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Depths and temperatures of <10.5 Ma mantle melting and the lithosphere-asthenosphere boundary below southern Oregon and northern California

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[1] Plagioclase and spinel lherzolite thermometry and barometry are applied to an extensive geochemical dataset of young (<10.5 Ma) primitive basaltic lavas from across Oregon's High Lava Plains, California's Modoc Plateau, and the central-southern Cascades volcanic arc to calculate the depths and temperatures of mantle melting. This study focuses on basalts with low pre-eruptive H₂O contents that are little fractionated near-primary melts of mantle peridotite (i.e., basalts thought to be products of anhydrous decompression mantle melting). Calculated minimum depths of nominally anhydrous melt extraction are 40–58 km below Oregon's High Lava Plains, 41–51 km below the Modoc Plateau, and 37–60 km below the central and southern Cascades arc. The calculated depths are very close to Moho depths as determined from a number of regional geophysical studies and suggest that the geophysical Moho and lithosphere-asthenosphere boundary in this region are located in very close proximity to one another (within 5–10 km). The basalts originated at 1185–1383°C and point to a generally warm mantle beneath this area but not one hot enough to obviously require a plume contribution. Our results, combined with a range of other geologic, geophysical, and geochemical constraints, are consistent with a regional model whereby anhydrous mantle melting over the last 10.5 Ma in a modern convergent margin and back arc was driven by subduction-induced corner flow in the mantle wedge, and to a lesser extent, toroidal flow around the southern edge of the subducting Juan de Fuca and Gorda plates, and crustal extension-related upwelling of the shallow mantle.

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1. Introduction

[2] Starting in the Oligocene, a significant plate reorganization occurred in the region that is now the western coast of North America, resulting in a period during which the relationship between the plate tectonic driving forces and the composition and location of volcanism is poorly understood [e.g., Atwater, 1970; Christiansen and McKee, 1978; Hart and Carlson, 1987]. The goal of this study is to use the chemical characteristics of primitive basalts to investigate the mantle processes that have led to the persistent and broadly distributed volcanism in the area from the central and southern Cascades volcanic arc to the Oregon-Idaho-Nevada border in the northwestern United States since 25 Ma (Figure 1). A significant number of the volcanoes in the Cascades volcanic arc are fed by hydrous flux melting in the mantle wedge above the subducting Juan de Fuca and Gorda plates [e.g., Grove et al., 2002]. The processes driving the formation of these hydrous mantle melts are not the focus of this study, although the patterns of mantle flow that contribute to their formation are likely inherently related to those driving anhydrous mantle melting in the northwestern U.S. Instead, the mafic lavas in question have been ascribed to a variety of tectono-magmatic processes including extension in the Basin and Range Province [Christiansen and McKee, 1978; Cross and Pilger, 1978; Christiansen et al., 2002], lithospheric downwelling [Hales et al., 2005; Camp and Hanan, 2008; West et al., 2009], and mantle flow related to the geometry of the subducting Juan de Fuca slab [Carlson and Hart, 1987; Humphreys et al., 2000; Faccenna et al., 2010]. Recent studies have also attributed some of the volcanic centers in question to mantle plume-related volcanism [e.g., Geist and Richards, 1993; Camp, 2004] because of the nearby eruption of >10⁵ km³ of evolved basaltic–basaltic andesite lavas ca. 16.5–14 Ma, which formed the Steens Mountain and Columbia River Plateau flood basalt provinces. However, <10.5 Ma basaltic volcanism in the area of interest is more limited in volume (600–1000 km³ in total) and consists predominantly of anhydrous high alumina olivine tholeiites (HAOT) to mildly

alkaline basalts (MAB) or calc-alkaline basalts (CAB) that have experienced very little crystal fractionation and crustal contamination [Hart et al., 1984; Hart, 1985; Bartels et al., 1991; Draper, 1991; Hart et al., 1997]. By utilizing primitive basaltic lavas that approximate liquid compositions and correcting for small amounts of fractional crystallization, we can infer the locations and conditions in the mantle from which these <10.5 Ma magmas segregated and subsequently constrain the tectonic and magmatic driving forces responsible for their formation.

[3] To calculate the depth and temperature of melt extraction from the mantle for the primitive basaltic lavas, we use a new model for melting variably depleted and metasomatized upper mantle calibrated with plagioclase and spinel lherzolite melting experiments [Till et al., 2012]. The calculated depths of melt extraction are then compared to the location of the Moho and the lithosphere-asthenosphere boundary (LAB) in the region as constrained by teleseismic receiver function data [Gashawbeza et al., 2008; Eagar et al., 2011], Rayleigh wave tomography [Wagner et al., 2010], and seismic refraction surveys [Leaver et al., 1984; Zucca et al., 1986], in order to place new constraints on the thickness of the mechanical lithosphere.

[4] In the broadest sense, the LAB can be defined as the boundary between the rigid material that composes the Earth's tectonic plates (i.e., the lithosphere) and the underlying material that undergoes solid-state creep such that it behaves as a viscous fluid on geologic timescales (i.e., the asthenosphere). However, in practice, there is no unified description of what comprises the lithosphere and it is alternately defined by its mechanical, seismic, electric, thermal, or chemical properties. Therefore, multiple depths for the LAB can be determined in a given area depending on the property chosen to define the LAB and the method used to calculate that property. The mechanical lithosphere is defined as the portion of the Earth's upper thermal boundary layer that behaves elastically and the depth of the mechanical LAB is usually inferred from plate flexure models.

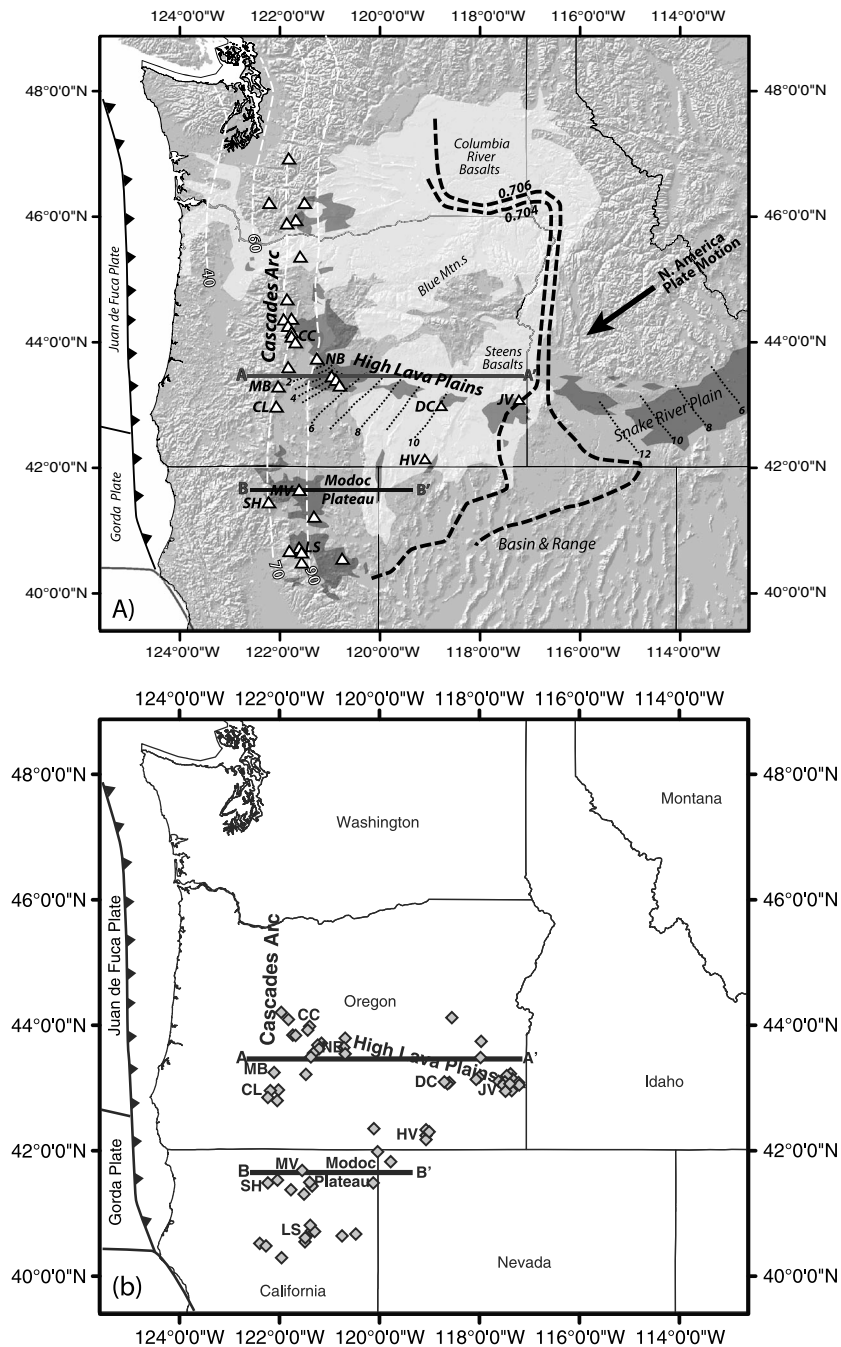


Figure 1. Geologic and tectonic map of the Pacific Northwest, United States. (a) Locations of Holocene volcanism indicated by the white triangles with locations of interest in this study labeled in bold italic: Crater Lake Volcanic Center (CL), Diamond Crater (DC), Jordan Valley Volcanic Field (JV), Hawks Valley (HV), Lassen Volcanic Field (LS), Medicine Lake Volcano (MV), Mount Bailey (MB), Mount Shasta (SH), and Cayuse Crater (CC). The location of Quaternary basaltic volcanism is shown by the dark gray-shaded regions and the mid-Miocene Steens and Columbia River flood basalts by the white-shaded area. The white-dashed lines represent depth contours to the top of the slab (km) [McCrory *et al.*, 2012] and the black-dashed lines the approximate location of the $^{87}\text{Sr}/^{86}\text{Sr}$ 0.704 and 0.706 lines [Armstrong *et al.*, 1977; Manduca *et al.*, 1992]. Isochrons (Ma) for the migration of rhyolitic volcanism in the High Lava Plains [Jordan, 2004] and Snake River Plain [Christiansen *et al.*, 2002] are denoted by the black-dotted lines. Gray bold lines A-A' and B-B' indicate the locations of the cross-sections in Figure 3. (b) Location of primitive basalt samples used in this study.

By calculating depths of melt extraction from the mantle in southern Oregon and northern California, we produce an independent estimate of the maximum thickness of the mechanical lithosphere that can be used to further evaluate the nature of the lithosphere in this region with a prolonged history of volcanism. Given the recognition of strong regional variations in isotopic composition of mantle-derived magmas in the Pacific Northwest [e.g., *Armstrong et al.*, 1977; *Hart and Carlson*, 1987] where at least one end member has been suggested to represent melts of ancient lithospheric mantle [e.g., *Carlson*, 1984], our use of the term “asthenosphere” in this paper follows the geophysical definition where the presence of partial melt defines a portion of the mantle as “asthenosphere” regardless of past history that may impose lithospheric chemical and/or isotopic characteristics on a given basalt mantle source region.

2. Tectonic and Volcanic History

[5] The western United States has a rich and varied volcanic and tectonic history (see review in *Christiansen and Lipman* [1972], *Lipman et al.* [1972], and *Humphreys and Coblenz* [2007]). The northwestern extent of the Basin and Range province including Oregon’s High Lava Plains (HLP) and California’s Modoc Plateau has been dominated by volcanism [*Walker*, 1977] since at least 25 Ma when a period of silicic volcanism [e.g., *Christiansen and Yeats*, 1992] was followed by a period of massive flood basalt eruptions near the Oregon/Idaho/Nevada border that produced the Steens Mountain basalts from 16.6 to 15.5 Ma [*Brueseke et al.*, 2007]. This flood basalt volcanism migrated north along the western edge of Precambrian North America to form the Columbia River Plateau basalts—basaltic andesites between 16.5 and 6 Ma [*Waters*, 1962; *Swanson et al.*, 1979; *Carlson*, 1984; *Tolan et al.*, 1989]. To the west, Cascadian volcanism produced by oblique subduction of the Juan de Fuca plate began ca. 37 Ma and continues to the present [*Christiansen and Lipman*, 1972]. Today, convergence of the Juan de Fuca and Gorda plates is occurring at 40–45 mm/yr and varies from orthogonal convergence in British Columbia to more oblique northwesterly directed convergence in southern Oregon and California [*Wilson*, 1988], where we focus in this study. Geophysical studies image the slab at depths of <100 km [*McCrory et al.*, 2012] below the central and southern extents of the Cascades. In the central and southern Cascades, abundant diffuse mafic scoria cones and small mafic volcanoes are found in close proximity to more iconic andesitic arc stratocones [*Sherrod and Smith*, 1990].

[6] This study in part focuses on lavas from the HLP of central and southeastern Oregon. This ≤10 Ma bimodal volcanic province seemingly connects the Yellowstone-Snake River Plain track to the east, with the Cascades arc to the west. Its tectono-magmatic origin has been a topic of debate since the 1970s [e.g., *MacLeod et al.*, 1976; *Draper*, 1991; *Humphreys et al.*, 2003; *Jordan*, 2004]. The volcanic rocks in the HLP consist predominantly of intercalated basalt lava flows and rhyolite ash flow tuffs and related sediments with scattered rhyolite dome complexes. Silicic volcanism in the HLP generally migrated in a northwesterly direction starting in southeastern Oregon at ca. 10 Ma and ending near Newberry Volcano at ca. 1 Ma where volcanism continues today [*MacLeod et al.*, 1976; *Jordan*, 2004]. Several periods of increased basaltic volcanism in the HLP occurred at 7.8–7.5 Ma, 5.9–5.3 Ma, and 3–2 Ma; however, the basaltic volcanism does not exhibit age progressive migration like that of the HLP rhyolitic volcanism [*Jordan*, 2004]. This study also includes basalts erupted in the Modoc Plateau, which is located in the extreme northeast corner of California, east of the Cascades arc, and south of the HLP. In general, the Modoc Plateau is composed of flat-lying pyroclastic rocks with intercalated and capping basalt flows that are believed to be Miocene and younger in age [e.g., *Duffield and Fournier*, 1974; *Carmichael et al.*, 2006]. The diffuse northwestern margin of the Basin and Range province impinges on the volcanic HLP and the Modoc Plateau, as well as the Cascades arc and back arc. These regions have experienced normal faulting oriented NW-SE to N-S and associated extension-related volcanism since at least ~12 Ma [*Blakely et al.*, 1997; *Colgan et al.*, 2004; *Scarberry et al.*, 2010].

3. Methods

3.1. Lava Sample Selection

[7] Primitive basaltic lavas from across the HLP, Modoc Plateau, and the central-southern Cascades (Figure 1) were selected for thermometric and barometric calculations, as these liquids are the least likely to have experienced significant modification via fractional crystallization and crustal assimilation since their origin in the mantle. We consider lavas to be primitive if they are rich in Mg relative to Fe ($Mg\# = \frac{Mg}{Mg + Fe^{2+}}$ with all Fe as $Fe^{2+} > 0.50$) with $SiO_2 < 52$ wt.% and relatively phenocryst poor (<3%). The samples that fit these criteria are HAOT to MAB and CAB that are typically aphyric, containing rare microphenocrysts of olivine and plagioclase.

Many samples approach compositions of primary magmas with $Mg\#s \geq 0.70$ (Figure 2, Auxiliary Table 1) and require very little correction to bring their compositions into equilibrium with a mantle mineral assemblage. The trace element compositions of the basalts suggest that they represent melts of depleted mantle, in some cases with the addition of a subduction component, which may consist of subduction-related fluids, melts, and/or previously

metasomatized lithosphere [Hart *et al.*, 1984; Baker *et al.*, 1994; Bacon *et al.*, 1997; Borg *et al.*, 1997; Jordan, 2004]. We restricted our calculations to <10.5 Ma basalts, with the majority of the samples erupted within the last 15,000 years, in order to constrain modern processes to the extent possible. In some cases, more than one sample was selected from a given volcanic vent or unit and therefore, each sample does not necessarily represent a distinct

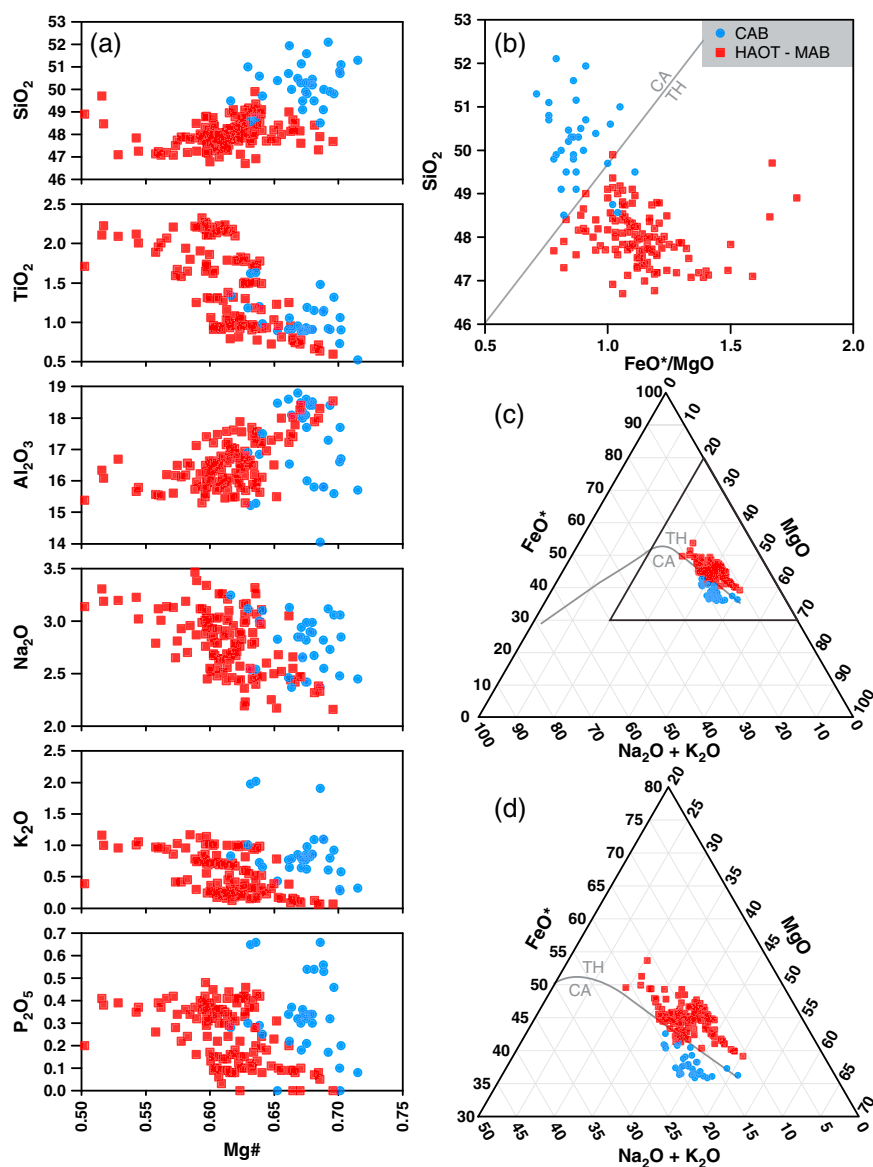


Figure 2. (a) Chemical characteristics of <10.5 Ma primitive basaltic lavas from southern Oregon and northern California selected for this study. Red squares designates nominally anhydrous high alumina olivine tholeiite to mildly alkaline basalts and blue circles designates samples determined to be calc-alkaline basalts based on their location in panels (b)–(d) as well as enrichment in trace elements consistent with the contribution of a slab-derived component (e.g., elevated Ba/Nb and Sr). Gray lines in Figures 2b–2d separate samples that follow calc-alkaline versus tholeiitic differentiation trends as designated by Miyashiro [1974]. Black triangle inset in Figure 2c shows region illustrated in Figure 2d.

melting event. These criteria are met by lavas erupted (1) across the HLP at Jordan Valley Volcanic Field, Diamond Crater, east of Newberry, and unnamed vents near Steens Mountain and the Blue Mountains (samples from east of Newberry and unnamed vents near Steens and Blue Mountains are labeled *HLP miscellaneous* in Supporting Information.¹ (2) in the central and southern Cascades arc at Crater Lake Volcanic Center, Mt. Bailey, Newberry Volcano, Mt. Shasta, Medicine Lake volcano, Lassen Volcanic Field, and smaller volcanic centers or vents in the High Cascades region of Oregon (Cayuse Crater, Foley Ridge, Sitkum Butte, and a vent south of South Sister); and (3) various unnamed basaltic vents on or adjacent to the Modoc Plateau, CA and Hawks Valley, OR, southwest of Steens Mountain. The major element compositions of all samples in this study along with estimates of their temperature and pressure of melt segregation are presented in Supporting Information.

3.2. Fractional Crystallization Corrections and Basalt Thermometry and Barometry

[8] The chemical analyses of the primitive rocks enable the examination of the pressure and temperature of origin for young basaltic volcanism in northern California and Oregon. For the thermometric and barometric calculations, the major element compositions of all samples were first corrected for minor fractional crystallization following the methods described by Till *et al.* [2012] until the bulk compositions could be approximated as a liquid in equilibrium with a mantle olivine with a forsterite content (Mg_2SiO_4) of Fo_{90} . The phase assemblage chosen for the fractional crystallization correction can have a significant effect on the calculated primary liquid composition and therefore the estimates of source pressure and temperature. For example, inaccurate phase assemblages can produce errors $>100^\circ\text{C}$ and 1.0 GPa in estimates of the pressure and temperature of origin for primary MORB liquids [Till *et al.*, 2012]. Rather than assuming that olivine-only fractionation occurred, the location of the sample bulk composition was compared to the location of the low-pressure olivine-plagioclase-liquid (OPM) and olivine-plagioclase-augite-liquid (OPAM) saturation boundaries predicted for that composition on a pseudo-quaternary projection scheme of Tormey *et al.* [1987] [see Till *et al.*, 2012, Figure 5] to determine the appropriate fractionating phase assemblage. Nearly all of the 155

lava compositions corrected in this study ($>86\%$) have $\text{Mg\#} > 60$ with 30% having $\text{Mg\#} > 65$, and there was little ambiguity in correcting these Mg-rich lava compositions for fractional crystallization. The high Mg\# compositions lay on the OPM boundary and the only variable in correcting them was to infer when the liquid leaves the OPM boundary, enters the olivine primary phase volume, and follows an olivine-only addition path to one of the spinel lherzolite multiple saturation points. Because the OPM boundary is orthogonal to and crosses the spinel multiple saturation points, the olivine-only paths are restricted by the rapid change in Mg\# that occurs when olivine alone is added. If olivine-only addition was started too soon, the liquid composition would not reach a spinel lherzolite boundary when it was saturated with Fo_{90} olivine. In the case of saturation with a plagioclase lherzolite residue, the trend formed by the multiple saturation points with increasing pressure lie close to and parallel the OPM boundary, again leaving little ambiguity for the back-fractionation path chosen to return the lava composition to a liquid in equilibrium with the mantle, since only a small amount of olivine ($<2\text{wt.}\%$) was needed. A more significant uncertainty is the assumption of Fo_{90} as the composition of olivine in the mantle residue. If the mantle contained a more magnesian olivine, then the estimates are a minimum pressure and temperature. For our fractional crystallization corrections, the locations of the OPM and OPAM boundaries for the range of crustal pressures were determined using the method of Grove *et al.* [1992] updated with the plagioclase and spinel lherzolite multiple saturation point recalibration of Till *et al.* [2012]. A more detailed explanation of these methods can be found in Grove *et al.* [1992, Appendix 2] for olivine and plagioclase fractionation corrections, and Yang *et al.* [1996] for the few cases ($\sim 14\%$) where corrections involved olivine, plagioclase, and clinopyroxene.

[9] Once the bulk compositions of the samples could be approximated as a liquid in equilibrium with mantle olivine, the plagioclase and spinel lherzolite melting model of Till *et al.* [2012] was used to calculate the pressure and temperature where the basaltic melts were last multiply saturated with an upper mantle assemblage of olivine, orthopyroxene, clinopyroxene, and plagioclase and/or spinel (i.e., at the multiple saturation point). These basaltic melts likely represent batch melts as demonstrated by Till *et al.* [2012] and therefore the calculated pressure represents the shallowest depth of mantle equilibration, or melting, for a given melt. The average absolute error is 0.15 GPa (~ 5 km) for the pressure calculations and 11°C for the temperature

¹All Supporting Information may be found in the online version of this article.

calculations. Density profiles for the crust and upper mantle below the HLP [Cox, 2011] and northern California [Zucca *et al.*, 1986] were used to determine a pressure-depth relationship at the basalt sample localities, yielding the depth of last equilibration.

[10] Of the 155 samples presented here, 33 have chemical compositions that fall within the calc-alkaline field on AFM or FeO*/MgO versus SiO₂ diagrams as identified by Miyashiro [1974] (Figure 2). All 33 of these CAB are from volcanic centers in the present-day Cascades arc axis or back arc; these are the subduction-influenced volcanic centers of Lassen, Mt. Shasta, Medicine Lake, Newberry, vents east of Newberry, and a cinder cone from the High Cascades region in Oregon, Cayuse Crater. All of these volcanic centers with the exception of Cayuse Crater, erupted HAOT or MAB also included in this study. H₂O has a large effect on the fractional crystallization path of a primitive basalt and results in the early iron depletion and silica enrichment characteristic of CAB erupted in continental arc settings [Sisson and Grove, 1993]. Therefore, these 33 CAB likely interacted with H₂O at some point during their genesis and their pressure and temperature of origin were corrected for the effects of H₂O following the H₂O correction included with the thermometer and barometer of Till *et al.* [2012]. Recent work by Walowski *et al.* [2012] suggests that olivine-hosted melt inclusions from mafic cinder cone tephra from the Lassen Volcanic Field are consistent with the volatile contents (~1.3–3 wt.% H₂O) and melt compositions of olivine-hosted melt inclusions from mafic tephra from the central Oregon Cascades [Ruscitto *et al.*, 2010]. Therefore, H₂O contents for the CAB from Lassen, Mt. Shasta, Medicine Lake, vents east of Newberry, and Cayuse Crater were approximated using the H₂O-melt composition scaling relationship determined by Ruscitto *et al.* [2010]. Twenty of the 33 CAB samples are from subduction-influenced Newberry Volcano, which erupted both dry tholeiitic and wet CAB [Donnelly-Nolan and Grove, 2009] since ~500 ka. Olivine-plagioclase hygrometry conducted on a representative subset of the 20 CAB from Newberry indicates they contained ~4 wt.% H₂O prior to eruption [Grove *et al.*, 2009], and the H₂O-lava composition scaling relationships for this subset of Newberry CAB were used to estimate the H₂O contents of the Newberry CAB. The H₂O-corrected CAB from Lassen, Mt. Shasta, Medicine Lake, vents east of Newberry, and Cayuse Crater produce temperature and depth estimates that are an average of $50 \pm 15^\circ\text{C}$ (1σ) cooler and 1.65 ± 0.27 km deeper than the calculated temperatures and depths of last

mantle equilibration for these basalts at anhydrous conditions. The H₂O-corrected CAB from the Newberry Volcano produce temperature and pressure estimates that are an average of $94 \pm 23^\circ\text{C}$ cooler and 3.11 ± 0.75 km deeper than at anhydrous conditions, as their average H₂O contents are higher than those from the other CAB samples. These results are consistent with the experimentally determined effect of H₂O on the temperatures and pressures of mantle melting [e.g., Gaetani and Grove, 1998]. The pressures and temperatures discussed below and illustrated in Figures 3 and 4 include the anhydrous depths and temperatures determined for the 122 HAOT-MAB samples and the H₂O-corrected depths and temperatures for the 33 CAB samples.

4. Pressure and Temperature of Mantle Melt Extraction

[11] The depths of melting for primitive southern Oregon and northern California lavas are illustrated in two E-W transects in Figure 3: one transect through the central Cascades arc and HLP in Oregon at ~43.5°N and another through the southern Cascades arc and Modoc Plateau in California at ~41.5°N, as shown in Figure 1. The calculated minimum depths of melt extraction are between 37 and 58 km depth for the studied basalts along the northern transect through Oregon, with the widest range of depths recorded by samples from Newberry Volcano. The calculated minimum depths of melt extraction are between 41 and 60 km along the southern transect through California, with the widest range of depths recorded below Lassen Volcanic Field. The overall average minimum depth of melt extraction for the entire dataset is 48.9 ± 4 km (1σ).

[12] Our thermometry indicates that the basalts originated at temperatures between 1185 and 1383°C. The nominally anhydrous HAOT and MAB samples originated between 1295 and 1383°C, while the CAB samples when corrected for H₂O originated at temperatures of 1185–1323°C (Figure 4). Phase equilibrium studies of primitive basaltic lavas from Medicine Lake [Bartels *et al.*, 1991] and Mt. Shasta [Baker *et al.*, 1994] indicate that these magmas were separated from the mantle at ~1300°C and 10–11 kbar, a pressure range equivalent to ~35–38 km depth in this region, consistent with our calculations. Our calculated temperatures and depths for the HAOT and MAB samples yield thermal gradients of 3.7°C/km at ~40–60 km depth beneath the Jordan Valley Volcanic Field in the eastern HLP or 3.4°C/km at ~40–50 km depth beneath the Hawks

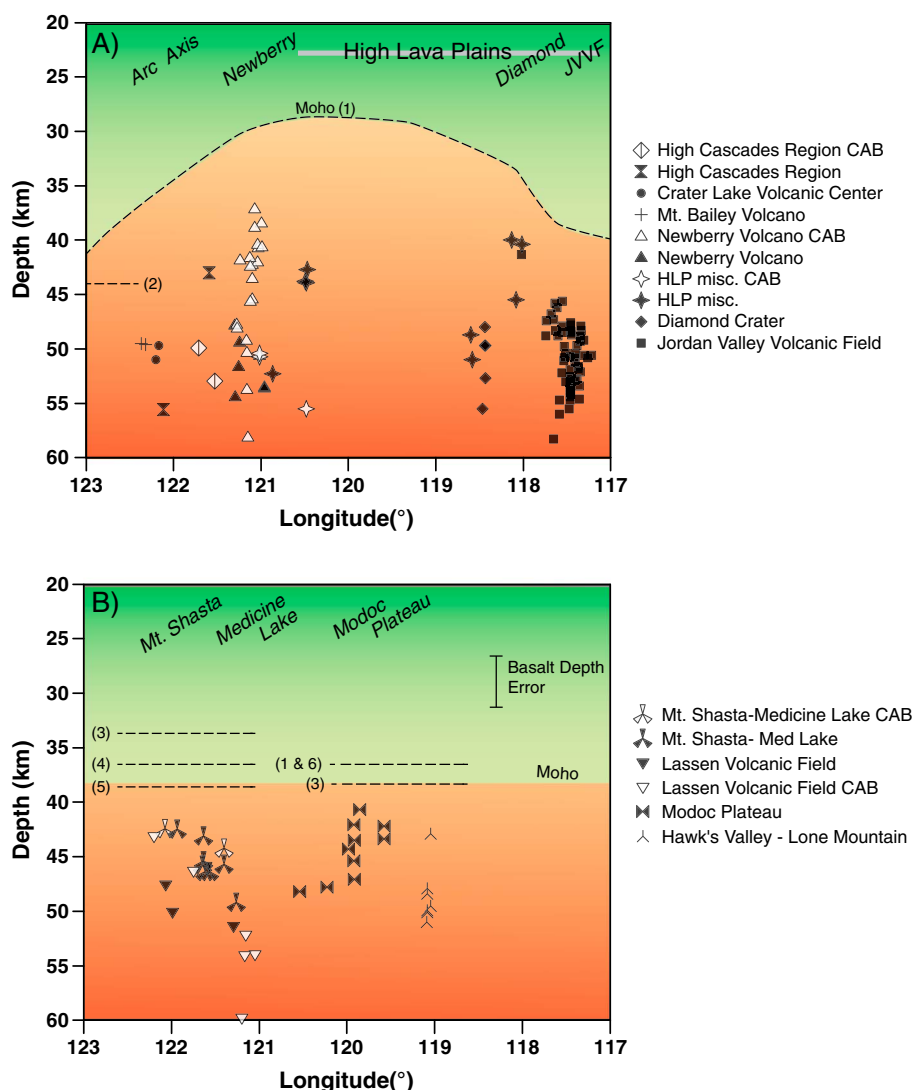


Figure 3. Depth of asthenospheric melt segregation calculated for primitive basaltic lavas from southern Oregon and northern California illustrated in two E-W cross-sections at $\sim 43.5^\circ\text{N}$ and $\sim 41.5^\circ\text{N}$ (locations illustrated in Figure 1). Samples are plotted at the location they were collected, as the vent location is not known for all samples. Black symbols denote HAOT or MAB and white symbols denote the CAB samples corrected for the effect of H_2O . The base of the green field illustrates the depth of the Moho based on regional geophysical estimates (black-dashed lines with the corresponding reference numbers: (1) *Eagar et al.* [2011], (2) *Leaver et al.* [1984], (3) *Zucca et al.* [1986], (4) *Ritter and Evans* [1997], (5) *Mooney and Weaver* [1989], and (6) *Gashawbeza et al.* [2008]).

Valley—Lone Mountain region in the eastern Modoc Plateau. These estimates are an order of magnitude larger than commonly assumed adiabatic gradients but are consistent with other petrologic estimates of the geothermal gradient below arcs [*Kelemen et al.*, 2003]. Super-adiabatic thermal gradients are expected, as they are imposed by adiabatic decompression melting of mantle lherzolite and the temperature-depth dependence of the mantle solidus. The shallowest depths of equilibration calculated for all regions (Figure 3) are within ~ 5 – 10 km of the present-day geophysical Moho, as discussed in the following section.

5. Constraints on Lithospheric Structure

[13] A number of geophysical studies have focused on the crustal and uppermost mantle structure in the study region and reveal remarkably consistent results. Here, we outline the results of a number of recent seismological studies and discuss their relationship to other geophysical constraints of lithospheric structure across the region, as well as our basalt thermometry and barometry.

[14] A number of seismological constraints on crustal structure are summarized with our results in

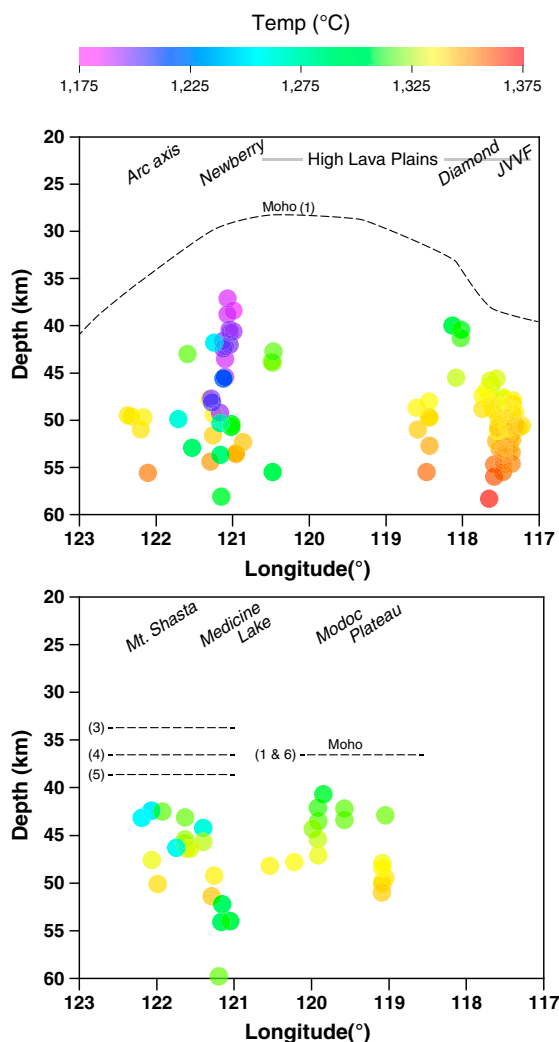


Figure 4. Temperature of asthenospheric melt segregation for primitive basaltic lavas illustrated in two E-W cross-sections, samples, and cross-section locations same as Figure 3: (average absolute error is 11°C).

Figure 3. *Eagar et al.* [2011] analyzed teleseismic *P*-to-*S* receiver functions to image the crustal structure below the HLP and determined Moho depths for the region using both a *H-κ* stacking and a new Gaussian-weighted common conversion point (GCCP) stacking technique. The techniques differ in that the *H-κ* method determines average crustal thickness and V_p/V_s values for a zone around each seismic station, whereas the GCCP method determines crustal thickness for a zone around common seismic raypath piercing points. *Eagar et al.* [2011] found that for the HLP and surrounding regions, GCCP Moho depths average ~5 km deeper than those determined from *H-κ* stacking, but the trend of Moho topography is very similar for both. The discrepancy is expected given the differences in

the techniques, the assumption of background V_p , and the nature of crustal structure variations across the region. Overall, *Eagar et al.* [2011] found Moho depths of ~40 km below the central Cascades arc and 31–36 km below the HLP and northern Great Basin. *Eagar et al.* [2011] also examined two stations in the Modoc Plateau that yield Moho depths of 35–36 km, a nearly identical result to that of *Gashawbeza et al.* [2008], who investigated the Modoc Plateau and areas to the east in the Great Basin. Scattered wave inversion images from a 2-D teleseismic migration [*Chen et al.*, 2011] also image a prominent Moho at 35 km below the HLP and thickening to 45 km beneath the Owyhee Plateau to the east.

[15] West of the HLP, a reversed seismic refraction profile ~30 km southwest of Crater Lake volcanic field determined a crustal thickness of 44 km [*Leaver et al.*, 1984]. A seismic refraction survey conducted by the USGS in 1981 characterized the crustal structure of the Klamath Mountains, Cascade Range, and Basin and Range province of Northern California [*Zucca et al.*, 1986]. This survey estimated Moho depths of 33–45 km below the southern Cascades arc, specifically 33–37 km beneath Mt. Shasta, and 38–45 km beneath the Modoc Plateau. In addition, a teleseismic tomography experiment to image Medicine Lake volcano [*Ritter and Evans*, 1997] inferred a Moho depth of ~36–37 km, similar to estimates by *Mooney and Weaver* [1989] of 38–40 km beneath Mt. Shasta to Medicine Lake and 38 ± 4 km beneath Lassen Volcanic Center. The variable depth of the Moho below the HLP and Oregon Cascades relative to the constant depth of the Moho below northern California likely reflects differences between the Cenozoic basement to the north and the Paleozoic-Mesozoic basement to the south. In addition, the amount of crustal extension in each region varies (see discussion in section 6) and likely contributes to at least a portion of the differences in crustal thickness.

[16] The shallowest depths of melting calculated for the primitive basalt samples from all regions in this study are within 5–10 km of the location of the Moho as determined from the seismological studies. A fundamental premise of the petrologic work presented here is that the studied primitive basalts originated in the asthenospheric mantle. This assumption is in part due to the relative inability of rigid mantle within the mechanical lithosphere to undergo adiabatic decompression melting unless a significant amount of lithosphere-scale extension occurs, which is not consistent with the observed degree of extension in southern Oregon

and northern California (see further discussion in section 6). The temperatures of melting in excess of $>1200^{\circ}\text{C}$ recorded by the primitive basaltic lavas in this study at 40–60 km depth also support an asthenospheric origin. We suggest that our calculated minimum depths of last melt equilibration are a proxy for the depth of the mechanical LAB in the region. The relationship between the petrologic constraints on variation in melting depths and the seismologic constraints on variations in Moho depths strongly suggest that the mechanical lithosphere is not significantly thicker than the continental crust across the region.

[17] Several other lines of geophysical evidence support the interpretation of very thin mechanical lithosphere in this region. For instance, results from a joint inversion [Wagner *et al.*, 2012] of surface wave tomography for upper mantle depths [Wagner *et al.*, 2010] and ambient noise tomography for crust and uppermost mantle depths [Hanson-Hedgecock *et al.*, 2012] reveal regions of very low seismic velocities at and immediately below our calculated depths for basaltic melt extraction in the HLP and Modoc Plateau, consistent with thin mantle lithosphere (Figure 5). Low S-wave velocities generally indicate high temperatures, the presence of partial melting, and/or the presence of water. Since the temperature recorded by the basalts exceeds the stability of hydrous minerals at these depths and any H_2O was likely partitioned into the melt, the observed low velocities are likely due to a combination of high temperatures and partial melting. The 2-D teleseismic migration of Chen *et al.* [2011] also images several pockets of extremely low velocities in the uppermost mantle beneath the HLP at the depths of origin for the basalts as calculated here, as well as beneath Steens Mountain and Newberry Volcano. Models from inversions of regional magnetotelluric data exhibit zones of high conductivity in the same regions as the extremely low seismic velocities, consistent with the presence of partial melt [Patro and Egbert, 2008; Kelbert *et al.*, 2012].

[18] Accounting for error in both the barometric calculations and geophysical observations, these complementary datasets are consistent with a model where no more than 5–10 km of mantle lithosphere exists beneath southern Oregon and northern California in the regions covered by our data. The surface wave tomography [Wagner *et al.*, 2010] and scattered wave inversion images [Chen *et al.*, 2011] suggest that the lithosphere deepens to depths of 60–70 km beneath the Owyhee Plateau to the east, paralleling the increase in regional crustal thickness. This distinction is important because the primitive basalts of the HLP

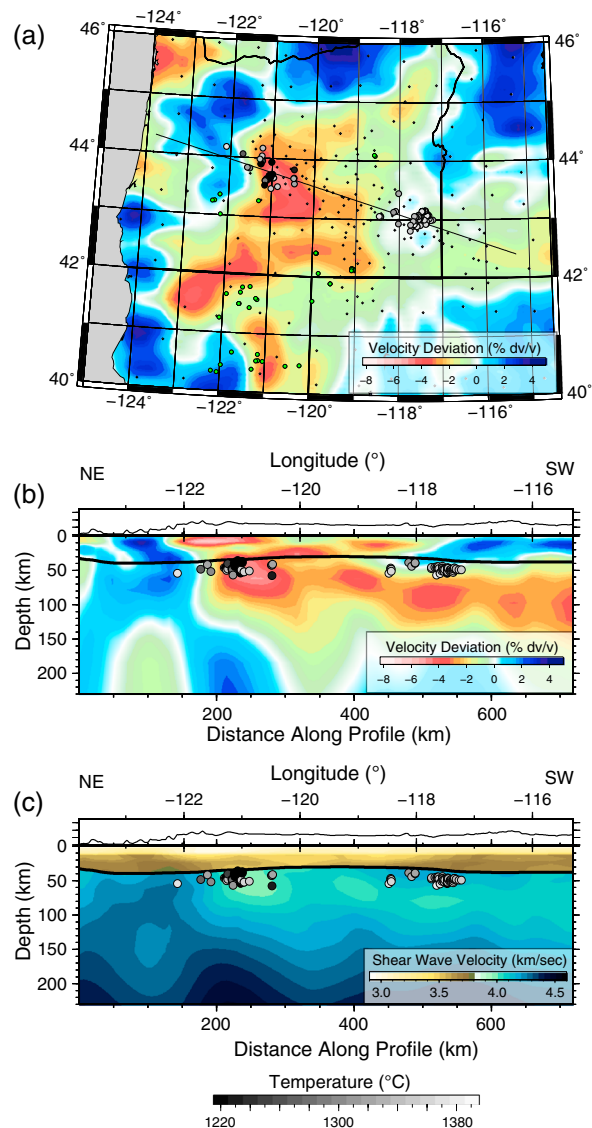


Figure 5. (a) Map slice at 45 km depth of a joint inversion [Wagner *et al.*, 2012] of surface wave tomography for upper mantle depths [Wagner *et al.*, 2010] and ambient noise tomography for crust and uppermost mantle depths [Hanson-Hedgecock *et al.*, 2012]. Also shown are the locations of basaltic samples in this study (gray circles denote samples in cross-sections in Figures 5a and 5c, and green circles denote samples not in cross-section in Figures 5b and 5c), the location of the NW-SE cross-section in Figures 5b and 5c (black line), and station locations used in the phase velocity inversions of Wagner *et al.* [2010] and Hanson-Hedgecock *et al.* [2012] (black diamonds). Basalt samples plotted in Figures 5a–c are scaled for temperature and plotted in gray scale. (b) NW-SE cross-section of the depths of origin for basalts from the High Lava Plains and central Oregon Cascades, velocity and basaltic temperature scale as in Figure 5a. Background colors indicate shear wave velocity deviations along the transect from the model of Wagner *et al.* [2012]. (c) Same as Figure 5b but showing in the background the absolute shear wave velocities along this transect from the model of Wagner *et al.* [2012].

show gradually increasing $^{87}\text{Sr}/^{86}\text{Sr}$ and decreasing $^{143}\text{Nd}/^{144}\text{Nd}$ from west (~ 0.703) to east (~ 0.704) with a very rapid increase crossing the Owyhee Plateau/Idaho border at 117.5°W longitude [Hart and Carlson, 1987; Leeman *et al.*, 1992]. Whether this isotopic variability is the result of crustal contamination [Carlson, 1984] or reflects variable input from ancient mantle lithosphere to the east [Carlson, 1984; Hart, 1985; Hart and Carlson, 1987] has been the subject of several studies. The mantle-like oxygen isotopic composition of HAOT across the HLP [Hart, 1985], and the fact that oxygen does not correlate with $^{87}\text{Sr}/^{86}\text{Sr}$ variation in these rocks, has been used to suggest that at least the eastern sources for HAOT in the HLP reside in old mantle lithosphere metasomatically enriched in incompatible elements through Proterozoic/Archean crust building east of the Idaho border. The results presented here suggest that if the sources of these lavas indeed are in old mantle lithosphere, the temperature of that lithosphere is sufficiently high so that this material behaves as geophysical and petrologic asthenosphere.

[19] Several other studies of lithosphere thickness in the overriding plate at subduction zones also find evidence for a thin mechanical lithosphere. The high temperatures required at ~ 40 – 60 km depth to generate the asthenospheric melts, as constrained in this and other petrologic studies (see review in Kelemen *et al.* [2003]), are difficult to produce in many geodynamic models of mantle flow at subduction zones [e.g., Wada and Wang, 2009; Syracuse *et al.*, 2010]. However, Kelemen *et al.* [2003] demonstrate that this problem can be resolved if the thermal boundary layer at the base of arc crust (i.e., the mechanical lithosphere) is thinner than originally thought (i.e., adiabatic mantle convection occurs to a depth of ≤ 50 km, rather than ~ 80 km used in most thermal models). Their mantle models incorporating temperature-dependent viscosity and widely accepted values for activation energy and asthenospheric viscosity are able to produce temperature-depth relationships consistent with the petrologic estimates, including those from this study. Similarly, the geodynamic models of Rowland and Davies [1999] find that mechanical lithosphere thickness at subduction zones is controlled by compositional buoyancy and therefore closely related to the thickness of the crust. Plank and Langmuir [1988] also find a correlation between crustal thickness and the major element composition of parental magma at subduction zones and suggest that this observation can be explained if crustal thickness controls the temperature and depth of asthenospheric melting, as observed here.

6. What Is Driving Mantle Melting Below Southern Oregon and Northern California?

[20] The results presented here indicate that < 10.5 Ma basalts from southern Oregon and northern California are generated by shallow mantle melting at temperatures that do not support the involvement of any unusually hot mantle, as might be contributed by a thermochemical plume rising from the deep mantle. Furthermore, the HLP basalts do not have the high $^3\text{He}/^4\text{He}$ values measured in Snake River Plain basalts, which is often taken as the best geochemical indication of a plume source [Graham *et al.*, 2009]. A number of recent geophysical studies also find little evidence for a mantle plume beneath the HLP [Lin *et al.*, 2010; Schmandt and Humphreys, 2010; Obrebski *et al.*, 2011; Schmandt *et al.*, 2012] where very low seismic velocities directly below the crust are restricted to the upper 100 km of the mantle [e.g., Roth *et al.*, 2008; Warren *et al.*, 2008; Wagner *et al.*, 2010].

[21] Instead, plate subduction likely provides the driving force for the upwelling and adiabatic decompression melting in the asthenosphere beneath southern Oregon and northern California [Long *et al.*, 2012]. Shear wave splitting results show strong E-W directed anisotropy beneath the HLP that is most easily explained by strong and well-organized flow in the mantle wedge induced by subduction and rollback of the Juan de Fuca plate [Long *et al.*, 2009; Long *et al.*, 2012]. This anisotropy reaches a maximum in areas of the HLP that have the lowest shear-wave velocities suggesting that mantle flow and partial melting in the mantle wedge are strongly coupled. In a previous study, Elkins-Tanton *et al.* [2001] calculated that the pressures of mantle wedge melting decrease from east to west below Medicine Lake and Mt. Shasta and inferred that these pressures of melting paralleled mantle flow in the upwelling limb of corner flow as calculated by Furukawa [1993]. In this study, we calculate a narrower range for the depths (this study: 42–49 km versus Elkins-Tanton *et al.*: ~ 36 – 66 km) and temperatures (this study: 1254 – 1345°C versus Elkins-Tanton *et al.*: ~ 1300 – 1450°C) of melting below Mt. Shasta and Medicine Lake volcanoes but come to the same conclusion regarding the driving force for mantle melting. Differences in our calculated depths and temperatures of melting are the result of using the Till *et al.* [2012] thermometer and barometer, which includes a reanalysis of the methodology used to calibrate

pressure in *Kinzler and Grove* [1992], the thermometer and barometer used by *Elkins-Tanton et al.* [2001]. The basalts from Lassen Volcanic Field also exhibit a similar trend in the depths of melting to those from Mt. Shasta and Medicine Lake. Furthermore, the close spatial and temporal association of nominally anhydrous basaltic lavas to products of hydrous flux melting at Lassen, Crater Lake, Newberry, Medicine Lake, and Mt. Shasta support the interpretation that plate subduction is simultaneously causing the formation of both these magma types. Other studies of primitive basalts erupted above a subduction zone find evidence for anhydrous adiabatic decompression melting induced by corner flow in the mantle wedge [*Sisson and Bronto*, 1998; *Righter*, 2000; *Cameron et al.*, 2003]. Geodynamic models that include realistic temperature-dependent viscosities [*Furukawa*, 1993; *Conder*, 2002; *Eberle et al.*, 2002; *Kelemen et al.*, 2003] produce the significant upwelling due to corner flow in the mantle wedge required to generate these anhydrous basalts, as discussed by *Wiens et al.* [2008].

[22] The southern boundary of the subducting slab is located just north of $\sim 40^\circ\text{N}$ latitude based on the location of the Mendocino Triple Junction and tomographic images of the slab [*Funiciello et al.*,

2006; *Roth et al.*, 2008]. Laboratory models of subduction illustrate toroidal flow around a slab edge, which includes a pronounced vertical component when slab rollback is occurring as in the northwestern U.S. [*Funiciello et al.*, 2006; *Druken et al.*, 2011]. Therefore toroidal flow around the southern edge of the slab could contribute to upwelling and decompression melting in the mantle below the southernmost volcanic centers in our study. This is consistent with the interpretation that at least two mantle sources with different isotopic characteristics produced the different types of primitive basalts erupted in the Lassen Volcanic Field [*Borg et al.*, 2002].

[23] Basin and Range extension is also the result of partitioning of the relative motion between the North American plate and the subducting Juan de Fuca and Gorda plates [e.g., *Humphreys*, 1995; *Atwater and Stock*, 1998; *Wesnousky*, 2005] and likely played a role in the formation of the mafic lavas in this study. Recent geological and geophysical studies suggest that low-magnitude ($\leq 20\%$) extension along high-angle normal faults [*Lerch et al.*, 2008] began ca. 12 Ma in the northwestern Basin and Range (see compilation in *Scarberry et al.* [2010]) with extension occurring at 0.01 mm/yr

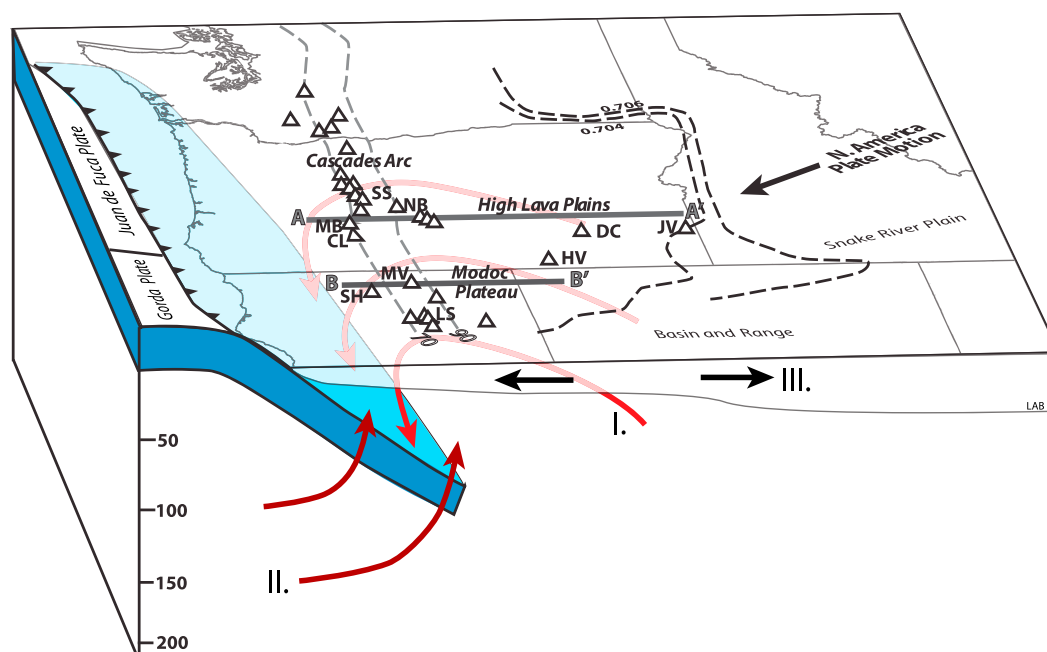


Figure 6. Cartoon illustrating three potential causes of mantle upwelling that likely produced <10.5 Ma primitive basalts with low pre-eruptive H_2O contents in southern Oregon and northern California: (I) subduction-induced corner flow in the mantle wedge, (II) toroidal flow around the southern termination of the subducting slab, and (III) northwest Basin and Range crustal extension. Gray-dashed lines represent depth counters for the subducting slab [*McCrory et al.*, 2012] and black-dashed lines are the approximate location of the $^{87}\text{Sr}/^{86}\text{Sr}$ 0.704 and 0.706 lines that coincide with the western margin of Precambrian North America [*Armstrong et al.*, 1977; *Manduca et al.*, 1992].

since 5.68 Ma [Trench *et al.*, 2012]. Paleomagnetic estimates at 42°N latitude suggest that eastern Oregon has extended ~17% since 15 Ma [Wells and Heller, 1988]. Eagar *et al.* [2011] use crustal thicknesses to estimate that the HLP has experienced ~16% extension since 10 Ma. Extension also occurred within the main Cascades arc during this time period, as indicated by discontinuous north-striking grabens and faults [Hughes and Taylor, 1986; Smith *et al.*, 1987]. Thus, crustal extension may have contributed to the amount of mantle upwelling in the mantle beneath southern Oregon and northern California and very likely played an important role in controlling when and where the mantle melts were concentrated in the crust and ultimately erupted.

[24] In conclusion, our results suggest that <10.5 Ma primitive basalts erupted in southern Oregon and northern California were produced by the mantle flow induced by the subduction of the Juan de Fuca and Gorda plates and not a thermochemical plume. Subduction-induced corner flow, and to a lesser extent toroidal flow around the southern edge of the slab and crustal extension, likely produced the upwelling and the warm temperatures at shallow depths necessary to generate mantle melts beneath southern Oregon and northern California over the last 10.5 Ma. These three causes of asthenospheric upwelling may have operated in unison or varied in importance through time (Figure 6). The thin nature of the mechanical lithosphere in this region also appears to play an important role in generating the conditions we calculate for the mantle below southern Oregon and northern California. Crustal extension in the diffuse northwest Basin and Range likely also constrained where mantle melts were emplaced in the crust and thus, the spatio-temporal patterns of volcanism in this region.

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