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# Volatile contents of mafic magmas from cinder cones in the Central Oregon High Cascades: Implications for magma formation and mantle conditions in a hot arc

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#### ABSTRACT

The Cascades arc is a warm-slab subduction zone characterized by the slow (3.5 cm/a) subduction of young (<10 Ma) oceanic crust. Elevated slab temperatures likely result in shallow dehydration beneath the forearc and reduced slab fluid addition to the mantle wedge beneath the volcanic front. Resulting calc-alkaline magmas should contain low volatile abundances relative to arcs with cooler subducting slabs. We test this hypothesis in Central Oregon, the region of the arc that has experienced the highest mafic output over the last 2 Ma. Naturally quenched basalt and basaltic andesite melt inclusions from tephra erupted from monogenetic vents record pre-eruptive magmatic volatile contents. After accounting for the effects of degassing, our data show that initial volatile contents from seven calc-alkaline reciped from the subducted slab ( $H_2O/Ce \sim 490-1700$ ; S/Nb ~110-310; and Cl/Nb ~37-190) are elevated above MORB values but are low compared to those from the southern Cascades (Mount Shasta) and other arcs (Mexico, Central America, and Kamchatka). Positive correlations between  $H_2O$ , Pb/Ce and Ba/Zr suggest that small additions of a slab-derived component to the mantle wedge are responsible for observed volatile contents.

Meet inclusion compositions are evolved ( $Po_{82-84}$  nost onlyines, 5.0–7.2 Wt.% MgO) and fractionated from primitive parental magmas by ~16% crystallization of predominantly olivine before being trapped as inclusions. Calc-alkaline basaltic and Sr-rich basaltic melts can be generated by 9–11% partial melting of spinel herzolite. H<sub>2</sub>O-rich basaltic andesite melts are highly depleted in incompatible elements and are probably generated from a more depleted mantle source by 12–16% melting. Subarc mantle temperatures inferred from H<sub>2</sub>O and trace elements are similar to or slightly warmer than temperatures in other arcs. Additionally, compositional proxies for slab surface temperatures in Central Oregon are at the high end of the global arc spectrum, corresponding to temperatures of 850–950 °C, and are consistent with a young, hot incoming plate.

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# 1. Introduction

Volatiles influence all aspects of subduction-related volcanic processes from mantle melting to eruption explosivity. Nearcontinuous dehydration reactions within the subducting slab release H<sub>2</sub>O and mobile elements into the overlying mantle (e.g., Schmidt and Poli, 1998), which depress the mantle solidus and induce melting (e.g., Grove et al., 2002). The Cascades volcanic arc is the global end member, warm-slab subduction zone ( $\phi/100 \sim 1.5-2.2$  km; where  $\phi$  is the slab thermal parameter; e.g., Wada and Wang, 2009) resulting from the slow subduction (~3.5 cm/a) of the young Juan de Fuca plate

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(~6–9 Ma at the trench) beneath Western North America (Wilson, 2002). Kinematic models predict slab–mantle interface temperatures as much as 300 °C hotter than those calculated for older subducting lithosphere (Peacock, 2003). Higher temperatures should result in eclogite formation and significant slab dehydration beneath the Central Oregon forearc (<45 km depth), consistent with geophysical and geochemical observations (e.g., Leeman et al., 1990, 2005; Brocher et al., 2003; Hurwitz, et al., 2005; Rondenay et al., 2008). Loss of most volatiles beneath the forearc should result in reduced slab contributions to the mantle wedge below the arc as has been inferred in Southern Washington and Northern Oregon (e.g., Leeman et al., 2005). In contrast, significantly higher volatile contents than expected are found in magmas from the southern part of the arc (e.g., Mt. Shasta; Anderson, 1974; Baker et al., 1994).

We measured pre-eruptive volatile (H<sub>2</sub>O, CO<sub>2</sub>, S, and Cl) contents in olivine-hosted melt inclusions from tephra deposits in the Central Oregon Cascades to examine the effects of a young subducting slab on magma generation in arcs. Volatile contents of three geochemically distinct mafic magma types are presented and compared with other

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arcs encompassing a wide range of slab thermal conditions. We also estimate the compositions of primary, mantle-derived magmas and mantle sources to infer how slab surface temperatures and degrees of mantle melting relate to  $H_2O$  flux into the mantle wedge.

## 1.1. Geologic setting and sample context

Intra-arc extension has promoted mafic volcanism in Central Oregon over the past ~5 Ma in the form of shield volcanoes and smaller monogenetic cones (Hildreth, 2007). The latter provides evidence for at least five distinct types of primitive magma (Hughes and Taylor, 1986; Bacon et al., 1997; Conrey et al., 1997). Low-K tholeiites (LKT; synonymous with MORB-like, high-Al olivine tholeiites, HAOT) and ocean-island-like basalts (OIB; also EIB or HFSE-enriched) are thought to represent decompression melts of depleted and enriched mantle domains, respectively (e.g., Hughes, 1990; Leeman et al., 1990; Conrey et al., 1997). Primitive calc-alkaline basalts (CAB) and rare, alkali-rich basalts (absarokites and shoshonites) represent melts generated from the addition of slab-derived fluids to the overlying mantle wedge (Leeman et al., 1990; Sisson and Layne, 1993; Bacon et al., 1997; Conrey et al., 1997; Rowe et al., 2009). Despite having elevated FeO<sup>T</sup>/MgO, subalkaline basalts in the Cascades have enrichments in Sr, K, and Ba, and depletions in Nb and Ta and have been classified as calc-alkaline by multiple authors on the basis of trace element ratios (e.g., Ba/Nb > 20; (Sr/P)<sub>PM</sub>>1.8; Leeman et al., 2005; Rowe et al., 2009). The timing of fluid addition and whether CABs are derived from lithospheric or asthenospheric mantle, however, are the subjects of debate (Leeman et al., 2005; Hildreth, 2007). Calc-alkaline basaltic andesites (BA) are also common and result from differentiation of basaltic parents or melting of highly depleted (harzburgitic) mantle fluxed by slab-derived fluids (Hughes and Taylor, 1986; Hughes, 1990; Conrey et al., 1997).

## 1.2. Sample descriptions and analytical methods

Samples studied here are from tephra deposits (mostly Holocene; Sherrod et al., 2004) erupted from monogenetic vents in the Santiam– McKenzie Pass area, Central Oregon Cascades (Fig. 1). We focused on



**Fig. 1.** Southern portion of the Cascadia subduction zone after Haugerud (2004). The Cascades volcanic arc is defined by the major peaks shown as labeled, red triangles (J: Mt. Jefferson; TS: Three Sisters; NB: Newberry volcano; S: Mt. Shasta; ML: Medicine Lake volcano). A yellow box between Mt. Jefferson and the Three Sisters volcances marks the study area shown as an inset. Cumulative mafic volcanic output along the arc over the past 2 Ma is shown in the histogram to the right (Sherrod and Smith, 1990). Inset is a shaded relief map of the Santiam–McKenzie Pass (study area) in Central Oregon with vent locations shown (symbols as in Fig. 2; TC: Twin Craters, GB: Garrison Butte, SM: Sand Mountain, IF: Island Fissure, Y: Yapoah, C: Collier, BM: Blue Lake Maar). Prominent volcanic peaks are labeled: TFJ (Three-Fingered Jack), MW (Mount Washington), and NS (North Sister). Graben-bounding faults (black), major highways (dark gray), and the town of Sisters (black square) are also highlighted.

calc-alkaline basalt and basaltic andesite samples to determine their relationships to subduction processes. Most of the tephras contain phenocrysts of olivine + plagioclase (and rarely pyroxene). Spinels occur as mineral inclusions inside most olivines. Tephra from Garrison Butte (GB072) contains only olivine (+ spinel inclusions). The tephra sample from the Island Fissure (PT05) was from a subglacial eruption and, despite strong palagonitization, contains unaltered olivine phenocrysts with glassy inclusions.

Loose olivine crystals were picked from tephra samples. Olivines were treated in HBF<sub>4</sub> to remove attached matrix and allow inspection under refractive index oil. Fully enclosed glassy inclusions were selected and host crystals were ground and polished into wafers. Melt inclusions were analyzed for H<sub>2</sub>O and CO<sub>2</sub> using a Thermo-Nicolet Nexus 670 FTIR spectrometer interfaced with a Continuum IR microscope and subsequently mounted in epoxy for electron microprobe and laser ablation ICP–MS analyses.

H<sub>2</sub>O and CO<sub>2</sub> concentrations were calculated from IR peak absorbances using the Beer–Lambert law and appropriate absorption coefficients. Electron microprobe analyses were performed using a Cameca SX-100 at the University of Oregon. Hydrous glass compositions were measured with a 10 nA, 10 µm diameter beam and 15 kV accelerating voltage (see details in Supplementary material). Individual inclusion analyses are averages of 3-5 spots and are reported in Table S1. Host olivine compositions were determined using a 15 kV accelerating voltage and a 50 nA, 2 µm diameter beam. Three to five point analyses of olivine (>100 µm away from inclusions and crystal edges) were averaged for each crystal to obtain a representative olivine composition. Trace elements in melt inclusion glasses were analyzed at Oregon State University with a VG PQ ExCell quadrupole ICP-MS equipped with a NewWave DUV 193 nm ArF Excimer laser system using a 70 µm spot size and a 4 Hz pulse rate. Measured trace element concentrations were determined by reference to BCR-2G glass as a calibration standard and using <sup>43</sup>Ca as an internal standard (see Kent et al., 2004 for details). BHVO-2G glass was analyzed to monitor accuracy and precision, and the analyzed values were within 10% of accepted values. Oxygen isotopes in olivine from several tephra samples were measured at the University of Oregon using CO<sub>2</sub>-laser fluorination and an MAT 253 mass spectrometer. Major and trace elements for bulk tephra and bomb samples (Table S1) were determined by XRF and ICP-MS analyses at the Washington State University GeoAnalytical lab.

# 2. Results

#### 2.1. Major and trace element chemistry

Olivine host crystals are mostly Fo<sub>82-84</sub> and have oxygen isotope values within the range of and slightly heavier than values for mantlederived basalts (Eiler, 2001; Table 1). The melt inclusions are medium-K calc-alkaline basalt and basaltic andesite in composition (Group II in Leeman et al., 2005). Inclusions from a given tephra sample have variable FeO<sup>T</sup> contents at similar MgO and are out of Fe-Mg equilibrium with host crystals. Therefore, melt inclusions have been numerically restored back to equilibrium with their host crystals using the method of Danyushevsky et al. (2000), which accounts for both olivine post-entrapment crystallization and diffusive Fe-Mg exchange. The maximum FeO<sup>T</sup> observed for each set of melt inclusions was assumed to be representative of the initial composition. A constant  $f_{O2}$  of NNO + 0.25 was assumed based on the average S–FeO<sup>T</sup> relationship in Fig. S1. Melt inclusions required an average of 2.8 wt.% (maximum 7.9 wt.%) olivine addition to be restored to equilibrium with host olivines (Table S2).

Major and trace element compositions of the melt inclusions are similar to whole-rock compositions of basalts and basaltic andesites in Oregon (Conrey et al., 1997; Rowe et al., 2009). Calc-alkaline melt inclusions are subdivided into three groups based on composition: (1) Table 1

Averaged primary melt compositions.

Source	Twin crater 121°52.808		Garrison butte		Island fissure		Sand mtn 121°44.370		Yapoah 121°46.603		Collier		Blue lake maar 121°46.448	
Long (W)														
Lat (N)	44°15.283		44°23.242		44°12.543		44°23.651		44°13.155				44°24.514	
wt.%	Average $n = 10$	1 s.e. <sup>a</sup>	8		11		3		4		5		12	
SiO <sub>2</sub>	50.01	0.19	48.03	0.17	50.36	0.06	49.73	0.05	52.19	0.51	52.18	0.49	51.66	0.16
TiO <sub>2</sub>	1.31	0.02	1.29	0.02	1.20	0.01	1.27	0.06	0.86	0.03	0.96	0.08	0.85	0.02
$Al_2O_3$	15.58	0.17	14.89	0.17	15.30	0.08	15.59	0.09	15.72	0.32	16.46	0.40	14.72	0.17
FeO <sup>1</sup>	9.24	0.02	10.75	0.02	9.50	0.01	9.12	0.05	8.23	0.10	7.75	0.02	8.99	0.02
MnO	0.12	0.01	0.13	0.01	0.13	0.01	0.11	0.01	0.11	0.02	0.09	0.01	0.10	0.00
MgO	11.04	0.04	12.89	0.03	11.37	0.03	10.89	0.08	11.24	0.15	10.51	0.06	12.44	0.06
CaU No. O	8.72	0.08	8.51	0.04	7.63	0.05	8.89	0.08	7.18	0.35	7.67	0.13	6.98	0.08
Nd <sub>2</sub> U	3.10	0.11	2.75	0.04	3.33	0.05	3.21	0.01	3.62	0.00	3.58	0.05	3.58	0.03
R <sub>2</sub> U	0.00	0.02	0.45	0.02	0.81	0.02	0.87	0.00	0.07	0.08	0.59	0.04	0.55	0.02
F205	0.201	0.01	0.34	0.01	0.37	0.01	0.55	0.01	0.18	0.02	0.20	0.04	0.140	0.003
Class	0.134	0.002	0.120	0.001	0.131	0.001	0.185	0.005	0.098	0.001	0.101	.005	0.145	0.007
H <sub>2</sub> O <sub>max</sub>	1 59	0.12	1 45	0.20	2.28	0.001	2.26	0.20	2.60	0.000	2.33	0.16	2.96	0.21
$CO_{2max}$ (ppm)	1234	74	823	75	938	63	943	71	738	58	718	27	694	50
T <sub>formation</sub> (°C)	1243	40	1282	40	1250	40	1229	40	1233	40	1222	40	1254	40
Mg# Ol	90		90		90		90		91		91		91	
% Ol add.	13.84	0.23	16.41	0.45	15.75	0.31	12.2	0.47	16.90	0.78	14.86	0.44	19.89	0.581
Lee T (°C)	1273	38	1331	40	1266	38	1260	38	1239	38	1227	38	1269	38
Lee P (GPa)	1.19	0.2	1.59	0.2	1.25	0.2	1.28	0.2	1.05	0.2	1.11	0.2	1.25	0.2
	n = 5		3		4		4		3		1		3	
$\delta^{18}\text{O ol}~(\text{‰})$	5.236	0.040	5.324	0.075	5.704	0.056	5.345	0.030	5.435	0.085	5.568	-	5.201	0.017
ppm	n = 7		4		7		2		4		4		6	
Sc	24.2	1.7	25.1	0.5	19.8	0.8	24.8	0.1	21.3	0.5	21.9	1.2	20.3	1.2
V	210	5	183	5	189	4	169	3	170	5	163	7	184	8
Cr	96	21	90	11	77	2	260	145	101	20	69	9	58	4
Rb	6.4	0.3	4.9	1.0	13.0	0.5	8.4	0.7	8.80	1.72	7.86	0.51	7.8	0.6
Sr	514	9	454	12	616	16	903	2	501.46	29.10	443.73	19.77	505	4
Y	21.4	0.5	20.7	0.6	19.1	0.5	21.4	0.6	14.36	0.64	14.26	2.19	12.1	0.9
Zr	111	2.5	120	8.5	119	3.3	135	2.3	83.00	5.36	79.31	11.18	62	5.4
ND	6.3	0.2	8.3	0.2	12.2	0.3	12.0	1.5	5.69	0.67	6.89	1.40	4.6	0.4
Ва	318 115	8 05	189	4	413	14	347	1	229.99	16.31	201.07	11.92	210	12
La	11.J 32.7	0.5	28.1	0.4	39.5	1.0	17.8	0.2	21.28	0.39	18.05	1,14	17.4	1.2
Pr	44	0.1	3.8	0.7	47	0.2	55	0.2	21.20	0.52	2.56	0.30	23	0.1
Nd	17.7	0.1	15.4	0.1	20.5	0.2	22.2	0.0	11.01	0.10	10.92	1 32	97	0.1
Sm	4.3	0.5	3.6	0.2	3.7	0.1	4.8	0.3	2.36	0.20	2.46	0.27	2.0	0.1
Eu	1.4	0.1	1.3	0.1	1.6	0.1	1.4	0.0	0.92	0.07	0.98	0.09	0.8	0.0
Gd	4.1	0.5	3.6	0.2	3.6	0.2	4.1	0.4	2.43	0.15	2.69	0.28	2.2	0.3
Dy	4.5	0.4	3.7	0.1	3.0	0.2	4.2	0.1	2.70	0.15	2.57	0.37	1.7	0.1
Er	2.2	0.1	2.2	0.1	2.0	0.1	2.1	0.0	1.44	0.13	1.30	0.22	1.1	0.1
Yb	1.9	0.1	1.8	0.0	1.7	0.2	1.8	0.0	1.31	0.14	1.21	0.16	1.1	0.1
Та	0.35	0.05	0.45	0.02	0.61	0.05	0.77	0.18	0.28	0.06	0.40	0.09	0.26	0.05
Pb	2.9	0.1	2.8	0.3	5.4	1.0	3.7	0.4	3.11	0.25	2.63	0.09	2.8	0.2
Th	1.02	0.06	0.62	0.04	1.15	0.05	1.28	0.05	0.80	0.13	0.77	0.05	0.61	0.06
U	0.38	0.02	0.26	0.03	0.62	0.04	0.40	0.02	0.32	0.06	0.33	0.02	0.36	0.05

basalts from Twin Craters (TC063A) and Garrison Butte (GB072), (2) Sr-rich basalts from the Island Fissure (PT05) and Sand Mountain (SM2), and (3) basaltic andesites (technically North Sister-type; Hughes and Taylor, 1986) from Yapoah (Y0061E and Y0065D), Collier (EJY02) and Blue Lake maar (BL1 and BL3) (Fig. 2). Restored inclusions contain 4.5–7.6 wt.% MgO and 48.2–55.5 wt.% SiO<sub>2</sub> (Table S2). Basalt and Sr-rich basalt compositions have similar major and high-field strength element (HFSE) abundances. Sr-rich basalts are distinguished from the basalts by higher alkali and large ion lithophile element (LILE) concentrations (Fig. 2). In addition to higher SiO<sub>2</sub> (>52 wt.%) and lower MgO (<5.5 wt.%), basaltic andesites are distinguished from the two types of basalts by lower incompatible element abundances, especially the HFSE and heavy rare earth elements (HREE). LILE abundances, however, are similar for both the basalts and the basaltic andesites. Basaltic andesites, on average,

have more pronounced Nb-Ta depletions than either of the two basalt groups.

## 2.2. Magmatic volatile contents

Dissolved H<sub>2</sub>O contents are higher in the basaltic andesites (3.6 wt.%  $H_2O_{max}$ ) than in either of the basalt types (1.8 [basalts] and 2.7 wt.%  $H_2O_{max}$  [Sr-rich basalts]). CO<sub>2</sub> contents range from below detection ( $\leq$  50 ppm) to 1400 ppm, with the highest concentrations in the basalts (Fig. 3). Maximum S contents are 1500 (basalts), 2100 (Sr-rich basalts), and 1800 ppm (basaltic andesites). An increase in maximum Cl is observed from the basalts (900 ppm) and Sr-rich basalts (1000 ppm) to the basaltic andesites (1300 ppm) (Fig. 3). Highly dissolved S contents in many samples require more oxidizing conditions ( $fo_2 \ge NNO$ ; Fig. 3c) than those inferred by Rowe et al. (2009) (NNO-1 to NNO). Basaltic



**Fig. 2.**  $K_2O$  and  $TiO_2$  contents of restored melt inclusions illustrating compositional differences between three types of mafic magmas studied (basalts [Twin Craters and Garrison Butte] as circles ['Bas'], Sr-rich basalts [Island Fissure and Sand Mountain] as squares ['Sr-Bas'], and basaltic andesites [Yapoah, Collier and Blue Lake Maar] as triangles ['Band']). Shown for comparison are high-Mg andesite melt inclusions from the Whaleback–Deer Mountain (WB–DM) cinder cone near Mt. Shasta (black circles in white field; Ruscitto and Wallace, 2008) and average melt inclusion compositions from Rowe et al. (2009) (OIB = ocean-island-like basalt, LKT = low-K tholeiite, and CAB = calcalkaline basalt).

andesite inclusions from Yapoah and Collier have the lowest  $\text{FeO}^{T}$  and S. Olivines from some Collier samples contain  $\leq$  50 µm diameter sulfide globules, demonstrating sulfide-saturation.

Primary  $H_2O-CO_2$  variability in melt inclusions results from either degassing or gas fluxing prior to melt inclusion formation (e.g., Metrich and Wallace, 2008; Johnson et al., 2009). Additional variations in  $H_2O$  and  $CO_2$  may be attributed to post-entrapment diffusive loss of  $H_2O$  through the host crystal (e.g., Portnyagin et al., 2008) and  $CO_2$  partitioning into shrinkage bubbles (e.g., Anderson and Brown, 1993). To constrain pre-eruptive volatile contents in each sample suite, only melt inclusions trapped above 100 MPa (Table S2) were considered, and of these inclusions, the maximum  $H_2O$ , S, and Cl contents identified in each suite were taken to represent the minimum pre-eruptive concentration for these mafic magmas (cf., Johnson et al., 2009).

## 3. Discussion

#### 3.1. Comparison of volatile contents in the Cascades with other arcs

Magmatic volatile contents in Central Oregon are elevated compared to MORB and overlap with values for many other arcs, including Kamchatka and the Marianas, which have considerably older subducting slabs ( $\phi/100 = 47.8$  [Kamchatka] and 36.7 km [Marianas]; Wada and Wang, 2009) (Fig. 3). Despite these overlaps, the maximum magmatic H<sub>2</sub>O contents in Central Oregon are not as high as maxima in other arcs. H<sub>2</sub>O and CO<sub>2</sub> in the Central Oregon basaltic andesite inclusions overlap those in Mt. Shasta high-Mg and esites determined by FTIR ( $\leq$  3.4 wt.%; Ruscitto and Wallace, 2008), but H<sub>2</sub>O contents are considerable lower than those inferred by Anderson (1974) using the H<sub>2</sub>O-by-difference method ( $\leq 6$  wt.%). We note that the data from Shasta come from inclusions hosted in Fo<sub>94</sub> olivines and are therefore not directly comparable to the Central Oregon melt inclusions until further corrections for fractionation are made (see below). Central Oregon magmatic volatiles are most similar to those in behind the volcanic front magmas from Guatemala and are intermediate between those for oceanic back-arc basin lavas and mafic magmas from other arcs (Fig. 3a and b). As in many other arcs, H<sub>2</sub>O and Cl contents in Central Oregon are well explained by mixing between a mantle source and small amounts of fluid with Cl/H2O similar to seawater (Fig. 3b; Kent et al., 1999; Wallace, 2005; Johnson et al., 2009). Sulfur contents are similar to those in melts from other arcs and back-arc basins (Fig. 3c).



Fig. 3. Dissolved volatile contents in restored olivine-hosted melt inclusions. (a) H<sub>2</sub>O and CO<sub>2</sub> contents from this study (symbols as in Fig. 2). Thick black lines are vapor saturation isobars (Papale et al., 2006). Fields enclose melt inclusion compositions from the following sources: CAVA (Central American Volcanic Arc) - Roggensack et al. (1997), Wade et al. (2006), Benjamin et al. (2007), Sadofsky et al. (2008); Mexico -Johnson et al. (2009); Guatemala BVF (behind the volcanic front) - Walker et al. (2003); K (Kamchatka) - Portnyagin et al. (2007), Auer et al. (2009); Marianas - Shaw et al. (2008); back-arcs - Shaw et al. (2004). Back-arc glass compositions are also shown (Stolper and Newman, 1994; Kelley et al., 2006; Bezos et al., 2009). (b) H<sub>2</sub>O and Cl, with the line labeled SW illustrating a mixing trajectory between a MORB mantle source and sea water (3.5 wt% NaCl) after Kent et al. (1999) and Kent and Elliot (2002). Data sources as in (a) with the Central American field separated into: Irazu and Arenal, Costa Rica, and CAVA = Guatemala, El Salvador, Nicaragua, (c) S vs.  $FeO^{T}$ , with the addition of MORB field (data from PetDB) and sulfide-saturation curves at NNO after Wallace and Carmichael (1992) modified using the temperature dependence from Mavrogenes & O'Neill (1999). Estimates of the total dissolved sulfur for NNO +0.25 and +0.50 were calculated by combining S<sup>Total</sup> calculated at NNO while assuming  $S^{6+}/S^{Total}$  ~0.21 (at NNO) and using equation (7) from Jugo (2009). Mexico and Kamchatka fields are shown for comparison; fields for other arcs are omitted for clarity but largely overlap with those shown.

## 3.2. Magmatic differentiation, mixing and crustal interaction

The mafic samples studied are not primitive and have evolved from parental magmas during transit through the crust. Dissolved volatile contents, therefore, cannot be related back to parent magmas or source compositions until differentiation processes are assessed. Minimum melt inclusion trapping pressures are 100–200 MPa (~3– 6 km depth; Papale et al., 2006) and constrain melt evolution to the mid- and lower-crust prior to melt inclusion formation (Moho ~40– 44 km; Trehu et al., 1994; Brocher et al., 2003; Hildreth, 2007). The extent to which the basalt and basaltic andesite magmas have interacted with the crust and/or mixed with one another is difficult to constrain. The diversity in primitive mafic magma types erupted at cinder cones throughout Oregon and Southern Washington suggests that homogenization by magma mixing and crustal assimilation are not dominant processes for small volume mafic eruptions, though they might be more important beneath larger, long-lived stratocones (Leeman et al., 2005; Hildreth, 2007; Mercer and Johnston, 2008). Evolved magmas in the Cascades could be heavily influenced by lower crustal melting (Hildreth and Moorbath, 1988; Conrey et al., 2001; Hildreth, 2007). However, studies of exposed island arc crustal sections (e.g. Kohistan and Talkeetna) demonstrate that intrusioninduced melting is subordinate to fractional crystallization processes (Kelemen et al., 2003; Greene et al., 2006; Jagoutz et al., 2009). U-series and Os isotope ratios of basalts from the Mt. Adams region also indicate that lower crustal interaction is relatively minor (<5 wt.%; Jicha et al., 2009). From these lines of evidence, we assume that the compositional evolution of basalts and basaltic andesites in Central Oregon is dominated by fractional crystallization.

It is important to consider the possible effects of crustal interaction on magmatic volatile contents, even if the extent of such interaction was minor. One possibility is a contribution from amphibole in lower crustal rocks. Minor assimilation (<10 wt.%) of bulk amphibolite (~50% modal amph; bulk rock 1 wt.% H<sub>2</sub>O) would contribute <0.1 wt. % H<sub>2</sub>O. Mixing between low-H<sub>2</sub>O mantle-derived magmas and dacitic partial melts of amphibolite (~6 wt.% H<sub>2</sub>O; Rapp and Watson, 1995) could potentially result in hybrid compositions similar to those of the basaltic andesite melt inclusions. However, high-pressure experiments on basaltic andesite compositions from the region imply last equilibration near the base of the crust between hydrous (~3.5 wt.%  $H_2O$ ) mafic melt and anhydrous gabbro (Mercer and Johnston, 2008). Therefore, any lower crustal interaction must have involved gabbro rather than amphibolite or dacitic partial melt. Low trace element abundances in the basaltic andesites have been interpreted to result from extensive interaction of mantle-derived LKT magma with gabbroic lower crust (Schmidt, 2005). LKT magmas, however, are interpreted to be volatile-poor (e.g. Sisson and Layne, 1993; Rowe et al., 2009), and gabbroic assimilation is unlikely to contribute significant  $H_2O$ ,  $CO_2$ , or Cl to the melt. Therefore, whatever the role of lower crustal input is, we conclude that the volatile contents in all three magma types are primarily controlled by mantle source variations.

#### 3.3. Mantle-derived parental melt compositions

Melt inclusions were restored to primary, mantle-derived compositions through incremental addition (0.1 wt%) of equilibrium olivine until the composition was in equilibrium with mantle olivine (Table 1). Primary basalt and Sr-rich basalt samples were assumed to have equilibrated with Fo<sub>90</sub> olivine. Basaltic andesite compositions were assumed to be from more refractory mantle (Fo<sub>91</sub> olivine) based on lower incompatible element abundances and higher SiO<sub>2</sub> compared to the basalts (Fig. 2). Mantle equilibration required an average



**Fig. 4.** Primary (mantle-derived) magma compositions (e.g.  $SiO_2^\circ$ ) calculated from melt inclusions in Central Oregon (major and trace elements are averages of restored, undegassed compositions; maximum H<sub>2</sub>O from each suite; see Table 1). Symbols as in Fig. 2, with the addition of Sisson and Layne (1993) data (averaged and restored to  $Fo_{90}$  equilibrium values as gray diamonds). (a, b) Linear correlations between H<sub>2</sub>O<sup>o</sup> and major elements (central OR samples from current study only with r<sup>2</sup> values shown) can be used to calculate H<sub>2</sub>O<sup>o</sup> and major elements (central OR samples from current study only with r<sup>2</sup> values shown for SiO<sub>2</sub><sup>o</sup>. OB and LKT He magma compositions of Rowe et al. (2009). Values shown for Rowe et al. samples are H<sub>2</sub>O<sup>o</sup> contents determined from averaging equations shown for SiO<sub>2</sub><sup>o</sup>. OB and LKT H<sub>2</sub>O<sup>o</sup> contents are extrapolated. (c, d) Correlations between H<sub>2</sub>O<sup>o</sup> and some common trace element proxies for water. H<sub>2</sub>O<sup>o</sup> shown for the Shasta High-Mg composition is based on the observed maximum concentration in olivine-hosted (Fo<sub>94</sub>) melt inclusions (6%, Anderson, 1974; 3.4%, Ruscitto and Wallace, 2008). Experiments suggest that H<sub>2</sub>O<sup>o</sup> in Shasta magmas may be as high as ~10 wt.% (e.g. Baker et al., 1994; Grove et al., 2003). Low H<sub>2</sub>O<sup>o</sup> value (~0.3 wt.%) from Sisson and Layne may represent a degassed sample. The two solid gray lines in panels (b), (c), and (d) represent flux-melting models of distinct enriched mantle sources using the relationship between amount of subduction component added to source (X<sub>SC</sub> in wt.%) and degree of melting (F as mass fraction) for back-arc basin basalts described in Kelley et al. (2006) (i.e. dF/dX<sub>SC</sub> = 0.48). The dashed gray line represents a flux-melting models begin at X<sub>SC</sub> = 0.0 wt.%, F = 0.05 and end at X<sub>SC</sub> = 0.06 wt.%, F = 0.18.

addition of 16 wt.% olivine. This approach may not fully account for effects of fractional crystallization on incompatible elements because some of the magmas were likely saturated in both olivine and Ca-rich pyroxene and potentially plagioclase (e.g., Sisson and Grove, 1993). However, hydrous, isobaric pMELTS simulations suggest that olivine is the dominant crystallizing phase until ~6 wt.% MgO (~13 wt.% crystallization) in the compositions of interest, and therefore the method of simulating olivine-only crystallization can approximate total extents of multi-phase crystal fractionation (Fig. S2). We emphasize that the MgO contents of the calculated primary melts are not as robust as the incompatible element concentrations.

Primary melt compositions (denoted by °) from Central Oregon have 1.5–3.0 wt.% H<sub>2</sub>O°, 1000–1500 ppm S°, and 450–1000 ppm Cl° (Table 1). The highest pre-eruptive H<sub>2</sub>O° and Cl° contents are in HFSEdepleted primary melts (parents of basaltic andesites), whereas the basalts contain the lowest H<sub>2</sub>O°. S° contents do not vary systematically between the compositions. Primary melt CO<sub>2</sub>° contents cannot be inferred from the melt inclusions because of pre-entrapment degassing but could be >0.6 wt.% (Wallace, 2005).

Correlations between H<sub>2</sub>O°, major elements, and indices of slab component addition are shown in Fig. 4. H<sub>2</sub>O° is positively correlated with SiO<sub>2</sub>°, Na<sub>2</sub>O°, Pb/Ce, and Ba/Zr and negatively correlated with TiO<sub>2</sub>°, CaO°, P<sub>2</sub>O<sub>5</sub>°, and HFSE°. Average primary (Fo<sub>90</sub>) melt compositions from Northern California are consistent with the Central Oregon trend (Sisson and Layne, 1993). We note that the low  $H_2O^{\circ}$ concentration (~0.3 wt.%) of the Black Crater sample (Sisson and Layne, 1993) is close to the solubility of H<sub>2</sub>O at 0.1 MPa and may reflect pre-entrapment degassing or post-entrapment diffusive loss. The high-Mg melt composition from the Mt. Shasta region extends the Central Oregon trend to higher SiO<sub>2</sub>° and H<sub>2</sub>O° and lower TiO<sub>2</sub>° (Fig. 4a and b). In contrast, estimated H<sub>2</sub>O° contents (see Table S3) of Central Oregon melt compositions from Rowe et al. (2009) plot at the opposite end of the trend. Less significant positive correlations are observed between H<sub>2</sub>O° and indices of slab addition (i.e., Pb/Ce and Ba/Zr) (Fig. 4c and d). Positive correlations observed between H<sub>2</sub>O°, Pb/Ce, and Ba/Zr appear consistent with a flux-melting model (Stolper and Newman, 1994; Kelley et al., 2006) and suggest that small additions of a slab-derived hydrous component into the mantle wedge are responsible for the observed volatile contents (Fig. 4). However, melt compositions require that either low (<1 wt.%) TiO<sub>2</sub>° melts are generated from high extents of melting (F > 0.15) of a single source or that the large variations in TiO<sub>2</sub>° result from distinct mantle sources (e.g., the high-Mg composition from Mt. Shasta region is in equilibrium with harzburgite rather than lherzolite; Baker et al., 1994; Ruscitto and Wallace, 2008). We investigate these competing models in more detail below.

#### 3.4. Mantle sources and melting conditions

Fig. 5 shows calculated pressures and temperatures of last equilibration between the primary melts and the mantle (Table 1; Lee et al., 2009). Most compositions have equilibrated near the base of the crust (~1.3 GPa), within the spinel stability field. LKT and OIB compositions (along with Garrison Butte) appear to have equilibrated slightly deeper (~1.6 GPa), and these results are consistent with Elkins-Tanton et al. (2001) and Leeman et al. (2005), though we note that pressure uncertainties are  $\pm 0.2$  GPa (Lee et al., 2009).

The diverse primitive magma types in the Cascades have been attributed to mantle heterogeneities that include an enriched OIB-like source, a depleted MORB-like source (with or without overprinting by a subduction component), and a less common ultra-depleted (harzburgite) source (Hughes and Taylor, 1986; Hughes, 1990; Conrey et al., 1997; Schmidt et al., 2008; Rowe et al., 2009). A range in mantle compositions is expected given the complex tectonic history of this region (Hildreth, 2007). However, unlike other segments of the arc, isotopic constraints (Sr and Nd) permit Central Oregon magmas



Fig. 5. Pressures and temperatures at which calculated primary magma compositions for Central Oregon (and WB-DM) are in equilibrium with lherzolite or harzburgite using the thermometer and barometer of Lee et al. (2009). Symbols as in Fig. 2, with the exception of the 6 wt.% H<sub>2</sub>O WB-DM magma (gray circle connected to black [3% H<sub>2</sub>O] with dashed line). Error bars on symbols are  $\pm$  0.2 GPa and  $\pm$  3% for calculated pressures and temperatures, respectively. Also shown are the P-T estimates of primitive melts in Northern California from Elkins-Tanton et al. (2001) (gray stars with error bars; note that the lowest pressure point from Elkins-Tanton et al., which represents a Shasta region high-Mg andesite, is within error of black circle WB-DM from Ruscitto and Wallace (2008). Light gray field represents Cascades whole-rock data (assuming 3 wt.% H<sub>2</sub>O) from Lee et al. (2009). Approximate depths to the Moho (vellow) and slab (blue) are also shown, depth axis was calculated assuming constant  $\rho = 2.8 \text{ g/cm}^3$ . Thick black line is the dry peridotite solidus (DPS) taken from Lee et al. (2009). Thick dashed lines represent hydrous peridotite solidi. Thin dot-dot-dash lines approximate melt fraction isopleths. Pressure of the Spinel (spl)garnet (gt) transition (horizontal dashed line) is labeled at 2.8 GPa. Thick red line (WW09) represents the geotherm calculated for Northern Cascadia in Wada and Wang (2009).

(except for shoshonites) to be derived from a single mantle source (Schmidt et al., 2008). Previous studies have demonstrated that fluidflux-melting models can reproduce the OIB, LKT, and CAB compositions using a single, enriched mantle source in which increasing amounts of slab-fluid addition drive increasing degrees of melting (Reiners et al., 2000; Rowe et al., 2009). We adopted a similar approach and calculated potential enriched mantle sources by assuming that primitive OIB compositions represent small (5%) degree melts (Conrey et al., 1997; Johnson et al., 2009). The inferred Central Oregon subarc mantle compositions have elevated Dy/Yb ~2, similar to the OIB mantle source of Sun and McDonough (1989). All melting calculations assumed modal batch melting as an approximation of pooled melts formed by polybaric fractional melting (Kelley et al., 2006; Portnyagin et al., 2007). Bulk mantle-melt partition coefficients were calculated from mineral-melt coefficients reported in Johnson et al. (2009). Mantle source compositions, modal mineralogy, and bulk partition coefficients are listed in Table S4.

We evaluate potential mantle sources (Fig. 6) using the approach of Portnyagin et al. (2007). Large variations in Nb/Y suggest that the Cascades subarc mantle is heterogeneous. OIB-type sources could produce the range of melts inferred from the melt inclusions, but more depleted sources (e.g. E-DMM; Workman and Hart, 2005) appear required to produce the range of primary melts inferred from the larger database of whole-rock compositions. Extents of mantle melting were calculated from the primary melt compositions by assuming that HFSE (Nb, Ti, and Zr), HREE (Dy, Er, and Yb) and Y are derived solely from the mantle wedge (Kelley et al., 2006; Portnyagin et al., 2007). These elements have relatively low fluid solubilities under most conditions relevant to slab dehydration (Kessel et al., 2005). The melting models suggest that the basalts and Sr-rich basalts represent melt fractions of ~10-15%. The more depleted primary melts (parents of basaltic andesites) require either higher extents of melting (~18-20%) than the basalts and Sr-rich basalts or similar extents of melting of a more depleted mantle source (dashed curve in Fig. 6a) (cf. Hughes and Taylor, 1986; Conrey et al., 1997).



**Fig. 6.** (a) Primary melt Nb and Y contents (denoted by °) for Central Oregon magmas, symbols as in Fig. 2. Gray fields enclosing small symbols are whole-rock compositions. Solid gray curves show batch modal melting models (from F = 0.05 to 0.30; increments of 0.05 are shown on lower curves; Nb and Y concentrations in the subduction component are assumed to be 0) for a spinel lherzolite source (see text). The double-headed arrow illustrates the range in mantle variability implied by the compositional variations of OIB samples. The dashed gray curve represents melting of a previously depleted source (2% PME). (b) H<sub>2</sub>O/Y vs. Nb/Y of Central Oregon magmas illustrating the effects of trace element enrichment from slab component addition (resulting in elevated H<sub>2</sub>O/Y) and the effects of previous melting and source depletion (resulting in decreased Nb/Y). Model curves are the same flux-melting models shown in Fig. 4. As in (a), the double-headed arrow illustrates mantle variability based on OIB samples and the dashed curve represents melting of a previously depleted mantle (2% PME from high Nb/Y source). Curves extend from X<sub>SC</sub> = 0 wt.%, F = 0.05 to X<sub>SC</sub> = 0.6 wt.%, F = 0.18. Field for Mexico MGVF (Michoachan–Guanajuato Volcanic Field) from Johnson et al. (2009).

Interestingly, the depleted basaltic andesite trace element compositions are similar to high-Mg andesites from the Mt. Shasta region that are interpreted to represent 15–20% melting (Grove et al., 2002). The Shasta high-Mg andesites, basaltic andesites of Central Oregon, and rare depleted basaltic andesites from the Western Cascades of Oregon may all represent melts that have equilibrated with refractory mantle (Conrey et al., 1997; Grove et al., 2003).

Given the similar P–T conditions inferred for all three magma types (Fig. 5) and the likelihood of mantle heterogeneities beneath the Central Oregon Cascades, we suggest a melting scenario that has resulted in similar overall extents of melting for each type. The primitive basaltic and Sr-rich basaltic magmas can be generated by 9–11% partial melting of an enriched mantle source (0.13–0.24 wt.% source H<sub>2</sub>O) with final equilibration temperatures of 1260–1330 °C at 1.2–1.6 GPa (Figs. 5 and 7; Table 1). Depleted basaltic andesite parent magmas can be generated by 12–16% partial melting of a more



**Fig. 7.** Primary melt H<sub>2</sub>O and calculated degree of partial mantle melting (F) in wt%. Central Oregon magmas (symbols as in Fig. 2), and fields encompass data from sources mentioned previously. Thick black line is the 1.5 GPa dry peridotite solidus and gray dashed lines represent equal temperature deviations from the solidus (Portnyagin et al., 2007). Blue lines contour equal wt.% H<sub>2</sub>O concentrations in mantle sources. H<sub>2</sub>O<sup>o</sup> contents were calculated using crystallization estimates derived from models, and degrees of melting were calculated as described in the text. Degrees of melting were calculated for basaltic andesites using mantle sources that were affected by 2% PME. Dashed gray line connects estimates for the WB–DM sample (Shasta region high-Mg composition) based on the observed maximum concentration in olivine-hosted (Fo<sub>94</sub>) melt inclusions (6%, Anderson, 1974; 3.4%, Ruscitto and Wallace, 2008).

refractory mantle source (0.40–0.56 wt.% source  $H_2O$ ) with final temperatures of 1240–1270 °C at 1.1–1.3 GPa. Based on these models, conditions beneath Central Oregon appear similar to or slightly warmer than other well-studied arcs but cooler and wetter than backarc basins (Fig. 7). Melting conditions for the primitive LKTs appear hotter than for the calc-alkaline melts, consistent with deeper equilibration pressures (cf., Leeman et al., 2005) and plot within the field defined for back-arc basin magmas. Higher degrees (>20%) of hydrous melting (>0.7 wt.% source H<sub>2</sub>O) are required to produce the high-Mg melt from the Shasta region, consistent with interpretations of previous workers (Baker et al., 1994; Grove et al., 2002; Ruscitto et al., 2008).

We conclude that LKT and OIB magmas form by decompression melting of variably enriched mantle domains. An apparent continuum exists between the calc-alkaline basalts and Sr-rich basalts, depleted basaltic andesites, and high-Mg andesites, representing melts derived from subduction component addition to increasingly refractory mantle sources. Because of the tectonic complexity beneath the region, this continuum may not imply a simple genetic relationship between the enriched and refractory mantle sources, though we note that radiogenic isotope data are consistent with such a relationship.

Inferred mantle temperatures and pressures are similar to those calculated for basalts and basaltic andesites in Northern California and SW Washington (Elkins-Tanton et al., 2001; Leeman et al., 2005; Lee et al., 2009) and inferred maximum temperatures of 1250–1260 °C predicted by geodynamic models (Currie et al., 2004; Wada and Wang, 2009) (Figs. 5 and 7). The apparent high mantle temperatures just beneath the crust cannot represent steady-state values as this would result in widespread melting of the lower crust (Elkins-Tanton et al., 2001). Such transient high temperatures could result from advection of heat by melts as they ascend from the hot core of the wedge.

Moderate input of slab-derived  $H_2O$  into the mantle wedge beneath Central Oregon is consistent with modest subduction-related geochemical signatures compared to other arcs (Leeman et al., 1990, 2005), geophysical evidence for shallow slab dehydration (Rondenay et al., 2008), and halogen outputs from spring waters (Hurwitz et al., 2005). Ratios reflecting volatile addition from a slab component (H<sub>2</sub>O/ Ce ~490–1700; S/Nb ~110–310; and Cl/Nb ~37–190) are low compared to those from the southern Cascades (Shasta) and other arcs (Mexico, Central America, and Kamchatka).

We calculated slab component compositions using the inferred mantle sources and average primary melt compositions (Table S5).



**Fig. 8.**  $H_2O/Ce$  and  $H_2O/K_2O$  ratios of the slab-derived fluid components calculated for the Central Oregon magmas (calculated using methods described in Portnyagin et al., 2007). Other arcs are shown for comparison, and symbols and fields are the same in Fig. 7. Field for Tonga is from Plank et al. (2009).  $H_2O/Ce$  and  $H_2O/K_2O$  ratios are also shown (gray box) for the mantle sources calculated by assuming that OIB compositions represent 5 wt.% partial melts (see text). Black line illustrates slab surface temperature concordance between the geothermometers of Plank et al. (2009). Very low Ce contents were calculated for the slab composition from WB–DM falls off of the geothermometer concordance line. This could indicate formation of phlogopite or another significant  $K_2O$ -bearing phase in the mantle (Plank et al., 2009). The thick line (WW09) represents the range of slab surface temperature s at subarc depths calculated for Northern Cascadia in Wada and Wang (2009).

 $H_2O/Ce$  and  $H_2O/K_2O$  of the slab components are shown in Fig. 8. Slab surface temperatures can be estimated using these ratios calibrated with fluid–rock partitioning experiments (e.g., Kessel et al., 2005; Plank et al., 2009). As expected for an arc with a young slab, inferred slab temperatures are high (850–950 °C) and at the high temperature end of the global range for arcs. Partitioning experiments indicate that Nb may preferentially partition into the slab component at elevated temperatures (>900 °C; Kessel et al., 2005), possibly undermining an assumption in our melting model. We emphasize, however, that our melting calculations used a combination of HFSE, HREE, and Y and that calculated melt fractions do not significantly change if Nb is excluded.

Based on geochemical data and inferred depths of magma generation, it has been proposed that CAB magmas in the Southern Washington and Northern Oregon Cascades form by remobilization of a 'stored' subduction component in the lithospheric mantle by ascending decompression-derived melts (LKT and OIB) from the asthenosphere (Leeman et al., 2005). This lithospheric component may be derived from the early stages of Cascadia subduction (~40 Ma) when subduction rates were faster. Our data do not permit us to address this issue in detail, but we note that the low volatile/HFSE ratios and high inferred slab surface temperatures for Central Oregon (Fig. 8) are consistent with an active but diminished role for volatile fluxing, as would be expected from a young oceanic plate.

## 4. Conclusions

Water contents of Central Oregon magmas are generally lower than those found in other arcs, but elevated compared to oceanic back-arc magmas. Sulfur, Cl, and CO<sub>2</sub> contents are comparable to those found in other arcs. The melt inclusions studied represent the differentiation products of mantle-derived melts that underwent ~16 wt.% fractional crystallization involving mostly olivine. Primitive basalt and Sr-rich basaltic melts were generated by 9–11% partial melting of an enriched mantle source. Primitive depleted melts were likely generated by slightly higher degrees of melting (12–16%) of a more refractory source. Calculated mantle equilibration temperatures are comparable to the hottest temperatures observed in geodynamic models. Calculated mantle equilibration pressures (~1.2 GPa), however, indicate shallow mantle equilibration near the Moho. Inferred mantle melting conditions beneath Central Oregon are similar to other arcs despite the modest subduction-related geochemical signatures expected from a hot slab. Concordance between slab surface geothermometers suggests high temperatures (850–950 °C) for the slab component compared to arcs with older subducting slabs, consistent with geodynamic calculations.

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