incompatible elements (Supplemental¹ Fig. A6); Cl/K ratios (ca. 0.9) are distinctly higher than observed in melt inclusions from Shasta basalts (<0.2; Le Voyer et al. 2010). We also note that PBA melt inclusions are heterogeneous in composition, with large variations in major elements (SiO₂, MgO, CaO, Al₂O₃, etc.), Cr (ca. 200-4000 ppm), Sr (860-1200 ppm), and B (2-40 ppm) (cf. Supplemental¹ Figs. A3 and A6). Considering the elements K, Ti, and P (all of which are incompatible in olivine), we note that the PBA inclusions have distinctly higher P₂O₅/TiO₂ (ca. 0.5) and lower K₂O/TiO₂ (ca. 0.6) ratios than observed in either the HMA-type inclusions or HMA bulk lavas or groundmass (ca. 0.2 and 1.2-1.5, respectively, Supplemental¹ Fig. A5). Because these ratios should not be affected by corrections for post-entrapment crystallization, they provide robust evidence for the exotic nature of the PBA component. The average PBA major element composition is considered similar to experimental hydrous melts in equilibrium with harzburgite (Ruscitto et al. 2011). Also, given their compositional distinctiveness, the source(s) of PBA-like melts must differ from sources for other known mafic magmas in the vicinity of Mt. Shasta (e.g., basaltic andesite (BA) and high alumina olivine tholeiite [HAOT) melt inclusions of Le Voyer et al. 2010).

According to the latter authors, the nearly anhydrous (<0.1% H_2O) HAOT inclusions are interpreted to represent melts of a relatively dry lherzolite mantle source. In contrast, the relatively hydrous (ca. \geq 1.5% H_2O) BA inclusions are attributed to melting of a more depleted mantle source that previously had been

metasomatized by subduction fluids and melts (thus stabilizing amphibole). Trapping Ts for HAOT- and BA-type inclusions are estimated to be near 1330 and 1250 °C, respectively. Leeman et al. (2005) deduce similar segregation Ts for Shasta basaltic lavas at uppermost mantle depths, and infer that HAOT source domains reside deeper in a compositionally stratified mantle. Ruscitto et al. (2011) estimate that PBA melts were trapped at Ts between 1200–1280 °C—consistent with our calculations as noted above. Thus, if BA- and HAOT-type magmas result from shallow melting of the mantle it is unclear where the alleged PBA source fits in spatially and compositionally.

We recognize that mantle heterogeneities may exist at relatively small scales allowing proximal magmas to acquire a spectrum of distinctive mantle signals (cf. Conrey et al. 1997; Borg et al. 2002; Leeman et al. 2005). But if source differences are attributed to variable enrichments by subduction-related fluids, a more important issue concerns why the PBA melt source is selectively enriched in some (e.g., Sr, Cl) but not all (e.g., Ba, K, Rb) fluid-mobile elements (cf. Supplemental¹ Fig. A6). A possible explanation is that the melt inclusions reflect inherent heterogeneities associated with accretion of arc terranes in this region, which formed intermittently since the early Paleozoic, such that specific melt compositions likely depend in detail on protolith composition. In this case, the PBA protolith appears to have affinities with some serpentinites and/or seawater-altered oceanic crustal rocks. Furthermore, we can envision that potential melt sources (mantle or crust) may evolve over time due to successive melt extraction (i.e., depletion) vs. metasomatic (i.e., selective enrichment) "events."

As an alternative origin for the PBA inclusions that is consistent with their unusual composition, we suggest that they could be products of melt interaction with accreted and hydrated ultramafic basement terranes comprising water- and Mg-rich phases like serpentine, tremolite, and magnesite. Melting of these materials at some scale could be triggered by intrusion of hot mafic magmas at appropriate crustal depths as documented in the roots of exposed volcanic arc/subduction complexes (e.g., DeBari et al. 1999; Greene et al. 2006; Horodyskyj et al. 2009). We predict that such melts could be moderately water-rich (if formed at sufficiently high pressure, say in the deep crust), highly magnesian, and could mix with the intrusive magma to produce PBA-like liquids that, in turn, could be trapped in highly forsteritic olivine formed under relatively oxidizing, prograde metamorphic conditions (cf. Vance and Dungan 1977) within a wall-rock aureole. Furthermore, such melts need not be volumetrically significant. Thus, we postulate that Type 1 forsteritic olivines with their unique PBA-type melt inclusions could be derived from disaggregated wall-rocks and entrained in the compositionally distinct HMA magma-perhaps shortly before its eruption (so as to allow growth of the thin reaction rims of Fo₈₇ olivine).

This scenario is consistent with several observed features. For example, if serpentinized ultramafic massifs undergo prograde metamorphism, it has been shown in numerous cases that the resulting olivine can be extremely forsteritic (e.g., Evans 1977, 2008; Pedrón et al. 2010); this could explain the unusually high Fo content and mostly anhedral habit of the olivine mineral hosts of the PBA inclusions. In this regard, and considering their heterogeneous nature, the melts may have never been in equilibrium with Fo₉₃₋₉₄ olivines; if so, correcting original PBA inclusions by

adding on average 20% olivine back to reach equilibrium with the host olivine (cf. Ruscitto et al. 2011) could be excessive. Furthermore, fluids derived from such massifs can be chlorine-rich if they originally were derived from seawater-altered rocks (e.g., Scambelluri et al. 2001; Garrido et al. 2005; Kodolányi and Pettke 2011), and this could explain the high Cl (and Sr) content of the high-Ca (PBA) inclusions found in Type 1 olivines (Supplemental¹ Fig. A6). It is noteworthy that there are no known occurrences of erupted PBA-like lava, whereas BA and HAOT lavas are relatively common near Shasta and elsewhere in the modern Cascades; thus, there is no compelling reason to believe that magmas with PBAlike compositions are volumetrically significant.

Generation of the silicic end-member (and other Shasta dacites)

Compositions of Fe-rich pyroxene cores and the melt and mineral inclusions they host (Anderson 1974; Stone et al. 1989; Streck et al. 2007a; Ruscitto et al. 2011; this study) implicate involvement of silicic (dacitic) magma as an essential component in forming HMA. Thus, the manner by which the dacites formed is relevant to discussion of the HMA. Specifically, what is the nature of the evolved melt component in HMA, and what is its relationship to other evolved magmas in the vicinity of Mt. Shasta? Although they clearly recognized that most Shasta dacites and andesites are to some degree mixed magmas, Grove et al. (2005) proposed that they are products of moderate pressure (near water-saturated) fractional crystallization predominantly derived from high magnesian andesite and, to lesser extent, more mafic mantle inputs. This interpretation is stressed in subsequent papers (e.g., Magna et al. 2006; Martin et al. 2011). As previously discussed (also see Fig. 1), the experimentally determined liquid lines of descent or phase assemblages do not provide conclusive support for this view, and several other processes are also inferred to be important-most notably, magma mixing as evidenced by heterogeneous mineral populations and textural relations in most samples.

The latter view is consistent with independent geochemical evidence. For example, U-series and trace element analyses of Shasta lavas (Newman et al. 1986; Volpe 1992) strongly implicate magma mixing processes in their formation, and define distinct mixing relations associated with each growth stage of the volcanic complex. The U-series data are notable in that they show relative enrichments of [230Th/232Th] with respect to [238U/232Th] (here, square brackets indicate activity ratios). This is significant because (1) the isotopic disequilibria imply recent disturbances in the U-series systematics of the magma sources, (2) the lack of significant ²³⁸U-enrichment in Shasta (and other Cascades) lavas seemingly precludes involvement of recently extracted subduction zone fluids (whereas many young lavas from other volcanic arcs do show significant ²³⁸U-enrichment; cf. Turner et al. 1997; George et al. 2004; Zellmer et al. 2005), and (3) varied ²³⁰Thenrichment in the respective mixing end-members may be related to dynamic melting processes unrelated to subduction per se. In particular, the second point implies that input of subduction zone fluid is relatively low for the Cascades in general, and this can be attributed to the unusually warm thermal structure of this arc (cf. Harry and Green 1999; Green and Sinha 2005; Leeman et al. 2005; Walowski et al. 2015).

As seen in our Figure 1, the Shasta lavas define relatively strong

near-linear arrays in plots of major element oxides vs. SiO₂, albeit there is notable diversity among the mafic lavas-as observed at most Cascades volcanoes (cf. Leeman et al. 1990, 2005; Bacon et al. 1997; Borg et al. 1997). Nevertheless, these arrays form "chords" connecting mafic (basaltic) and silicic (dacitic) endmember compositions on otherwise curvilinear trajectories defined by loci of relevant experimentally produced fractionation/melting liquids. These chords approximate linear mixing arrays analogous to those observed for the 1915 Lassen eruption, that are attributed to mixing of basaltic and dacitic magma end-members (Clynne 1999). These relations imply the coeval existence of distinct reservoirs of mafic and evolved magma that can comingle and mix in a fairly reproducible manner. Some scatter is to be expected in plots that represent multiple eruptive stages over extended duration (ca. 250 ka for Shasta; Christiansen et al. 1977), because the nature of the mixing end-members may vary to some degree over time. Importantly, the HMA groundmass (or near-equivalent melt inclusions) lies close to the main Shasta arrays as defined by basalts, andesites, and dacites, whereas the bulk HMA lava compositions diverge from the Shasta arrays in a manner consistent with entrainment of xenocrystic materials (largely olivines and pyroxenes) of ultramafic provenance as noted earlier (cf. Streck et al. 2007a). In our view, the relict vestiges of dacite affinity (e.g., the Fe-rich pyroxene cores) are inherited from dacitic magma(s), the bulk of which were otherwise assimilated by hot mafic liquids to produce the HMA magma.

However, the compositional spectrum of Shasta dacitic lavas is characterized by a wide range in trace element concentrations that precludes a simple petrogenesis for such magmas. For example, over a narrow range of SiO₂ (62-64 wt%) Sr and Ba concentrations in Shasta dacites range more than twofold, and are negatively correlated (Fig. 17a); andesitic compositions overlap but are more scattered. To model the dacite array by fractional crystallization (FC), a high degree of fractionation (>50%) is required and plagioclase must contribute throughout to maintain a sufficiently high Sr bulk partition coefficient (>1) to account for decreasing content of this element. This process could also lower Sr/Y progressively (Fig. 17b) because Sr content decreases while Y content remains near constant or increases slightly. Thus, if the dacites evolved in this manner, it is implicit that parental dacitic magmas must initially be characterized by elevated Sr and Sr/Y ratios. However, there would also be a significant concomitant increase in SiO₂, which is inconsistent with the narrow SiO₂ range for the dacite array. There is no quartz, alkali feldspar, or low An plagioclase ($<An_{20}$) in any of the Shasta dacites (cf. Grove et al. 2005) that could contribute to buffering the SiO₂ of bulk rock within the 62-64% silica window during the required high degrees of fractionation to explain the dacite array. It has been proposed that such magmas could form either by partial melting of subducted oceanic crust (Kay 1978; Drummond and Defant 1990), or by melting mafic amphibolite protoliths in the deep crust (e.g., Smith and Leeman 1987; Baggerman and DeBari 2011; Qian and Hermann 2013).

Alternatively, the nearly fourfold range in dacite (or andesite) Sr/Y values could partly reflect source heterogeneities (i.e., different protoliths or different depths of melting; cf. Gao et al. 2016 for review), resulting in high- and low-Sr variants (also seen at Crater Lake; Bacon and Druitt 1988). In this case, the near-linear



FIGURE 17. Plots showing (**a**) Ba vs. Sr and (**b**) Sr/Y vs. Sr of dacite and andesite lavas of Mt. Shasta area (Baker et al. 1994; Grove et al. 2002) and bulk, groundmass, and melt inclusion data for HMA (see Supplemental¹ Table A1). Arrows show effects of 50% fractional crystallization using the indicated bulk distribution coefficients (see text for discussion).

trends in Figure 17 could reflect mixing of diverse dacitic magmas within a complex storage/feeder system at depth.

Other evidence against a sole fractionation control on compositional variations of Shasta dacites is the sub-vertical trend in a Sr vs. Al_2O_3 diagram defined by dacites and all basaltic Shasta magmas that could potentially serve as fractionating parental magmas (Supplemental¹ Fig. A6). Plagioclase-free fractionation at greater crustal depth could enhance Sr but, in this case, derivative compositions would not trend vertically but toward increased Al_2O_3 content. Second, if plagioclase is removed at shallower depth, vertically decreasing Sr trends would be unlikely as well (cf. Grove et al. 2005). Considering these difficulties in explaining the dacite arrays solely in terms of magma fractionation, it seems that partial melting processes are implicated in generating Mt. Shasta dacitic magmas.

Dacitic magmas at other Cascades volcanoes (and many other places) commonly are interpreted to be partial melts of crustal lithologies such as amphibolites (e.g., Smith and Leeman 1987; Borg and Clynne 1998; Baggerman and DeBari 2011). Hence, partial melting scenarios to produce dacite magmas at Shasta are a viable alternative (or at least complementary) to fractional crystallization (cf. López et al. 2005; Qian and Hermann 2013), particularly if large volumes of mafic magma stagnate in the lower crust (cf. Tiepolo et al. 2011, 2012; Yu and Lee 2016). A notable feature of Mt. Shasta dacites is that they mostly have high Mg# (ca. 54-64, except a few at Mg#~25; Grove et al. 2002, 2005). It could well be that high Mg# is a reflection of the source protolith, possibly enhanced by higher oxygen fugacity (cf. Sisson et al. 2005; Gao et al. 2016). Protolith and partial melting conditions may vary so as to produce relatively silicic melts with varied trace element contents. For example, slight positive correlation between Mg# and Sr, and a negative correlation with Ba in Shasta dacites might suggest that high Sr, low Ba dacites are derived from higher Mg# protoliths and vice versa.

Oxygen isotope constraints on formation of HMA

Controversy over the significance of Shasta HMA has encouraged alternative approaches to understand the origin of this (and similar) magma. For example, Martin et al. (2011) measured oxygen isotopic composition of olivines and pyroxenes from several HMA samples as well as other Shasta basaltic and dacitic lavas. Unfortunately, mineral bulk composition (i.e., Mg#) is not available for the specific samples (usually aggregates of several grains) analyzed for δ^{18} O. Melt-equivalent δ^{18} O values calculated from the mineral analyses for HMA (6.6-6.9%) partly overlap with those for HAOTs (6.7-6.9‰) and BAs (6.3-6.7‰), and are lower than values for dacites (7.2-7.4%). All of these values are considered higher than expected (ca. 5.5-6‰) for "normal" mantle-derived magmas, and this is taken to reflect input of 18Oenriched components from the subducting slab in this region (Martin et al. 2011). In contrast, a more detailed investigation of the nearby Mt. Lassen volcanic field concluded that elevated δ^{18} O values there (~5.8–6.6‰ for basaltic lavas and ~6.0–8.3‰ for more evolved lavas) primarily reflect assimilation or melting of preexisting crustal rocks, and that intrusion of mafic magmas resulted in hybridization of the preexisting crust (Feeley et al. 2008). Of particular relevance to our study, the overlap of estimated compositions for mafic melts seems permissive of some mixing between basaltic and dacitic magmas, particularly if the precursors to the basaltic lavas had normal mantle δ^{18} O values (ca. 5.5-6‰). Data reported for only two "xenocrystic" HMA olivines (4.8-5.0%) are lower than for olivines from other Shasta mafic lavas (5.3-5.8‰ for BAs, and 5.9-6.1‰ for HAOTs). Because compositions of rocks and minerals from the Trinity Ophiolite range widely (4-10%; Lécuyer and Fourcade 1991; Martin et al. 2011, and references therein), the xenolithic olivines could be derived from such sources. However, given the latitude in end-member compositions, we conclude that the oxygen data cannot adequately resolve mixing relations in the Shasta HMA. Furthermore, considering the work of Feeley et al. (2008), the role of crustal contributions at Shasta may be underestimated.

Possible mixing scenarios to yield Mt. Shasta HMA

In an attempt to quantify the mixing relations that produced HMA magma, several scenarios were evaluated using major element compositions of hypothetical mixing components. In reality, the number of components that have been discerned as well as variability (and in some cases, uncertainty) in their compositions makes it difficult to precisely accomplish this goal. The basic model to be tested concerns the proposition that HMA liquid is a mixture of some basaltic magma with a dacitic magmatic component. The latter appears to be the source of very high Sr in HMA. Also, the bulk HMA magma appears to have contributions of an ultramafic nature to account for observed xenocrystic crystals and also high contents of transition metals. In practice, for this component we used a composite ultramafic component comprising the average of Trinity Ophiolite samples with more than 37% MgO using data from the literature (Jacobsen etal. 1984; LaPierre et al. 1987; Gruau et al. 1998) combined with three samples from Castle Lake area (cf. Supplemental¹ Table A7). Initial models were tested by systematically involving all variants of basaltic lavas from the Shasta region, coupled with average dacite, and then the dacite components were broken down into averages for each of the four eruptive phases of Mt. Shasta (cf. Grove et al. 2002, 2005). Conceptually, we envisioned that injection of a specific type of basalt into a reservoir of dacitic magma could trigger a

mixing process and perhaps related eruption event as proposed for many other volcanoes. Also included in the models were dominant crystal types observed in HMA samples, either as phenocrysts or xenocrysts. Combinations of these components were incorporated into a linear mixing algorithm and mixing proportions determined that provided optimal fits to the target composition of HMA, its groundmass, and the HMA-type melt inclusions. Optimization was achieved by minimizing the sum of the squared residuals using the SOLVER tool in Excel (cf. Herrmann and Berry 2002). A drawback to this approach is that the matrix solutions can be rapidly overdetermined if the number of components exceed ~50% of the number of elements used in the models, in which case solutions are not reliable. So, after initial screening of potential mixing components, the number used in any given model was restricted to five or less. In practice, this was accomplished by omitting minor mineral phases that contributed little to the target composition, or that effectively mimicked contributions of similar components (e.g., olivine or pyroxene variants).

Initial results indicate that optimal fits could be obtained using a calcalkalic basalt variant (represented by sample 85-44; cf. Grove et al. 2002); the HAOT variants produced inferior fits. For the dacite end-member, all of the aforementioned averages were tested, as they varied in subtle ways and it is clear from the similarity in REE (and other incompatible element) contents of HMA and the dacites, that they likely were a dominant ingredient in HMA. The most successful models are summarized in Table 2. All have low values for ΣR^2 (most <0.5), and thus closely reproduce the target compositions. However, nature is vastly more complex that our simple models. Thus, we stress that these solutions are sensitive to the specific choice of mixing end-members and to the hypothetical scenarios being tested. Principal observations are that the HMA groundmass (i.e., devoid of phenocrysts or xenocrysts) can be modeled essentially by mixing basaltic (CAB) and dacitic magmas, with roughly 80-84% dacite depending on the average composition used. Here and with other scenarios, use of the average dacites of Sargents Ridge age produces distinctive results with greater proportion of dacite, but also poorer least-squares fits; for this reason they are considered least likely (also note that they have distinctively higher ⁸⁷Sr/⁸⁶Sr than than HMA; cf. Supplemental¹ Fig. A1). Basalt proportions range from about 13-20%, and the most successful models include addition of Cpx (~8%) and removal of xenocrystic olivine (~5%). Models for the HMA-type melt inclusions indicate that they can be reproduced by mixing of basalt and dacite in subequal proportions with small additions of Cpx and loss of olivine. However, these models have consistently poorer fits, and this could reflect systematic differences in the compositional data (obtained by different methods) as well as the heterogeneity of the melt inclusions. Finally, HMA whole rock compositions are best modeled by combination of dacite (~56%) and basaltic (\sim 30%) magmas plus ultramafic material (\sim 5%) and Type 2 pyroxene (~8%); ΣR^2 values for these models are among the best obtained. In summary, the proposed mixing scenarios can reasonably account for bulk compositions of HMA products, and confirm a dominant contribution of dacitic material to the mixtures.

As an independent test of these results, we used the predicted mixing proportions and concentrations of nominally incompatible elements in the mixing components to simulate the trace element compositions of HMA products. We avoided use of transition metals in this exercise because their concentrations are highly variable and difficult to constrain for many of the model end-members. Comparisons between calculated and observed compositions are shown for all models (excluding those with Sargents Ridge dacite) in Figure 18. Overall, average misfits are about 15% for the HMA WR and groundmass models, and 22% for the melt inclusion model. Because uncertainties on the target compositions are almost as large, these are considered to be successful fits. The largest misfits are for Rb, Sr, Th, and U, and they could be rectified in large part by using a handpicked dacite composition in the models. In conclusion, compositions of the HMA variants are readily reproduced by plausible mixing scenarios.

Summary interpretation of HMA origin

The high-magnesian andesite from the Mt. Shasta area contains a complex array of equilibrium and disequilibrium mineral components. Based on the aggregate information now available, the HMA clearly represents a hybrid mixed magma as proposed by Streck et al. (2007a). It cannot be considered as a primitive direct melt from the mantle, but how far removed it is remains an important point of contention. Nor does the petrologic or experimental evidence support the formation of HMA simply by differentiation of mafic parental magmas. Its origin is most plausibly the result of mixing between mafic and silicic magmatic components coupled with contamination of crustal wallrocks. The felsic magmatic components could in part be products of melting of previously emplaced or underplated mafic magmas in the lower crust or uppermost mantle.

The earliest discernable components are attributed to formation and crystallization of silicic (dacitic?) magma as recorded by pyroxene crystals with low Mg# cores and by rare melt inclusions in these cores. Normally zoned cores of Type 1 pyroxenes could signify that the host magma evolved from a more mafic liquid and/or interacted or mixed with more evolved magmas. Considering the low-Al contents of the pyroxenes and evidence for entrainment of disaggregated melt-infiltrated country rocks,

TABLE 2. Summary of least-squares mixing models for HMA based on major elements

Model	Dacite	Dacite	BA 85-44	Trinity	Срх	Fo87	An60	Σ(Res) ²				
	variant			UM	Mg#85			,				
Type: HMA whole-rock average												
	Avg. all	0.569	0.306	0.051	0.083	-0.006	-	0.240				
	Hotlum	0.583	0.277	0.050	0.091	-	-	0.166				
	Shastina	0.563	0.311	0.048	0.077	-	-	0.191				
	Misery	0.554	0.306	0.050	0.091	-	-	0.229				
	Sargents	0.622	0.234	0.054	0.097	-	-	0.370				
HAOT 85-38												
	Avg. all	0.636	0.230	0.042	0.096	-	-	0.311				
Type: HMA low-Ca melt inclusions (HMA type)												
	Avg. all	0.515	0.507	-	0.041	-0.052	-0.006	0.467				
	Hotlum	0.526	0.500	-	0.046	-0.051	-0.016	0.460				
	Shastina	0.507	0.543	-	0.031	-0.055	-0.023	0.415				
	Misery	0.499	0.541	-	0.043	-0.055	-0.025	0.463				
	Sargents	0.563	0.442	-	0.054	-0.045	-0.006	0.573				
Type: HMA groundmass												
	Avg. all	0.821	0.137	_	0.080	-0.045	0.008	0.379				
	Hotlum	0.838	0.126	-	0.087	-0.044	-0.008	0.295				
	Shastina	0.806	0.196	-	0.063	-0.050	-0.020	0.385				
	Misery	0.794	0.191	-	0.083	-0.049	-0.022	0.296				
	Sargents	0.898	0.030	-	0.100	-0.033	0.009	0.456				

we infer that these magmas resided in a crustal reservoir and that the country rock comprised abundant Opx and olivine (i.e., harzburgite to dunite protoliths). Some of the dacitic melt (or precursors thereof) infiltrated ultramafic(?) country rock leading to pervasive resorption, patchy zoning, and occasionally low-Mg# overgrowths on orthopyroxene. Mobilization of selvages of altered and resorbed country rock dispersed abundant xenocrysts (estimated as 1/3 of all single crystals in HMA), micro-xenoliths and any partial melt that resulted from mineral breakdown reactions. Many of the entrained crystals have been disaggregated and comminuted during transport. This mobilization was possibly facilitated by recharge of new mafic magma. Subsequent mixing between resident evolved magmas, country rock-derived crystal debris, and possibly newly emplaced mafic magma can account

and comminuted during transport. This mobilization was possibly facilitated by recharge of new mafic magma. Subsequent mixing between resident evolved magmas, country rock-derived crystal debris, and possibly newly emplaced mafic magma can account for the final bulk composition and petrologic character of the erupted HMA magma. Late-stage cooling of interstitial liquid in this magma can explain the presence of skeletal to more euhedral Type 2 olivines, nearly unzoned Type 3 pyroxene phenocrysts, and high-Mg# overgrowths on previously entrained "dacite-derived" and "country-rock-derived" pyroxenes and olivines. This final crystallization interval must have occurred shortly prior to (and perhaps contemporaneous with) final magma ascent, and preceded eruption by months at most.



FIGURE 18. Calculated incompatible trace element compositions for (a) HMA whole rock (WR) average, (b) HMA Groundmass, and (c) average HMA melt inclusion (MI) using mixing model proportions calculated from major element data (Table 2; see text). All are normalized to the observed compositions of these materials. Multiple profiles shown in each case correspond to different assumed dacite end-members (see text). Thickness of horizontal gray lines approximates analytical uncertainty on the HMA components (cf. Supplemental¹ Table A1 and Table 1).

IMPLICATIONS

Basaltic magmatism in volcanic arcs is widely attributed to melting of mantle domains variably modified by inputs of fluids or melts derived from subducting oceanic crust and/or sediments. Globally, there is considerable diversity among such magmasranging from almost MORB-like low-K tholeiitic compositions to a spectrum of more potassic calcalkalic and even alkalic compositions (cf. Gill 1981). In the Cascades and several other arcs, basalts of OIB affinity are common (Leeman et al. 1990, 2005; Bacon et al. 1997; Conrey et al. 1997; Righter 2000; Green 2006; Rowe et al. 2009). Such diversity may stem from many factors, but for the Cascades two distinct generic interpretations have been favored. One camp sees the compositional spectrum as essentially the product of flux-melting with the strength of calcalkalic signatures increasing with magnitude of aqueous flux addition to the source (cf. Borg et al. 1997; Reiners et al. 2000; Rowe et al. 2009). Alternatively, Leeman et al. (2005) suggested that the mantle wedge may be compositionally stratified, with low-K tholeiites and OIB-like basalts derived from little-fluxed domains in the deeper convecting asthenospheric part of the mantle, whereas the calcalkalic magmas formed by melting shallower, more rigid lithospheric mantle that had been modified previously by additions of melts and fluids over the 40 Ma lifetime of the arc. The presence of more refractory olivine and spinel compositions in primitive calcalkalic basalts supports the notion that they are derived from distinct sources (Smith and Leeman 2005).

More evolved andesitic to dacitic and even rhyolitic rocks comprise significant volumes in most volcanic arcs, particularly those sited on relatively mature continental margins (cf. Leeman 1983). Origins of these magmas are undoubtedly varied. They may be derivative from more mafic magmas, represent partial melts of local crust, or be produced by some combination of these processes. Ongoing debate on this topic is motivated by interest in using magma chemistry to understand physical and chemical processes in subduction zones, their potential contributions to crustal formation or modification, and the effects of subduction on recycling of evolved materials to the deeper mantle. For example, if primary arc magmas are predominantly basaltic, then net removal of a mafic to ultramafic component from such material is required to produce the inferred "andesitic" composition of average continental crust (cf. Kay and Kay 1993; Lee et al. 2006).

Alternatively, the occurrence of high Mg-number (Mg#=molar 100 Mg/[Mg+Fe]) and esitic (or "HMA") lavas in some arcs suggests the possibility that significant volumes of arc crust may be derived directly from the mantle (Kelemen et al. 2003; Tatsumi 2006; Straub et al. 2011). It has also been proposed that some HMA magmas could be melting products of subducted basaltic oceanic crust (Kay 1978; Yogodzinski et al. 2015), particularly if subduction zone temperatures are sufficiently high (e.g., where young lithosphere is being subducted or tectonic conditions favor advective heating; e.g., Defant and Drummond 1990; Yogodzinski et al. 2001). The latter scenario predicts melting of eclogitic rocks under conditions such that the melts would attain distinctive garnet-dominated chemical signatures (notably, MORB-like or primitive radiogenic isotopic compositions, enrichment of LREE and elevated Sr/Y, La/Yb, etc.-so-called "adakitic" signatures; Drummond and Defant 1990; Martin et al. 2005; Moyen 2009; Zellmer 2009). The occurrence of such lavas has been documented

in many arcs (Yogodzinski and Kelemen 1998; Kelemen et al. 2003; Danyushevsky et al. 2008; cf. Wood and Turner 2009), including rare examples in the Cascades (Anderson 1974; Grove et al. 2002).

However, the importance of such magmas to crustal growth is uncertain for the following reasons. First, the occurrence of significant volumes of basaltic lavas in virtually all intraoceanic arcs (Gill 1981; Leeman 1983) implies that these are the dominant product of mantle melting, whereas HMA magmas are subordinate in volume in most arcs. This appears to be the case for the Cascade arc. Moreover, Tolstykh et al. (2017) document the rarity of adakitic melt inclusions in adakitic (HMA-like) rocks worldwide. Second, HMA-type rocks occur in various geologic settings (including greenstone belts, collision zones, etc.) other than active arcs, which suggests that they may be produced in multiple ways (cf. Richards and Kerrick 2007; Moyen 2009; Qian and Hermann 2013). Third, many HMA-type magmas bear evidence of a hybrid origin (e.g., Tatsumi et al. 2006; Rodriguez et al. 2007; Qian and Hermann 2010; Straub et al. 2011), in which case it is unclear to what extent they represent primary magmas.

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Endnote:

¹Deposit item AM-18-26151, Supplemental Figures and Tables. Deposit items are free to all readers and found on the MSA web site, via the specific issue's Table of Contents (go to http://www.minsocam.org/MSA/AmMin/TOC/2018/Feb2018_data/Feb2018_data.html).