Melt Generation in the Earth's Mantle at Convergent Plate Margins

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Abstract

The five geologic studies presented in this thesis document how the recycling of tectonic plates at subduction zones has a profound effect on the melting behavior of the Earth’s mantle. Two experimental studies (Chapters 1 and 2) of the melting behavior of mantle peridotite demonstrate that the forefathers of arc magmas are formed at extremely low temperatures in the mantle wedge at convergent plate margins in the presence of excess H₂O or following the breakdown of the hydrous mineral chlorite. A new petrologic model that simulates anhydrous melting of variably metasomatized upper mantle is developed to investigate the petrogenesis of primitive basaltic lavas erupted in continental back-arc and ocean island settings (Chapter 3). This model suggests that <10.5 Ma anhydrous basaltic lavas erupted east of the Cascades arc were formed by mantle upwelling near the lithosphere-asthenosphere boundary caused by plate subduction, not by a mantle plume (Chapter 4). A geodynamic and petrologic study (Chapter 5) of asthenospheric flow at the margins of thick continental lithosphere reveals small degree melts of the mantle may be responsible for the large negative gradient in seismic wave speed observed at the lithosphere-asthenosphere boundary below eastern North America. These studies together advance our knowledge of how the recycling of tectonic plates on Earth affects melt generation in the Earth’s mantle and subsequently the unique differentiation of our planet.

Thesis Supervisor: Timothy L. Gove
Title: Professor of Geology
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“The pieces that I am, [they] gather them and give them back to me in all the right order.”
- Toni Morrison

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Introduction

The plate tectonic processes unique to Earth have played an important role in the chemical differentiation of the planet since early in its history, such that it is the only terrestrial planet with a highly differentiated continental crust. Consequently, knowledge of how plate tectonics affects melt generation in the Earth’s interior is fundamental to understanding the differentiation history of Earth. The link between convergent plate margins and the formation of highly differentiated magmas at volcanic arcs was recognized as early as 1892, when the “Volcanic Girdle of the Pacific” was first identified, which later became known as the “Andesite Line” or the “Ring of Fire.” But it was not until 1949, when it was demonstrated that earthquakes are located at progressively greater depths beneath the volcanic arc, that the presence of a subducting plate below arc volcanoes was clearly established. Since then, petrologic, geochemical, geophysical and geodynamic studies reveal thermal and chemical exchange between the subducting plate and the overlying mantle wedge are responsible for forming a physically and chemically complex melting regime. Volatiles such as H₂O and CO₂ liberated from the subducting plate and the accompanying soluble elements alter the composition and melting behavior of the overlying mantle, a process known as “metasomatism.” Viscous coupling between the hot mantle and the cool subducting plate induces downward flow at the base of the wedge and draws mantle into the top half of the wedge, a process known as “corner flow.” Despite these advances in our knowledge, many uncertainties remain regarding the specifics of melt generation in subduction-modified mantle. This thesis focuses in advancing our understanding of melt generation in mantle modified by both
modern and historical subduction. It is comprised of five distinct studies that employ experimental, petrologic, geochemical, geodynamic, and geophysical tools to accomplish this goal.

The first two chapters are experimental investigations of the high pressure melting behavior of mantle peridotite in the presence of excess H₂O. These chapters together establish that the forefathers of highly differentiated arc magmas are born at the base of the mantle wedge as a result of the high flux of H₂O off the subducting plate. In Chapter 1, we resolve a long-standing controversy regarding the melting temperature, or solidus, of H₂O-saturated mantle peridotite and demonstrate that the H₂O-rich mantle melts at the extremely low temperature of ~800°C over the pressure range relevant for subarc mantle melting. This study also suggests the hydrous mineral chlorite is stable in the mantle wedge, as well the subducting plate. Furthermore, chlorite contains sufficient H₂O to trigger H₂O-undersaturated melting or perpetuate ongoing H₂O-saturated mantle melting when it breaks down. When these findings are coupled with thermal and geodynamic models of subduction zones, we observe that the beginning of hydrous melting in the mantle wedge directly below arc volcanoes is primarily the result of chlorite breakdown and the low melting temperature of H₂O-saturated peridotite. Chapter 2 examines the composition of the low extent melts of the H₂O-saturated mantle peridotite generated in a subset of the experiments in Chapter 1. The compositions of hydrous experimental melts are notoriously difficult to analyze because of H₂O-exsolution and the growth of “quench” minerals when the experiments are depressurized. In this chapter we explore a variety of techniques to filter chemical analyses of the quenched material and we are able to determine the compositions of both a high-pressure melt and vapor phase in these
experiments. These compositions are used to constrain the melting reactions and the rates of melt production applicable to H$_2$O-saturated mantle at the base of the mantle wedge. As these melts ascend through the mantle wedge and crust, modification by crystallization and assimilation will result in the formation of the magma types erupted at arc volcanoes.

A significant number of primitive anhydrous basaltic lavas erupted in continental back-arc settings and at ocean islands exhibit geochemical characteristics that suggest they are the product of melting previously metasomatized mantle. Chapters 3 and 4 shift the focus from hydrous to anhydrous melting to explore the compositions and conditions in the mantle required to generate these lavas. Although quantitative models exist for the type of anhydrous mantle melting that occurs at mid-ocean ridges, they do not accurately predict the melting behavior of mantle previously modified in subduction zones. In **Chapter 3**, we develop a new petrologic model to simulate anhydrous melting of variably metasomatized upper mantle and explore the effects of mantle pressure, temperature, and composition on melt composition. The model is calibrated with a compilation of anhydrous partial melting experiments on a variety of compositions relevant to the mantle. This model can predict the temperatures and compositions of melts formed from variably metasomatized mantle, or calculate the temperature and pressure at which a primary basaltic magma of interest was produced. For example, **Chapter 4** applies our model to anhydrous melt generation in southern Oregon and northern California, where the mantle has been metasomatized by subduction of the Farallon plate for approximately 50 Ma. A notable number of volcanic centers have formed east of the Cascades volcanic arc within the last 25 Ma and scientists have
ascribed their formation to a variety of tectono-magmatic processes, such as crustal extension, lithospheric delamination, or mantle flow related to the ongoing subduction to the west. We calculate the temperatures and pressures at which primitive basalts from these volcanic centers originated in the mantle in order to draw conclusions about the tectonic and magmatic driving forces responsible for their formation. The calculated depths of melt extraction are also compared to the geophysical estimates of the location of the crust-mantle boundary in this region to place new constraints on the thickness of the continental lithosphere.

In the broadest sense, the lithosphere is the rigid material that makes up the Earth’s tectonic plates and it is underlain by the asthenosphere, which is the part of the Earth’s mantle that undergoes solid-state creep such that it exhibits fluid-like behavior on geologic time scales. Yet the thermal, mechanical and chemical variations of the mantle that create the lithosphere-asthenosphere boundary (LAB) are poorly understood. Scientists have previously attributed the negative gradient in seismic velocity observed at the LAB to variations in chemical composition, grain size, volatile content and/or temperature between the lithosphere and the asthenosphere. However, none of these mechanisms are capable of producing the large negative seismic velocity gradient observed at the LAB below eastern North America. Thick continental lithosphere, such as found below the North American continent, is constructed through many generations of plate subduction. In Chapter 5, we conduct finite element models of asthenospheric flow at an abrupt lateral decrease in lithospheric thickness, similar to the geometry of the eastern edge of the North American continent. These models suggest this geometry, together with lateral plate motions, produces edge-driven convection and asthenospheric
upwelling capable of triggering adiabatic decompression melting. Whether the upwelling asthenospheric mantle will melt depends on its chemical composition and volatile content, as demonstrated in Chapters 1 and 3. Using observations from these studies, we develop a parameterization of the H2O-undersaturated peridotite solidus that can be applied to our finite element models. This study illustrates that mantle with average temperatures and average to slightly elevated volatile contents will produce small extent melts below eastern North America, suitable for creating the seismic wave speeds observed at the LAB.

In summary, the recycling of tectonic plates at subduction zones has a profound effect on the composition and melting behavior of the Earth’s mantle. Volatiles fluxed from a subducting plate trigger melting in the mantle wedge at temperatures ~600°C lower than would otherwise occur, as demonstrated in Chapter 1. This melting process produces H2O-rich basaltic magmas that evolve to form the highly differentiated continental crust unique to Earth, as examined in Chapter 2. The chemical alteration of the overlying mantle in subduction zones has lasting effects on its melting behavior, as illustrated in Chapter 3. This melting behavior is evident in the basalts from southern Oregon and northern California studied in Chapter 4. Finally, Chapter 5 investigates how an abrupt change in lithospheric thickness combined with the chemical and volatile budget of the mantle affect the rheologic structure of the Earth. Together, the work presented in thesis advances our knowledge of how plate tectonics drives magma genesis and consequently, the chemical differentiation of our planet.

The chapters in this dissertation are intended to be five free-standing manuscripts, which are published (Chapter 5), accepted for publication pending revision (Chapter 1) or
in preparation for submission (Chapters 2, 3, 4). The corresponding publication reference information is listed below in the order of the chapters presented here.


Chapter 1:
The Beginnings of Hydrous Mantle Wedge Melting
Abstract

This study presents new phase equilibrium data on primitive mantle peridotite (0.33 wt.% Na₂O, 0.03 wt.% K₂O) in the presence of excess H₂O (14.5 wt.% H₂O) from 760-1200°C at 3.2-6 GPa. Based on textural and chemical evidence, we find that the H₂O-saturated peridotite solidus remains isothermal between 800 and 820°C at 3–6 GPa. We identify both quenched solute from the vapor phase and quenched silicate melt in supersolidus experiments. Chlorite is stable on and above the H₂O-saturated solidus from 2 to 3.6 GPa and chlorite peridotite melting experiments (containing ~6 wt.% chlorite) melt at the chlorite-out boundary over this pressure range, which is within 20°C of the H₂O-saturated melting curve. Chlorite can therefore provide sufficient H₂O upon breakdown to trigger melting in the mantle wedge or perpetuate ongoing H₂O-saturated melting. Constraints from recent geodynamic models suggest in a hot subduction zone like Cascadia, there is significantly more H₂O fluxed from the subducting slab near 100 km depth than can be bound in a layer of chloritized peridotite ~1 km thick at the base of the mantle wedge. Consequently in hot subduction zones, H₂O is available to trigger melting at the H₂O-saturated solidus in the lowermost mantle wedge. Alternately in cool subduction zones like the Northern Marianas, a layer of chloritized peridotite up to 1.5 km thick could contain all the H₂O fluxed from the slab every million years near 100 km depth, which suggests the dominant form of melting below arcs in cool subduction zones is dehydration melting. Slab $P$-$T$ paths from recent geodynamic models also allow for melts of subducted sediment, oceanic crust and/or sediment diapirs to interact with hydrous mantle melts within the mantle wedge at intermediate to hot subduction zones.
1. Introduction

Central in our pursuit to understand the processes that give rise to arc magmas, are the observations of the petrologic, chemical and volatile content of the erupted rocks. Work over the past 30 years has resulted in a preponderance of evidence to suggest arc parental magmas commonly contain up to 4 to 6 wt.% H₂O (e.g., Sisson and Grove, 1993a; Carmichael, 2002; Grove et al., 2003; Pichavant and MacDonald, 2007) and some arc andesites contain up to 8 to 10 wt. % H₂O (e.g., Sisson and Grove, 1993b; Carmichael, 2002; Grove et al., 2005; Wallace, 2005). However, considerable uncertainty remains about the physically and compositionally complex processes that lead to the generation of hydrous arc magmas with these observed water contents. Furthermore, a comprehensive understanding of the temperatures of hydrous melting in the mantle and the properties of these melts is a fundamental constraint for geodynamic models as they develop an integrated, three-dimensional picture of the flow of fluid and melt through the mantle wedge at subduction zones (e.g., Cagnioncle et al., 2007; Billen, 2008).

Previous work on garnet peridotite melting in the presence of H₂O has established that H₂O dramatically lowers the peridotite solidus (Kushiro et al., 1968b; Green, 1973; Millhollen et al., 1974; Mysen and Boettcher, 1975; Kawamoto and Holloway, 1997; Grove et al., 2006; Green et al., 2010). One challenge facing experimental studies of hydrous peridotite melting is the exsolution of H₂O from hydrous melts during depressurization, which transforms the melt and/or vapor solute into a froth of micron-thick bubbles and/or quench phases. Other challenges include, attaining chemical equilibrium, maintaining bulk composition, and distinguishing between quench of a silicate-rich vapor phase and a melt phase to determine if a near solidus experiment is
melted. Here we address these challenges and document the systematics of melting of H₂O-oversaturated and chlorite-bearing undepleted peridotite from 3 to 6 GPa. These experiments are used to understand the temperatures and chemical reactions of mantle wedge melting that constitute the primary controls on 1) the location of arc volcanoes, 2) the width of the volcanic arc, and 3) the abundance of wet vs. anhydrous primary magmas at arcs.

2. Experimental and Analytic Methods

2.1 Starting Materials

The primitive upper mantle composition of Hart and Zindler (1986) was synthesized by two different methods as experimental starting materials for this study (Table 1). The starting material for the H₂O-saturated experiments (“H & Z + H₂O” in Table 1) was created from a mixture of high-purity synthetic oxides or metasilicates. Brucite, Mg(OH)₂, was added to the synthetic oxide mixture instead of MgO, which provided the starting composition with ~14.5 wt% H₂O. The chlorite peridotite experimental starting material (“H & Z natural” in Table 1) is a mixture of natural olivine, orthopyroxene, clinopyroxene, and garnet grains separated from a garnet lherzolite xenolith found in Pali-Aike, Patagonia (Stern et al., 1989; Koga et al., 1999) and chlorite from New Idria, California (Van Baalen, 2004). The chlorite (Mg#=0.94) is near the Mg-end member clinochlore (Mg₃Al)((AlSi₃)O₁₀(OH)₈) which contains ~12 wt. % H₂O (Deer et al., 1962). Thus the chlorite peridotite starting material for these
experiments contains ~0.7 wt.% H\textsubscript{2}O. All starting materials were weighed out and ground for 3 hours under ethanol, using an agate mortar and pestle.

2.2 Piston Cylinder Experimental Methods

The H\textsubscript{2}O-oversaturated experiments from 3.2 to 4 GPa and the chlorite peridotite experiments were performed in a 0.5” end-loaded solid-medium piston cylinder apparatus (Boyd and England, 1960) in the MIT Experimental Petrology Laboratory (Table 2). All piston cylinder experiments were conducted in Au capsules to minimize the loss of Fe from the starting material to the capsule and because Au is impermeable to hydrogen at these experimental conditions. The experimental samples were prepared by packing 30 mg of powdered starting material into an Au capsule fabricated from 0.375” long Au tubes with a 0.01” wall thickness. The Au tube was triple-crimped and welded at both ends and flattened to make a uniformly thick capsule top and bottom. The welded capsule is ~0.15” in length and able to retain the H\textsubscript{2}O added to the starting material as brucite or chlorite. The sealed capsule was then placed in a nonporous Al\textsubscript{2}O\textsubscript{3} sleeve and positioned in the center of a graphite furnace using MgO spacers. The pressure medium consisted of sintered BaCO\textsubscript{3} and the pressure was calibrated using the breakdown reaction of Ca-tschermak pyroxene to anorthite + gehlenite + corundum and the spinel to garnet transition in CMAS (Hays, 1966; Longhi, 2005). The run pressures for these experiments are thought to be accurate to ± 0.05 GPa. The temperature was monitored and controlled using W\textsubscript{97}Re\textsubscript{3}/W\textsubscript{75}Re\textsubscript{25} thermocouples with no correction for the effect of pressure on thermocouple EMF. The thermal gradient in our piston cylinder assembly has been determined by two independent methods: direct measurement with offset
thermocouples and temperature mapping using the kinetics of the MgO + Al₂O₃ = MgAl₂O₄ reaction (Watson et al., 2002; Medard et al., 2008). Both methods indicate that the hot spot in the graphite furnace containing the experimental capsule is ~20°C (18 ± 6 °C: Medard et al., 2008) hotter than the thermocouple location, 1.5 mm above the capsule. The temperature difference across the capsule is ~10°C (9 ± 5 °C: Medard et al., 2008), with the hottest temperatures at the top of the capsule. The run temperatures for our experiments are thus accurate to ± 10°C. To raise the piston cylinder to pressure, experiments were pressurized to 1 GPa at room temperature, then the temperature was raised at 100°C/min to 865°C or the final run temperature if below 865°C. The experiment was held at these conditions for 6 minutes before raising the pressure to the final experimental pressure (reported in Table 2) and the temperature at 50°C/min to the final temperature if >865°C. The sample was held at isothermal and isobaric conditions for the duration of the experiment (24-338 hours; Table 2) and then quenched by turning off the power. Following each experiment, a hole was drilled in the capsule with a small diameter hand drill and the presence or absence of liquid water was noted (Table 2). The capsule was then sliced in half, dried, vacuum-impregnated with epoxy and polished for electron microprobe analysis.

2.3 Multi-anvil Experimental Methods

Six H₂O-saturated experiments at 5 and 6 GPa (Table 2) were conducted in the 1000-ton Walker-type multi-anvil apparatus at the University of Minnesota Experimental Petrology Lab. Au tubing 2 mm in diameter and 2.5 mm in length was quadruple-crimped, welded and flattened on one end to form the experimental capsule. The capsule
was then packed with 10 mg of powdered starting material and the open end was single-
crimped and welded in a water cooled fixture to form a closed capsule ~2 mm in length. 
For these experiments, WC anvils with 12 mm truncations, cast MgO-Al₂O₃-SiO₂-Cr₂O₃ 
octahedra with integrated gasket fins and straight-walled graphite heaters (comprising the 
“12-TEL assembly”) were used. Further details of the experimental design are given by 
Withers et al. (in press). The pressure calibration for the 12-TEL assembly was 
established by bracketing multiple high temperature fixed points (Dasgupta et al., 2004). 
On the basis of 6 brackets and 2 half-brackets at temperatures that range from 700 to 
1350°C, the pressure uncertainty is estimated to be ± 0.3 GPa. The force-pressure 
relationship shows no observable dependence on experimental temperature within the 
700-1350°C temperature range. The temperature uncertainty, based on orthopyroxene-
clinopyroxene thermometric measurements and brackets of the congruent sphene melting 
reaction, is estimated to be ± 15°C. Experiments were quenched by switching off the 
heating circuit and subsequently were depressurized over a period of ~10 hours.

2.4 Analytical Methods

The major element concentrations of all experimental products were analyzed by 
wavelength dispersive spectrometry on the 4- or 5-spectrometer JEOL 733 and 8200 
JEOL electron microprobes (EMP) at the Massachusetts Institute of Technology. All 
analyses of minerals presented in Table 3 were conducted with a 15kV accelerating 
voltage, a 10 nA beam current and a 1 μm spot size. Online data reduction utilized the 
CITZAF correction package (Armstrong, 1995) and the atomic number correction, the 
absorption coefficients, and the fluorescence correction available in CITZAF.
3. Experimental Results

3.1 Mineral Phase Relations & Textures

Extensive EMP analyses of all minerals present in our experiments were conducted to assess the homogeneity of each phase, and the extent of any systematic changes in composition with temperature or pressure (Table 3). All experiments in this study run below 1100°C contain the minerals olivine, orthopyroxene, clinopyroxene and the anhydrous aluminous phase, garnet (Figure 1; Table 2). At temperatures greater than 1100°C at 3.2 GPa olivine, orthopyroxene and spinel are the dominant solid phases.

Textures in our experiments are identical to those described in Grove et al. (2006). Olivines tend to be euhedral to subhedral (10-150 μm) and orthopyroxenes are small (≤20 μm long) needle-like laths in the lower temperature experiments and more prismatic at higher temperatures (≤40 μm). Clinopyroxenes also tend to be elongate and prismatic and increase in size with temperature (20-100 μm). Garnets tend to be larger (100-200 μm) and poikilitic in the experiments where chlorite is absent (Figure 2). Where garnet and chlorite coexist, garnet tends to be smaller (<40 μm), likely limited in size by Al-partitioning between these two Al-bearing phases.

We observe a marked change in texture in our experiments at ~810 °C from 3 - 6 GPa. Experiments conducted at or below 800°C tend to be fine grained and homogeneous in texture with no segregation of the phases over the height of the capsule, whereas those at temperatures above 800°C are coarse grained, exhibit significantly more void space and are mineralogically-zoned, with orthopyroxene, clinopyroxene and garnet
segregated toward the cool region (bottom) of the capsule and large olivine grains and melt segregated toward the hot end (top) (Figure 2). These textures are similar to some textures documented in other fluid-present melting experiments (e.g., Stalder and Ulmer, 2001; Schmidt and Ulmer, 2004; Kessel et al., 2005; Grove et al., 2006; Luth, 2006; Medard and Grove, 2006). Our experiments also exhibit changes in mineral composition that occur with increasing temperature above 800°C at 3.2 and 4 GPa. Figure 3 illustrates a persistent increase in the Mg# of the solid phases with increasing temperature at constant pressure. These experiments at constant pressure also reveal trends in the minor element concentrations of the clinopyroxene, orthopyroxene and garnet with increasing temperature as shown in Figure 4. Additional abrupt changes in texture or composition of the experimental phases are not observed with increasing temperature, including the temperature range from 1000 to 1300°C at 3.2 and 4 GPa. In experiments at or above 880°C, small wisps of inter-granular glassy quenched material are visible (Figure 2).

These observations lead us to interpret the abrupt change in texture between 800 and 820°C and the progressive change in mineral composition above this temperature, as evidence of the presence of the H2O-saturated solidus at ~810°C from 3 to 6 GPa. Experiments bracket this critical behavior within 20°C at 3.2, 3.6, 4, and 5 GPa and 30°C at 6 GPa and reveal the steep slope of the solidus in P-T space (Figure 1). We attribute the modal zoning in experiments at or above 820°C as the result of thermal compaction facilitated by the presence of melt, a small thermal gradient and the long experimental durations. The modal zoning does not occur in experiments at temperatures less than 820°C, likely because only crystals and aqueous fluid are present and the solubility of the silicates in solution is not sufficient for dissolution and recrystallization to occur.
Mineral phases in subsolidus experiments with durations longer than 168 hours are compositionally uniform throughout the experiment and homogeneous (Figure 2; Table 3). Supersolidus experiments, with durations of 168 hours or longer near 800°C and 24 hours or longer for T >1050°C, also exhibit homogeneous mineral compositions with orthopyroxene crystals displaying a more Ca-rich rim, which is likely the result of interaction with the melt during quenching.

Chlorite is stable to temperatures just above the H₂O-saturated solidus between 2 and 3.6 GPa (Figure 1). Above 3.6 GPa, the H₂O-saturated solidus and the chlorite stability fields diverge. The composition of the chlorite in the experimental products is closest to the Mg-rich chlorite end-member, clinochlore (Table 3). Chlorite is distributed throughout the experimental charges, and not restricted to the edges or cracks where hydrous fluids segregate. A similar high-pressure chlorite phase has been observed in experiments in the MgO-Al₂O₃-SiO₂-H₂O (MASH) (Pawley, 2003) and Na₂O-CaO-FeO-MgO-Al₂O₃-SiO₂-H₂O (NCFMASH) systems (Fumagalli and Poli, 2005; Dvir et al., 2011). This work contributes 10 new experiments to further define chlorite stability between 3.2 and 4 GPa and better constrains this phase boundary at 3.6 and 4 GPa relative to previously published experiments (Table 2).

A small amount of carbonate appears as a quench phase in experiments conducted more than a year and a half after the preparation of our H & Z + H₂O starting material, although no carbon was used in the synthetic oxide mixture and no melt traps were used in the experiments. We interpret the appearance of carbonate as the result of progressive carbonation of the synthetic oxide mixture over time from atmospheric CO₂. This has been observed in other experimental studies when the starting mix is not kept under
vacuum (M. Schmidt, personal communication, 2010). These experiments are still valuable, as a small amount of CO\textsubscript{2} added into our starting material to create a mixed CO\textsubscript{2}-H\textsubscript{2}O vapor is more realistic for simulating mantle melting processes than a pure H\textsubscript{2}O volatile component, as CO\textsubscript{2} concentrations of the mantle are estimated to be 50-2000 ppm below mid-ocean ridges (Saal et al., 2002) and 1000-4000 ppm below oceanic islands (Trull et al., 1993; Dixon et al., 1997; Pineau et al., 2004).

### 3.2 Chlorite Peridotite (H & Z Natural) Melting Behavior

Six chlorite peridotite melting experiments were conducted using the natural mineral starting material, which contains no H\textsubscript{2}O except that bound in the hydrous phase chlorite (chlorite contains 12 wt.% H\textsubscript{2}O, ~6 wt.% chlorite in starting material = ~0.7 wt.% H\textsubscript{2}O in starting material). In these experiments, the solidus is displaced to slightly higher temperatures, relative to the H\textsubscript{2}O-saturated case, and follows the chlorite-out boundary between 2 and 3.6 GPa (Figure 5). This is significant because it suggests there is sufficient H\textsubscript{2}O released from the breakdown of chlorite to allow H\textsubscript{2}O-saturated melting. The chlorite-out boundary, and therefore the solidus for chlorite peridotite, is 20-30°C higher in temperature than the H\textsubscript{2}O-saturated solidus between 2 and 3.6 GPa. Above 3.6 GPa and below 2 GPa, chlorite breaks down at temperatures below the H\textsubscript{2}O-saturated solidus and the first melts occur at the H\textsubscript{2}O-saturated solidus.

### 3.3 Approach to Equilibrium

The H\textsubscript{2}O-saturated piston-cylinder and multi-anvil experiments reported in this paper are all synthesis experiments (i.e., minerals grew from a synthetic oxide mixture or...
are reacted natural minerals selected from a garnet peridotite and natural chlorite sample) with durations from 24 to 338 hours (median = 168 hours) (Table 2). The combination of the H$_2$O-saturated experiments and experiments using natural mineral powders as starting material, are analogous to reversal-type experiments. Both types of experiments yield similar mineral compositions and solidus temperatures, with the exceptions of experiments where chlorite is stable above the H$_2$O-saturated solidus. In addition, three observations indicate that in both types of experiments an approach to equilibrium was attained: (1) the growth of compositionally homogeneous chemically unzoned minerals with triple-junction contacts to one another, (2) maintenance of constant bulk composition, and (3) temperatures calculated from coexisting orthopyroxene and high-Ca clinopyroxene in 3.2 GPa H$_2$O-saturated experiments yield temperatures similar to the experimental conditions (Grove et al., 2006). Two experiments that bracket the H$_2$O-saturated solidus at 3.2 GPa were run for 24 hours (D169 and D167) for comparison to longer experiments at these conditions. Melting textures are detected in the higher temperature of these short duration experiments, but mineral compositions are very heterogeneous and individual grains are strongly zoned. An experiment run for only six hours at 880°C and 3.2 GPa (D212; Table 2) remained synthetic powder starting material with the exception of the growth of several small garnet porphyroblasts.
4. Discussion

4.1 Comparison with Previous Experimental Studies

Seven prior experimental studies have investigated the H2O-saturated melting of peridotite over a range of similar pressures (Kushiro et al., 1968b; Green, 1973; Millhollen et al., 1974; Mysen and Boettcher, 1975; Kawamoto and Holloway, 1997; Grove et al., 2006; Green et al., 2010) and are compared in Table 4. The P-T conditions of H2O-saturated melting in the 1960’s and 1970’s studies fall into two groups based on the duration of the experiment. Shorter experimental durations (<24 hours) find solidus temperatures of ~1000°C at 3 GPa and longer experimental durations (>24 hours) yield solidus temperatures of ~800°C at 3 GPa (Table 4). Modern experimental techniques allow us to avoid problems encountered in the earlier studies, for example short experimental durations, which lead to problems attaining equilibrium, and non-ideal capsule material (e.g., Pt or Mo) that lead to loss of H2 and Fe. Short experimental duration peridotite melting experiments yield quantities of solidus depression similar to those observed for H2O-saturated diopside, while longer experimental duration experiments yield quantities of solidus depression that are ~200°C greater, similar to those observed for H2O-saturated forsterite (Kushiro, 1969; Hodges, 1974; Eggler and Burnham, 1984; Luth, 1993) and confirms the likelihood that shorter experimental durations record the solidus depression of the phase with the fastest melting kinetics rather than equilibrium melting. The results presented here are most similar in the temperature of the H2O-saturated solidus to those of Mysen and Boettcher (1975), whose
experiments contained ~20 wt.% H₂O, lasted up to 64 hours, and employed AgPd capsules, which are inert to exchange of Fe with the starting material (Table 4).

Scientists have long debated whether aqueous fluids in equilibrium with silicate minerals at high pressures (> 3 GPa) contain high concentrations of silicates in solution (Bowen and Tuttle, 1949; Kushiro, 1972a; Nakamura and Kushiro, 1974; Tatsumi et al., 1986; Zhang and Frantz, 2000; Newton and Manning, 2002) and some recent work suggests fluid solute concentrations may be on the order of 15-40 wt. % when in equilibrium with model mantle compositions for the pressure range considered here (Stalder et al., 2001; Mibe et al., 2002; Kessel et al., 2005; Dvir et al., 2011). Also relevant are the many studies that examine the location of the second critical end point (SCEP) in simple compositional systems (e.g., 1.5 GPa in albite-H₂O: Bureau and Keppler, 1999; Stalder et al., 2000); ~12 GPa in MSH-H₂O: Stalder et al., 2001). However the location of the SCEP is not well understood in the peridotite-H₂O system due to the difficulty of these experiments. Direct observations of the SCEP in the peridotite-H₂O system were made using in situ X-ray radiography techniques in a multianvil apparatus and locate the SCEP by the disappearance of distinctive round shapes, thought to represent spheres of either aqueous fluid or silicate melt in the other phase formed from interfacial tension below the SCEP (Mibe et al., 2007). Using this technique the SCEP was found to be ~3.8 GPa and 1000°C. Based on the presence of both quench spheres and wisps in our supersolidus experiments with distinctive compositions (Figure 6, Table 3) that we interpret to be quenched fluid and melt, respectively, we conclude all of our supersolidus experiments are located below the
SCEP. Therefore the question of how to determine quenched melt from quenched vapor phase in our experiments logically arises.

Using recent observations of element solubility in high-pressure fluids in concert with our experimental data, we can examine the quenched vapor versus melt question for our experiments. Na, Al, are found to be the most soluble major elements in a high pressure fluid phase in equilibrium with peridotite over the pressure range 2-6 GPa, followed by Ca and Si, while Mg and Fe are found to be relatively insoluble (Brenan et al., 1995; Adam et al., 1997; Ayers et al., 1997; Stalder et al., 1998; Dvir et al., 2011). These fluids are also rich in large ion lithophile elements relative to the high field strength elements. Figure 3 reveals a steady increase in the Mg# of the solid phases in our experiments and Figure 4 illustrates the change in minor element concentrations in the two pyroxene phases and garnet, both with increasing temperature at 3.2 GPa. As demonstrated by the predicted batch melting curves for these elements, the changes in the composition of the solid phases are not consistent with simple partitioning of elements into a vapor phase. Similar changes in the clinopyroxene composition with increasing degree of melting are observed in Tenner et al. (2009) and Kawamoto and Holloway (1997) used an increase in the Mg# of olivine crystals with increasing temperature as a criterion for determining when experiments were above the solidus (Table 4). Therefore we conclude the experiments above 810°C exhibit evidence for quenched melt, in addition to a quenched vapor phase.

In addition to the persistent increase in Mg# of the solid phases, two abrupt changes in the Mg# of several of the solid phases can also be identified: a jump in the Mg# of all solid phases between 840 and 880°C corresponds to the breakdown of chlorite
and a drop in the Mg# of olivine and clinopyroxene between 940 and 980°C corresponds to the breakdown of clinohumite (Figure 3). The breakdown of chlorite can be easily observed in our experiments as discussed in section 4.4. The abrupt change in Mg# between 940 and 980°C is consistent with the location of the clinohumite breakdown reaction in the MASH system (Pawley, 2000; Stalder and Ulmer, 2001). Small amounts of titanoclinohumite are observed in our experiments at temperatures below 900°C at both 3.2 and 3.6 GPa (Table 2 & 3). Breakdown of clinohumite may occur in a reaction proposed by Stalder and Ulmer (2001) where,

\[
0.5 \text{ enstatite} + \text{clinohumite} = 5 \text{forsterite} + H_2O. \tag{1}
\]

Thus the decrease in the modal abundance of orthopyroxene and the simultaneous increase in the abundance of olivine observed in our experiments between 880 and 980°C also suggests clinohumite breakdown is occurring in our experiments between these two temperatures (Table 2). Because our experiments are not conducted in the simplified MASH system, the iron in the orthopyroxene may then be partitioned into olivine and clinopyroxene and drive their Mg# down at clinohumite breakdown. A similar “leveling” of the Mg# of the solid phases between phase boundaries is observed in hydrous peridotite melting experiments by Mysen and Kushiro (1977) at constant pressure and increasing temperature.

An alternative interpretation of the change in Mg# of two solid phases at either temperature is that they represent the solidus boundary. For example, if there is an increase in the Mg/Si ratio of the solute, either at the solidus or with increasing pressure (Stalder et al., 2001; Dvir et al., 2011), this could produce a corresponding drop in the Mg-content of solids. However, these compositional shifts do not correspond to any
determinations of the temperature H₂O-saturated peridotite solidus, at either ~800°C (Mysen and Boettcher, 1975; Grove et al., 2006) or ≥1000°C at 3 to 6 GPa (Kushiro et al., 1968b; Green, 1973; Millhollen et al., 1974; Kawamoto and Holloway, 1997; Green et al., 2010). Furthermore, no abrupt change in Mg# is observed in garnet or orthopyroxene between 940 and 980°C. Our starting material has sufficient H₂O to have an H₂O vapor phase present in addition to the H₂O partitioned into a melt to more than 55 wt.% melting. Therefore changes in the Mg# of the solid phases above 840°C are either the result of phase transitions, such as chlorite and clinohumite breakdown, or require a better understanding of element partitioning between solid, melt, and fluid to resolve.

In the low-K₂O, -Na₂O, -TiO₂ and high H₂O (14.5 wt.%) bulk mantle composition used in our experiments, amphibole is stable to ~2 GPa and 1000°C as presented by Grove et al. (2006). Other experimental studies find that alkali and TiO₂ contents, as well as oxygen fugacity, have a significant effect on amphibole stability and can stabilize amphibole to pressures up to 3 GPa in lherzolite compositions (Wyllie, 1978; Mengel and Green, 1989; Niida and Green, 1999; Fumagalli and Poli, 2005). Green et al. (2010) suggest pargasite is stable to 3 GPa for a bulk composition similar to ours with significantly lower H₂O concentrations (≤ 1.45 wt.%), which they argue prevents leaching of highly soluble elements such as Na₂O. They present P-T conditions and the phase assemblage for two experiments with pargasite stable at 2.5 GPa and 0.3 wt.% H₂O and for an experiment at 3.0 GPa and 0.5 wt.% H₂O. Pargasite is also listed as a stable phase in a 2.5 GPa experiment at anhydrous conditions. The high H₂O content of our experiments may affect amphibole stability, as the mole fraction of H₂O in a mixed CO₂ + H₂O fluid phase has been observed to affect the thermal stability of pargasite at
constant pressure in the system pargasite-H₂O-CO₂ (Holloway, 1973) and phlogopite in a
spinel lherzolite composition (Yoder and Kushiro, 1969; Wendlandt and Eggler, 1980).
However, changes in hydrous phase stability with bulk H₂O content did not affect the
solidus temperature in the above cases and likewise would not affect the determination of
the H₂O-saturated peridotite solidus presented here.

Instead, the method of determining whether an experiment is sub- vs. supersolidus
may play a large role in the determination of solidus temperature. Green et al. (2010)
uses the presence of interstitial patches of amphibole and/or quench clinopyroxene, mica
and glass, all with Mg#'s of 70-85 within the lherzolite and melt trap layers as the
diagnostic criterion for the presence of quenched hydrous silicate melt in an experiment.
Conversely, the lack of the quench mafic silicates, as well as the presence of hydrous
phlogopite or pargasite with Mg#'s greater than or equal to olivine is used to diagnose
their subsolidus experiments. This choice of criterion for sub- and supersolidus
experiments may bias their identification of the solidus to high temperatures where melts
are mafic enough to grow amphibole ± clinopyroxene upon quench. Experimental work
thus far on the composition of hydrous spinel and garnet lherzolite melts suggests the
melt is andesitic at lower pressures and temperatures and becomes more mafic as melting
continues (Kushiro et al., 1968b; Kushiro, 1972b; Kawamoto and Holloway, 1997;
Gaetani and Grove, 1998; 2003; Grove et al., 2006). The coexistence of melt with
olivine in the hotter portion of hydrous melting capsules indicates peritectic melting is
occurring to produce a SiO₂-rich melt composition (Kushiro et al., 1968b). Furthermore,
Gaetani and Grove (1998; 2003) found that the SiO₂ increases by ~1 wt.% and FeO and
MgO decrease by ~2 wt.% in hydrous experimental melts with the addition of 3-6 wt.%
dissolved H$_2$O. Preliminary analyses of the melt in our experiments support these findings (Table 3). Our textural and chemical diagnostic criterion as discussed above, do not bias our detection of melt to any melt composition. Green et al. (2010) also observes euhedral olivine or pyroxene crystals “coated with fragmented glass ‘froth’ or with films of silicate glass” in high water content (> 5 wt.%) experiments, which they interpret to be subsolidus. We observe similar features in experiments we interpret to be supersolidus based on the Mg# and minor element composition of the solid phases and textures. The lack of compositional data for the minerals, melts and fluids in the experiments presented in the Green et al. (2010) study make it difficult to further resolve our conflicting interpretations of the solidus temperature.

Other experimentally-determined hydrous solidi have similarly steep slopes to that presented here for the H$_2$O-peridotite system. The H$_2$O-saturated solidus for a KLB-1 model mantle of Kawamoto and Holloway (1997) as well as that for the system forsterite+ H$_2$O (Luth, 1993) both remain isothermal between 3 and 12 GPa. The critical curves for both hydrous sediment and basalt also exhibit steep slopes over several GPa near their inflection point (e.g., Hermann and Spandler, 2008; Hill and Boettcher, 1970b), as does the MgSiO$_3$ + H$_2$O system (Kushiro et al., 1968a; Hodges, 1974). Theoretical analysis of wet melting relations in a MSH system suggests that such steep slopes are common as $P$-$T$ minima are reached in these types of critical curves and represent an inflection in density where $\Delta V=0$ for the phases involved (Boettcher and Wyllie, 1969; Hack et al., 2007). Eventually at a pressure to be determined, the H$_2$O-saturated peridotite solidus will turn over to have a positive slope as the change in volume of the vapor phase and solids outpaces the change in volume of the melt phase for a given $\Delta P$. 

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4.4 Chlorite Stability in the Mantle Wedge

Overall the chlorite-out boundary from our experiments agrees with that from the MASH system (Pawley, 2003) and is ~50°C higher at 3.2 to 4 GPa than experiments in the NCFMASH system (Fumagalli and Poli, 2005; Dvir et al., 2011), which exhibit less magnesian chlorite than our experiments. Figure 3 illustrates iron’s affinity for garnet relative to the other solid phases including chlorite in our experiments. The chlorite breakdown reaction calculated at 4 GPa between 760 and 800°C is,

\[ 0.06 \text{ enstatite} + 0.05 \text{ clinopyroxene} + 0.19 \text{ chlorite} = 0.19 \text{ olivine} + 0.11 \text{ garnet} \] (2)

Thus in more iron-rich peridotite compositions chlorite stability is shifted to lower temperatures in favor of olivine and garnet. This thermal depression in the stability of chlorite with increasing iron content has been observed in other hydrous melting experiments on Martian mantle compositions, which contain two to three times as much wt.% iron as the Earth’s mantle (Medard and Grove, 2006).

A variety of rocks exhumed from high pressures in the Earth support our experimental observations of chlorite stability in peridotite at upper mantle pressures. Chlorite has been found in relatively high-grade metamorphosed ultramafic rocks from the lower crust, such as in the Western Gneiss Region, Norway (Medaris, 1984; Hacker et al., 2010) and in peridotite inclusions in kimberlites from the Colorado Plateau, USA (Smith, 1979). New evidence from exhumed exposures of the sub-arc mantle, like the Higashi-akaishi peridotite of the Sanbagawa belt in SW Japan (Enami et al., 2004; Hattori et al., 2009; Till et al., 2009), also support the conclusion that chlorite is stable in the mantle wedge. High-Mg chlorite is found as a stable prograde mineral and occurs as
euhedral crystals intergrown with and as inclusions in garnet or pyroxene crystals in the garnet peridotites from Higashi-akaishi, where the garnet rims represent the peak metamorphic conditions of 2.9-3.8 GPa and 700-810°C ca. 110 Ma, similar to the $P$-$T$ conditions of experiments in this study containing chlorite.

The experiments in this study support the hypothesis of Schmidt and Poli (1998) and Grove et al. (2006) that chlorite is a stable mineral in the $P$-$T$-$X$ range appropriate for the base of the upper mantle wedge. Chlorite, as well as serpentine, likely form with fluids released during the dehydration of serpentine and amphibole in the subducting oceanic crust and lithosphere (Schmidt and Poli, 1998; Iwamori, 1998; Pawley, 2003; Grove et al., 2006; Iwamori, 2007) and/or from the dehydration of subducted sediments in the mantle wedge nose at <80 km depth. Seismic images of the Honshu subduction zone reveal a low velocity layer that is interpreted to result from the presence of hydrous phases in the shallow wedge (Kawakatsu and Watada, 2007). The chlorite and serpentine can then be advected downward in the mantle wedge by corner flow. Provided all the Al$_2$O$_3$ present in the fertile mantle composition used here is available to form chlorite, the mantle above the slab can contain up to 6.5 wt.% chlorite, which equates to 2 wt.% H$_2$O in the mantle. The degree of serpentinization in an upper mantle with forsteritic olivine depends on the extent to which the system is open and fluids can transport elements such as magnesium away from the reacting volume. In an open system where complete serpentinization can occur, the maximum H$_2$O content of the upper mantle will be that of serpentine; 13 wt.% bulk H$_2$O. Therefore, 1 km$^3$ of serpentinitized mantle (up to 13 wt.% H$_2$O) with a density of 3.3 g/cm$^3$ can hold up to can hold up to 390 Tg of H$_2$O and 1 km$^3$ of chloritized mantle (2 wt.% H$_2$O) can hold up to ~66 Tg of H$_2$O. In a more depleted or
Fe-rich bulk composition, the addition of water can completely metamorphose olivine to serpentine and magnetite in a closed system and chlorite will break down at slightly lower temperatures as discussed above. Our experiments illustrate chlorite is stable at the undepleted peridotite solidus between 2 and 3.6 GPa and thus has the ability to supply water during its breakdown to trigger H₂O-saturated melting of the mantle, like in the experiments presented here with natural chlorite peridotite starting material, or to perpetuate H₂O-saturated melting previously triggered by H₂O fluxed from the hydrous phases in the subducting mantle lithosphere.

4.5 Implications for the Generation of Hydrous Melts in the Mantle Wedge

When combined with numerical models, our observations on chlorite stability and the H₂O-saturated peridotite solidus can provide constraints on melt generation in the mantle below the more than 40,000 km of arcs on Earth today. Forward numerical models of subduction zones have provided insights into how plate tectonic variables, such as convergence rate, slab age, dip, and structural setting (i.e., trench advance or rollback), affect the temperature distribution, dynamics and structure of the mantle wedge at subduction zones (e.g., Davies and Stevenson, 1992; Peacock et al., 2005; Kincaid and Griffiths, 2003; van Keken et al., 2002; Cagnioncle et al., 2007; van Keken et al., 2008). Here we focus on recent geodynamic models that explore slab surface temperature distributions calculated for a global suite of arc systems and therefore cover the global variability in plate tectonic variables. These models pay particular attention to the location and nature of the transition from decoupling to coupling between the slab and the overlying mantle wedge near the cold corner (e.g., Wada and Wang, 2009; Syracuse et
al., 2010). For all of the different sets of assumptions regarding the location of the slab-mantle coupling, global slab surface $P-T$ paths cross the serpentine-out curve (Ulmer and Trommsdorff, 1995), the chlorite-out curve (this study), and the $H_2O$-saturated peridotite solidus (this study) in the range of pressures and temperatures that are present at the slab surface directly below the positions of volcanoes in the global arc data set (2.5-6 GPa, 650-900°C: the region of overlap of the gray rectangles in Fig. 7). Therefore according to the numerical models, the thermal conditions just above the slab and directly below all global arc volcanoes are ideal for triggering the breakdown of hydrous minerals in the mantle wedge and subsequently $H_2O$-saturated mantle wedge melting.

Using these thermal models and recent estimates of $H_2O$ flux from subducting slabs at depth (van Keken et al., 2011), we can estimate the percentage of $H_2O$ fluxed from the slab that can be bound in chlorite in the base of the mantle wedge at subarc depths. Our constraints on the $P-T$ conditions of chlorite breakdown and the slab surface temperatures from Syracuse et al. (2010) where slab-mantle coupling occurs at 80 km depth predict chlorite would be stable in a region ~1 km thick above a slab surface at 100 km depth in the hottest subduction zones and a region up to 8 km thick in the coldest subduction zones, if the thermal gradient at the slab surface varies between 25°C/km - 50°C/km (Cagnioncle et al., 2007). The $P-T$ paths for the crust-mantle boundary at 7 km beneath the slab surface in the Syracuse et al. (2010) models predict that the subducting lithospheric mantle in all except the coolest subduction zones cross the serpentine dehydration curve but remain below the chlorite dehydration curve at subarc depths. Therefore only $H_2O$ from serpentine is released from the subducting mantle lithosphere at subarc depths. The calculations of van Keken et al. (2011) indicate that for the warm
Cascadian subduction zone, 11 Tg H$_2$O/Myr/m (along strike) is fluxed from the slab between 55 and 102 km depth if the slab has a serpentinized upper mantle that is 2 wt.% H$_2$O to 2 km depth. At the other end of the spectrum, they determine 3 Tg H$_2$O/Myr/m is fluxed from the cool North Marianas slab between 79 and 114 km depth given the same set of assumptions regarding slab hydration. Therefore near 100 km depth, ~11000 Tg H$_2$O are released over an area of 47 km$^2$ on the Cascadian slab surface due to serpentine dehydration, or ~3000 Tg H$_2$O over 35 km$^2$ on the Northern Marianas slab, every million years. Recall 1 km$^3$ of chloritized mantle peridotite can hold up to 66 Tg of H$_2$O, so a layer of chlorite peridotite ~1 km thick over 47 km$^2$ could hold ~3102 Tg H$_2$O.

Therefore in a hot subduction zone like Cascadia, there is significantly more H$_2$O fluxed from the subducting slab near 100 km depth every 1 million years than can be bound in a layer of chloritized mantle wedge peridotite ~1 km thick. This suggests, free H$_2$O is available to trigger melting at the H$_2$O-saturated solidus in the lowermost mantle wedge (Fig 8a & b) and the subsequent dehydration of chlorite formed at the base of the mantle wedge can perpetuate ongoing H$_2$O-saturated melting or ensure that H$_2$O is present above the H$_2$O-saturated solidus at subarc depths. In the Northern Marianas, a layer of chloritized peridotite up to 1.5 km thick could contain all the H$_2$O fluxed from the slab every million years near 100 km depth, which is well within the 8 km region above the slab where we predict chlorite to be stable in cool subduction zones. Therefore in a cool subduction zone like the Northern Marianas, no free H$_2$O is available to trigger melting until chlorite breakdown in the lowermost mantle wedge and the dominant form of melting at subarc depths is dehydration melting (Fig. 8c & d). This highlights the significance of chlorite as a stable phase in the mantle wedge and as a provider of the
H$_2$O necessary to trigger melting below arcs in cool subduction zones. Only the coldest outlier $P$-$T$ path (i.e., Tonga) in Figure 7, which is slightly too cold at subarc depths to cross the H$_2$O-saturated peridotite solidus, does not fit a model where the inception of hydrous melting below arcs is caused by some combination of the low temperature of the H$_2$O-saturated solidus and chlorite dehydration. We thus envisage a continuum in the way in which melting is initiated in the base of the mantle wedge at global arcs between a Cascadia-type melting scenario and a Marianas-type melting scenario.

Many arc lava trace element and isotope compositions suggest a subducted sediment and/or altered oceanic crust component is also added to hydrous melts of the mantle wedge at subduction zones. However, the trace elements such as Ba, Th, Be and Pb that are indicative of a subducted sediment component are relatively insoluble in the aqueous fluids and thus cannot be carried by the fluids fluxed from the subducting slab alone. A number of the $P$-$T$ paths for the intermediate (e.g., Sunda and Solomon) to hot (e.g., Chile and Cascadia) global arcs cross both the water-saturated sediment (Hermann and Spandler, 2008; Poli and Schmidt, 2002; Nichols et al., 1994; Vielzeuf and Schmidt, 2001) and oceanic basalt solidus (Lambert and Wyllie, 1970; Hill and Boettcher, 1970a; Liu et al., 1996) between 1.5 and 4 GPa, followed by the H$_2$O-saturated peridotite solidus (Figure 7). Thus above 1.5 GPa at these arcs, melts of both sediment and basalt may form at mantle wedge conditions and then ascend into the mantle wedge and interact with early melts of the H$_2$O-saturated mantle, just above the slab surface to form a composite melt with the trace element characteristics of many arc lavas. An alternate observation is that global slab $P$-$T$ paths when combined with estimates of the maximum water contents of hydrated MORB at subarc P-T conditions (Hacker, 2008) suggest most slab crust has
fully dehydrated when it reaches subarc depths (Syracuse et al., 2010). Recent numerical models find that at temperatures less than ~850°C, subducted sediment layers form buoyant instabilities (i.e., diapirs) that rise into the overlying mantle wedge within ± 40 km of the slab depth below the arc at 13 or 17 global subduction zones (Behn et al., in press). Therefore alternatively, sediment diapirs may advect relatively fluid-poor sedimentary material into the mantle wedge where they contribute their chemical signature to hydrous mantle wedge melts in the hotter core of the mantle wedge, which reaches temperatures in excess of ~1050°C, the temperature required to deplete metasediments of Th, Sr, Pb and Nd (Bebout et al., 1999; Becker et al., 2000; Busigny et al., 2003; Behn et al., in press). In either case, the H2O-saturated peridotite solidus presented here provides a mechanism for hydrous mantle melts to be generated at relatively low temperatures in the base of the mantle wedge and then interact with either sediment melts or diapirs and perhaps in some cases melts of the oceanic crust, to form the chemical composition of arc volcanics worldwide.
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Medaris, L.M., 1984, A geothermobarometric investigation of garnet peridotites in the Western Gneiss Region of Norway: Contributions to Mineralogy and Petrology, v. 87, p. 72–86.


Figure Captions

Figure 1. Phase equilibria for Hart and Zindler (1986) primitive mantle composition with the P-T conditions of experiments from this study (black) and Grove et al. 2006 (gray). Olivine, clinopyroxene and orthopyroxene are stable everywhere in the diagram with the exception of the highest temperature experiments at 3.2 GPa where clinopyroxene is absent. The aluminous and hydrous phase stability with the exception of chlorite breakdown between 2 - 4 GPa are from Ulmer and Trommsdorff (1995), Schmidt and Poli (1998), and Fumagalli and Poli (2005). The H2O-saturated solidus from this study (3.2-6 GPa) and Grove et al. (2006) (<3.2 GPa) is indicated by the bold black line.

Figure 2. Electron back-scatter images of experiments bracketing the change in texture we interpret as the H2O-saturated peridotite solidus. Experiments at ≤800°C are homogeneous and fine-grained in texture, whereas experiments at >800°C are more coarse-grained, porous, and exhibit mineralogical-zoning with olivine and melt segregated toward the hot (top) and opx and cpx toward the cold (bottom) end of the capsules (orientation as shown for piston cylinder experiments), garnet is distributed throughout. Thermal gradients are ~10°C across the gold capsule, which is the white area in each picture. Photos for experiments D179, D183, M338, M337 and D214 (listed CW from lower left) are all shown at the same scale with the scale bar in M338 equivalent to 900 µm, and photos of experiments C237, D184, D180, D194, D165, and D164 (CW from lower left) are shown with the scale bar in D164 equivalent to 100 µm.
Figure 3. Mg# of solid phases in experiments above 840°C at 3.2 GPa. The increasing Mg# of all solid phases is interpreted to be the result of solid-melt partitioning. Changes in Mg#s between 840 and 880°C, and 940 and 980°C likely result from the breakdown of chlorite and clinohumite, respectively. Two sigma errors for the Mg# based on the electron microprobe compositions and number of analyses are shown for all phases possible.

Figure 4. Minor element concentrations in garnet, clinopyroxene and orthopyroxene from experiments at 840-1200°C at 3.2 GPa. Trends in oxide concentrations with increasing temperature are consistent with partitioning between solid and melt, rather than solid and a vapor phase. The modeled behavior of an element during batch melting is illustrated by the bold lines and were calculated using solid-melt partition coefficients determined from the hydrous melting experiments of Gaetani et al. (1998; 2003). The larger of the standard deviation or the error determined from the EMP counting statistics is plotted as the error. Oxide concentrations and their standard deviations determined from EMP analyses are also given in Tables 2 and 3.

Figure 5. Phase diagram for chlorite peridotite. The illustrated experiments were conducted using a Hart and Zindler (1986) primitive mantle composition synthesized from natural peridotite minerals + chlorite for a starting material with a total of 0.7 wt.% H₂O (H & Z natural; Table 1).
Figure 6. Scanning electron backscatter images of experiments D206 and D214. Both images show the two quench phases with distinct chemical compositions (Table 3) observed in our experiments above the H$_2$O-saturated solidus. We interpret the spherical features to be quenched of the vapor phase and the wisps to be quench of silicate melt.

Figure 7. Comparison of Subduction-Relevant Solidi, Dehydration Reactions and Global Slab P-T paths. Slab P-T paths for global subduction zones are plotted from the numerical models of Syracuse et al. (2010) where full coupling between the slab and mantle occurs at 80 km depth. The majority of slab surface P-T paths are illustrated as a gray field and cross the serpentine dehydration curve from Ulmer and Trommsdorf (1995) (black curve labeled “serp-out”), the chlorite dehydration curve (black curve labeled “chl-out”), and the H$_2$O-saturated peridotite solidus (bold black curve labeled “peridotite solidus”) as determined from our experiments within the range of temperatures and pressures of the slab surfaces directly below global arcs, as shown by the overlapping gray boxes (Syracuse et al., 2010). The colored dashed lines illustrate the P-T paths for a couple of the hottest and coolest subducting slabs. The approximate location of the solidus for subducted oceanic basalt (Lambert and Wyllie, 1970; Hill and Boettcher, 1970a; Liu et al., 1996; Vielzeuf and Schmidt, 2001) is shown by the black dashed line labeled “OC,” and the approximate location of the subducted sediment solidus (Nichols et al., 1994; Poli and Schmidt, 2002; Hermann and Spandler, 2008) is shown by the dotted black curve labeled “sed.” The P-T paths for warm subduction zones cross the sediment and oceanic basalt solidi in the coolest and shallowest region of the subarc mantle.
Figure 8. Diagrams of the locations of H$_2$O release and the initiation of melting in the mantle wedge below arc volcanoes. a) and c) cartoons of the mantle wedge illustrating chlorite and serpentine stability and the locations where the breakdown of these hydrous phases produces free H$_2$O available to trigger H$_2$O-saturated melting in the mantle wedge. The locations of the isotherms are taken from the geodynamic models of Grove et al. (2009) using a dip of 30° and a convergence rate of 40 mm yr$^{-1}$. b) and d) hydrous undepleted peridotite phase diagrams illustrating the locations where the breakdown of these hydrous phases produces free H$_2$O available that triggers H$_2$O-saturated melting in the mantle wedge. At warm subduction zones such as Cascadia, the quantity of H$_2$O fluxed from the slab at subarc depths is greater than that which can be bound in a ~1 km thick layer of chloritized peridotite at the base of the mantle wedge, and will therefore trigger melting at the temperature of the H$_2$O-saturated solidus. The dehydration of chlorite in the mantle wedge will perpetuate ongoing H$_2$O-saturated melting. In cool subduction zones such as the Northern Marianas, the lesser quantity of H$_2$O fluxed from the slab at subarc depths can be bound in a ~1.5 km thick layer of chloritized peridotite and melting therefore will not occur until the P-T conditions of chlorite dehydration.

Table 1. Composition of experimental starting materials.

Table 2. Experimental run conditions and phases present. Mass balance calculations are included for selected experiments.
Table 3. Normalized electron microprobe analyses of run products from H$_2$O-saturated experiments. "Analyses" indicates number of individual electron microprobe analyses included in average.

Table 4. Compilation of H$_2$O-rich peridotite melting experiments.
Figure 1.
Figure 2.
Figure 3.
Figure 4.
Figure 5.
Figure 6
Figure 7.
Warm Subduction Zones (e.g., Cascadia)

Cool Subduction Zones (e.g., N. Marianas)

Figure 8

Key

- Free H$_2$O released from hydrous mineral breakdown in the slab
- Free H$_2$O released from hydrous mineral breakdown in the mantle wedge
- Initiation of Melting
Table 1. Composition of starting materials.

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Proportions in H&Znatural

*Pali-Aike Mineral Analyses are from Stern et al., 1989*
Table 2. Experimental run conditions and phases present. Mass balance calculations are included for selected experiments.

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*Intentionally short run times to test equilibration.
### Table 3. Normalized electron microprobe analyses of run products from H2O-saturated experiments. Analyses indicate number of individual electron microprobe analyses included in average.

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67
Pyroxenes in these samples are very small and we were not able to analyze any points with confidence around and below the wisp analyzed.

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*a*Quenched melt is small wisps of intergranular material that are difficult to analyze and composition likely reflects the composition of quench crystals around and below the wisp analyzed.

*b*Pyroxenes in these samples are very small and we were not able to analyze any points with confidence.

*c*Pyroxenes are small and difficult to analyze and therefore composition listed represents single points of pyroxene with lowest Al and Ti contents.
Chapter 2:
Compositions of Low Extent Melts of H$_2$O-Saturated Peridotite
Abstract

We present a study of the nature and composition of quench products in a subset of the super-solidus H₂O-saturated peridotite melting experiments of Till et al. (2011)¹. Electron microprobe analyses of the quench products substantiate the presence of both a low alkali tholeiitic melt and a high-silica vapor-phase quench. When multiple filtering techniques are applied to the tholeiitic quench analyses, we are able to infer equilibrium melt compositions at 3.2-4 GPa. These melt compositions suggest a switch in melting reaction at clinopyroxene- and garnet-out (~1125°C, 3.2 GPa), from oliv + cpx+ garnet → opx + melt, to oliv+ opx → melt. The isobaric melt production rates in these experiments are extremely low (dF/dT_p = 0.05-0.13%/°C) and may represent a minimum in peridotite melt production rates, which occur when a melt is H₂O-saturated. These melting reactions and melt production rates compare favorably with previous studies of hydrous and anhydrous peridotite melting behavior and are compatible with the solidus temperature of ~800°C at 3.2 GPa interpreted by Till et al. (2011). Further work is required to refine these estimates of the equilibrium melt compositions and determine the cause of anomalously high K_D^Mg-Fe^{Oliv-Liq} values present in a number of the experiments. These preliminary results support previous interpretations that low extent melts of H₂O-saturated peridotite at 3-4 GPa are the forefathers of magmas erupted at volcanic arcs.

1. Introduction

Scientists have long sought to understand the link between convergent margins and the formation of hydrous andesitic magmas erupted at volcanic arcs (e.g., Gill, 1981). Geochemical and experimental studies of the erupted lavas, quenched magmatic inclusions, and mineral-hosted melt inclusions found at volcanic arcs have been successful in reconstructing the composition of parental arc magmas (e.g., Yogodzinski et al., 1995; Jicha et al., 2009). Parental mantle-derived magmas are found to be primitive magnesian andesites and to a lesser extent basaltic andesites that contain ~52-58 wt.% SiO₂, 9-11 wt.% MgO, a minimum of ~4.5-6 wt.% H₂O and trace element abundances inherited from slab-derived fluids (e.g., Grove et al., 2002; 2005). These melts tend to experience fractional crystallization in the shallow crust to produce the more evolved andesites and dacites that often form the bulk of arc volcanoes (Grove et al., 2005; Krawczynski and Grove, 2011a). Although the crustal history of parental arc magmas can be well constrained, considerable uncertainty remains about the processes that lead their generation in the mantle prior to their emplacement in the base of the crust.

It is generally accepted that partial melting of peridotite in subduction zones is initiated by an influx of volatiles from the subducted ocean lithosphere (e.g., Gill, 1981; Tatsumi, 1986; Davies and Stevenson, 1992). However the specifics of the mantle melting process, as well as the composition of the hydrous partial melts generated, have been the subject of experimental investigations and scientific debate for more than 30 years. Much of our understanding about these melts comes from the early experimental work of Kushiro and co-workers, which focuses on using simplified compositional systems, such as Mg₂SiO₄-SiO₂-H₂O (Kushiro et al., 1968; Kushiro 1969, 1972). This work suggests hydrous peridotite melting produces SiO₂-rich melt; consistent with the hypothesis that andesite magmas erupted at arc volcanoes are the descendants of partial melts of hydrous mantle peridotite (Poldervaart, 1955; Kushiro, 1972). Later experimental studies on natural peridotite compositions indeed produce hydrous glasses high in SiO₂ and low in FeO and MgO (Kushiro et al., 1972; Mysen et al., 1974; Nehru and Wyllie, 1975) but raised questions regarding the modification of melt composition via growth of amphibole and pyroxene upon quench (e.g., Green, 1973; 1976). Careful experimental studies by Gaetani and Grove (1998, 2003) at 1100-1345°C and 1.2-2 GPa
demonstrate that the addition of 3-6 wt.% H2O in spinel lherzolite melts increases SiO2 by ~1 wt.% and decreases FeO + MgO by ~2 wt.% relative to analogous anhydrous peridotite melts because of the lower temperature of melting in the presence of H2O. Their results also support the interpretation that arc parental magmas with high pre-eruptive H2O contents form when fluid-saturated peridotite partial melts percolate upward in the mantle wedge and maintain equilibrium with the hotter overlying peridotite (i.e., reactive porous flow).

Within the last five years, new petrologic evidence has arisen that resolves the long standing discrepancy in the solidus temperature of the H2O-saturated mantle (BVSP, 1981) and demonstrates that mantle peridotite melts at extremely low temperatures when in the presence of excess water. New long-duration high-pressure mantle melting experiments by Grove et al. (2006) and Till et al. (2011) indicate that the H2O-saturated peridotite solidus is located between 800° and 820°C at 3-6 GPa. Likewise, studies of a peridotite massif exposed in the Japanese Sanbagawa belt (Hattori et al., 2009; Till et al., 2009) find evidence for hydrous melting processes at ~800°C and ~3 GPa in a paleo-mantle wedge. Consequently, analysis of the composition of low extent melts produced in the experiments of Till et al. (2011) can shed new light on the composition of hydrous partial melts and the specifics of hydrous melting behavior in the mantle wedge at subduction zones.

Quenching experimental melts with a high H2O-content has long been recognized to be problematic, as the exsolution of H2O during the rapid cooling and depressurization of the experiment causes the melt to quench as a froth of bubble walls and/or as “quench” mineral phases, rather than as a glass. Here we present a study of the nature and composition of quench products in a subset of the Till et al. (2011) experiments where they were of sufficient size and quantity to be chemically analyzed on an electron microprobe. We find evidence for multiple quench phases and discuss methods for filtering the chemical analyses of the quench products. The melting reactions and rates of magma production implied by the quench compositions are discussed. We also explore additional techniques to determine an equilibrium melt composition.
2. Experimental and Analytical Methods

Till et al. (2011) present a comprehensive study of piston cylinder and multi-anvil experiments conducted between 3.2 - 6 GPa and 740-1200°C on the primitive upper mantle composition of Hart and Zindler (1986) in the presence of excess H$_2$O (Table 1). The experimental and analytical methods for these experiments are discussed in detail in Till et al. (2011). The starting material for these experiments was created from a mixture of high-purity synthetic oxides or metasilicates. Brucite, Mg(OH)$_2$ was added to the synthetic oxide mixture instead of MgO, which provided the starting material with ~14.5 wt.% H$_2$O.

The experiments considered here were conducted at the MIT Experimental Petrology Laboratory in a 0.5" solid medium piston cylinder apparatus (Boyd and England, 1960) using the hot piston-in technique (Johannes et al., 1971). The experiments were conducted in Au capsules to minimize the iron loss and prevent hydrogen loss from the starting material. The pressure medium was BaCO$_3$ and pressure was calibrated using the anorthite + gehlenite + corundum = Ca-tschermak pyroxene reaction and the spinel to garnet transition in the CMAS system (Hays, 1966; Longhi et al., 2005). Experimental pressures are thought to be accurate to ± 0.05 GPa. Temperature was monitored and controlled using W$_9$Re$_3$-W$_7$Re$_{25}$ thermocouples with no correction applied for the effect of pressure on the thermocouple EMF. The run temperatures for the experiments are accurate to ± 10°C (Médard et al., 2008). To raise the piston cylinder to pressure, experiments were pressurized to 1 GPa at room temperatures then the temperature was raised at 100°C/min to 865°C. The experiment was held at these conditions for 6 minutes before raising the temperature at 50°C/min to the final temperature and pressure. The sample was held at isothermal and isobaric conditions for the duration of the experiment and then quenched by turning off the power. Following each experiment a hole was drilled in the capsule with a small diameter hand drill and the presence or absence of liquid water was noted. The capsule was then sliced in half, dried, vacuum-impregnated with epoxy and polished for electron microprobe analysis.

The major element concentrations of all experimental products were analyzed by wavelength dispersive spectrometry on the 4- or 5-spectrometer JEOL 733 and 8200 electron microprobes (EMPs) at the Massachusetts Institute of Technology. All analyses
of minerals were conducted with a 15 kV accelerating voltage, a 10 nA beam current and a 1 μm spot size. Quenched melt analyses were conducted with a range of beam currents (3 to 10 nA), and spot sizes (1-30 μm) and measurement times (5-40 sec), depending on the size and fragility of the quench product. Online data reduction utilized the CITZAF correction package (Armstrong, 1995) and the atomic number correction, the absorption coefficients, and the fluorescence correction available in CITZAF.

3. Determination of Quench Compositions

Experimental determinations of the H₂O solubility in silicate melt indicate at 30 kbar, 27.0 ± 1.0 wt.% H₂O is required to saturate forsterite melt and 21.5 ± 1.0 wt.% H₂O to saturate diopside melt (Hodges, 1974). This data suggests the olivine(±diopside)-saturated melts in the Till et al. (2011) experiments are H₂O-saturated until >55% melted and will suffer from extensive H₂O-exsolution during depressurization for the entire range of melting extents relevant to the lower half of the mantle wedge (<50%). In near solidus experiments, the melt wisps tend to coat the crystalline mineral phases making them especially difficult to chemically analyze on the EMP. However, in the higher temperature experiments (>250-300°C above solidus) of Till et al. (2011), there is a sufficient fraction of melt such that it segregates toward the hot end of the capsule making EMP analyses of the quench products possible (Figure 1). Therefore we focus on this subset of the Till et al. (2011) experiments conducted between 1100°C and 1200°C at 3.2 and 4 GPa. The compositions of the quench products in these experiments are illustrated in Figure 2.

In experiments between 1100-1200°C, EMP major element totals for the quench vary between 48-100 wt.%. Totals of 65-100 wt.% may reflect the volatile content of the quench products. We dismiss analyses with totals <65 wt.% from consideration here, as they are likely poor analyses resulting from the high aspect ratio of the individual melt wisps and the size of the EMP beam (1-30 μm) or the EMP beam puncturing a thin layer of melt and penetrating a crystalline mineral phase. The quench with totals >65 wt.% are normalized on an anhydrous basis and the resulting average quench composition for each experiment is presented as “Filter 1 Results” in Table 1. These quench compositions exhibit a significant amount of intra-experiment variability, as well as inter-experiment
variability (Figure 2, Table 1). Several compositional filters are applied to these data in an attempt to hone in on the compositions of the melts in equilibrium with the crystalline mineral phases.

The exchange coefficient for iron and magnesium between olivine and silicate liquid ($K_{D}^{Mg-Fe}_{Oliv-Liq}$) is widely used to assess equilibrium between phases in experimental studies, as well as for natural minerals and melts. $K_{D}^{Mg-Fe}_{Oliv-Liq}$ is commonly assumed to have a constant value of 0.3 over a wide range of temperatures, pressures and compositions, as originally proposed by Roder and Emslie (1970). The $K_{D}^{Mg-Fe}_{Oliv-Liq}$ for quench analyses with totals > 65 wt.% varies between 0.10-0.80 (Figure 3). Published values for $K_{D}^{Mg-Fe}_{Oliv-Liq}$ in experiments are between ~0.17-0.45 (Toplis, 2005) and the extent to which $K_{D}^{Mg-Fe}_{Oliv-Liq}$ is actually constant has been a subject of debate (e.g., Mysen 1975; Toplis, 2005). Pressure, temperature, olivine composition, melt composition, and water content are potential sources of variation in $K_{D}^{Mg-Fe}_{Oliv-Liq}$ (e.g., Ulmer, 1989; Kushiro and Walter, 1998; Toplis, 2005). Toplis (2005) illustrated that the effects of increasing temperature and changing olivine composition tend to counteract one another as melting proceeds. The effect of pressure on $K_{D}^{Mg-Fe}_{Oliv-Liq}$ is fairly minimal (0.01/GPa), which equates to an acceptable range of 0.33-0.34 in the experiments considered here. Melt composition is a likely source of variation in $K_{D}^{Mg-Fe}_{Oliv-Liq}$ in the Till et al. (2011) experiments. Melt composition can account for variations in $K_{D}^{Mg-Fe}_{Oliv-Liq}$ between ~0.2-0.42 (Toplis, 2005). As such, we use these as maximum and minimum values for $K_{D}^{Mg-Fe}_{Oliv-Liq}$ to filter the analyses of the quench with totals > 65 wt.% since the correct value for a given experiment is unknown. The remaining quench compositions are normalized to 100%, averaged for each experiment, and presented as “Filter 2 Results” in Table 1. Experiments by Gaetani and Grove (1998) and Ulmer (1989) suggest that H2O likely increases the value of the $K_{D}^{Mg-Fe}_{Oliv-Liq}$ relative to anhydrous melting experiments (Toplis, 2005). The H2O-content of the experiments examined here is constant and would therefore affect all of the experiments uniformly. But the magnitude of this H2O-effect is uncertain and we choose not to interpret quench analyses with $K_{D}^{Mg-Fe}_{Oliv-Liq}$ > 0.42 as equilibrium melts.

In addition, the composition of the quench products was compared to those of the crystalline mineral phases in a given experiment (Figure 2). Humayun et al. (2010)
demonstrate that laser ablation ICP-MS (LA-ICP-MS) analyses can yield major and minor element compositions of similar or better accuracy and precision than EMP analyses of USGS glass reference materials. They also collect transects of major element LA-ICP-MS analyses across large pools of the quench products in anhydrous partial melting experiments of KLB-1 peridotite at 3 GPa and 1450°C. The LA-ICP-MS results illustrate that analyses dominated by crystalline minerals form excellent correlations in MgO vs. SiO₂ plots, whereas analyses dominated by quench do not exhibit simple mixing lines between the compositions of olivine and pyroxene, as determined by EMP analysis. When our EMP analyses of the quench products are compared on MgO vs. SiO₂ plots, they too reveal that a number of the analyses appear to plot on mixing lines between a crystalline mineral and what is presumably the equilibrium melt composition. We adopt the technique of Humayun et al. (2011) and remove EMP analyses we interpret as analysis sampling crystalline phases from the dataset and the resulting average quench compositions are presented under “Filter 3 Results” in Table 1.

There is a significant variability in the average quench composition produced by the three filters, especially for the 1100 and 1200°C experiments at 3.2 GPa. Overall, the average quench compositions can be described as silica-undersaturated low-alkali olivine tholeiites, as evidenced by their positions on the oxygen-normalized pseudo-ternary projection scheme of Tormey et al. (1987) (Figure 4). Because of the variability in the average quench compositions, an additional technique for determining the quenched melt compositions was explored. For this technique, the quench products were carefully removed from half of experiment D194 (1200°C, 3.2 GPa) and the remaining half capsule was mounted in epoxy and analyzed on the microprobe. The removed quench was melted in a platinum foil packet in a 1 atm furnace at ~1500°C for 3 hours and quenched to an anhydrous glass. The composition of this anhydrous glass is very similar to the hydrous glass for experiment D194 (Table 1), with slightly higher Al₂O₃, FeO, Na₂O, K₂O and lower TiO₂, Cr₂O₃ and CaO when both are normalized on an anhydrous basis. This remelted quench composition also plots in a location on the Oliv-Cpx-Qtz ternary consistent with the formation of a melt similar to the bulk composition at increased extents of melting. Because of their internally consistent variations in composition with increasing temperature, the average melt compositions produced by
Filter 3 at <1200°C and the remelted quench at 1200°C are our best estimates of the equilibrium melt compositions (referred to hereafter as the "preferred melts").

In addition to the tholeiitic quench, there is evidence for a vapor phase quench in the experiments. Spheres found in the voids between quenched melt wisps in experiments at 1100°C (Figure 1) produce compositions with >70 wt.% SiO₂ at <2 wt.% MgO (Table 1). There is chemical evidence for this high-silica quench phase in experiments at 1100, 1125°C and 1175°C at 3.2 GPa, where several quench analyses plot on mixing lines between the high-silica composition and what we interpret to be the equilibrium melt composition. Instead of the high-silica spheres, there is a carbonate quench phase present in the voids spaces between the quenched melt at 4 GPa (Figure 1). The 4 GPa experiments were conducted more than a year and half after the preparation of the starting material and the carbonate quench is likely the result of progressive carbonation of the synthetic oxide mixture over time by atmospheric CO₂, as observed in other experiments where the starting material is not kept under vacuum (M. Schmidt, personal communication, 2010). This small amount of CO₂ in the starting material (<100 ppm) is not sufficient to change the solidus temperature of peridotite by more than 20°C (Dasgupta and Hirschmann, 2007). We interpret the spheres and carbonate as quench products of a vapor phase that was either present with the melt phase at pressure or exsolved during cooling, similar to the interpretation of White and Wyllie (1992).

4. Discussion

4.1 Phase Proportions & Inferred Melting Reactions

The proportions of the crystalline phases and melt in each experiment were calculated by fitting the bulk composition of the starting material to a linear combination of the analyzed phase compositions. The phase proportions produced using the three different quench filters are illustrated in Figure 5, in addition to the proportions of the equilibrium minerals calculated without a melt composition for experiments at <1100°C and 3.2 GPa from Till et al. (2011). Both the ternary positions and the phase proportions produced with our preferred melt compositions (Filter 3 compositions at 1100-1175°C and the 1200°C remelted quench composition) imply a melting reaction prior the exhaustion of
clinopyroxene and garnet (≤1100°C) where olivine, clinopyroxene and garnet are consumed to produce melt and orthopyroxene. This is illustrated in Figure 6, where the tie line between the 1100°C orthopyroxene composition and the 1100°C average quenched melt composition pierce the plane defined by coexisting clinopyroxene, olivine and garnet (the garnet in these experiments plots above the Plag apex of the Olv-Cpx-Qtz pseudo-quaternary). The proportions of the crystalline phases in all the experiments at <1100°C have also been renormalized in Figure 6 to include the predicted proportion of melt. At temperatures above 1100°C, only a harzburgitic residue remains and the location of the preferred melts on the pseudoternary suggest olivine and orthopyroxene are consumed in a eutectic melting reaction.

4.2 Comparison to Previous Experimental Studies

The preferred melts are lower in FeO and MgO and higher in SiO₂ and CaO than similar extent anhydrous pyrolite melts at 3 GPa (Walter, 1998) (Figure 7; Table 1). The increased SiO₂/FeO+MgO ratio in the hydrous preferred melts is a result of the lower melting temperature relative to anhydrous peridotite as demonstrated by Gaetani and Grove (1998; 2003) and Parman and Grove (2004). The melting reactions and the switch in melting reactions at clinopyroxene-out observed here are identical those observed in anhydrous melting experiments at 3 GPa, as well as anhydrous melting experiments at lower pressure (Kinzler and Grove, 1997; Longhi, 2002). This similarity in melting behavior is encouraging, as it lends credibility to the equilibrium melt compositions and melting reactions we interpret.

The fact that our melt compositions are silica-undersaturated tholeiites is significant. Early work on low extent H₂O-saturated peridotite melts by Kushiro (1968, 1969, 1972) suggests that melts coexisting with olivine, orthopyroxene, clinopyroxene (± amphibole) at 0.75 GPa, 1000°C and 1.95 GPa, 1100°C are silica-saturated quartz tholeiites (~60 wt.% SiO₂, Mg# 0.79-0.83) with compositions similar to calc-alkaline andesites (Figure 7). The composition of H₂O-saturated peridotite melts in equilibrium with olivine, orthopyroxene, clinopyroxene and spinel at 1.6 GPa, 940°C in the Till et al. (2011) study and at 1.2 GPa, 1020°C by Grove et al. (2006) are both also silica-saturated quartz tholeiites (58-59 wt.% SiO₂, Mg# 0.75-0.91). These observations suggest that low-extent
melts at lower pressures are silica-saturated but become increasingly silica-undersaturated at higher pressures.

4.3 Melt Productivity

The melting extents predicted by our preferred melt compositions suggest the rate of isobaric melt production \((dF/dT)_p\) at 1100-1200°C and 3.2 GPa is \(\sim 0.13\% \text{ melt}/\text{°C}\) (Figure 8). Gaetani and Grove (1998) also calculate \((dF/dT)_p = \sim 0.05\%/\text{°C}\) at 2% partial melt for a peridotite with 0.15 wt.% H₂O at 1.5 GPa and find the productivity increases continuously to 0.80%/°C at 20% partial melt. Studies of anhydrous peridotite melting determine an average melt productivity of \(\sim 0.30\%/\text{°C}\) for \(\sim\)20-40% partial melt at both 1 and 3 GPa (Baker and Stolper, 1994; Walter, 1998; Falloon and Danyushevsky, 2000) and values of 0.8-1.27%/°C for fertile peridotite at \(\sim\)2 GPa (Kinzler, 1997). These observations indicate the concentration of dissolved H₂O in the melt exerts an important control on melt productivity. The extremely low melt production rate determined here for the H₂O-saturated 20-30% melts of Till et al. (2011) and the H₂O-saturated 2% partial melts of Gaetani and Grove (1998), suggest that a melt production rate of \(\sim 0.05-0.13\%/\text{°C}\) may be a minimum bound on peridotite melting rates common to H₂O-saturated melts. These low rates of melt production suggest a significantly longer melting column is required to produce the same amount of partial melt generated in anhydrous partial melting environments. When extrapolated from the 1100-1200°C experiments of Till et al. (2011) to 0% melt, the H₂O-saturated melt productivity predicts a H₂O-saturated solidus temperature of \(\sim 800°C\), consistent with that documented by Till et al. (2011) (Figures 6 & 8).

4.4 Comparison of the Experimental Melts to Arc Lavas

H₂O-saturated partial melts formed at the base of the mantle wedge buoyantly ascend into the hotter overlying mantle, where these superheated melt will re-equlibrate by dissolving the surrounding lower pressure peridotite minerals (Grove et al., 2002). As these melts subsequently ascend through the top half of the wedge and lower crust, they will experience fractional crystallization and to varying extents, assimilation and magma mixing. Krawczynski and Grove (2011) conducted an experimental study of mantle-
derived parental arc magma compositions from Mt. Shasta in the presence of H$_2$O that demonstrates near liquidus crystallization of amphibole and clinopyroxene and olivine at lower crustal pressures (500-800 MPa) can produce calc-alkaline liquids with compositions typical of arc lavas.

Our preferred melt compositions are compared to one of the parental arc magmas studied by Krawczynski and Grove (2011) in Figure 9, as well as quenched mafic inclusions and calc-alkaline basalts-andesitic lavas from Cascades arc volcanoes. The preferred melts compositions are only slightly lower in silica and alkali contents than the parental arc magma and the most mafic of the arc lavas. If the preferred melt compositions are approximations of the melts formed at the base of the mantle wedge, their ascent to the shallow crust will likely cause them to evolve to the more SiO$_2$-rich compositions typical of arc magmas. Thus the melts of H$_2$O-saturated peridotite at ~3 GPa generated in the experiments of Till et al. (2011) may be the forefathers of arc lavas.

4.5 Where Do We Go From Here?

Although we are able to determine potential equilibrium melt compositions from the quench products in the experiments of Till et al. (2011) and make cogent inferences about H$_2$O-saturated peridotite melting behavior, there is still uncertainty regarding the exact composition of the equilibrium melt. The three different methods for filtering the quench compositions produce average compositions that vary by as much as 6 wt.% SiO$_2$ and 11 wt.% MgO for the experiment at 1100°C, and 1 wt.% FeO and 4 wt.% CaO for the experiment at 1175°C, for example. As a result, several of the average quench compositions plot at locations on the pseudoternary diagrams that are untenable, as well as inconsistent with the other quench compositions (e.g., the Filter 1 1100°C average quench). Na$_2$O is highly volatile and notoriously difficult to analyze on the EMP. Our mass balance calculations suggest that up to 75% of the bulk Na$_2$O was lost in the EMP analyses of the experimental phases, the majority likely from the quench analyses.

One explanation for these discrepancies could be that the analyzed quench in these experiments is not an equilibrium melt, but instead represents a complicated super-critical fluid with a high dissolved silicate content. Till et al. (2011) demonstrate that the changes in the composition of the equilibrium minerals in these experiments with
temperature could only be the result of solid-melt partitioning and not solid-fluid partitioning. The presence of both the high-silica quench and the more mafic quench wisps further suggests these experiments are below the second critical end point, which is poorly constrained in the peridotite-H$_2$O system.

Therefore to determine the equilibrium melt composition of H$_2$O-saturated peridotite at pressures $>$3 GPa with greater precision and accuracy, further work is required. Experimentalists have developed and struggled with a number of methods to segregate and analyze low extent ($<$10%) hydrous melts, these methods include sandwich experiments, melt traps, and thermal segregation (Stopler, 1980; Takahashi and Kushiro, 1983; Hirose and Kushiro, 1994; Baker and Stolper, 1994; Walter, 1998; Laporte et al., 2004). The simplest of these, the sandwich experiment, consists of a layer of the assumed melt composition (the meat) to be surrounded by two thicker layers of the bulk rock composition (the bread). During the experiment, the melt will approach equilibrium with the minerals in the bread through diffusion and dissolution and may quench to a homogenous material that can be chemically analyzed at the end of the experiment. Several problems can arise using this method, as the length scales of the capsule can be too great and the dissolution rates too slow to reach chemical equilibration, a layer of reaction minerals can form between the two materials, impeding chemical exchange and/or the addition of the melt can shift the bulk composition of the experiment such that the resulting melt composition is not in equilibrium with the original bulk composition. An iterative version of the sandwich experiment, where the melt composition from each experiment is used as the initial meat of the next experiment can relieve the first two sets of pitfalls. Hirschmann and Dasgupta (2007) propose a new modified iterative sandwich experiment technique that approaches equilibrium in a small number of experiments and successfully produces the composition of finite near-solidus melt fractions. This technique may be an important next step in determining the equilibrium melt compositions in <1100°C Till et al. (2011) experiments. Using a diamond wire saw to thinly slice the experimental capsules could create additional surfaces to expose additional analyzable quench pockets in some experiments, as well. The use of LA-ICP-MS to analyze the major element composition of quench by Humayun et al. (2010) suggests this technique may also be useful in experiments with pooled melt to further
refine estimates of the equilibrium melt composition of H₂O-saturated peridotite at pressures >3 GPa.

5. Concluding Remarks

EMP analyses of the quench products in the H₂O-saturated peridotite melting experiments of Till et al. (2011) are suitable to document the presence of both a low alkali tholeiitic melt and a high-silica vapor-phase quench. The composition of the equilibrium melts can be determined if the Mg-Fe partition coefficients and co-variations in oxide abundances are used to filter the quench compositions. The melting reactions and melt production rates compare favorably with previous studies of hydrous and anhydrous peridotite melting behavior and are compatible with the solidus temperature of ~800°C at 3.2 GPa interpreted by Till et al. (2011). The low extent hydrous melts at 3.2-4 GPa are silica-undersaturated and do not resemble the SiO₂-rich compositions of melts documented at lower pressures. Several additional techniques are recommended to refine the estimates of the equilibrium melt compositions. These preliminary results support previous interpretations that low extent melts of H₂O-saturated peridotite at 3-4 GPa are the forefathers of magmas erupted at volcanic arcs.
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Figure and Table Captions

**Figure 1.** Electron back-scatter images of experiments with analyzed quench. Experiment D214 at 1100°C, 3.2 GPa exhibits a pool of the segregated melt at the hotter end of the experimental capsule, where both the quench wisps and high-silica quench spheres are visible. Experiment D230 at 4 GPa, exhibits similar features but with quench carbonate in place of the high-silica quench.

**Figure 2.** SiO$_2$ vs. MgO for quench analyses with EMP totals >65 wt.%. All compositions were normalized as anhydrous prior to plotting and averaged for to produce the average Filter 1 compositions in Table 1. Only the analyses with 0.20 < $K_D^{Mg-Fe}_{Oliv-Liq}$ < 0.42 (red squares) are averaged to produce the Filter 2 quench compositions. In Filter 3 (green circles), the quench analyses that we interpret sampled multiple phases (i.e., quench + equilibrium mineral or high-silica quench) are removed.

**Figure 3.** $K_D^{Mg-Fe}_{Oliv-Liq}$ vs. MgO for quench analyses with EMP totals >65 wt.%. Symbols are the same as in Figure 2.

**Figure 4.** Average quench compositions for the experiments at 3.2 GPa determined using the three filters plotted on the oxygen-normalized Oliv-Cpx-Qtz pseudoternary (projected from garnet) of Tormey et al. (1987).

**Figure 5.** Proportions of melt and the crystalline phases in the 3.2 GPa experiments calculated with each of the average quench compositions determined from the three filtering techniques. The proportions of the equilibrium minerals calculated without a melt composition for experiments at <1100°C and 3.2 GPa from Till et al. (2011) are also shown in each plot.
Figure 6. Phase proportions and melting reactions predicted by our preferred melt compositions. The proportions of the crystalline phases in all the experiments at <1100°C have also been renormalized to include the predicted proportion of melt.

Figure 7. Anhydrous peridotite melts of Walter (1998) at 3 GPa (gray) and low-pressure hydrous peridotite melts of Mysen et al. (1974) (green), Grove et al. (2006) (blue) and Till et al. (2011) (orange) plotted on the oxygen-normalized Oliv-Cpx-Qtz pseudoternary (projected from garnet) of Tormey et al. (1987). The bulk composition for all of these experiments resembles that from Till et al. (2011) (black).

Figure 8. Isobaric melt production rates \((dF/dT)_p\) as determined from peridotite melting experiments at ~1 or 3 GPa.

Figure 9. Comparison of volcanic rocks from the Cascades volcanic arc and the average quench compositions from the experiments of Till et al. (2011). The compositions of quenched mafic inclusions in andesite lavas from Mt. Shasta are from Krawczynski and Grove (2011b). Star indicates the composition of a parental arc magma from Mt. Shasta from Grove et al. (2005). Samples are plotted on the IUGS classification for extrusive rocks. The range of melt compositions illustrated are corrected for the loss of alkalis during EMP analyses.

Table 1. Average quench compositions determined from H2O-saturated peridotite melting experiments of Till et al. (2011). EMP quench analyses were filtered using three different methods to determine their average compositions for each experiment. The composition of similar extent anhydrous pyroline melts at 3 GPa from Walter (1998) are shown for comparison.
Figure 1.
Figure 2.

**LEGEND**

- Filter 1: all analyses with totals >65 wt. %
- Filter 2: analyses from Filter 1 with $0.20 < K_{o}Fe-MgO_i < 0.42$
- Filter 3: analyses with >65 wt. % totals and reasonable $K_o$'s, that also do not fall on mixing lines with the equilibrium minerals or the high-silica quench
Figure 3.
Figure 4.

Filter 1: all quench analyses with totals >65 wt.% (normalized first)

Filter 2: Filter 1 analyses with $0.20 < K_{\text{Fe-Mg}}^{\text{Oliv-Liq}} < 0.42$

Filter 3: Filter 2 analyses we interpret do not fall on mixing lines with the equilibrium minerals or high-silica quench
Figure 5.
Figure 6.
Figure 7.
Figure 8.

Legend

Anhydrous Experiments
- Falloon et al., 1999 (1 Gpa)
- Baker & Stolper, 1994 (1 GPa)
- Falloon & Danyushevsky, 2000 (1.5 GPa)
- Walter, 1998 (3 GPa)

H₂O-Saturated Experiments
- Grove et al., 2006 (1.2 GPa)
- This study (3.2 GPa)

H₂O-Undersaturated Experiments
- Gaetani & Grove, 1998 (1.2 GPa)
Figure 9.
### Table 1. Average composition of quench in experiments at 1100-1200°C and 3.2-4 Gpa from Till et al. (2011).

<table>
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<th></th>
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<th>Cr₂O₃</th>
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<td>0.16</td>
<td>13.91</td>
<td>13.31</td>
<td>0.28</td>
<td>0.12</td>
<td>0.76</td>
<td>0.30</td>
<td></td>
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<tr>
<td>3.2 Gpa, 1150°C D200</td>
<td>47.95</td>
<td>0.73</td>
<td>11.50</td>
<td>0.16</td>
<td>7.88</td>
<td>0.19</td>
<td>17.01</td>
<td>14.35</td>
<td>0.21</td>
<td>0.02</td>
<td>0.78</td>
<td>0.35</td>
<td></td>
</tr>
<tr>
<td>3.2 Gpa, 1175°C D209</td>
<td>51.97</td>
<td>0.48</td>
<td>9.80</td>
<td>0.34</td>
<td>7.62</td>
<td>0.20</td>
<td>21.93</td>
<td>7.53</td>
<td>0.09</td>
<td>0.02</td>
<td>0.79</td>
<td>0.38</td>
<td></td>
</tr>
<tr>
<td>3.2 Gpa, 1200°C D194</td>
<td>48.65</td>
<td>0.60</td>
<td>9.75</td>
<td>0.41</td>
<td>8.35</td>
<td>0.16</td>
<td>19.99</td>
<td>11.97</td>
<td>0.12</td>
<td>0.00</td>
<td>0.80</td>
<td>0.34</td>
<td></td>
</tr>
<tr>
<td>3.2 Gpa, 1200°C D194 remelted</td>
<td>48.27</td>
<td>0.43</td>
<td>11.09</td>
<td>0.14</td>
<td>10.38</td>
<td>0.16</td>
<td>20.65</td>
<td>8.37</td>
<td>0.30</td>
<td>0.15</td>
<td>0.78</td>
<td>0.28</td>
<td></td>
</tr>
<tr>
<td>4 Gpa, 1100°C D230</td>
<td>53.26</td>
<td>0.49</td>
<td>12.66</td>
<td>0.07</td>
<td>7.37</td>
<td>0.08</td>
<td>21.65</td>
<td>4.12</td>
<td>0.23</td>
<td>0.07</td>
<td>0.84</td>
<td>0.53</td>
<td></td>
</tr>
</tbody>
</table>

**Filter 1 Results: Quench with totals >65 wt.%**

- 3.2 Gpa, 1100°C D214
- 3.2 Gpa, 1125°C D207
- 3.2 Gpa, 1150°C D200
- 3.2 Gpa, 1175°C D209
- 3.2 Gpa, 1200°C D194
- 3.2 Gpa, 1200°C D194 remelted
- 4 Gpa, 1100°C D230

**Filter 2 Results: Quench filtered by Kd only**

- 3.2 Gpa, 1100°C D214
- 3.2 Gpa, 1125°C D207
- 3.2 Gpa, 1150°C D200
- 3.2 Gpa, 1175°C D209
- 3.2 Gpa, 1200°C D194
- 4 Gpa, 1100°C D230

**Filter 3 Results: Quench filtered by Kd and mixing lines**

- 3.2 Gpa, 1100°C D214
- 3.2 Gpa, 1125°C D207
- 3.2 Gpa, 1150°C D200
- 3.2 Gpa, 1175°C D209
- 3.2 Gpa, 1200°C D194
- 4 Gpa, 1100°C D230

**Filter 4 Results: Quench filtered by Kd and mixing lines**

- 3.2 Gpa, 1100°C D214
- 3.2 Gpa, 1125°C D207
- 3.2 Gpa, 1150°C D200
- 3.2 Gpa, 1175°C D209
- 3.2 Gpa, 1200°C D194
- 4 Gpa, 1100°C D230

**High Silica Quench**

- 3.2 Gpa, 1060°C D206
- 3.2 Gpa, 1100°C D214

**Anydrous Peridotite Melts of Walter (1998)**

- 3 Gpa, 1540°C 24% melt
- 3 Gpa, 1580°C 37% melt
Chapter 3:

Primary basaltic magmas from variably metasomatized mantle: The effects of pressure, temperature, and composition on melting plagioclase and spinel lherzolite
Abstract

Here we develop a new model that simulates melting of variably metasomatized mantle peridotite and explore the effects of variation in mantle pressure, temperature and composition on melt composition. This new model combines the approaches of Kinzler and Grove (1992a,b) and Kinzler (1997) to predict the temperature and major element composition of a broad spectrum of primary basalt types produced under anhydrous conditions at upper mantle pressures. The model can also be used to calculate the temperature and pressure at which primary magmas were produced in the mantle, as well as to model both near-fractional adiabatic decompression and batch melting. Our experimental compilation definitively locates the pressure interval of the plagioclase to spinel transition on the solidus and shows that it is narrow (~0.1 GPa) for melting of natural peridotite compositions. A comparison of our melting models to other mantle thermometers reveals that the choice of fractionated phase assemblage(s) used to correct primitive liquids back to primary magmas has a significant effect on the calculated source temperature (>200°C). Our model is used to examine the petrogenesis of a suite of Holocene basaltic lavas from Diamond Crater in Oregon’s High Lava Plains (HLP) and demonstrate that the primary magmas inferred for Diamond Crater are ~6 wt.% batch melts of a spinel lherzolite mantle. After correction for fractional crystallization, our model predicts these basaltic liquids were last in equilibrium with the mantle at 1.2-1.6 GPa and 1322-1340°C.
1. Introduction

Primary basaltic magmas are the principal instrument for understanding the depth, temperature and style of melt generation in the Earth’s upper mantle. Beneath mid-ocean ridges (MOR’s), near-fractional adiabatic decompression melting is the dominant process of melt generation (Klein and Langmuir, 1987; 1989; Kinzler and Grove, 1992b, Niu and Batiza, 1993). In contrast, petrologic and geophysical constraints show that melting beneath the thick crust of continental margins and interiors occurs under near-batch melting conditions (Bartels et al., 1991; Holbig and Grove, 2008), often close to the crust-mantle boundary. An additional important but little explored factor is the effect of variations in the composition of mantle lherzolite. A number of major element mantle melting models have brought us closer to a rigorous quantitative description of melting focused on the formation of mid ocean ridge basalts (MORB) (e.g., Klein and Langmuir, 1987; McKenzie and Bickle, 1988; Niu and Batiza, 1991; Kinzler and Grove, 1992a, b; Langmuir et al., 1992; Kinzler, 1997). However, a significant portion of primitive basalts not erupted at MOR’s, including many that form in continental back arc settings and at ocean islands, exhibit a greater range of K$_2$O (up to 2.0 wt.%), and P$_2$O$_5$ (up to 0.50 wt.%) relative to MORB, which is thought to result from melting variably metasomatized peridotite. Here we seek to build on this prior work and examine mantle melting behavior over a wide range of upper mantle compositions and specifically constrain the effects of K$_2$O on mantle melting equilibria. Small extent melts of the upper mantle are saturated with the complete upper mantle phase assemblage of olivine + orthopyroxene + clinopyroxene + plagioclase and/or spinel at genesis, and our models are restricted by incorporating only these multiply-saturated mantle melts in their calibration. We will
compare our estimates of the temperature and pressure of mantle melting for MORB and ocean island basalts (OIB) to those produced by other popular thermometers and barometers. The compilation of experimental melt compositions in equilibrium with an upper mantle phase assemblage also facilitates observations of the location of the plagioclase to spinel lherzolite boundary and we explore how the stability of plagioclase + spinel lherzolite varies with pressure, Mg# (Mg/(Mg+Fe\textsuperscript{T})), Na\textsubscript{2}O, and K\textsubscript{2}O. Our models for melting the upper mantle can also be used to identify at what pressure and temperature a primitive basaltic magma was last in equilibrium with the mantle and the extent to which it crystallized on the way to the surface. We provide a worked example of these calculations by examining a suite of primitive basaltic lavas erupted in the Diamond Crater area of Oregon’s High Lava Plains (HLP).

II. Experimental Data

a. Experimental Data From the Literature

A query was conducted in the Library of Experimental Phase Relations (LEPR) database (Hirschmann et al., 2008) for experiments containing the phases glass, olivine, orthopyroxene, clinopyroxene, and plagioclase or spinel. Starting materials for the experiments from the literature range from synthetic analogs of fertile or depleted mantle to natural MORB and high alumina olivine tholeiite (HAOT) lava samples (Table 1). The experiments returned by our query were divided into two categories: one for the plagioclase-bearing experiments, the other for spinel-bearing experiments (Table 1). Experiments that contained both plagioclase and spinel were included in both categories for model development. In addition to the experiments downloaded from LEPR, we
added to our datasets spinel and plagioclase lherzolite melting experiments from an
electronic supplement to Kinzler (1997), 0.1 MPa experiments from Grove and Juster
(1989), CMAS experiments from Presnall et al. (1979), and CMASN experiments from

Several different criteria were used to evaluate the data. First, experiments saturated
with garnet were removed from the dataset, these were experiment L92 from Kinzler
(1997), experiments DPI-48 and DPI-44 from Draper and Johnston (1992), and
experiments 3402.11, 3403.11, 3503.11 from Walter and Presnall (1994). Second,
experiments conducted in the presence of water were also removed from the datasets and
include experiments B277g, B304, B329, B330, B333, B342, B348, B359 from Gaetani
and Grove (1998), 85-41-7b from Grove et al. (2003) and experiments 71,72 and 76 from
Feig et al. (2006). Third, single experiments that were clear outliers from the surface
defined by the bulk of the data in pressure, temperature and composition space were
removed. These experiments include a 2 kilobar experiment from Takagi et al. (2005)
and four 10 kilobar experiments from Meen (1990).

**b. New Experiments**

New experiments were also undertaken as part of this study (Tables 2, 3, 4) to
determine the phase relations and compositions for a high K$_2$O basaltic liquid in
equilibrium with a both a plagioclase and spinel lherzolite mantle phase assemblage. To
this aim three sets of experiments were conducted. The first set of experiments utilized a
high K$_2$O basaltic lava composition from Jordan Valley Volcanic Field in eastern Oregon
(JC-30B), which between 8-14 kbars and 1250-1310°C saturated with olivine,
clinopyroxene, and plagioclase but failed to crystallize orthopyroxene. A second set of experiments, conducted with a derivative of the original starting material (JC30B+opx), were multiply-saturated with olivine, orthopyroxene, clinopyroxene and plagioclase between 8-10 kbar. The third set of experiments was conducted on a composition calculated to be in equilibrium with a spinel lherzolite mantle composition at 17 kbar using our model equations (SpLM). This starting composition was saturated with olivine, clinopyroxene, and spinel at 17 kbar and 1313°C but again failed to crystallize orthopyroxene between 1310-1353°C. Here we report the results of the second and third set of experiments but use only the second set of experiments in our model that yield insights relevant for our understanding melts saturated with a complete upper mantle phase assemblage.

c. Starting material

A primitive basaltic composition with 48.43 wt% SiO₂, 0.96 wt% K₂O and a Mg# of 0.64 from Coffee Pot Crater in the Jordan Valley Volcanic Field in eastern Oregon was chosen as the first starting composition for our experiments (JC-30B; Table 2). A synthetic oxide mixture of this composition was prepared and homogenized in an automatic agate mortar and pestle for three hours. A pellet of the sample was then prepared using polyvinyl alcohol as a binder and attached to a Pt loop for use in a DelTech 1 atm gas mixing furnace. The sample was held at 1000°C at the quartz-magnetite-fayalite buffer in the furnace to ensure the iron in the starting material was dominantly FeO, following the methods of Grove (1981) and Grove and Juster (1989). The sample was then reground into a powder with the automatic agate mortar and pestle.
A second starting composition was subsequently fabricated using the original starting material and Kragerø orthopyroxene (JC30B+opx; Table 2). 15 wt. % hand-picked, cleaned, and crushed orthopyroxene was added to the original starting material and homogenized for three hours in the automatic agate mortar and pestle. A third starting material with 47.09 wt% SiO₂, 0.62 wt% K₂O and a Mg# of 0.73, which is an estimate of a liquid in equilibrium with a spinel lherzolite at 1.7 GPa calculated with our model (SpLM; Table 2) was prepared following the methods described above for JC-30B. Prior to each experiment, approximately 4 mg of the starting material was packed into a graphite capsule and held at ~125°C for at least 24 hours to assure the experiments were conducted under volatile-free conditions.

d. Experimental methods

Experiments were performed in a 0.5” end-loaded solid-medium piston cylinder apparatus (Boyd and England, 1960) in the MIT Experimental Petrology Laboratory using the setup detailed by Medard et al. (2008). All experiments were conducted in graphite capsules that totaled ~0.15” in length. The capsules were placed in a nonporous Al₂O₃ sleeve, positioned in the center of a graphite furnace using MgO spacers, and encased in a sintered BaCO₃ pressure medium. The temperature was monitored and controlled using W₉₇Re₃-W₇₅Re₂₅ thermocouples positioned ~1.5 mm above capsule. The experimental temperatures reported in Table 2 are corrected for the 18 ± 6°C temperature difference between the thermocouple temperature (colder) and the average capsule temperature (hotter) (Medard et al., 2008). The pressure was calibrated using the breakdown reaction of Ca-tschermak pyroxene to anorthite + gehlenite + corundum and
the spinel to garnet transition in CMAS (Hays, 1966; Longhi et al., 2005). Experimental conditions are reported in Table 3.

e. Analytical methods

Major element concentrations of the experimental products were analyzed with the JEOL 733 or 8200 microprobes at Massachusetts Institute of Technology. All analyses of the individual phases presented in Table 3 were conducted with a 15kV accelerating voltage, a 10 nA beam current and a beam diameter of 1 μm for solid phases and 10 μm for silicate glasses. On-line data reduction used the CITZAF correction package (Armstrong, 1995) and analytical precisions were estimated from replicate analyses of a basalt glass working standard reported in Gaetani and Grove (1998).

g. Composition of Experimentally Produced Melts Saturated With a Mantle Residue Assemblage

The composition of the melts in equilibrium with spinel lherzolite from the previously published and new experiments is illustrated in Figure 1. The lower pressure experiments (i.e., 1-1.5 GPa) cover a wider range of liquid composition space relative to the higher pressure experiments because of the greater diversity of starting compositions used for the lower pressure experiments. At a given pressure, melts coexist with an upper mantle assemblage over a temperature range of ~50-150°C, which narrows toward higher pressures. The dataset is well suited to explore the effects of oxides such as K₂O and TiO₂ on mantle melting as highlighted by the range of these oxides in the experimental liquids relative to CMAS liquids, which plot at an Mg# of unity. The experimental
liquids in these compositionally complex peridotitic systems exhibit an overall covariation in NaK# ((Na₂O+K₂O)/(Na₂O+K₂O+CaO)), wt.% TiO₂, and wt.% Al₂O₃ with Mg#.

The composition of the minerals and melts from our experiments are given in Table 4. Three liquids in equilibrium with a plagioclase lherzolite assemblage have 1-1.3 wt.% K₂O and Mg#'s of 0.48-0.55. The liquids are very similar in their oxide content to other plagioclase lherzolite liquids compiled from the literature (Figure 1).

III. Parameterization of Plagioclase and Spinel Lherzolite Melting Equilibria

Melting of plagioclase and spinel lherzolite is a 5-phase equilibrium involving liquid + olivine + orthopyroxene + clinopyroxene + plagioclase or spinel. The thermodynamic variance of the melting equilibrium can be obtained from the Gibbs phase rule, \( F = C + 2 - \phi \), where \( F \) is the number of degrees of freedom or variance, \( C \) is the number of components and \( \phi \) is the number of phases. In the CMAS system, these melting reactions have one degree of freedom and are univariant (\( F = 1 = c + 2 - \phi \)). Therefore, specifying the pressure of melting fixes the temperature and composition of all the phases. If an additional chemical component is added (e.g., Na₂O: Walter and Presnall, 1994), the thermodynamic variance increases by one and the Gibbs method (Spear, 1993) can be applied to constrain the compositions of all coexisting phases. We choose to extend the univariant CMAS composition space into the natural system in our model by exploring the effects of additional compositional degrees of freedom on the melting equilibria, following Longhi (1991), Grove and Juster (1989), Kinzler and Grove (1992a) and Kinzler (1997). The number of chemical components was chosen to be nine.
(this includes the dominant major and minor oxides required to calculate mineral components: SiO$_2$, TiO$_2$, Al$_2$O$_3$, FeO, MgO, CaO, Na$_2$O, K$_2$O, P$_2$O$_5$) and there are five mantle phases involved (olivine, orthopyroxene, clinopyroxene, plagioclase/spinel, melt), therefore the thermodynamic variance of our system is six. Thus five compositional variables in addition to pressure should ideally be fixed to constrain the variability of upper mantle melts in $P$-$T$-$X$ space. In previous studies, wt.% TiO$_2$, K$_2$O, Na$_2$O and Mg$\#$ have been identified as important compositional variables that systematically influence the composition of multiply-saturated liquids (e.g., Kushiro, 1975; Grove and Juster, 1989; Walter and Presnall, 1994). We chose to express the composition of the spinel and plagioclase lherzolite melts using oxygen-based mineral components, because these vary systematically from their CMAS univariant values in response to changes in compositionally important exchange vectors like Mg-Fe and Ca-Na. We used the same component choice as Kinzler and Grove (1992a): Olivine (Oliv), Clinopyroxene (Cpx), Plagioclase (Plag) and Quartz (Qtz). After extensive testing, we settled on an empirical model with only four compositional variables in addition to pressure. The five independent variables in our model are 1-Mg$\#$, NaK$\#$, wt.% TiO$_2$, wt.% K$_2$O and pressure (in GPa) (Table 5). Principal component analysis of these variables indicates that $>97\%$ of the chemical variance in the dataset is accounted for in two principal components in which K$_2$O and TiO$_2$ are the dominant coefficients.

The results of linear least squares regressions of the experimental dataset using the five independent variables are illustrated in Figure 2. Here we show the difference between the observed value of the mineral components Oliv, Cpx, Plag and Qtz for experimental liquids in equilibrium with spinel lherzolite and the same value calculated
using our equations and the pressures and temperatures of the experiments. Figure 2a-d illustrate the mineral component values returned by the regressions of the full dataset, while Figure 2e-h illustrate the mineral component values returned by regressions of only those experiments analyzed on the microprobe at MIT and experiments done in the simple systems CMAS and CMASN. There is significantly more scatter in the mineral component values calculated from the full dataset equations relative to the MIT + CMAS(N) equations. One explanation for the scatter in Figure 2a-d is likely systematic differences in the pressure calibration and the microprobe standardization used to produce the experimental liquid compositions at different labs. For example, the synthetic peridotite melting experiments of Kinzler (1992a) and Kinzler (1997) were both analyzed using the same microprobe and standards at MIT but conducted at different piston cylinder labs. When the pressure of these experiments are calculated using the spinel lherzolite olivine equation from the MIT + CMAS(N) regression, pressure is systematically underestimated for experiments run at the MIT experimental lab and overestimated for the experiments conducted at Lamont Doherty experimental lab (Figure 3a), pointing toward discrepancies in the pressure calibration between these two labs. Here we report only the regression of experiments in the CMAS(N) system plus those analyzed at MIT (Table 5; Figure 2e-h), which more closely reproduces the known experimental glass composition in each case. The regressions of the full experimental dataset are presented for comparison in Appendix 1.

The expressions in Table 5 are qualitatively similar to the plagioclase and spinel lherzolite melting equilibria expressions of Kinzler and Grove (1992a). All the coefficients have the same sign and relative effect on temperature and the composition of
the liquid expressed in mineral components with the exception of the 1-Mg# coefficient in the Cpx expression, which is 0 in the Kinzler and Grove (1992a) expression, and the TiO₂ coefficient in the Plag expression, which is the most poorly resolved coefficient in all expressions. Our calculated 1 atm solidus temperature for the metastable spinel lherzolite assemblage is ~40°C higher than that calculated by Kinzler and Grove (1992) for the regression of the MIT + CMAS(N) experiments (Table 5) and ~30°C higher for the full dataset regression (Appendix 1), while our 1 atm plagioclase lherzolite solidus temperatures are very similar for both regressions (1220°C, 1216°C) and ~20°C lower than Kinzler and Grove’s (1992a). In our expressions, 1-Mg#, NaK# and K₂O lower the solidus temperature for both plagioclase and spinel lherzolite saturated liquids, and all other components serve to increase the solidus temperature. Pressure exerts a positive control on temperature and the Oliv and Plag mineral components for a liquid in equilibrium plagioclase lherzolite, as well as a positive control on temperature and the Oliv and Cpx mineral components for a liquid in equilibrium with a spinel lherzolite.

Principal component analysis of the mineral component variance reveals that ~99% of the variance resides in two principal components, one in which Plag has the dominant coefficient, the other in which Oliv has the dominant coefficient. Of the four potential barometers, both the Plag and the Qtz equations are the most sensitive to small variations in analytical error, which manifests itself in the NaK# coefficient. Because of the coordination of sodium to three silica ions in feldspar, small changes in the sodium content of the melt have large effects on the calculated Plag and Qtz components. This is evident in the larger average errors returned by the Plag and Qtz equations relative to the Oliv equation. In spite of the lower $r^2$ for the spinel lherzolite Oliv equation ($r^2 = 0.63$),
this equation recovers pressure with the lowest average error (0.25 GPa) and thus it is our preferred equation for calculating the pressure of a spinel lherzolite primitive melt. The Cpx component exhibits the poorest fit to the experimental liquid dataset, and has the least variance, suggesting the majority of variability may stem from interlab calibration issues. Therefore a single value for the Cpx component may be an adequate approximation for the range of bulk compositions and pressures considered here.

The plagioclase and spinel lherzolite thermometers (Figure 3b, Table 5) have \( r^2 \) values of 0.97 and 0.94, respectively, and reproduce the experimental temperatures with average errors that are equivalent to the temperature uncertainty of the experiments (~10°C). The plagioclase thermometer is calibrated over the temperature range of ~1100-1300°C and overlaps for ~50°C with the calibrated 1250-1550°C range of the spinel lherzolite thermometer. The overlap in these calibrations reflects the range of temperatures where both spinel and plagioclase are stable in the experimental dataset.

**IV. Plagioclase Lherzolite to Spinel Lherzolite Boundary**

The compilation of experimental plagioclase and spinel lherzolite liquids allows us to revisit the location and primary controls on the plagioclase to spinel transition (PST) in the upper mantle. Presnall et al. (1979) locate the invariant PST in the CMAS system at 0.9 GPa and 1300°C. Additional experiments by Walter and Presnall (1994) in CMAS have both spinel and plagioclase present at 0.93 GPa, with only plagioclase present at 0.7 GPa and only spinel at 1.1 GPa. Because there is a negligible temperature-dependence of the PST (Kushiro and Yoder, 1966), the experiments in the CMASN system can be used
to define the location of the univariant PST in $P$-$NaK#$ space. The equation of the resulting line is,

$$P = 1.56 \ (NaK#) + 0.999$$

where the pressure of the PST is in GPa. This result reinforces earlier observations that the Na$_2$O/CaO content of the liquid, a proxy for the degree of depletion of the source, is positively correlated with the pressure of the PST (Walter and Presnall, 1994; Falloon et al., 2008).

The PST expands from a line in two dimensions into a field where both plagioclase and spinel are stable in three dimensions. This can be visualized in Figure 4 where the plagioclase$\rightarrow$plagioclase+spinel and the plagioclase+spinel$\rightarrow$spinel boundaries are depicted as two bent screens. We also can approximate the plagioclase + spinel$\rightarrow$spinel boundary as a plane for practical use, where the pressure of this boundary in GPa is given by the expression,

$$P = (0.615 \ NaK# + 0.3255)/0.35$$

Although the experimental liquid compositions suggest an apparent positive correlation between pressure and Mg#, the third dimension in our illustration, it is actually an artifact of the Na$_2$O/CaO control on the location of the PST, as evidenced by the covariance in NaK# and Mg# in our dataset of experimental liquids (Figure 1). Therefore Mg# does not appear in this planar approximation of the PST. Experimental liquids in equilibrium with both spinel and plagioclase with the highest NaK#’s and lowest Mg#’s in the dataset suggest the maximum width of the PST field in this composition space is ~0.3 GPa. Therefore low extent mantle melts are in equilibrium with both plagioclase and spinel over a relatively narrow pressure interval. Typical mantle melts with an Mg# equal to
0.75 plot close to the Mg-axis, therefore plagioclase + spinel are stable in these liquids over a pressure range much smaller (~0.1 GPa: 1.1-1.2 GPa) than 0.3 GPa. Metasomatic enrichment of mantle alkali contents will increase the NaK# of the mantle and slightly increase the pressure interval over which plagioclase and spinel are stable.

V. Implementation of the Plagioclase and Spinel Lherzolite Models

Our mantle melting model can be used in two ways. The first is to estimate the depth and temperature of origin of a fractionation-corrected magma composition. The second is to model mantle melting processes given a mantle bulk composition and melting model. The second step can provide useful information on the compositional characteristics of the mantle residue and the processes that can lead to the production of a melt with a given set of characteristics. Ideally, both of these tasks should be carried out when one is trying to understand the origin of mantle melts.

a. Fractional Crystallization Corrections

The first step in estimating the composition of melts of plagioclase or spinel lherzolite is to correct for the effects of fractional crystallization. This step requires knowledge of the conditions under which the melt was modified by fractionation, which in turn requires knowledge of the phase equilibria that control fractional crystallization under those conditions. The phase assemblage chosen for the fractional crystallization correction can have a significant effect on the corrected liquid composition and therefore the estimates of source pressure and temperature, as evidenced by the comparison of the thermometers and barometers for MORB compositions (see Section VI below). For our fractionation corrections, we used the methods of Grove et al. (1992) (see their Appendix
2). Yang, Kinzler and Grove (1996) updated this methods and we recommend using their approach if clinopyroxene is a fractionating phase. To chose the appropriate fractionating phase assemblage, the bulk sample compositions are plotted in the oxygen-normalized Olivine-Plagioclase-Clinopyroxene-Quartz pseudo-quaternary projection scheme of Tormey et al. (1987). The location of the sample is compared to the location of the multiple saturation points (MSP) for that sample (i.e., where a liquid of that composition would be multiply-saturated with a spinel or plagioclase lherzolite mantle phase assemblage) with varying pressure, which can be calculated using our model equations (Table 5). This allows the fractionation path of the sample to be visualized in 2D in the Plag-Cpx-Oliv pseudo-ternary projection (Figure 5a). The trace element composition of a sample, especially the sign of the Eu-anomaly, is a good indication of whether fractional crystallization has shifted a liquid away from the plagioclase or spinel lherzolite MSP. Figure 5a illustrates that an olivine-only fractionation correction will drive a liquid toward the spinel MSP for a much higher pressure than olivine + plagioclase fractionation. This is exemplified by our pressure estimates for the olivine-only and olivine + plagioclase corrected MORB compositions (Figure 6). Therefore to use our model to calculate the pressure and temperature a liquid was last saturated with a mantle lherzolite, an independent constraint on the type of fractionation experienced by the liquid is required. For example, the presence of olivine + plagioclase or olivine + plagioclase + clinopyroxene phenocrysts in a basalt sample are a good indication of the fractionating phase assemblage, as are the chemical composition of these minerals. This type of data is not available for the MORB glass compositions used in Section VI from the PetDB database, therefore we plot these compositions on a series of pseudo-ternary
projections. All the MORB glass compositions plot on the Oliv-Plag cotectic on the Plag-Cpx-Oliv ternary (Figure 5a) within the analytical uncertainty, which suggests they all experienced olivine + plagioclase fractionation, with the exception of the EPR and Azores compositions which may have experienced olivine + plagioclase + clinopyroxene fractionation. Correcting for olivine + plagioclase fractionation moves the liquid composition in a vertical direction along the Oliv-Plag cotectic toward the Oliv-Plag axis as shown in the Figure 5a. The relative abundance of plagioclase and olivine for the fractionation correction can be estimated by projecting the liquid onto the Oliv-Plag axis and converting from volume percent to weight percent using the density of the crystallizing phases. This estimate of the crystallized phase abundances can then be used to add increments of the crystallized phases to the bulk composition of each sample until it can be approximated as a liquid in equilibrium with a mantle olivine with a forsterite content (Fo) of 90.

Here we emphasize that the choice of fractionating mineral assemblage (oliv; oliv + plag; oliv + cpx; or oliv + plag + cpx), any changes in the fractionating minerals with decreasing pressure and temperature, and when the liquid reaches equilibrium with the mantle, all have a significant effect on the thermometric and barometric calculations for a primitive basalt. Liquids calculated by Kinzler (1997) are used to illustrate this in Figure 7. As shown in Figure 7, a liquid that has experienced multi-phase fractionation but is corrected for only olivine fractionation until it is in equilibrium with a Fo90 olivine significantly overestimates the FeO* and the MgO content of the parental liquid. When corrected for olivine-only fractionation, the 1.5-0.3 GPa post fractional crystallization liquid composition from Kinzler (1997) yields a temperature and pressure estimate from
our regression that is 145°C and 1.1 GPa greater than the actual parental liquid and the 3.0-0.3 GPa liquids yield temperature and pressure estimates that are up to 225°C and 1.75 GPa more than the parental liquid. This figure also illustrates how the choice of olivine Fo-content that represents equilibrium with the mantle can have a significant effect on the FeO and MgO of the fractionation corrected liquid and thus any temperature and pressure calculations using that composition.

b. Calculating Melts of Mantle Lherzolites

The forward model of plagioclase and spinel lherzolite melting we use here is an update of the model presented by Kinzler and Grove (1992a,b) that incorporates the improvements of Kinzler (1997). In order to use the melting model, the Mg# and abundances of minor components in each melt increment must be inferred. Estimates of the minor components are obtained using the non-modal batch melting equation,

\[ C_i = \frac{C_o}{D_B} + F(1-P_B) \]  

where \( C_i \) is the concentration of the minor component (e.g., K2O) in the melt, \( C_o \) is the concentration of that component in the bulk solid, \( D_B \) is the partition coefficient for the component between solid and melt, \( F \) is the extent of melting, and \( P_B \) is the bulk solid-melt partition coefficient for the component weighted by the stoichiometric coefficients in the melting reaction (wt.% units). Partition coefficients have been calculated using the experimental data that is incorporated in the model. The partition coefficient models are temperature- and pressure-dependent, following the method of Kinzler (1997), with the important exception of our modeling being conducted in weight units.

Our model uses the pressure-dependent spinel lherzolite melting reaction calibrated by Kinzler (1997) and the plagioclase lherzolite melting reaction of Kinzler
and Grove (1992b). The Mg\# of a melt coexisting with a lherzolite composition is determined by solving equations for the conservation of mass for FeO and MgO between lherzolite and a given fraction of melt, following the method of Kinzler and Grove (1993). These equations require knowledge of the partitioning of Fe and Mg between melt and all the coexisting phases, and the exchange $K_D$'s for Fe and Mg are calculated from the experimental dataset. Since the value for $K_D^{Fe-Mg}$ can vary with pressure and melt composition (Ulmer, 1989; Toplis, 2005), the model sets the $K_D^{Fe-Mg}$ values for clinopyroxene, orthopyroxene and spinel to 0.94, 0.97 and 1.5 times the $K_D^{Fe-Mg}$ value for olivine-melt.

Melting in the model can be specified to be either batch or polybaric near-fractional adiabatic decompression melting and the melt production rate for polybaric melting can be varied depending on preference for the heat of fusion, we recommend a value of ~0.01 wt.% per 0.1 GPa of decompression. In the case study presented in Section VII, we use this forward model to test models of both batch and fractional upper mantle melting to generate basalts from Oregon’s HLP.

VI. Comparison of Parameterization to Other Thermometers and Barometers

Previous major element models of upper mantle melting focus on the generation of MORB (Klein and Langmuir, 1987; McKenzie and Bickle, 1988; Niu and Batiza, 1991; Kinzler and Grove, 1992a, b; Langmuir et al., 1992; Kinzler, 1997) and do not consider melting of variably metasomatized peridotite. Other thermometers, barometers and melting models aim to investigate the production of a wider range of basalt compositions (e.g., OIB) but are calibrated on experimental liquids that are in equilibrium
with olivine only (e.g., Ford et al., 1983; Putirka, 2007, 2009) or olivine + orthopyroxene only (Lee et al., 2009). To explore the differences that might arise in estimating melting conditions for basaltic liquids with a model calibrated exclusively on liquids saturated with the complete upper mantle phase assemblage of olivine + orthopyroxene + clinopyroxene + plagioclase and/or spinel, we compare our thermometric and barometric estimates for five different locations of MORB genesis and three locations of OIB genesis to those from a variety of popular mantle thermometers and barometers.

MORB compositions were compiled from PetDB ocean rock database using the latitude and longitude of the selected areas, filtered for fresh glass analyses (MgO > 8 wt.%) and averaged before they were corrected for olivine-only and alternatively olivine + plagioclase fractionation until they were in equilibrium with a Fo90 mantle olivine following the methods described in Section V. Pressure estimates are required for the thermometers of Ford et al. (1983), Kinzler and Grove (1992a), Langmuir et al. (1992), Putirka et al. (2007) and the thermometer presented here (Table 5), and were calculated for the fractional crystallization corrected compositions at each location using the barometers of Lee et al. (2009) and the spinel lherzolite olivine expression from this study. In the case of MORBs, which are thought to be polybaric near fractional melts (McKenzie, 1984; McKenzie and Bickle, 1988; Johnson et al., 1990), the calculated pressure represents an average pressure for the entire melting column. Our barometric estimates for the olivine + plagioclase corrected MORB compositions are within 0.3 GPa of the Lee et al. (2009) pressures for the olivine only corrected compositions and both are within the range of accepted values for MORB generation (Kinzler and Grove, 1992b) (Figure 6). Our pressure estimates for the olivine-only corrected compositions are quite
high relative to average MORB generation pressures, because our Oliv barometer is sensitive to the increase of 0.02-0.04 in NaK# and 0.02-0.03 in Mg# for the olivine-only corrected liquid relative to the olivine + plagioclase corrected liquid. The temperature estimates for each location vary up to 185°C between thermometers when calculated with olivine-only corrected compositions (Figure 6) or 117°C when calculated with the olivine + plagioclase corrected compositions. It is interesting to note that all temperatures for the MORB localities cluster around 1300°C when the olivine + plagioclase corrected compositions are used with the exception of the Ford et al. (1983) thermometer (Fig. 6b). This suggests the variety of methods employed by the different thermometers compared here are all similarly sensitive to the chemical variability of the liquid produced by different types of fractionation corrections. It also appears that when corrected for the accurate type of crystal fractionation the various thermometers converge on similar source temperature estimates. This comparison of thermometers also reveals that calculations of mantle temperature for MORB depend heavily on the type of fractionation correction and the source pressure.

Determining the temperature and pressure for OIB genesis is significantly more complicated due to the paucity of primitive OIB glasses that can be positively identified as liquids and the abundance of compositions that have experienced complex fractional crystallization and other pre-eruptive processes after they segregated from a peridotitic source. Glasses from Kilauea volcano in Hawaii represent some of the only picritic glasses from a hotspot locality and work by Wagner and Grove (1998) reveals these liquids were never in equilibrium with garnet lherzolite assemblage on their liquidus. Rather, these tholeiites appear to have been in equilibrium with a harzburgitic source at
lithospheric depths. Wagner and Grove (1998) propose a model where the Kilauea tholeiites formed by melting of garnet lherzolite in a mantle plume, followed by equilibration of the liquids with a harzburgitic mantle at shallow depths. In addition, many primitive OIBs, such as those from Samoa (Workman et al., 2004), Marquesas (Desonie et al., 1993) and the Ontong Java plateau (Herzberg et al., 2007), exhibit evidence for olivine and/or clinopyroxene accumulation. Therefore reconstructing the parental liquid composition of an OIB in order to calculate the source pressure and temperature is not simply a matter of determining the right phase assemblage for the fractionation assemblage, as is the case for MORB. A number of techniques have been developed to circumvent these issues and calculate the parental liquid for a given OIB compositions (e.g., Putirka et al., 2007; Herzberg and Asimow, 2008). Parental liquid compositions for selected OIBs calculated by Putirka et al. (2007) plot in a location on a series of pseudo-ternary ternary projections that are not consistent with liquids in equilibrium with plagioclase or spinel lherzolite as determined by our model, or garnet lherzolite as determined by the work of Walter (1998). Rather they appear to have experienced olivine addition since last in equilibrium with a spinel lherzolite. However for the sake of comparison, we choose to use the Putirka et al. (2007) calculated compositions to contrast thermometric and barometric estimates, rather than attempt to correct a fictitious liquid composition for olivine addition.

Pressures for the calculated parental OIB liquids determined with our olivine spinel lherzolite barometer and that of Lee et al. (2009) are quite different (Figure 6f). Both sets of pressure estimates yield temperature estimates that vary by ~150°C at a given locality (Figure 6c & 6d). All the thermometers yield similar temperature estimates
for the Hawaiian and Samoan compositions and temperatures that are ~100-200°C lower for the Icelandic composition depending on the pressure. The thermometer of Lee et al. (2009) is not pressure-dependent; therefore the OIB source temperature estimates are identical at either pressure. The parameterization of Putirka et al. (2007) is the least pressure sensitive after Lee et al. (2009), and an increase in source pressure of ~2 GPa for the Hawaii and Iceland compositions results in an increase of temperatures of ~50°C. At the other end of the spectrum, the parameterization of Kinzler and Grove (1992a) is the most pressure-sensitive and an increase in source pressure of ~2 GPa results in an increase of temperatures of ~300°C. The pressure-dependence of the remaining thermometers falls between these two end members, with our model tending toward more pressure-dependent. It is difficult to draw conclusions about the actual source temperature and pressure for the three OIB localities from these calculations, which reinforces the observation that the determination of the parental liquid compositions and estimates of source pressure have profound effects on the temperature estimates for OIB genesis.

VII. Case Study of Basaltic Lavas from Oregon’s HLP

A suite of primitive basaltic lavas from Diamond Crater and surroundings in Oregon’s eastern HLP were selected to demonstrate the utility of our model for calculating the pressure and temperature where a primitive basaltic liquid was last in equilibrium with an upper mantle phase assemblage. Diamond Crater (43°05’, 118°43’) is a suite of basaltic lava flows emplaced around a central vent between 8,400-6,400 yr. ago (Sherrod, written comm., 2009). The lavas are relatively pristine diktytaxitic high
alumina olivine tholeiites (HAOT) with varying proportions of olivine and plagioclase phenocrysts visible in hand sample. Major and trace element geochemistry of all samples presented here were conducted at the Washington State University GeoAnalytical Lab (Table 6). In addition, electron microprobe analyses of the phenocrysts and groundmass of samples from Diamond Crater were conducted at the Massachusetts Institute of Technology following the same analytical procedures outlined for the experiments presented above.

The petrology and geochemistry of the Diamond Crater samples demonstrate they are low extent anhydrous melts of spinel lherzolite. The trace element abundances of our most primitive samples are compared to modeled 1-15% batch melts of a primitive spinel lherzolite mantle in Figure 8. We modeled batch melting of primitive upper mantle using the appropriate equations from Shaw (2006), major element abundances from Hart and Zindler (1986), trace element abundances from Hoffman (1988), partition coefficients from Halliday et al. (1995) with additional spinel partition coefficients from Canil (2004) and Kinzler and Grove (1992a) and mantle modes and melting reactions from Kinzler (1997). This model suggests samples ch8329 and ch8343 are 2 - 6 wt.% melts of a primitive spinel lherzolite mantle, samples DC 07-16 and DC 07-12 are 6 -10 wt. % melts and DC 07-13 and ch8357b are 10-15 wt.% melts. All samples are relatively enriched in Ba, Sr and La and depleted in Th and Pb relative to equivalent degree model melts. The positive or negligible Eu-anomaly demonstrates these samples were not a product of melting plagioclase lherzolite. The bulk rock compositions, as well as the chemical composition of the plagioclase and olivine phenocrysts in thin sections of the Diamond Crater samples indicate these liquids did not form in the presence of hydrous fluids.
Thus these samples are excellent candidates for our regression because they appear to be anhydrous low extent mantle melts.

For use in our model, the major element compositions of the selected samples were first corrected for minor fractional crystallization following the methods discussed in Section V. The presence of either olivine + plagioclase phenocrysts or olivine + plagioclase + minor clinopyroxene microcrysts in the Diamond Crater samples and the chemical composition of these minerals imply these samples experienced a period of low pressure olivine + plagioclase fractionation, in some cases followed by a late episode of olivine + plagioclase + clinopyroxene fractionation. Alternatively, we tested correcting the samples for a period of olivine-only fractionation, followed by a period of lower pressure olivine + plagioclase fractionation. However, correcting for this scenario drives the liquids toward the olivine apex on a Plag-Cpx-Oliv ternary and beyond the spinel MSP long before they reach equilibrium with Fo90 olivine. Therefore we conclude, the most logical fractional crystallization history for these samples is a period of low-pressure olivine + plagioclase fractionation, and in some cases followed by a period of olivine + plagioclase + clinopyroxene fractionation.

Once corrected for olivine + plagioclase fractionation, the corrected compositions and their corresponding mineral component values were plugged into our spinel lherzolite Oliv barometer and the spinel lherzolite thermometer (Table 5) and the results are presented in Table 7. The samples were generated at temperatures of ~1323-1370°C and pressures of 1.22-1.6 GPa, which corresponds to depths of ~42-54 km based on density estimates for the crust and mantle from recent seismic refraction profiles near Diamond Crater (Cox, personal communication, 2011). Recent work on the crustal
structure of Oregon’s HLP using teleseismic P-to-S receiver functions finds the Moho is located at \( \sim 30 \) km in the region of Diamond Crater when the H-\( \kappa \) method is applied to individual stations, or \( \sim 36 \) km if the maximum amplitudes from GCCP stacks are used (Eager et al., 2011). In either case, the primitive basalts erupted in the Diamond Crater area appear to be formed by shallow decompression melting of spinel lherzolite, with the shallowest depths of melting almost immediately below the Moho.

We can also apply our forward model of spinel lherzolite melting to model both batch and near-fractional mantle melting over the range of calculated depths for the Diamond Crater parental liquids. The trace element modeling indicates Diamond Crater basalts are relatively enriched in several of the large ion lithophile elements typically carried by subduction zone fluids and/or melts, such as Ba and Sr, but are depleted in others, such as Rb and Th, relative to the primitive mantle (Figure 8), suggesting they may have originated from variably metasomatized mantle. Therefore, we specified the bulk mantle composition in our models to be either that of the primitive mantle from Hart and Zindler (1986) or a xenolith of metasomatized depleted mantle from southern Patagonia (Ntaflos et al., 2007). Models were run with initial pressures of 1.0-1.8 GPa, an olivine-melt \( K_{D}^{Fe-Mg} \) of 0.3, and either 100% of the melt was removed for batch melting or 90% in the case of near-fractional melting. Batch melting models were run for 2-10 wt.% melting. For near-fractional melting, we chose the melt production rate \( \frac{dF}{dP} \) to be 1% per 0.1 GPa of decompression and track the composition of the melt produced at each step, as well as the accumulated melt. The modeling results that best replicate the composition of the fractionation-corrected Diamond Crater basaltic lavas are presented in Figure 9 and Table 8. Models that use the Patagonia xenolith mantle
composition exhibit a much better fit to the Diamond Crater liquids than those that use the bulk primitive mantle composition (Figure 9). Overall, the 6% batch melts generated at either 1.3 or 1.6 GPa and the accumulated melt generated during near-fractional melting between 1.6-1.0 GPa are closest to the Diamond Crater liquids. The 1.3 GPa 6% batch melt probably exhibits the single best fit, falling within <0.5 wt.% of all the key oxides, as shown in Figure 9. This is an exciting result, as two completely independent sets of calculations (i.e., the thermo-barometric calculations and the forward mantle melting model) both suggest the Diamond Crater basalts were generated during small degrees of melting of a metasomatized, depleted spinel lherzolite at 1.3-1.6 GPa. This example also illustrates the power of our spinel lherzolite forward melting model to constrain the melting style, in addition to the pressure, temperature and composition of the bulk mantle during melting. Our result that Diamond Crater HAOTs likely formed under batch melting conditions immediately below the Moho in Oregon’s HLP agrees with an increasing number of studies on HAOT magma generation at continent margins and interiors (e.g., Bartels et al., 1991; Bacon et al. 1997; Holbig and Grove, 2008). Thus the plagioclase and spinel lherzolite melting models presented here represent two new tools to understand the melting of variably metasomatized mantle and the petrogenesis of HAOT magmas.
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Figure & Table Captions

**Figure 1.** Composition of the experimental liquids in equilibrium with a plagioclase (white circles) or spinel lherzolite (black circles) phase assemblage (olivine + cpx + opx ± plag ± spinel) modeled in our regression (see Table 1 for references). Additional experiments conducted as part of this study are shown for comparison (triangles).

**Figure 2.** Observed versus calculated mineral component values for spinel lherzolite experimental liquids empirically fit by our regression. Dashed line represents 1:1 correspondence. Plots a-d show the mineral component values returned by equations calculated from the full dataset (see Appendix 1), while plots e-h show only the mineral component values returned from equations calculated with only those experiments done in the simple systems CMAS and CMASN and those analyzed on the microprobe at MIT.

**Figure 3. a)** Experimental versus calculated pressure for experiments from Kinzler and Grove (1992a) and Kinzler (1997) using the spinel lherzolite olivine expression. These two sets of experiments were conducted in different labs but analyzed on the same microprobe with the same standards, therefore the systematic differences in pressure between the two studies likely result from interlab pressure calibration, not analytical, differences. **b)** Experimental versus calculated temperature for the plagioclase and spinel lherzolite experimental liquids determined with our plagioclase lherzolite and spinel lherzolite thermometers.
Figure 4. Illustration of the plagioclase-spinel lherzolite boundary in Pressure-Mg#-(1-NaK#) space. The plagioclase-spinel transition (PST) can be conceptualized as two bent screens (the plag-only to plag + spinel and plag+spinel to spinel-only boundaries) that define a three dimensional field with a height of ~0.3 GPa at high NaK#s and low Mg#s. This field collapses to a line in both the P-(1-NaK#) and P-Mg# planes. We plot 1-NaK#, rather than NaK#, to allow the CMASN experiments that constrain the slope of the univariant PST to be visualized on the plane where Mg#=1. Also shown are experiments in the CMAS system that constrain the pressure of the PST at an Mg# and 1-NaK# of one. A typical upper mantle primitive liquid is illustrated by the red X’s.

Figure 5. Rational for fractional crystallization corrections depicted on pseudo-ternary projections. a) Uncorrected average primitive glass compositions from the five MORB localities plotted with the plagioclase and spinel lherzolite multiple saturation points calculated for the av. Reykjanes MORB composition. Insert illustrates how different types of fractional crystallization corrections drive the glass compositions in very different directions with regards to the spinel multiple saturation points. Gray numbers indicate pressures for the spinel multiple saturation points. b) Uncorrected MORB compositions on Oliv-Cpx-Qtz pseudo-ternary. Grey dashed line illustrates the 1 atm Oliv-Cpx cotectic. The EPR and Azores average compositions lie on the cotectic indicating the possibility of 1 atm oliv+plag+cpx fractional crystallization for these liquids. Liquids below the cotectic experienced oliv+plag fractionation. c) Average MORB glass compositions corrected for olivine and plagioclase crystallization using the methods described in Section IV.
Figure 6. Comparison of temperature estimates from popular thermometers and this study for five MORB and three OIB locations. MORB compositions are primitive glasses from the PetDB database and OIB compositions are calculated by Putirka et al. (2007). Temperatures were calculated using pressure estimates from the Lee et al. (2009) barometer and our spinel lherzolite olivine expression.

Figure 7. Effect of olivine-only fractionation correction on mantle melts that experienced a multiphase fractionation history. Parental liquids are polybaric, near fractional aggregate magma compositions calculated by Kinzler (1997). The composition of these parental liquids (black lines and dots) are shown during olivine fractionation, followed by olivine + plagioclase fractionation and in the case of the high-pressure parental liquid, a third period of olivine + plagioclase + clinopyroxene fractionation, until MgO in the liquid equals 8 wt.%. The fractionated liquid is then corrected for olivine-only fractionation until it is in equilibrium with Fo90 olivine using a Herzberg and Asimow (2008) olivine-liquid Fe-Mg Kd. Also shown are contours of the liquid composition using an olivine-liquid Fe-Mg Kd of 0.3 (dashed gray lines).

Figure 8. a) Trace element abundance of basalts from the Diamond Crater region, Oregon and modeled 1-15% melts of a spinel lherzolite mantle. b) Rare earth element abundances of the same samples and modeled melts.
**Figure 9.** Results of our forward spinel lherzolite melting model relative to fractionation-corrected basaltic liquids from the Diamond Crater area, Oregon.

**Table 1.** Experiments used in our plagioclase and spinel lherzolite melting model calibration. Star indicates experiments used in the MIT + CMAS(N) regression presented in Table 5.

**Table 2.** Composition of experimental starting materials for experiments conducted in this study.

**Table 3.** Experimental run conditions and phases present for experiments conducted in this study.

**Table 4.** Normalized major element composition of experimental phases from our experiments, as determined by electron microprobe analysis.

**Table 5.** Model equations produced by multiple linear regression of experimental plagioclase and spinel lherzolite liquids. Number in italics indicates the standard error for the coefficient above.

**Table 6.** Geochemistry of primitive basalts from Diamond Crater area, Oregon. Major element XRF and trace element ICPMS analyses were conducted at the WSU GeoAnalytical Lab.
Table 7. Pressure and temperature estimates for Diamond Crater area samples produced by our spinel lherzolite thermometer and barometer (Oliv expression).

Table 8. Starting compositions, run conditions and results from our spinel lherzolite forward melting model, as used to replicate the parental liquids for basaltic lavas from the Diamond Crater area, Oregon. Starting bulk peridotite compositions used in the modeling are a xenolith of depleted metasomatized mantle from Patagonia (Ntaflos et al., 2007) and the primitive mantle composition of Hart and Zindler (1986), renormalized to exclude MnO, NiO and P<sub>2</sub>O<sub>5</sub>. The Diamond Crater compositions were corrected for olivine + plagioclase fractionation using the methods discussed in Section V until they reached equilibrium with Fo90 olivine.
Appendix 1.

Both the plagioclase and spinel lherzolite expressions presented in Section III and Table 5 are the product of regressing only those experiments analyzed on the microprobe at MIT and experiments done on the simple systems CMAS and CMASN. Here we present an additional set of lherzolite regressions determined from the full dataset minus several experiments as discussed below. There are only five additional experiments that can be included in the full dataset for plagioclase lherzolite melting. These experiments are from Dunn and Sen (1994), Falloon et al. (1997), and Falloon et al., (1999).

Figure 2a-c reveals that the suite of experiments from Schwab and Johnston (2001) and Draper and Johnston (1992) define a slope <1, that does not correspond to the slope of ~1 defined by the majority of the data. For this reason, we have chosen to exclude the data of Draper and Johnston (1992) and Schwab and Johnston (2001) from our regressions of the full dataset.

Full Dataset Regression:

Plag Lherzolite

\[ T = 1216 + 98.8 \, (P) - 73.5 \, (1-Mg#) - 168.0 \, (NaK#) + 24.8 \, (TiO_2) - 9.70 \, (K_2O) \]  
\[ r^2 \]

Oliv = 0.128 + 0.077 (P) + 0.165 (1-Mg#) - 0.271 (NaK#) - 0.009 (TiO_2) + 0.017 (K_2O)  
\[ (0.004) \quad (0.005) \quad (0.014) \quad (0.014) \quad (0.003) \quad (0.004) \]

Cpx = 0.244 - 0.064 (P) - 0.053 (1-Mg#) - 0.081 (NaK#) + 0.005 (TiO_2) - 0.025 (K_2O)  
\[ (0.008) \quad (0.010) \quad (0.029) \quad (0.030) \quad (0.007) \quad (0.009) \]

Plag = 0.411 + 0.139 (P) - 0.130 (1-Mg#) + 0.639 (NaK#) + 0.014 (TiO_2) - 0.035 (K_2O)  
\[ (0.009) \quad (0.006) \quad (0.031) \quad (0.032) \quad (0.008) \quad (0.009) \]

Qtz = 0.216 - 0.152 (P) + 0.018 (1-Mg#) - 0.287 (NaK#) - 0.010 (TiO_2) + 0.043 (K_2O)  
\[ (0.009) \quad (0.011) \quad (0.032) \quad (0.034) \quad (0.008) \quad (0.010) \]
Spinel Lherzolite

\[ T = 1187 + 135.8 \, (P) - 41.9 \, (1-Mg\#) - 97.1 \, (NaK\#) - 1.63 \, (TiO_2) - 13.92 \, (K_2O) \]

\[ \text{Oliv} = 0.116 + 0.090 \, (P) + 0.195 \, (1-Mg\#) - 0.262 \, (NaK\#) - 0.009 \, (TiO_2) + 0.011 \, (K_2O) \]

\[ \text{Cpx} = 0.191 - 0.003 \, (P) + 0.038 \, (1-Mg\#) - 0.243 \, (NaK\#) + 0.015 \, (TiO_2) + 0.012 \, (K_2O) \]

\[ \text{Plag} = 0.559 - 0.046 \, (P) - 0.199 \, (1-Mg\#) + 0.945 \, (NaK\#) + 0.007 \, (TiO_2) - 0.042 \, (K_2O) \]

\[ \text{Qtz} = 0.134 - 0.042 \, (P) - 0.034 \, (1-Mg\#) - 0.439 \, (NaK\#) + 0.001 \, (TiO_2) + 0.019 \, (K_2O) \]
Figure 1.
Figure 2.
Figure 4.
Figure 5.
Figure 6.
Figure 7.
Figure 8.
Figure 9
Table 1. Experiments used in our plagioclase and spinel lherzolite melting model calibration. Star indicate the experiments used in the MIT + CMAS(N) regression presented in Table 5.

<table>
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<td><strong>Plagioclase Lherzolite</strong></td>
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<td>Baker, M.B., Grove, T.L., Price, R. (1994)*</td>
<td>basalt, basaltic andesite, andesite, Mt. Shasta, CA</td>
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<td>Falloon, T.J., Green, D.H., O'Neill, H.St.C., Hibberson, W.O. (1997)</td>
<td>peridotite and corresponding low degree melts synthesized from oxides</td>
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<td>basaltic andesite, Medicine Lake, CA and seeds of natural minerals</td>
</tr>
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<td>Kinzler, R.J., Grove, T.L. (1992)*</td>
<td>MORB, natural samples and synthetic oxide mixtures</td>
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<td>Presnall et al. 1979*</td>
<td>CMAS synthesized from oxides</td>
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<td>Walter and Presnall (1994)*</td>
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<td>Bartels, K.S., Kinzler, R.J., Grove, T.L. (1991)*</td>
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<td>Falloon, T.J., Danyushevsky, L.V., Green, D.H. (2001)</td>
<td>peridotite and corresponding low degree melts synthesized from oxides</td>
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<td>Kinzler, R.J. (1997)*</td>
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<td>Pichavant, M., Mysen, B.O., MacDonald, R. (2002)</td>
<td>high-MgO basalt, Soufriere, St. Vincent</td>
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<td>Presnall et al. 1979*</td>
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<td>Villiger, S., Ulmer, P., Muntener, O., and Thompson, A.B. (2004)</td>
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Table 2. Composition of experimental starting materials for experiments conducted in this study.

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Table 4. Normalized major element composition of experimental phases from our experiments as determined by electron microprobe analyses.
Table 5. Model equations produced by multiple linear regression of experimental plagioclase and spinel lherzolite liquids. Number in italics indicates the standard error for the coefficient above.

|                       | Plagioclase Lherzolite |  
|-----------------------|------------------------|-------|
|                       |                        |       |
| T                     | 1220 + 90.8 (P) - 21.1 (1-Mg#) - 160.9 (NaK#) + 6.28 (TiO2) - 13.4 (K2O) | 0.97  |
|                       | 7.63                   | 10.7  |
| Oliv                  | 0.130 + 0.076 (P) + 0.151 (1-Mg#) - 0.272 (NaK#) - 0.003 (TiO2) + 0.016 (K2O) | 0.96  |
|                       | 0.016                  | 0.009  |
|                       | 0.011                  | 0.051  |
|                       | 0.008                  | 0.011  |
| Cpx                   | 0.252 - 0.088 (P) - 0.047 (1-Mg#) - 0.005 (NaK#) + 0.005 (TiO2) - 0.005 (K2O) | 0.88  |
|                       | 0.01                    | 0.014  |
|                       | 0.09                    | 0.039  |
|                       | 0.03                    | 0.044  |
| Plag                  | 0.410 + 0.138 (P) - 0.135 (1-Mg#) + 0.547 (NaK#) + 0.016 (TiO2) - 0.036 (K2O) | 0.97  |
|                       | 0.01                    | 0.014  |
|                       | 0.01                    | 0.039  |
|                       | 0.09                    | 0.044  |
| Qtz                   | 0.213 - 0.144 (P) - 0.005 (1-Mg#) - 0.303 (NaK#) - 0.003 (TiO2) + 0.050 (K2O) | 0.97  |
|                       | 0.009                   | 0.012  |
|                       | 0.009                   | 0.034  |
|                       | 0.034                   | 0.009  |
|                       | 0.01                    |

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<td>T</td>
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Table 6. Geochemistry of samples from Diamond Crater area, Oregon. Major element XRF and trace element ICPMS analyses were conducted at the WSU GeoAnalytical Lab.

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<td>21.000</td>
<td>10.140</td>
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Table 7. Pressure and temperature estimates for Diamond Crater area samples produced by our spinel Iherzolite thermometer and barometer (Oliv expression).

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<th>P (GPa)</th>
<th>T (°C)</th>
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<tr>
<td></td>
<td>Total F</td>
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<tr>
<td><strong>Patagonia Xenolith</strong></td>
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<tr>
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Chapter 4:
Constraints on the Depths and Temperatures of Anhydrous Asthenospheric Melting and the Location of the Lithosphere-Asthenosphere Boundary Beneath southern Oregon and northern California
Abstract

An extensive geochemical dataset of young (<10.5 Ma) primitive basaltic lavas from across the northernmost Basin and Range and the central - southern Cascades is used to calculate the depths and temperatures of asthenospheric melting in this region. In conjunction with recent geophysical observations, these calculations place primary constraints on the tectono-magmatic processes driving anhydrous melting in a modern convergent margin and back-arc. The basalts have high Mg#s (100Mg/(Mg + FeT)) at low SiO₂ (<52 wt.%), low phenocryst abundances, and trace element concentrations such that we consider them to be primitive. The minimum depth of melting below Oregon’s High Lava Plains, a geographic sub-province of the Basin and Range, decreases towards the west, with melting occurring at ~45 km below Jordan Valley volcanic center, ~30 km below Newberry volcano, and ~25 km below Crater Lake volcano on the Cascades arc axis. To the south, the minimum depth of melting also decreases towards the west, from ~40 km below the Modoc Plateau to 33 km below Medicine Lake and Mt. Shasta volcanoes. Further south at Mt. Lassen, the southernmost Cascades volcanic center, the minimum depth of melting is 35 km. All basalts in this study originated at 1250-1320°C and the calculated minimum depths of asthenospheric melting are very close to Moho depths as determined from a number of regional geophysical studies. These observations point to the hot nature of the mantle immediately below the Moho and suggest the Moho and lithosphere-asthenosphere boundary occur at approximately the same depth in this region. <10.5 Ma anhydrous mantle melting in southern Oregon and northern California was likely driven by a combination of corner flow in the mantle wedge, toroidal flow around the southern edge of the subducting Juan de Fuca and Gorda plates, and crustal
extension-related upwelling, not a mantle plume. Geodynamic models of mantle flow also indicate the thin mechanical lithosphere is an important factor in the generating the observed conditions in the mantle.
1. Introduction:

The goal of this study is to use the chemical characteristics of primitive basalts from southern Oregon and northern California in the northwestern United States to further our understanding of the processes driving anhydrous mantle melting beneath convergent margins and extending into the back-arc. 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 where 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). Specifically, a notable number of volcanic centers have formed since 25 Ma in the area between the southern Cascades volcanic arc and the Oregon-Idaho border that 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, 1982; Christiansen et al., 2002), lithospheric delamination (Hales et al., 2005; West et al., 2009), back-arc extension (Christiansen and McKee, 1978) 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 the volcanic centers in question to mantle plume-related volcanism (e.g., Geist and Richards, 1993; Camp and Ross, 2004) because of the nearby eruption of $>10^5$ km$^3$ 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 km$^3$) and consists predominantly of anhydrous high alumina olivine tholeiites (HAOT) that have experienced very little crystal fractionation and crustal contamination.
(Hart et al., 1984; Draper, 1991; Hart et al., 1997). By utilizing primitive basaltic lavas from these volcanic centers that approximate liquid compositions and correcting for the effects of 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 identify the tectonic and magmatic driving forces responsible for their formation.

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 (Till et al., 2011*). 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 functions data (e.g., Gashawbeza et al., 2008; Eager et al., 2011) and Rayleigh Wave tomography (Wagner et al., 2010) and other seismic refraction surveys (e.g., Leaver et al., 1984; Zucca et al., 1986), in order to place new constraints on the thickness of the mechanical lithosphere.

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 viscous fluid on geologic timescales (i.e., the asthenosphere). However, in practice there is no unified theory for what comprises the lithosphere and it is alternately defined by its mechanical, seismic, electric, thermal, or chemical properties. As such multiple depths for the LAB can be calculated in a given area depending on the property used 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 the depths of intraplate earthquakes
and plate flexure models. By calculating depths of melt extraction from the
asthenosphere in southern Oregon and northern California, we produce an independent
estimate of the maximum thickness of the mechanical lithosphere in the region that can
be used to further evaluate the nature of the lithosphere, especially in regions with a
prolonged history of volcanism.

2. Tectonic and Volcanic History

The western United States has a rich and varied volcanic and tectonic history (see
reviews in Humphreys and Coblentz, 2007). Extensive volcanism in the western United
States began in the early Cenozoic with the ignimbrite flare-up, a period of large volume
silicic magmatism that swept through the western United States between 50 and 25 Ma
(e.g., Lipman et al., 1972; Christiansen and Yeats, 1992). In eastern Oregon, a period of
silicic volcanism was followed by a period of massive flood basalt eruptions near the
Oregon/Idaho/Nevada border that produced the Steens Mountain basalts from 16.6-15.5
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 between
16.5 and 6 Ma (Waters, 1961; Swanson and Wright, 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. The Basin and Range province
impinges on the volcanic highlands of southern Oregon and Cascades arc, and as such
these regions have been characterized by NW-SE to N-S normal faulting and associated extension-related volcanism since ~14 Ma.

This study in part focuses on lavas from the High Lava Plains (HLP) of central and southeastern Oregon, a geographic sub-province of the northern Basin and Range. This ≤10 Ma bimodal volcanic field seemingly connects the Yellowstone-Snake River Plain (YSRP) track to the east, with the Cascades arc to the west and its tectono-magmatic origin has been a topic of debate since the 1970’s (e.g., MacLeod et al., 1975; Draper, 1991; Jordan et al., 2004). The volcanic rocks in the HLP are intercalated basalt lava flows and rhyolite ash flow tuffs and related sediments with widely distributed rhyolite dome complexes. Silicic volcanism in the HLP migrated in a northwesterly direction starting in southeastern Oregon at ca. 10 Ma and ending near Newberry volcano at ca. 2 Ma where volcanism continues today (Jordan et al., 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 did not exhibit age progressive migration similar to the rhyolitic volcanism in the HLP (Jordan et al., 2004). HLP basaltic lavas are dominantly olivine and plagioclase phryic high-alumina olivine tholeiites, which often exhibit a diktytaxitic texture (Hart et al., 1984; Draper, 1991; Jordan et al., 2004; Grove et al., Petrology of the HLP, special volume). The abundant volcanic and sedimentary cover in the HLP make it difficult to unravel the basement geology. Exposures of oceanic and arc terranes in the Klamath Mountains to the southwest and Blue Mountains to the north, suggest the HLP is built on accreted Mesozoic-Cenozoic basement composed of island-arc assemblages, geosynclinals deposits and paleo-microplates assembled on the western margin of the North American plate over time (e.g., Dickinson, 1979). The eastern margin of the HLP
is located at the Oregon-Idaho border, which roughly coincides with western margin of Precambrian North America and an abrupt change in the initial $^{87}\text{Sr} / ^{86}\text{Sr}$ of Mesozoic and Cenozoic magmatic rocks from $<0.706$ west of the boundary to $>0.706$ to the east (Armstrong et al., 1977; Manduca et al., 1992).

The Cascades arc is a subduction-related volcanic arc extending 1300 km N-S from southern British Columbia to northern California. Today, the Cascades are characterized as a “hot” subduction zone due to the subduction of the young (<28 Ma) Explorer, Juan de Fuca and Gorda Plates and high heat flow within the central portion of the arc (e.g., Schmidt et al., 2008). Convergence of these plates is occurring at 40-45 mm/yr and varies from orthogonal convergence in British Columbia to more oblique northwesterly convergence in southern Oregon and California (Wilson, 1988), where we focus in this study. Geophysical studies image the slab at ~100-200 km depth below this southern extent of the Cascades (Zucca et al., 1986; Harris et al., 1991; Benz et al., 1992). In northern California, subduction of the Mendocino triple junction suggests the presence of a slab window below the southernmost Cascades (Beaudoin et al., 1996). The Tertiary volcanic deposits exposed in the southern Cascades arc can be divided into a western faulted Eocene-Miocene sequence, unconformably overlain by the High Cascades Pliocene-Holocene sequence. In the southern Cascades, abundant diffuse mafic scoria cones are erupted in close proximity to more proto-typical andesitic arc volcanic rocks (Sherrod and Smith, 1990). Mt. Shasta is a andesite stratovolcano near the southern end of the Cascades arc that grew in four eruptive stages between 250 ka and 2 ka along with isolated flank eruptions and cinder cones (Christiansen et al., 1977). Work by Baker et al. (1994) reveals Quaternary HAOTs erupted in proximity to Mt. Shasta represent
anhydrous low degree (6-10%) partial melts of a depleted mantle source that has been slightly enriched by a fluid component derived from the subducted slab. Similar to the HLP, the structure and composition of the southern Cascades basement can be difficult to unravel but appears to be composed of thrust sheets of Paleozoic to Eocene accreted oceanic and immature continental terrains, such as the Paleozoic ultramafic Trinity peridotite, and the Paleozoic-Mesozoic marine metasedimentary and volcanic rocks exposed in the Klamath Mountains (Griscom, 1980; Blakely et al., 1985; Zucca et al., 1986; Snoke and Barnes, 2006).

Newberry Volcano in central Oregon and Medicine Lake Volcano in northern California are both broad, shield-shaped volcanoes > 1 km in height with central calderas, flanks dotted with multiple vents and extensive aprons of basaltic lava (Donnelly-Nolan, 2008). Both volcanoes are located just east of the Cascades arc axis at the intersection of major tectonic features. Newberry is located at the westernmost terminus of the HLP volcanic trend and the Brother’s Fault zone, the northern bounding fault of the Basin and Range (Lawrence, 1976), while Medicine Lake volcano is located at the northwestern extension of the Walker Lane fault zone (Grose et al., 1989; Hildreth, 2007), the southern extension of the Klamath Graben, and the volcanic highlands that extend WSW toward Mt. Shasta (Donnelly-Nolan et al., 2008). Both volcanoes have also erupted primitive basaltic through rhyolitic lavas during their ~half million year histories. Newbery Volcano is the larger of the two volcanoes (~3000 km²) and has erupted >500 km³ of bimodal basalt and high-silica rhyolite since ~500 ka (Jensen et al., 2009), including at least three significant ash-flow tuffs, the most recent at ~80 ka. Overall Newberry lava compositions are unlike those from elsewhere in the HLP and primitive basalts at
Newberry are notably higher in K$_2$O that those from Medicine Lake. Volcanism at Medicine Lake (~2000 km$^2$) began at ~475 ka (Donnelly-Nolan and Lanphere, 2005; Donnelly-Nolan et al., 2008) and the early eruptive history of the volcano produced abundant rhyolitic to dacitic lava until 300 ka. The majority of the volcano was built by eruptions of basalts between 100 and 13 ka that was followed by postglacial episodic bimodal silicic and basaltic volcanism. Petrologic and isotopic work indicates arc fluids are present in both volcanic systems, such that both erupted wet calc-alkaline basalts, as well as dry HAOT’s throughout their histories, often in temporal and spatial proximity to one another. Studies of HAOT’s erupted at Medicine Lake suggest they are anhydrous partial melts of a MORB-like mantle lherzolite source enriched by a fluid component derived from the subducted slab (Bartels et al., 1991).

The Modoc Plateau is a block-faulted Tertiary volcanic terrain in the northwest Basin and Range (NWBR), located in the extreme northeast corner of California, east of the Cascades arc and south of the HLP. Although the Basin and Range province (BRP) of eastern, central and western Nevada has been well studied, comparatively little is known about the diffuse northern margin of the province that extends into northeastern California and southern Oregon. Limited data suggest that BRP-related extension migrated northwestward from central Nevada since the late Oligocene (see summary by Wernicke, 1992) and began after 12 Ma in the NWBR of northeast California and southern Oregon (e.g., Colgan et al., 2004). The numerous models for the driving forces for BRP extension largely fall into two categories: (1) partitioning of the relative motion between the North American plate and the subducting Farallon plate (e.g., Humphreys, 1994; Atwater and Stock, 1998; Wesnousky, 2005) or (2) collapse of the Laramide-age
crust in the region (e.g., Glazner and Bartley, 1984; Humphreys and Coblentz, 2007; West et al., 2009). 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 (Duffield and Fournier, 1974). The Warner Range, located in the eastern Modoc Plateau, is a N-S horst bounded by the Surprise Valley fault, the westernmost large-offset fault of the BRP. The Warner Range exposes over 3000 m of rhyolitic to basaltic pyroclastic deposits and lava flows erupted between 32 and 14 Ma (Duffield and McKee, 1986; Fosdick et al., 2005).

3. Methods

3.1 Lava Sample Selection

Primitive basaltic lavas from across the northernmost Basin and Range 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 asthenosphere. We consider lavas to be primitive if they are rich in Mg relative to Fe (Mg# (Mg/(Mg+Fe\textsuperscript{T})) >0.50) with SiO\textsubscript{2} <52 wt.% and relatively phenocrystal poor. Most of the samples are HAOT and they are typically aphyric, containing rare microphenocrysts of olivine and plagioclase. Many samples, approach compositions of primary magmas with Mg#s > 0.70 (Figure 2) and require very little correction to bring their compositions into equilibrium with a mantle mineral assemblage. In addition, we restricted our calculations to Pliocene or younger 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 melting event. These criteria are met by lavas from across the HLP, including Jordan Valley Volcanic Field (Hart, unpublished data), Diamond Crater (Till et al., 2011), unnamed vents near Steens Mountain and the Blue Mountains (Carlson, unpublished data), Newberry Volcano (Donnelly-Nolan, unpublished data) and the central and southern Cascades arc, including South Sister (Schmidt et al., 2008), Crater Lake (Bacon, 1990; Bacon et al., 1997), Mt. Bailey (Barnes, 1992), Mt. Shasta (Baker et al., 1994; Elkins-Tanton et al., 2001), Medicine Lake (Baker et al., 1991; Bacon et al., 1997; Elkins-Tanton et al., 2001; Donnelly-Nolan et al., 2008; Donnelly-Nolan, 2010) and Mt. Lassen (Bacon et al., 1997) volcanoes, as well as various unnamed basaltic vents in the Modoc Plateau (Carlson, unpublished data). The basaltic lavas considered here are HAOTs to calc-alkaline basalts. The trace element compositions of the basalts suggest 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.

3.2 Basalt Thermometry and Barometry

These data 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. (2011) until the bulk compositions could be approximated as a liquid in equilibrium
with a mantle olivine with a forsterite content (Mg$_2$SiO$_4$) of 90. The spinel lherzolite melting model of Till et al. (2011) was then 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, also known as the multiple saturation point (MSP). The olivine equation of Till et al. (2011) was demonstrated to most closely reproduce the pressure of experiments and is therefore used here for our pressure calculations. These basaltic melts likely represent batch melts as demonstrated by Till et al. (2011) and therefore the calculated pressure represents the shallowest depth of mantle equilibration, or melting, for a given melt. Density profiles for the crust and upper mantle below the HLP (Cox and Keller, unpublished data) and northern California (Zucca et al., 1986) were used to determine a pressure-depth relationship at the basalt sample localities and convert to depth.

Experiments by Gaetani and Grove (1998) illustrate that silicate melts formed in the presence of water are multiply saturated with a mantle assemblage at greater pressures than analogous anhydrous melts. Specifically, these experiments demonstrate the multiple saturation point for a liquid in equilibrium with a spinel lherzolite mineral assemblage shifts to greater pressure by 0.4 GPa with the addition of 6 wt.% H$_2$O. The melting model of Till et al. (2011) was calibrated with only anhydrous plagioclase and spinel lherzolite melting experiments, therefore the pressures of multiple saturation calculated with the Till et al. (2011) barometer are inaccurate for hydrous melts. Olivine-plagioclase hygrometry suggests basalts from Newberry Volcano with ~1000 ppm Sr contained ~4 wt.% H$_2$O prior to eruption (Barr, 2010). The Sr/H$_2$O ratio from the Newberry basalts was therefore used to estimate the magnitude of the pressure correction.
required for HLP and Cascades samples that likely involved H$_2$O in their formation based on their petrology (e.g., olivine and plagioclase compositions) and/or geochemistry (e.g., calc-alkaline basalts with $>400$ ppm Sr).

4. Pressure and Temperature of Mantle Melt Extraction for Primitive HLP and Cascadian Basalts

The depths of melting for primitive southern Oregon and northern California lavas are illustrated in two E-W transects: one transect below the HLP and Oregon Cascades at $\sim 43.5^\circ$N that covers $\sim 5^\circ$ longitude and another below the California Cascades and Modoc Plateau at $\sim 41.5^\circ$N that covers $\sim 3^\circ$ longitude (Figure 3). The depth of Quaternary melt extraction along the HLP track overall shallows to the west, from $\sim 45$ km below Diamond Crater and Jordan Valley Volcanic Field in the east, to $\sim 30$ km below Newberry volcano and $\sim 25$ km below Crater Lake in the west. The minimum depth of melting is $\sim 35$ km for a suite of basalts from unnamed vents near Steens Mountain and the Blue Mountains in eastern Oregon, approximately 10 km shallower than the Quaternary basalts at Jordan Valley and Diamond Crater. Two primitive basalt samples from the South Sisters area in the Cascades arc (Schmidt et al., 2008), which is found at the far western end of the HLP track, originated at 45-53 km similar to the average depth of basalts from Newberry (Donnelly-Nolan, unpublished data; Barr, 2010). The maximum depth of melt extraction is located at $\sim 55$ km for the northern transect and $\sim 50$ km for the southern transect.
Our thermometry indicates that the basalts originated at temperatures between 1250-1380°C in the asthenosphere. Our temperature calculations are dependent on both pressure and the compositional characteristics of the fractional crystallization-corrected liquid, including its Mg# and wt.% K₂O. Therefore the temperatures shown in Figure 3 are an average for the liquids that originated at the specified depth. Phase equilibrium studies of primitive basalts lavas from Medicine Lake (Baker et al., 1991), Mt. Shasta (Baker et al., 1994), and Jordan Valley (Till et al., 2011) find that HAOT’s from these localities were separated from the mantle at ~1300°C and 10-11 kbar, a pressure range equivalent to ~35-39 km depth in this region, corroborating our calculations.

5. Constraints on Crustal and Lithospheric Structure

A number of geophysical studies have focused on the crustal structure and Moho depths for the regions of interest in this study and reveal remarkably consistent results. Most recently, Eager 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 H-κ and the GCCP stacking techniques overall yield similar results for the topography of the Moho with ≥ 40 km thick crust below the Cascades and thinner 31-36 km crust below the HLP and northern Great Basin, where GCCP Moho depths average ~5 km deeper than those determined from H-κ stacking (Figure 3). A reverse seismic refraction profile 30 km southwest of Crater Lake volcanic field defined a crustal thickness of 44 km (Leaver et al., 1984). Gashawbeza et al. (2008) also analyzed
teleseismic data using H-κ stacking to determine Moho depths below the Modoc Plateau and the areas to the east in the Great Basin. Two stations in the Modoc Plateau yield Moho depths of 35 and 37 km, compared with H-κ Moho depths of 36-37 km determined by Eager et al. (2011) for the same region. 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 estimates Moho depths of 36-45 km below the southern Cascades arc, specifically 33-37 km below Mt. Shasta, and 38-45 km below the Modoc Plateau. In addition, a teleseismic tomography experiment focusing on 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 below Mt. Shasta to Medicine Lake and 38±4 km below Lassen volcano. These results are summarized in Figure 3. 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 in the nature of the Cenozoic basement to the north and the Paleozoic-Mesozoic basement to the south, as well as variability in the amount of crustal extension experienced by both regions.

The shallowest depths of melting for the primitive basalt samples from all regions in this study appear to hug the location of the Moho as determined from the various geophysical studies. In the northern transect, primitive mantle melts from Crater Lake, unnamed vents to the east, and Newberry volcano originated at such shallow depths according to our calculations that they plot above some or all estimates of the present-day geophysical Moho. A fundamental premise of this work is that the studied primitive basalts originated in the asthenospheric mantle. Therefore our calculated depths can be
used to infer an alternative petrologic Moho. These depths also suggest that the mechanical lithosphere cannot be significantly thicker than the continental crust here. Several other studies of the thickness of the mechanical lithosphere in the overriding plate at subduction zones draw a similar conclusion. 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 parental magma at subduction zones that can be explained if crustal thickness controls the temperature and depth of asthenospheric melting, as observed here.

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

The extremely low S-wave velocities observed in the shallow mantle below a significant portion of the Pacific Northwest support our finding of anomalously warm sub-lithospheric mantle in these regions (Wagner et al., 2010). The high temperatures required to generate magmas at such shallow depths in this region are startling and raise the obvious question, what is driving asthenospheric melting?

We find no evidence for hotspot or plume-related melting processes to generate the <25 Ma basalts or the thin mantle lithosphere in the region of interest. Geochemically, these basalt compositions do not resemble ocean island or flood basalts that are attributed
to hot spots (Figure 2; Carlson, 1984). The basalts studied here are relatively depleted in the high field strength elements, such as U, Th, and Zr, and last equilibrated with the mantle at shallow depths, inconsistent with an origin in a mantle plume (Figure 3).

Furthermore, the HLP HAOTs 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, where very low seismic velocities directly below the crust are restricted to the upper 100 km of the mantle (e.g., Warren et al., 2008; Wagner et al., 2010). Therefore plate motions likely provide the driving force for the observed conditions in the mantle beneath southern Oregon and northern California.

In a study similar to ours, Elkins-Tanton et al. (2001) found the pressures of mantle wedge melting decrease from east to west below Medicine Lake and Mt. Shasta and inferred these pressures of melting paralleled mantle flow in the upwelling limb of corner flow as calculated by Furukawa (1993). In this study we observe a similar pattern in the depth of melting below Mt. Shasta and Medicine Lake volcanoes, but at overall shallower depths and corresponding lower temperatures. Differences in our calculated depths and temperatures of melting are the result of using the new lherzolite melting model of Till et al. (2011), relative to the model of Kinzler and Grove (1992) used in their study. In our northern transect, the depth of melting also appears to decrease from east to west. Other studies of primitive basalts erupted above a subduction zone also 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; Eberle et al.,
2002; Conder 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). Furthermore, the close spatial and temporal association of nominally anhydrous basaltic lavas to products of hydrous flux melting at South Sister, Crater Lake, Newberry, Medicine Lake and Mt. Shasta volcanoes support the interpretation that shared patterns of mantle flow are responsible for their production.

If melting below the HLP and southern Cascades is initiated by upwelling due to corner flow in the mantle wedge, an important consideration is the oblique nature of the North America-Juan de Fuca (NA-JdF) and North America-Gorda (NA-G) plate convergence. A component of horizontal flow in the mantle wedge may therefore be aligned with the SW-direction of NA plate motion, rather than solely perpendicular to the NA-JdF and NA-G plate boundaries. Measurements of seismic anisotropy in the region do in fact suggest an ESE-directed component of mantle flow that becomes more pronounced north of the HLP (Long et al., 2009). The oblique nature of mantle flow vectors relative to our E-W cross-sections could also explain why the minimum melting depth across southern Oregon has such a shallow slope relative to the steep dip of mantle flow lines in the top half of the wedge predicted by geodynamic models with temperature-dependent viscosity (e.g., Condor et al., 2002; Kelemen et al., 2003).

The southern boundary of subducting slab is located just north of ~40°N latitude based on the location of the Mendicino Triple Junction and tomographic images of the slab (James et al., EPSL, in revision). Laboratory models of subduction illustrate a 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
Therefore toroidal flow around the southern edge of the slab could contribute to upwelling and decompression melting in the mantle below northern California. The complex interplay of corner and toroidal flow at the southern edge of the slab may also produce the narrow E-W surface expression volcanism in California relative to a more proto-typical pattern of corner flow producing the wider E-W location of volcanism in southern Oregon.

An alternative interpretation is that anhydrous melting in southern Oregon and northern California is caused by extension-driven mantle upwelling. Recent geological and geophysical studies suggest that the evolution of the NWBR was characterized by voluminous volcanism without significant tectonism, followed by low-magnitude (≤20%) extension along high-angle normal faults (Lerch et al., 2008) that began ca. 12 Ma (see compilation in Scarberry et al., 2010). Extension also occurred within the main Cascades arc during this time period, as indicated by discontinuous north-striking graben faults (Hughes and Taylor, 1986; Smith et al. 1987). Paleomagmatic estimates at 42°N latitude suggest that eastern Oregon has extended ~17% since 15 Ma (Wells and Heller, 1988) and Eager et al. (2011) use crustal thicknesses to estimate that the HLP has experienced ~16% extension since 10 Ma. The depths of melting we calculate for basalts from the Modoc Plateau and unnamed vents near Steens Mountain in Oregon, are inconsistent with the overall pattern of decreasing melting depths from east to west in both of our transects. Because several of these basalts appear to originate at or above the present-day geophysical Moho, they are good candidates for melting triggered by crustal extension. Thus there is evidence for extension-driven asthenospheric upwelling in southern Oregon and northern California, which is likely the result of partitioning of the relative motion.
between the North American plate and the subducting JdF-G plates (e.g., Humphreys, 1994; Atwater and Stock, 1998; Wesnosky, 2005).

Therefore anhydrous melt generation below southern Oregon and northern California, could be the result of, 1) upwelling due to corner flow, 2) toroidal flow around the southern edge of the slab, and/or 3) extension-driven upwelling. Ultimately, all these processes are driven by plate subduction and indicate the <10.5 Ma primitive basalts erupted in the region are the result of purely tectonic processes. As such, it is probable these three processes operate in unison or vary in importance through time. In many geodynamic models of mantle flow at subduction zones, it is extremely difficult to produce the high temperatures required at 30-50 km depth required to generate asthenospheric melts, as constrained in this and other petrologic studies (see review in Kelemen et al., 2003). However, Kelemen et al. (2003) show that if the thermal boundary layer at the base of arc crust (i.e., the mechanical lithosphere) is thinner than originally thought, mantle models incorporating temperature-dependent viscosity and widely accepted values for activation energy and asthenospheric viscosity produce temperature-depth relationships consistent with the petrologic estimates, including those from this study. Therefore, no matter which process triggers mantle melting, the thin nature of the mechanical lithosphere in this region also plays an important role in generating the conditions we observe in the mantle below southern Oregon and northern California, which may be more prevalent at convergent margins than previously thought.
Acknowledgements

The ideas and geochemical data presented in this paper are an outgrowth of the Continental Dynamics Project on Oregon's High Lava Plains and earlier foundational research by a number of the authors. A great deal of thanks is extended to Rick Carlson, Julie Donnelly-Nolan, Matt Fouch, Anita Grunder, Randy Keller, David James, Chris Kincaid, Bob Duncan, Jenda Johnson and all the other scientists and personalities that brought this project to fruition.

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Figure Captions

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 Center (JV), Lassen Volcanic Field (LS), Medicine Lake Volcano (MV), Mount Bailey (MB), Mount Shasta (SH), and South Sister Volcano (SS). 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) (Harris et al., 1991) 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 et al., 2004) and Snake River Plain (Christiansen et al., 2002) 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.

Figure 2. Chemical characteristics of <10.5 Ma primitive basaltic lavas from southern Oregon and northern California selected for this study.

Figure 3. Depth of asthenospheric melt segregation calculated for primitive basaltic lavas from southern Oregon and northern California. White symbols denote basalt samples corrected for the effect of H$_2$O. The base of the green field illustrates the depth of the lithosphere-asthenosphere boundary as predicted from our calculations and the dashed black lines indicate geophysical estimates of the depth of the Moho in the area and the corresponding reference number. Location of the cross-sections as shown in Figure 1.

Figure 4. Cartoon illustrating the three types of asthenospheric flow that could be responsible for the generation of <10.5 Ma primitive basalts in southern Oregon and northern California: (I) ESE-directed corner flow in the mantle wedge, (II) toroidal flow around the southern termination of the subducting slab, and (III) additional upwelling
triggered by Basin and Range crustal-extension. Gray dashed lines represent depth counters for the subducting slab (Harris et al., 1991) and black dashed lines are 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).

Table 1. Primitive Basalt Samples used in this study. Samples pressures in italics were corrected for the effect of H$_2$O as described in the text. Sample pressures were converted to depth using crustal density profiles from regional seismic refraction experiments (Zucca et al., 1986; C. Cox, personal communication, April 2011).
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4. Schmidt et al., 2008
5. Barnes
6. Bacon et al., 1997
7. Elkins-Tanton et al., 2001
8. Till et al., Submitted

*italics corrected for the effect of H2O*
Chapter 5:
A Mechanism for Low Extent Melts at the Lithosphere-Asthenosphere Boundary
Abstract

Recent studies have imaged sharp vertical drops in shear wave velocity at the lithosphere-asthenosphere boundary (LAB). In some regions, the magnitude of the negative velocity gradient at the LAB is too large to be explained by changes in temperature alone. This study demonstrates that small amounts of partial melt in the shallow asthenosphere are a viable model for this sharp seismic boundary. In particular, we examine melting in the upper asthenosphere at the edge of thick cratonic lithosphere, using the example of eastern North America where a sharp LAB velocity gradient has been observed. Finite element modeling of asthenospheric flow at an abrupt lateral decrease in lithosphere thickness indicates that this geometry, together with lateral plate motions, produces edge-driven convection and asthenospheric upwelling at the continental margin. A key component of this work is a comparison of the locations and extents of melting produced by using different models for the depression of the peridotite solidus with varying H$_2$O content. In addition, we develop a simplified parameterization of the H$_2$O-undersaturated peridotite solidus for a constant degree of H$_2$O saturation in nominally anhydrous minerals. The patterns of mantle flow produced by our numerical modeling and various solidus parameterizations predict less than 0.1 wt% to 2.8 wt% (0.01-3.3 vol%) melting at depths between 102 and 126 km for an asthenosphere with a mantle potential temperature of 1350°C and 150 ppm H$_2$O, or between 91 km and a maximum of 200 km for a mantle at 1350°C and 450 ppm H$_2$O. If the asthenosphere has a mantle potential temperature ≤1340°C or contains less than 150 ppm H$_2$O at 3 GPa, no melting will occur. This process of generating melt in the asthenosphere to produce a sharp vertical velocity gradient at the LAB is viable in other locations where convective upwelling occurs in the
shallow asthenosphere although it is dependent on asthenospheric potential temperature, composition, and H$_2$O content. Because the asthenosphere may be heterogeneous in composition and H$_2$O content, the onset of melting below the LAB may fluctuate with time and space, as may the magnitude of the shear velocity drop at the LAB.

1. Introduction

In the most general sense, the lithosphere-asthenosphere boundary (LAB) is defined as the boundary between the rigid material that composes the Earth’s tectonic plates (i.e., the lithosphere) and the underlying material which undergoes solid-state creep such that it exhibits fluid-like behavior on geologic times scales (i.e., the asthenosphere). Yet the first-order physical and chemical properties of the mantle, which change at the LAB to produce this rheological boundary, are poorly understood.

The LAB is marked by a negative gradient in seismic velocity from the fast lithosphere to the slow asthenosphere. Slow velocities in the asthenosphere have been explained with a variety of models, some of which employ temperature gradients with or without grain size variations [e.g., Faul and Jackson, 2005; Stixrude and Lithgow-Bertelloni, 2005; Priestley and McKenzie, 2006], while others invoke a contrast between a dry depleted lithosphere and a hydrated fertile asthenosphere [e.g. Hirth and Kohlstedt, 1996; Karato and Jung, 1998; Gaherty et al., 1999] or the presence of melt within the asthenosphere [e.g. Anderson, 1989; Kawakatsu et al., 2009].

The sharpness of the velocity gradient at the LAB in many regions poses a challenge for purely thermal models [Fischer et al., 2010]. In the last decade, resolution of the LAB has been improved with a variety of seismic data, including teleseismic phases such as $Ps$ and $Sp$ [Farra and Vinnik, 2000; Levin and Park, 2000; Li et al., 2000;
Collins et al., 2002; Oreshin et al., 2002; Li et al., 2004; Rychert et al., 2005; Vinnik et al., 2005; Kumar et al., 2005a; Kumar et al., 2005b; Chen et al., 2006; Kumar et al., 2006; Mohsen et al., 2006; Sodoudi et al., 2006; Hansen et al., 2007; Heit et al., 2007; Kumar et al., 2007; Li et al., 2007; Rychert et al., 2007; Wittlinger and Farra, 2007; Chen et al., 2008; Snyder, 2008; Hansen et al., 2009; Rychert and Shearer, 2009; Abt et al., 2010; Ford et al., 2010], ScS reverberations [Bagley and Revenaugh, 2008] and multiple S surface reflections [Tan and Helmberger, 2007] (reviews may be found in Rychert et al. [2010] and Fischer et al. [2010]). In some cases, the LAB velocity gradient has been explicitly modeled. Rychert et al. [2005; 2007] utilized inversions of P-to-S (Ps) and S-to-P (Sp) receiver functions to resolve a 5-10% decrease in shear wave velocity over less than 11 vertical km at depths near 100 km beneath the northeastern U.S. and southeastern Canada. Using experimentally-derived relationships between temperature and shear velocity [Faul and Jackson, 2005] and a minimum olivine grain size of 1 mm [Evans et al., 2001], to produce this velocity gradient with temperature alone, a temperature gradient of more than 20°C/km is required at the LAB [Rychert et al., 2007]. In contrast, temperature gradients are less than 10°C/km in numerical models for comparable tectonic settings (the boundary of thick cratonic and thinner continental lithosphere) in which viscosity varies with temperature and pressure but is not affected by other factors such as composition [King and Ritsema, 2000; Cooper et al., 2004]. Moreover, in these and other models for this type of tectonic setting [Lenardic and Moresi, 1999; Korenaga and Jordan, 2002], thermal gradients from the lithosphere to the asthenosphere are distributed over 60-80 km or more, as opposed to the much smaller depth range spanned by the gradients in seismic velocity. Rychert et al. [2005; 2007;
therefore conclude that a mechanism that involves more than temperature is required to explain the observed velocity gradients: a more hydrated asthenosphere, perhaps in combination with a downward decrease in chemical depletion or grain size or with a coincident change in anisotropy, or simply a small amount of partial melt in the asthenosphere.

Evidence for sharp LAB velocity gradients has also been found in other regions. Using $Ps$ phases in eastern China, Chen et al. [2006] inferred shear velocity decreases of 3-7% over 10 km or less and reached conclusions similar to those of Rychert et al. [2005; 2007]. Kawakatsu et al. [2009] obtained velocity drops of 7-8% and argued for horizontal melt layers in the asthenosphere beneath the Pacific and Philippine Sea plates. Using $Sp$ phase alone, Li et al. [2007], Abt et al. [2010], and Ford et al. [2010] were able to limit LAB velocity gradients to depth ranges of less than 30-40 km beneath the western U.S., large portions of the non-cratonic eastern U.S., and eastern Australia. Although constraints on vertical velocity gradients from $Sp$ phases alone are looser than those possible with $Ps$ data, 30-40 km is still much smaller than the depth range of lithosphere-asthenosphere thermal gradients in numerical models [Lenardic and Moresi, 1999; King and Ritsema, 2000; Korenaga and Jordan, 2002; Cooper et al., 2004]. Thus these results also point to compositional differences between the lithosphere and asthenosphere or the presence of partial melt in the asthenosphere.

To further explore the origins of the LAB, this study examines the case of the LAB below eastern North America. In this region, surface wave tomography studies indicate that the lithosphere has a thickness of approximately 200 km below the westernmost Appalachians and beneath the craton to the west [van der Lee, 2002; Li et
al., 2003; Nettles and Dziewonski, 2008; Bedle and van der Lee, 2009; Yuan and Romanowicz, 2010]. Moving east, the lithosphere abruptly thins to thicknesses of 90-110 km beneath New England, based on $Sp$ and $Ps$ analyses [Rychert et al., 2005; Rychert et al., 2007; Abt et al., 2010] and most surface wave tomography [van der Lee, 2002; Li et al., 2003; Nettles and Dziewonski, 2008; Bedle and van der Lee, 2009] (in one exception, Yuan and Romanowicz [2010] argue for a deeper LAB beneath New England based on vertical variations in azimuthal anisotropy). An abrupt eastward decrease in lithospheric thickness together with a WSW absolute motion for the North American plate [Gripp and Gordon, 2002] is ideal to produce edge-driven convection. The work of King and Anderson [1995; 1998] originally modeled small-scale convection in an isothermal mantle resulting from discontinuities in the thickness of the lithosphere (i.e., edge-driven convection). Newer models of edge-driven thermal convection suggest it could be a local mechanism for melt formation or an alternative explanation to mantle plumes for intraplate volcanism [King and Ritsema, 2000; Korenaga and Jordan, 2002].

We have conducted numerical experiments using an abruptly thinning lithosphere to test whether edge-driven convection can produce a small melt fraction at depths below the seismically-imaged LAB [Rychert et al., 2005; Rychert et al., 2007; Abt et al., 2010]. We examine the range of compositions and $H_2O$ contents required to produce melting, given the range of calculated mantle flow patterns, by comparing a variety of solidi relevant to the asthenosphere, including the anhydrous peridotite solidus and several parameterizations for peridotite solidus depression with varying $H_2O$ content. These models for the melting behavior of peridotite in the presence of $H_2O$ express solidus depression as a function of the concentration of water in the melt based on the bulk
partition coefficient for H_2O between solid peridotite and melt. In addition, we derive a simplified parameterization for solidus depression at a constant degree of H_2O-saturation in nominally anhydrous minerals, which is only a function of pressure. Our parameterization is based on the method of Hirth and Kohlstedt [1996] and includes new experimental data on the H_2O-saturated solidus and on H_2O partitioning between nominally anhydrous peridotite minerals and silicate melts. The parameterization yields comparable melting temperatures for a given H_2O content, between 2 and 7 GPa, to pre-existing H_2O-undersaturated melting models, including those of Hirschmann et al. [2009], Katz et al. [2003] and a model for olivine liquidus depression by Medard and Grove [2008]. The similarity of these expressions for solidus depression is not surprising given they are all derived from or calibrated with high-pressure melting experiments on peridotite or primitive basalts. We also explore the implications of the similarity of these models for solidus depression in the presence of H_2O and their implications for the presence of small amounts of partial melt in the asthenosphere in generalized tectonic settings.

2. Methods

2.1 Solids Used in Melting Calculations

For our modeling we use the anhydrous peridotite solidus parameterization of Hirschmann [2000], which is based on an up-to-date compilation of the experimental data. The experiments used to bracket the solidus are depicted in Figure 1, along with additional experiments at 0.9 - 1.6 GPa by Kinzler and Grove [1992b] and at 1 atm by Grove and Juster [1989]. As noted by Hirschmann [2000], the bulk composition of a
peridotite (specifically the alkali content, Mg\# and cpx mode) plays an important role in
determining the solidus temperature. Therefore we use the Hirschmann [2000] equation
that has been filtered to exclude enriched peridotites and includes the experiments
conducted prior to 1988, which is as follows,

\[
T(\degree C) = -5.104 P^2 - 132.899 P + 1120.611
\]

(1)

where \(P\) is in GPa. We choose to include experiments conducted prior to 1988 because of
the lack of more recent experiments to constrain the solidus at <1 GPa. The above
equation yields a 1 atm solidus temperature of 1120.661\degree C, whereas the parameterization
filtered for enriched and depleted peridotites, using only data that post-dates 1988, yields
a 1 atm solidus temperature of 1107\degree C. Using the equation of Grove and Juster [1989]
for the variation in the 1 atm anhydrous peridotite solidus temperature with composition,
typical glass compositions from MORB melting experiments at 0.9 GPa by Kinzler and
Grove [1992b] predict 1 atm solidus temperatures of 1150 - 1177\degree C; H\textsubscript{2}O-saturated
peridotite melting experiments on a Hart and Zindler [1986] primitive mantle
composition by Grove et al. [2006] at 1.2 GPa and Till et al. [2007] at 3.2 GPa predict 1
atm solidus temperatures of 1196\degree C and 1205\degree C, respectively. Therefore, the
Hirschmann [2000] parameterization with an intercept of 1120.611\degree C most closely
matches the constraints from 1 atm and other low pressure melting experiments. An
updated version of the Hirschmann [2000] parameterization for the pressures beyond that
required in our models (>10 GPa) can be found in Hirschmann et al. [2009].

H\textsubscript{2}O-saturated peridotite melting experiments by Grove et al. [2006] from 1.2 to
3.2 GPa and Till et al. [2007] from 3.2 to 6 GPa are illustrated on Figure 1 and are the
source of the H\textsubscript{2}O-saturated solidus used for our parameterization. All regressions for the
following equations were calculated using the method of least squares. Peridotite solidi used here represent the temperature of the first melt at a given pressure in a typical undepleted upper mantle composition. In general, the solidi are suitable for a typical MORB-source upper mantle composed of approximately 60% olivine, 15% clinopyroxene, 15% orthopyroxene and 0-10% garnet (depending on pressure). A fourth-order polynomial fit most closely matches the H₂O-saturated experimental data (R²=0.9816), with pressure in units of GPa. This solidus is applicable over the range of experimental data it is derived from, which is 0 to 6 GPa.

\[
T(°C) = -0.0842P^4 - 1.7404P^3 + 36.777P^2 - 191.69P + 1120.7
\]  

Although experimental liquids from experiments by Grove et al. [2006] at 1.2 GPa and Till et al. [2007] at 3.2 GPa predict 1 atm solidus temperatures of ~1200°C, we prefer to use the 1120.7°C value from Hirschmann [2000], as it is infeasible that the anhydrous and H₂O-saturated solidus cross at low pressures. For extrapolation to pressures between 6-12 GPa, a second order polynomial better describes the predicted (i.e., not tested experimentally) solidus behavior relative to the fourth order polynomial but does not fit the data as accurately at pressures less than 6 GPa (R²=0.9647). This quadratic is used to determine the nominally anhydrous mineral surface in three dimensions as discussed below.

\[
T(°C) = 16.777P^2 - 149.5P + 1120.7
\]
Partial melting experiments from 3-7.5 GPa by Kogiso et al. [2003] are the basis for the garnet pyroxenite solidus used in our calculations (Figure 1). In our melting calculations this parameterization is extrapolated to surface conditions.

\[ T(\degree C) = -3.4524P^2 + 120.95P + 1096.8 \]  

(4)

H₂O-undersaturated peridotite melting is not as well understood experimentally as the anhydrous and H₂O-saturated cases. To calculate H₂O-undersaturated solidi it is necessary to know the H₂O storage capacity of nominally anhydrous mantle minerals. Here we define the H₂O storage capacity as the maximum abundance of H₂O that can be accommodated in nominally anhydrous mantle minerals (we assume that all of the H₂O resides in olivine, clinopyroxene, orthopyroxene and garnet), before the stabilization of hydrous minerals, and use the H₂O-saturated peridotite solidus as a reference P-T path for the upper bound on the storage capacity of nominally anhydrous minerals, after the studies of Hauri et al. [2006] and Hirschmann et al. [2005]. Techniques to measure the partition coefficients for H₂O in nominally anhydrous minerals have largely developed within the last 15 years [e.g., Bell and Rossman, 1992; Hirth and Kohlstedt, 1996; Koga et al., 2003; Hirschmann et al., 2005]. Here we use the pressure-dependent partition coefficients and mineral modes of upper mantle minerals near the LAB presented by Hirschmann et al. [2009] and references within.

These partition coefficients can then be used to calculate the maximum concentration of H₂O dissolved in a peridotite melt at incipient melting using the expression for batch melting at zero melt fraction,

\[ C_{H₂O}^{\text{Solids}} = \frac{C_{H₂O}^{\circ}}{D_H^{\text{perd/lig}}} \]  

(5)
where \( C''_{\text{H}_2\text{O}} \) is the bulk concentration of \( \text{H}_2\text{O} \) in the peridotite, \( D_{\text{perid}/\text{liq}} \) is the bulk partition coefficient for \( \text{H} \) between peridotite solid and melt and \( C^\text{Solids}_{\text{H}_2\text{O}} \) is the concentration of \( \text{H}_2\text{O} \) dissolved in the first infinitesimal fraction of melt. Peridotite solidus depression was then calculated using both the cryoscopic approach benchmarked against hydrous melting experiments of Hirschmann et al. [2009] (illustrated in their Figure 10), and the mathematical formulation of Katz et al. [2003] based on constraints from thermodynamic modeling and high-pressure experiments (see their equation 16). Solidus depression follows a power-law relationship with \( C^\text{liq}_{\text{H}_2\text{O}} \) in both of these expressions due to the greater depolymerizing effect and larger quantity of hydroxyl ions in silicate melt compositions at low water contents (1-2 wt.% \( \text{H}_2\text{O} \)), relative to molecular water [Stolper, 1982; 1989; Inhinger et al., 1999]. Therefore, both equations account for the larger effect of \( C^\text{liq}_{\text{H}_2\text{O}} \) on solidus depression at low total water contents and the decreasing proportion and effect of \( C^\text{liq}_{\text{H}_2\text{O}} \) on solidus temperature at higher total water contents. The solidi for a bulk peridotite with \( C''_{\text{H}_2\text{O}} \) of 150 ppm and 450 ppm \( \text{H}_2\text{O} \), shown in Figure 2, were calculated by subtracting the solidus depression (\( \Delta T \)) calculated from Hirschmann et al. [2009] and Katz et al. [2003] from the anhydrous solidus (given in (1)).

Medard and Grove [2008] conducted experiments to determine the effect of \( \text{H}_2\text{O} \) on the liquidus temperature of olivine-saturated primitive melts. They conclude that liquidus depression for primitive basaltic melts can be approximated as a linear function of \( C^\text{liq}_{\text{H}_2\text{O}} \) up to 1.3 wt \% \( \text{H}_2\text{O} \) and as a third-order polynomial up to 12 wt. \% \( \text{H}_2\text{O} \). Furthermore, for more silica-rich melt compositions, including low-degree mantle melts,
they find that the effect of H$_2$O is smaller than for basaltic melt, and can be better approximated as a linear function. For comparison to the previous two expressions for solidus depression, $\Delta T$ for bulk peridotite melting was calculated using both the linear and third-order expressions of Medard and Grove [2008] for liquidus depression. The third-order polynomial produces $\Delta T$ within ±9°C of the expressions for solidus depression of Katz et al. [2003] and Hirschmann et al. [2009] at 150 ppm H$_2$O and the linear approximation yields $\Delta T$ within ±3°C at low $C_{H_2O}^{liq}$ (< 2 wt. %) and ±11°C at higher $C_{H_2O}^{liq}$.

For comparison to the solidus depression at constant bulk H$_2$O contents, we have developed a simplified parameterization for solidus depression at a constant degree of H$_2$O-saturation in nominally anhydrous minerals, which is only a function of pressure. Our parameterization is based on the method of Hirth and Kohlstedt [1996] and includes new experimental data on the H$_2$O-saturated solidus. Mantle melting in the presence of H$_2$O can be described using a phase diagram for a constant pressure such as that shown in Figure 3 after Silver and Stolper [1985] and Grove et al. [2006], with bulk H$_2$O content increasing on the x-axis and temperature increasing on the y-axis. The green curve illustrates the maximum storage capacity of nominally anhydrous peridotite minerals based on data from Hauri et al. [2006]. The blue curve illustrates the simplified melting boundary (i.e., the liquidus) for peridotite, extending from a eutectic-like H$_2$O-saturated solidus with the maximum H$_2$O content of the melt based on H$_2$O solubility in silicate melt [Grove et al., 2006] to an anhydrous liquidus [Zhang and Herzberg, 1994].

We calculate the H$_2$O-undersaturated solidi for our melting calculations by fitting a line to the nominally anhydrous curve between the anhydrous solidus and the H$_2$O-
saturated solidus temperatures, for a given pressure, these equations are given in Table 1. We approximate the nominally anhydrous curve as a line based on an equivalent formulation for the liquidus in the thermodynamic model for hydrous silicate melts developed by Silver and Stolper [1985] for diopside-H$_2$O and albite-H$_2$O, as well as the similarity of the linear expression for liquidus depression with H$_2$O of Medard and Grove [2008] to the power-law expressions for solidus depression of Katz et al. [2003] and Hirschmann et al. [2009]. The model of Silver and Stolper [1985] is derived from the assumption of ideal mixing of hydroxyl groups, H$_2$O molecules and oxygen in the melt, which is an adequate first-order approximation at low H$_2$O contents but not likely adequate for higher-order models (see discussion of thermodynamic models of silicate melts in Medard and Grove [2008]). Following the model of Silver and Stolper [1985], as well as the work of Hirth and Kohlstedt [1996] and Hauri et al. [2006], we assume an activity of unity for H$_2$O in the fluid. Therefore at the saturation of the nominally anhydrous minerals, the peridotite has a water-activity (a$_{H2O}$) equal to unity and melts at the H$_2$O-saturated solidus temperature. When a peridotite contains no H$_2$O, it has an a$_{H2O}$ equal to zero and melts at the anhydrous solidus temperature. And when the peridotite contains an amount of water less than that required for mineral saturation, the melting temperature scales linearly with water-content between 0<a$_{H2O}$<1. To assume an activity of unity for water in the fluid is a simplifying assumption, as the amount of dissolved silicate in hydrous fluids increases with increasing pressure [e.g., Stalder et al., 2001; Hermann et al., 2006], causing the activity of water to constantly decrease with increasing pressure (i.e., max a$_{H2O}$<1). Therefore our estimates of the temperature of the H$_2$O-undersaturated solidus at the higher end of the range of pressures examined (~3-6...
GPa) are maximum estimates, and in fact mantle peridotite may reach water-saturation at even lower temperatures for a given pressure. A simple H$_2$O-undersaturated melting scenario for a peridotite with 150 ppm H$_2$O (0.015 wt%) is illustrated by the red points in Figure 3. This H$_2$O-undersaturated peridotite will melt at a temperature of 1409°C, where the H$_2$O content intersects the nominally anhydrous boundary. A peridotite containing >1500 ppm H$_2$O (0.15 wt%) at 3 GPa will melt at the temperature of the H$_2$O-saturated solidus (825°C) and a completely anhydrous peridotite will melt at the anhydrous solidus (1473°C).

Here we choose four relative values for peridotite mineral saturation (represented as a$_{H2O}$ = 0.1, 0.3, 0.5, and 0.7) and calculate the temperature of the corresponding H$_2$O-undersaturated solidus and how it varies with pressure. Each of the values of mineral saturation corresponds to an H$_2$O content determined from the partitioning data from Hauri et al. [2006], which changes with pressure, as H$_2$O solubility is pressure-dependent as described in Table 2. For example, in the original melting scenario illustrated by the red points in Figure 3, the peridotite contains 50 ppm H$_2$O at 1 GPa, 150 ppm H$_2$O at 3 GPa (the pressure illustrated in Fig. 3) and 300 ppm at 6 GPa, which corresponds to an a$_{H2O}$=0.1. By choosing to parameterize our model in terms of constant H$_2$O activity, we do not mean to imply that there is a depth gradient in the H$_2$O content of the asthenosphere. Rather by holding H$_2$O activity constant in these expressions, it facilitates a qualitative assessment of the degree of H$_2$O saturation required to trigger melting at a given depth in the asthenosphere.

\[
A_{H2O} = 0.1 : T(°C) = -2.848P^2 + 104.312P + 1120.7 \tag{5}
\]

\[
A_{H2O} = 0.3 : T(°C) = 1.659P^2 + 47.152P + 1120.7 \tag{6}
\]
The water content of melts at the H$_2$O-saturated eutectic can be calculated using the expression

\[ 7290P - 810T - 24600H_2O + 1093500 = 0 \]  

(9)

from Grove et al. [2006] where \( P \) is pressure in kbars, \( T \) is the temperature of the H$_2$O-saturated solidus in °C and \( H_2O \) is the water content of the melt in wt%.

2.2 Numerical Models of Mantle Flow and Temperature

To explore the implications of the melting model, we applied it in the context of a model of possible mantle flow beneath the northeastern U.S.. Two-dimensional Cartesian numerical experiments on convection were conducted using the finite-element code FEMcont [Kelemen et al., 2003; Zaranek et al., 2004]. The aim of these experiments was to produce predictions of asthenospheric flow patterns and temperature fields near the LAB at the edge of a lithospheric keel (Fig. 4). The model box for the numerical experiments consists of a mesh of 201 by 101 nodes. The top of the model box has a no-slip boundary, where \( v_x = v_z = 0 \), and temperature is held fixed at \( T=0 \). The bottom of the box has a fixed temperature and velocity boundary conditions such that \( v_x = V \) and \( v_z = 0 \). The vertical sides of the computational domain allow flow-through, such that \( dv_x/dx = dv_z/dz = 0 \). The model material is an infinite Prandtl number, Boussinesq fluid. In our Eulerian code, the model asthenosphere moves relative to the fixed reference frame of the continental lithosphere. Thus absolute plate motion for eastern North America is represented as a “mantle wind” in the asthenosphere and applied as a
constant horizontal velocity boundary condition at the base of the mantle in the model, as illustrated in Figure 4. The models were run with mantle wind velocities from 0-3 cm/year in a west-to-east direction to represent the range of predictions for the east-west component of North American plate motion [Conrad et al., 2004].

Initial temperatures are defined linearly within the lithosphere and as a mantle potential temperature within the asthenosphere. The experiments used an LAB profile that approximates the results of Sp and Ps analyses [Rychert et al., 2005; Rychert et al., 2007; Abt et al., 2010] and some surface wave tomography studies [van der Lee, 2002; Li et al., 2003; Nettles and Dziewonski, 2008; Bedle and van der Lee, 2009] across the edge of the craton in Pennsylvania and the continental margin in New England. The model includes a 200 km thick continental lithosphere, a 90 km thick lithosphere beneath the continental margin and a horizontal transition between these two lithospheric thicknesses of 50 km. The width of this horizontal transition is difficult to precisely constrain, but a combination of regional surface wave tomography [Li et al., 2003] and Sp and Ps constraints [Rychert et al., 2005; Rychert et al., 2007; Abt et al., 2010] indicate that values of 50-100 km are reasonable. Mantle flow patterns at the continental margin were affected by changing the thickness of the continental lithosphere and the width of the horizontal transition in lithosphere thickness (Fig. 5), but not continent length.

The model solves non-dimensional equations composed of the conservation of mass, momentum, and energy. The thermal buoyancy of the model fluid was determined by a dimensionless Rayleigh number,

$$Ra = \frac{\rho g \alpha \Delta T h^3}{\mu \nu}$$

(10)
where \( p \) is fluid density (kg/m\(^3\)), \( g \) is gravitational acceleration (m/s\(^2\)), \( \alpha \) is thermal expansivity (K\(^{-1}\)), \( \Delta T \) is temperature difference across the region in question (K), \( h \) is the height of the model box (m), \( \mu_0 \) is the reference fluid viscosity (Pa\(\cdot\)s) when \( T=T_p \), and \( \kappa \) is thermal diffusivity (m\(^2\)/s). Below a critical Rayleigh number, conduction is more efficient than convection. Rayleigh numbers in the numerical experiments were varied over the range of the following variables, where \( p=3300 \) kg/m\(^3\), \( \alpha=2.4-3\times10^{-5} \) K\(^{-1}\), \( \Delta T=1573-1673 \) K, \( h=410-660 \) km, \( \mu_0 \) for the asthenosphere is \( 10^{18-21} \) Pa\(\cdot\)s and \( \kappa=10^{-6} \) m\(^2\)/sec. The lower bound on the reference viscosity of the upper mantle is based on deformation in response to local loads (i.e., earthquakes and lake filling) in the western U.S. [e.g., Kaufmann and Amelung, 2000], whereas the higher estimates are from glacial isostatic adjustment models at cratonic or global scales [Lambeck and Johnston, 1998; Peltier, 2001b; Dixon et al., 2004].

Asthenosphere viscosity in the model deviates from the reference viscosity according to temperature in an Arrhenius relationship,

\[
\mu = \mu_0 \times e^{-\frac{Q}{RT} (\frac{T_p - T}{T_p})}
\]

(11)

where \( \mu_0 \) is the reference fluid viscosity (Pa\(\cdot\)s), \( T_p \) is the mantle potential temperature and \( Q \) is activation energy. \( Q/RT \) varied between 30 (a small increase in temperature resulting in a large decrease in viscosity) and 15 in our experiments by using experimentally-derived values of \( Q \) that vary from \(~200\) kJ/mol for diffusion creep to \(~500\) kJ/mol for dislocation creep in the mantle [Karato and Wu, 1993]. To create a more viscous lithosphere, a multiple of the asthenosphere reference viscosity was applied to the lithosphere; lithospheric viscosity was allowed to deviate from the reference value again using equation 11. No evidence exists for lithospheric instability beneath eastern North
America in the last 100 m.y., therefore a relative lithosphere-asthenosphere viscosity of 1000 was used in the numerical experiments as it produced the most stable lithosphere geometry [Lenardic and Moresi, 1999]. Lower relative viscosities produced lithospheric instabilities within 20 million years.

2.3 Melt Calculations: Criteria for Melt Production

After solving for the temperature and velocity fields using FEMcont, we identify regions of upwelling and compare the temperature and pressure of each node to the solidi relevant to the asthenosphere described in section 2.1 to determine if melting could occur. Particular attention is paid to vertical slices through the model box with significant convective upwelling at the edge of the thicker continental lithosphere. The mantle is considered to contain partial melt if it is above the solidus (i.e., temperature of first melt) in a given time step. No compositional information is built into the FEMcont numerical code. Composition is only expressed in the solidus chosen for the melting calculations. We acknowledge that interaction between melting and solid flow are not considered explicitly here, as the melting calculations are not fed back into viscosity. To incorporate these effects in the computation requires a significant step in the complexity of the model, which we have chosen to exclude in our effort to simply demonstrate the plausibility of mantle melting near the LAB.

The melting calculations were run for a range of mantle potential temperatures (T_p=1300-1550°C) based on olivine-liquid equilibria and olivine phenocryst compositions in primary magmas from a range of tectonic settings [Kinzler and Grove, 1992a; Putirka, 2005; Herzberg et al., 2007]. Anderson [1982] pioneered the idea that
continents can act to insulate the mantle by preventing dissipation of heat from the Earth’s interior, causing broad thermal anomalies, sometimes known as “bottom heating.” Today this idea has been substantiated in modern 2D and 3D mantle convection models [Gurnis, 1988; Lowman and Jarvis, 1996; Lowman and Gable, 1999; Phillips and Bunge, 2005; Korenaga, 2007; Phillips and Coltice, 2010]. The possibility of subcontinental insulation was considered in our models by calculating the maximum range of additional heat retained below the North American continent [-20-40°C; Phillips and Bunge, 2005] and applying this additional heat to mantle that has traveled under the continent before upwelling.

To calculate melting, we used mantle adiabats ranging from 0.29-0.39°/km (10-13°/GPa) to reflect different estimates of the coefficient of thermal expansion and the specific heat of the asthenosphere. Changing the adiabatic gradient had only a small effect on the location of melting in the asthenosphere; a mantle adiabat of 13°/GPa located the maximum depth of melting at 4-8 km deeper than an adiabat of 10°/GPa for a model with the same mantle potential temperature and initial conditions. No other effect on the location and H2O content of melting was observed based on varying the adiabatic gradient alone.

Once a mantle potential temperature, adiabatic gradient, composition and H2O content are assigned, the occurrence of asthenosphere melting in our calculations depends only on pressure. If the asthenosphere is above the solidus at a given pressure, the presence of edge-driven convection cells will only increase the quantity of melt produced over time.
3. Results

The calculated asthenosphere flow fields produce upwelling regions viable for adiabatic decompression melting at the margin of the lithospheric keel. The lateral extent of upwelling is affected by both the width of the step transition from the keel to thinner lithosphere and the thickness of the keel itself. While in eastern North America the keel appears to be ~200 km thick, and the transition to thin lithosphere to occur over ~50 km, both thinner and thicker keels and transition widths will produce upwellings with the potential for melting. Thicker keels and more abrupt transitions to thinner lithospheres produce larger convective upwellings, extending almost 200 km across the thin lithosphere (see Fig. 5). When the transition from the keel to thinner lithosphere is more gradual (approaching a fluid dynamic shape), the convective eddy is located more under the transition step from keel to thinner lithosphere, and extends less under the thin lithosphere. When the mantle wind velocity is 0 cm/year, convective upwelling still occurs via simple edge-driven convection immediately adjacent to the change in lithospheric thickness. As the mantle wind velocity is increased, the location of maximum upwelling migrates upwind (Fig. 7). Thus the magnitude of the mantle wind also acts to relocate the site of the greatest chance of melting adiabatically. The size of the melting region predicted by a given solidus also decreases with an increase in the velocity of the mantle wind.

Our modeling and melting calculations suggest melting of anhydrous peridotite will not occur in the asthenosphere for mantle potential temperatures less than 1425°C. The location of melting produced by the 150 ppm H₂O solidi calculated using the models of Hirschmann et al. [2009] (Fig. 6e), Katz et al. [2003] (Fig. 6d) and Medard and Grove...
[2008] (Fig. 6c), as well as the solidus for aH2O=0.1 or 10% mineral saturation (Fig. 6b) are virtually identical due to their overlap in temperature between 2 to 7 GPa (Fig. 2, Table 3). Our melting calculations with any of these solidi suggest a mantle potential temperature greater than approximately 1350°C is required to produce melting near the LAB. The depth of melting in our experiments with these solidi for an asthenosphere with a potential temperature of 1350°C is between 102-126 km, with each solidus producing a melting region that is 21 or 24 km in height. Melting in our numerical models occurs over a narrower range of pressures (102-121 km vs. 99-141 km) than predicted by the original Hirth and Kohlstedt [1996] solidus at H2O concentrations of 150-200 ppm (Fig. 6a, Table 3).

The location of melting for experiments with the same mantle potential temperature and initial conditions varies more for the calculated 450 ppm H2O solidi, relative to those for 150 ppm H2O, because of more substantial differences in solidus temperature between the models at the relevant pressures. The 450 ppm H2O solidi of Katz et al. [2003] (Fig. 6i) and Medard and Grove [2008] (Fig. 6h) predict melting between ~91-130 km for a mantle potential temperature of 1350°C (also see Table 3). Our solidus for an aH2O=0.3 or 30% mineral saturation (Fig. 6g) is closest in temperature to the calculated 450 ppm H2O Hirschmann et al. [2009] solidus (Fig. 6i), and these two models predict similar melting regions between ~91-200 km for an asthenosphere with a potential temperature of 1350°C.

Composition (i.e., peridotite vs. garnet pyroxenite) has a small effect on the temperature of the solidus at low pressures in this model. The anhydrous garnet
pyroxenite solidus [Kogiso et al., 2003] triggers a small region of melting just below the LAB at mantle potential temperatures $\geq 1375^\circ$C.

4. Discussion

4.1 The LAB beneath eastern North America

As is clear from the temperature structure of the flow models presented here, vertical gradients in temperature alone are too gradual to produce an LAB consistent with seismic constraints. In our models, temperature gradients from the lithosphere to the asthenosphere are distributed over more than 60 km in depth, and the maximum temperature gradient does not exceed 14.5$^\circ$C/km. In contrast, a temperature gradient of more than 20$^\circ$C/km over less than 11 km is required to produce the combination of $Sp$ and $Ps$ phases observed at the LAB beneath stations in the northeastern U.S. and southeastern Canada [Rychert et al., 2007]. $Sp$ phases from the LAB observed more widely in the eastern U.S. [Abt et al., 2010] indicate a velocity gradient that is localized over less than 40 km in depth, and even those looser constraints are not met by an LAB that is created purely by temperature. This conclusion is consistent with earlier comparisons to the work of models by King and Ritsema [2000], Lenardic and Moresi [1999], Korenaga and Jordan [2002], and Cooper et al. [2004].

Small degrees of asthenosphere melting may be a viable mechanism to explain the observed drop in shear wave speeds across the LAB in eastern North America. The temperature and H$_2$O contents of the upwelling asthenosphere at the edge of the continental lithosphere are the main factors that determine if melting will occur in our numerical experiments. These asthenosphere source conditions are dependent on the
direction and pattern of the mantle wind (equivalent to plate motion direction) under North America. Here we examine a range of possible source conditions because of the uncertainty of the pattern of a North American mantle wind.

Asthenospheric mantle material saturated with H$_2$O is most likely to melt since its solidus is at the lowest temperature for a given pressure. However it is unlikely the $>500-3000$ ppm H$_2$O required to produce melt at H$_2$O-saturated conditions is present in the asthenosphere, except perhaps above a downgoing oceanic slab at a subduction zone. However, moderate amounts of H$_2$O are likely in the asthenosphere, perhaps heterogeneously distributed throughout [Bell and Rossman, 1992]. Estimates of H$_2$O content for a normal MORB-source upper mantle converge on a value of $100 \pm 50$ ppm [Dixon et al., 1988; Michael, 1988; Hirth and Kohlstedt, 1996; Saal et al., 2002; Salters and Stracke, 2005; Workman and Hart, 2005] and $300-900$ ppm H$_2$O for enriched MORB- and OIB-source upper mantle [Jambon and Zimmermann, 1990; Sobolev and Chaussidon, 1996; Dixon et al., 1997; Aubaud et al., 2005; 2006]. One plausible scenario to produce melting in our numerical experiments is a mantle potential temperature of $\sim 1350^\circ$C (or a potential temperature of 1310-1330$^\circ$C combined with 20-40$^\circ$C of subcontinental insulation) for an asthenosphere with $\sim 150$ ppm H$_2$O. If the mantle is significantly colder or contains significantly less H$_2$O, no melting will occur. If there are small regions of higher H$_2$O content (e.g., $\sim 450$ ppm H$_2$O) within the convecting asthenosphere, for example asthenosphere that was hydrated during the subduction of the Farallon slab or another subduction event, melting is possible at mantle potential temperatures as low as $1190^\circ$C for our parameterization at 30% mineral saturation, $1200^\circ$C for the Hirschmann et al. [2009] 450 ppm H$_2$O solidus, and $1215^\circ$C for the 450
ppm H_2O Katz et al. [2003] solidus. Likewise, CO_2 combined with H_2O will increase the likelihood of small degrees of partial melting in the MORB-source upper mantle [Dasgupta et al., 2007]. Alternatively, if there is garnet pyroxenite in the convecting asthenosphere (e.g., remnant pieces of subducted slab), small regions of melting occur at mantle potential temperatures ≥1375°C. Mantle potential temperatures above 1420°C (equivalent to what we estimate for plume source regions) are required to melt an anhydrous asthenosphere source with no H_2O in the nominally anhydrous minerals.

How much melt would be produced in the numerical models of this study? A simplified expression for melt production (\(f\)) per degree of temperature above the mantle solidus is:

\[
\frac{df}{dT} = \frac{dT}{dT_{solidus}}
\]

where \(df\) is the fraction of melt (wt\%) produced over \(dT\), an increment of temperature, and \(T-T_{solidus}\) is the temperature interval above the solidus. We use estimates for \(df/dT\) for H_2O-undersaturated peridotite melting at 1.2 GPa by Gaetani and Grove [1998], which yield melt productivity rates of 0.03 wt\%/1°C near the solidus to 0.173 wt\%/1°C at increased extents of melting, which are much lower than experimentally-determined melt productivity rates from anhydrous peridotite melting of 0.26 wt\%/1°C at 1.5 GPa [Falloon and Danyushevsky, 2000] or 0.45 wt\%/1°C at 3 GPa [Walter, 1998]. Using this highly simplified expression for melt fraction, we calculate the range of in situ melt fractions for our most reasonable case (mantle temperature=1350°C, 150 ppm H_2O) to be <0.1 – 2.8 wt\% (or 0.01-3.3 vol\% assuming a hydrous silicate melt density of 2.7g/cc) for an adiabatic gradient of 13°C/GPa.
Once generated, a multitude of potential fates await these melt fractions. Any initial melt fraction will grow at a rate proportional to its upwelling rate. The permeability and porosity of the host mantle will depend on deformation. Lab experiments on stress-driven melt segregation demonstrate as little as 0.5 vol% MORB melt can segregate from an olivine host and 2-3 vol% MORB melt can overcome surface tension at mantle grain sizes to form networks [Holtzman and Kohlstedt, 2007]. Likewise, work by Takei [2002; 2005] indicates that deformation can reduce grain contact area, create high porosity melt sheets with a strong preferred orientation, and dramatically reduce shear velocity. An in-depth analysis of the distribution of melt, accounting for melt buoyancy and deformation driven melt migration, is beyond the scope of this paper. However, in the simplest scenario, for these melt fractions to adequately explain the sharp velocity gradient at the LAB below eastern North America, a significant percent of the melt must remain trapped in the uppermost asthenosphere. If the small amount of melt predicted by our model for a potential temperature of 1350°C and 150 ppm H₂O does indeed segregate, the work of Sparks and Parmentier [1991], Katz et al. [2006; 2008], and Holtzman and Kohlstedt [2007] suggests this melt will form networks in the uppermost asthenosphere adjacent to the LAB due to some combination of dilation of the porous matrix and the high shear strains located here from corner flow and the bordering rheological transition. Therefore, Kohlstedt and Holtzman [2009] state that “based on the simple point that stress-driven segregation and organization will be most effective where strain rates are highest, we hypothesize that the LAB is marked by the location of melt-enriched shear zones that lubricate the LAB and thus the plate boundary system.” This configuration of melt would be ideal to produce the observed
shear wave velocity profiles across the LAB. These melt networks would quickly freeze if the melt penetrates the cooler overlying lithosphere [Rubin, 1995]. As such we would not expect any surface expression of these melts.

Given that the melt fractions produced in situ likely represent a lower bound on the actual melt fraction just below the solidus-defined LAB, how do they compare to the seismic observations? Rychert et al. [2007] constrained the seismic shear wave velocity drops beneath two seismic stations in eastern North America to be 5-7% (HRV) and 6-10% (LMN), and similar velocity drops are consistent with the Sp phases observed more widely in eastern North America by Abt et al. [2010]. According to Hammond and Humphreys [2000] melt fractions of 1-2% are capable of producing velocity reductions within these ranges, assuming realistic melt distribution at grain boundaries. Following Takei [2002], Takei and Holtzman [2009ab] and Kawakatsu et al [2009], for texturally equilibrated melt at grain boundaries, melt fractions required to produce the observed velocity reductions would exceed 3%, but if melt occurs in horizontal melt-rich bands, significantly smaller average mantle melt fractions would be sufficient. Therefore, although predicted velocity drops strongly depend on deformation and the inferred geometry of melt distribution, the in situ melt fractions at all but the lowest end of the range extrapolated from the models in this study appear to be capable of matching the velocity drops observed at the LAB in eastern North America. In addition, if melt produced deeper in the super-solidus zones collects at depths just below the solidus-defined LAB, even larger seismic velocity reductions would be predicted.

4.2 The Global LAB
The models presented here show that small degrees of asthenospheric melting are a possible mechanism for the sharp vertical velocity gradients observed in eastern North America, but they also imply that asthenospheric melting may play a role in defining the LAB in other regions where similar ranges of temperature, composition and H2O content apply. Several recent studies have found evidence for a sharp seismic discontinuity at depths interpretable as the LAB beneath relatively thin Phanerozoic continental lithosphere, but not beneath thick cratonic lithosphere, on a global basis [Rychert and Shearer, 2009], and beneath North America [Abt et al., 2010] and Australia [Ford et al., 2010]. Our work suggests that this observation may be explained by favorable conditions for asthenospheric upwelling and melting in regions of thin lithosphere and the unfavorable conditions for asthenospheric upwelling and melting below thick cratonic lithosphere. While the modeling results in this study show that melting produced by upwelling is a viable explanation for sharp LAB velocity gradients in regions of relatively thin continental lithosphere downwind of lithospheric keels, in some cases at distances of up to almost 200 km from the edge of the keel, comparably sharp LAB velocity gradients have been observed beneath continents outside this zone of influence [e.g. Li et al., 2007; Heit et al., 2007; Chen et al., 2008; Rychert and Shearer, 2009; Abt et al., 2010; Ford et al., 2010]. In some of these regions, different forms of upwelling are possible, but other mechanisms to produce low degrees of melt also exist. Other than upwelling, melting can be produced by raising temperature, possibly by gradual asthenospheric heating from radiogenic elements while advecting beneath a thick insulating lithosphere (i.e., bottom heating). The degree of heating from this process is limited and would only operate under specific circumstances, when a long
sublithospheric traverse occurred and asthenospheric material was already near its solidus. Melting can also be triggered by injection of incompatible elements such as water, perhaps through dewatering of small parcels of downwelling enriched lithosphere melting in small-scale convection at the LAB. Alternatively, the possibility that sharp LAB velocity gradients reflect dry, depleted lithosphere over hydrated, fertile asthenosphere cannot be ruled out.

Because the melting temperature of the asthenosphere is highly dependent on the asthenosphere temperature and H\textsubscript{2}O content, the conditions required for melting below the LAB may be met at some locations and times but not others. Therefore, as temperature or H\textsubscript{2}O content fluctuate within the region of upwelling asthenosphere, the fraction of melt produced may increase or decrease with time. Thus the magnitude of the shear wave speed drop at the LAB produced by small degrees of partial melt in the asthenosphere may change with time and space. For example, the amplitudes of the Sp phases interpreted as originating from the LAB in eastern North America vary by a factor of two, and, as a whole, LAB Sp phases observed in the tectonically active western U.S. are significantly larger than their eastern counterparts [Abt et al., 2010]. In addition, LAB Sp phases in Phanerozoic eastern Australia show a marked correlation between shallower depths, larger amplitudes, and the locations of most recent, voluminous magmatism [Ford et al., 2010].

4.3 Comparison to Other Melting Models

The location and extent of melting in our numerical experiments for an upper mantle with 150 ppm H\textsubscript{2}O are nearly identical when the expressions for solidus
depression from Hirschmann et al. [2009], Katz et al. [2003], the expression for liquidus depression from Medard and Grove [2008], or our own simplified parameterization for solidus temperature at a given degree of H2O mineral saturation, are used (Table 2, Fig. 2 & 6).

The expression for solidus depression from Hirschmann et al. [2009] predicts that a peridotite with 100 ppm H2O begins to melt 80 km along an adiabat with a potential temperature of 1323°C, or 15 km deeper than the case of a truly dry peridotite. Despite the differences in the formulation of our simplified parameterization, which uses a linear approximation for the nominally anhydrous surface, this result is almost identical to our predictions for a peridotite with 10% mineral saturation (aH2O=0.1), which melts at 73 km along an identical adiabat, or 13 km deeper than the case of a truly dry peridotite using the Hirschmann [2000] anhydrous solidus. In another example, using our solidus parameterization and a potential temperature of 1350°C, a peridotite with 10% mineral saturation melts at 84 km, or 18 km deeper than the truly anhydrous case. Likewise, when the expression for solidus depression of Katz et al. [2003] is used with updated bulk partition coefficients for H2O between peridotite and melt, rather than the constant bulk partition coefficient used in their original paper, it is remarkably similar to both the Hirschmann et al. [2009] and 10% mineral saturation expression presented here.

At higher H2O contents (e.g., 450 ppm H2O) differences between the various expressions for solidus depression with H2O increase. The Medard and Grove [2008] expression is derived from experiments on basaltic liquidus depression and was not originally intended to apply to solidus depression, as the expressions for Katz et al. [2003] and Hirschmann et al. [2009] are. The predictions of our parameterization for
solidus temperature at 30% mineral saturation varies from the Hirschmann et al. [2009] solidus for 450 ppm bulk H₂O by only 2-14°C between 2-7 GPa.

4.4 Implications of the similar expressions for solidus and liquidus depression with H₂O

The power-law expressions for solidus depression from the cryoscopic approach of Hirschmann et al. [2009] and the mathematical model of Katz et al. [2003], although calibrated on high-pressure experiments, do not fully account for the speciation of water in silicate melts or the non-ideal mixing between the anhydrous and hydrous components. Although the activity of water in silicate melt can be approximated as these power-law expressions [Burnham, 1975; Stolper, 1982], the non-linearity of liquidus (or solidus) depression cannot be explained by an ideal solution model for the speciation of OH and H₂O molecules in silicate liquids because it is likely related to the interaction between H₂O and other melt components [Nicholls, 1980; Medard and Grove, 2008].

Medard and Grove [2008] are able to use Margules parameters in their thermodynamic model for liquidus depression after the method of Nicholls [1980] to account for the non-ideality of H₂O in silicate melt. Although we recognize both the linear and cubic parameterization of liquidus depression by Medard and Grove [2008] were never intended to apply to solidus depression, the similarity between the quantity of liquidus depression calculated from their model and the quantity of solidus depression calculated from models that do not account for non-ideality, is intriguing. It appears the effect of small amounts of H₂O (<200 ppm bulk H₂O) on solidus depression, as calculated using the various expressions here, may not vary significantly from its effect on liquidus depression. Alternatively, the amount of solidus depression calculated from a
yet to be determined non-ideal solution model may be dramatically different from the expressions used here and from the amount of liquidus depression.

In the absence of sufficient high precision experimental data and a derivative model that accounts for the non-ideality of solidus depression, there is a qualitative agreement between the power-law expressions for solidus depression and our linear approximation of solidus depression at low H$_2$O contents and upper mantle pressures ($< 7$ GPa) when using the bulk H$_2$O-peridotite partitioning data now available.

The advantage of our parameterization is that it can be applied to any scenario where there is a designated temperature and H$_2$O content of mantle peridotite to test the plausibility of mantle melting at pressures $\leq 7$ GPa, or to determine an approximate H$_2$O content required for melting at a given temperature and pressure. But because this model does not take into account phase equilibria, employ a more sophisticated activity-composition model or account for reductions in temperature due to the heat of fusion, it is not intended to replace a complete treatment of melting conditions or to determine the extent of melting at a given P and T, except as a maximum estimate given by the simple approximation in equation 12.

5. Conclusions

Melting of the asthenosphere at the edge of the continental keel may be responsible for shear wave speed gradients observed at the LAB in eastern North America. Numerical experiments and melting calculations predict $<0.1$-2.8 wt% (0.01-3.3 vol%) melting at a depth of 102-126 km at the edge of the continental keel for an asthenosphere with average mantle potential temperatures and H$_2$O contents equivalent to
the average MORB-source. This melting is sufficient to generate the 5-10% drop in shear wave speed observed at the LAB below eastern North America. Such melt could in part explain the rheological contrast observed at the LAB without producing a surface signature.

Production of low extent melts via convective upwelling in the shallow asthenosphere is a viable process in a variety of locations in addition to the one examined here. Melting is likely to vary in time and space with H₂O content, composition and temperature of the asthenosphere. For example, lenses within the mantle of remnant subducted slabs or hydrated mantle wedge will have lower melting temperatures than the nominally anhydrous mantle. Similarly, geologic time periods with a higher mantle potential temperature relative to today are more likely to produce melt fractions near the LAB. Thus if small degrees of asthenospheric melting are a primary cause of the LAB, the magnitude of shear wave gradients across this boundary are likely to be variable in space and time.

A comparison of different expressions for peridotite solidus depression in the presence of small amounts of H₂O all produce very similar predictions for the location and extent of melting, including the simplified parameterization at constant mineral saturation presented here. Until more sophisticated models of peridotite solidus depression that use a non-ideal solution model for H₂O in silicate melts are developed, our simple parameterization is a good starting point to test the feasibility of melting in the mantle anywhere P, T and H₂O content are specified.
Acknowledgements

This research was supported by NSF EAR 0538155 and a NSF Graduate Research Fellowship. We thank Sarah Zaranek for help with the FEMcont code, Timothy Grove for guidance with the solidus parameterization and four anonymous reviewers for their thoughtful comments.

Figure Captions

Figure 1. Anhydrous and H2O-saturated Peridotite Solidi Used in Our Melting Calculations. Blue bold line: fourth-order parameterization of the H2O-saturated solidus based on the illustrated experiments by Grove et al. [2006] and Till et al. [2007] (light blue circles) and Mysen and Boettcher [1975] (dark blue circles). Closed circles represent super-solidus experiments, and open circles, sub-solidus experiments. Black bold line: second-order parameterization of the H2O-saturated solidus for extrapolation between 6-12 GPa. Red bold line: anhydrous peridotite solidus parameterization from Hirschmann [2000], which is based on the compilation of anhydrous peridotite melting experiments shown by the black circles. Also shown are anhydrous peridotite melting experiments from Kinzler and Grove [1992b] and Grove and Juster [1989] in gray. Green bold line and circles: garnet pyroxenite melting experiments and solidus from Kogiso et al. [2003]. Thin gray line: mantle adiabat of 13°C/GPa with a mantle potential temperature of 1350°C.

Figure 2. H2O-undersaturated Peridotite Solidi Used in Our Melting Calculations. The H2O-undersaturated peridotite solidi calculated with the expressions of Hirschmann et al.
[2009], Katz et al. [2003], and Medard and Grove [2008] as described in section 2.1 for 150 and 450 ppm H$_2$O are illustrated by the purple and blue bold lines, respectively. The bump visible in these solidi at pressures near 2.5 GPa is related to the change in partition coefficient for H between bulk peridotite and liquid immediately preceding and following the onset of garnet stability. The H$_2$O-undersaturated peridotite solidi calculated at a constant degree of saturation of the nominally anhydrous minerals from expression presented here and by Hirth and Kohlstedt [1996] are illustrated by the orange and red dashed lines, respectively. Black bold curves denote the anhydrous and H$_2$O-saturated peridotite solidi used in our melting calculations and the thin gray line represents a mantle adiabat of 13°C/GPa with a mantle potential temperature of 1350°C. The solidi for 10% mineral saturation ($a_{H_2O}=0.1$) overlap with the solidi calculated for 150 ppm H$_2$O for pressures between 2 and 7 GPa and those for 30% ($a_{H_2O}=0.3$) with the 450 ppm H$_2$O solidi over the same pressure range.

**Figure 3.** Temperature-Water Content Phase Diagram for Peridotite Melting at Constant Pressure after Grove et al. [2006]. Temperature and H$_2$O contents are illustrated for 3 GPa. The green curve illustrates the maximum storage capacity of nominally anhydrous peridotite minerals (NAM) based on partitioning data from Hauri et al. [2006] and the blue line the simplified melting boundary for peridotite from a eutectic-like H$_2$O-saturated solidus to an anhydrous liquidus, after Silver and Stolper [1985]. A peridotite containing 150 ppm H$_2$O, illustrated by the red points, will melt at 1409°C where the H$_2$O content intersects the NAM curve. This H$_2$O content corresponds to an activity of H$_2$O for this peridotite of ~0.1 (as calculated in section 2.1). At 3 GPa the H$_2$O-saturated
peridotite solidus is from Grove et al. [2006] and Till et al. [2007], the anhydrous solidus Hirschmann [2000] and the liquidus from Zhang and Herzberg [1994]. Note, in this phase diagram, the liquidus and solidus curves do not show phases that are known to be in equilibrium except at the H₂O-saturated solidus.

**Figure 4.** FEMcont Convection Model Setup. Dimensions of the lithospheric step were taken directly from seismic studies of eastern North America [Li et al., 2003; Rychert et al., 2007; Abt et al., 2010]. μ. refers to the reference viscosity of the asthenosphere or lithosphere.

**Figure 5.** Sensitivity of Finite Element Modeling of Asthenospheric Flow to the Depth of the Lithospheric Keel and the Width of the Step to Thinner Marginal Lithosphere. While in eastern North America the keel appears to be ~200 km thick, and the transition to thin lithosphere to occur over ~50 km, both thinner and thicker keels and transition widths will produce upwellings with the potential for melting. Thicker keels and more abrupt transitions to thinner lithospheres produce larger convective upwellings, extending almost 200 km across the thin lithosphere. When the transition from the keel to thinner lithosphere is more gradual (approaching a fluid dynamic shape), the convective eddy is located more under the transition step from keel to thinner lithosphere, and extends less under the thin lithosphere.

**Figure 6.** Locations of Melting near the LAB Predicted by Our Numerical Experiments and Melting Calculations. The color contours denote the temperature of the lithosphere
and asthenosphere in degrees C. Supersolidus regions in the asthenosphere lie within a white contour line at which the asthenosphere temperature equals that of a given solidus. These calculations are shown after 13 million years and use a plate velocity of 3 cm/year, a mantle potential temperature of 1350°C, an adiabatic gradient of 13°C/GPa, a Rayleigh number of 9.4x10⁶ (μ₀=10¹⁹ Pa•s) and a relative lithosphere-asthenosphere viscosity of 1000.  

a) Melting region predicted using the solidus of Hirth and Kohlstedt [1996] for nominally anhydrous mantle minerals (NAM) that are 10% saturated with H₂O.  
b) Melting region predicted using our solidus for 10% H₂O saturation of the NAM (a_{H₂O}=0.1).  
c) Melting region predicted using the solidus for a peridotite with 150 ppm H₂O calculated from the expression for liquidus depression of olivine-saturated basaltic melts in the presence of H₂O from Medard and Grove [2008] as described in section 2.1.  
d) Melting region predicted using the solidus for a peridotite with 150 ppm H₂O calculated from the expression for solidus depression of Katz et al. [2003].  
e) Melting region predicted using the solidus for a peridotite with 150 ppm H₂O calculated from the expression for solidus depression of Hirschmann et al. [2009].  
f) Calculated with the solidus of Hirth and Kohlstedt [1996] for NAM that are 30% saturated with H₂O.  
g) Calculated with our solidus for 30% H₂O saturation of the NAM (a_{H₂O}=0.3).  
h) Calculated for a peridotite with 450 ppm H₂O using Medard and Grove [2008].  
i) Calculated for a peridotite with 450 ppm H₂O using Katz et al. [2003].  
j) Calculated for a peridotite with 450 ppm H₂O using Hirschmann et al. [2009].
Figure 7. Comparison of Model Runs With and Without the Mantle Wind. 
A) Same model run as illustrated in Figure 5, panel b). 
B) Model with identical initial conditions except with a mantle wind velocity of 0 cm/year.

Table 1. Equation for the H\textsubscript{2}O storage capacity of the nominally anhydrous peridotite minerals at different pressures. Y-intercept is the anhydrous solidus for a given pressure. Temperature in °C, pressure in GPa.

Table 2. Predicted H\textsubscript{2}O contents of mantle peridotite with pressure for different degrees of H\textsubscript{2}O saturation of the nominally anhydrous minerals based on the partitioning data of Hauri et al. (2006). All H\textsubscript{2}O in ppm, P in GPa.

Table 3. Location and extent of melting for the vertical slice across the melting region in our numerical models that includes the node with the maximum temperature above a given solidus. The location and extent of melting is also given for the node with the minimum temperature above the solidus. All values are calculated from model runs with a mantle potential temperature of 1350°C. Maximum extent of melting calculated using a \(\frac{df}{dT}=0.173\) (wt%/°C) and minimum extent of melting calculated for \(\frac{df}{dT}=0.03\) as discussed in section 4.1. Horizontal distances are measured from the right margin (see Fig. 6).
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Figure 1.
Figure 2.
Figure 3.
Figure 4
Figure 5.
Figure 7
Table 1. Equation for the H2O storage capacity of the nominally anhydrous peridotite minerals at different pressures. Y-intercept is the anhydrous solidus for a given pressure. Temperature in °C, pressure in GPa.

<table>
<thead>
<tr>
<th>Pressure</th>
<th>Equation</th>
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<tbody>
<tr>
<td>1</td>
<td>$T = -6080P + 1248.5$</td>
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<td>2</td>
<td>$T = -490P + 1366$</td>
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<td>3</td>
<td>$T = -4322.7P + 1473.4$</td>
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<td>4</td>
<td>$T = -3808P + 1570.6$</td>
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<td>5</td>
<td>$T = -3390P + 1657.6$</td>
</tr>
<tr>
<td>6</td>
<td>$T = -3081.3P + 1734.4$</td>
</tr>
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Table 2. Predicted H$_2$O contents of mantle peridotite with pressure for different degrees of H$_2$O saturation of nominally anhydrous minerals based on the partitioning data of Hauri et al. (2006). All H$_2$O in ppm, P in GPa.

<table>
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<tr>
<th>Pressure</th>
<th>$a_{H_2O}=1.0$</th>
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<th>$a_{H_2O}=0.5$</th>
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<td>3000</td>
<td>2100</td>
<td>1500</td>
<td>900</td>
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</table>

* equivalent to XH$_2$O at NAM saturation
Table 3. Location and extent of melting for the vertical slice across the melting region in our numerical models that includes the node with maximum temperature above a given solidus. The location and extent of melting is also given for the node with the minimum temperature above the solidus. All values are calculated from models with a mantle potential temperature of 1350°C. Maximum extent of melting calculated using a df/dT=0.173 (wt%/°C) and minimum extent of melting calculated for df/dT=0.03 as discussed in section 4.1. Horizontal distances are measured from the right margin (see Figure 5).

<table>
<thead>
<tr>
<th>Depth (km)</th>
<th>Horizontal Location (km)</th>
<th>T above solidus (°C)</th>
<th>f (wt%)</th>
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<tr>
<td>150 ppm H₂O</td>
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<td>184</td>
<td>4.3</td>
<td></td>
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<tr>
<td>122</td>
<td>216</td>
<td>0.12 (min)</td>
<td>0.0036</td>
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