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Eruption of the Grande Ronde Basalt lavas, Columbia River Basalt Group: Results of numerical modeling

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ABSTRACT

The Grande Ronde Basalt lavas constitute ~63% of the Columbia River Basalt Group, a large igneous province in the NW United States. The lavas are aphyric or contain less than 5% phenocrysts of plagioclase, augite, pigeonite, and olivine (altered). Plagioclase hygrometry shows that the erupted lavas generally contained less than 0.3% dissolved H_2O; however, the presence of rare disequilibrium An_{96} plagioclase phenocrysts suggests that some magmas may have originally had 4.5 wt% dissolved H_2O at depth, but they all degassed during ascent and eruption. The size of plagioclase phenocrysts suggests an average plagioclase phenocryst residence time in the magmas of 160 yr. Ignoring hiatuses between eruptions, we estimate that the ~110 flows of the Grande Ronde Basalt erupted over a cumulative time of 17,600 yr, with an average eruption rate of ~8.6 km^3/yr. The average interval between eruptions is estimated to be 3658 yr. It is envisaged that a shallow intrusive dike-sill complex, rather than large kilometer-sized magma chamber(s), fed the Grande Ronde basalt lavas.

We performed model simulations using the COMAGMAT software to retrace the pre-eruption histories of the Grande Ronde Basalt lavas. Based on such simulations and petrological reasoning, we find that the primary melts could have been generated from a spinel peridotite source at 1.5 GPa, perhaps under hydrous conditions. Extensive melting of lithospheric eclogite may have played an important role as well; however, this is not constrained by our simulations. All lavas were contaminated by the crust, and they were last processed (mixing, crystallization) during their short residence within shallow (6 km) intrusive rocks prior to eruption. Our petrologic conclusions lead us to present a petrotectonic model that supports the hypothesis that the Columbia River Basalt Group magma generation was greatly aided by a thinned lithosphere and H_2O that may have come off the asthenospheric wedge.

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INTRODUCTION

Large igneous provinces on Earth were formed by the eruption of large volumes of lava at different times in geological history. Many of these volcanic events lasted less than a million years. Generation of such large-volume magma over very short times is commonly attributed to melting of large plume "heads" that rise from the deep mantle, perhaps the D″ layer (e.g., Farnetani and Hofmann, 2009; Garnero and McNamara, 2008; Jellinek and Manga, 2004, and references therein).

In this study, we examine the petrogenesis of the youngest, best preserved, and best studied large igneous province—the Columbia River Basalt Group of the Pacific Northwest (Fig. 1; e.g., Waters, 1961; Watkins and Baksi, 1974; McKee et al., 1977; Swanson et al., 1979; Hooper et al., 2002; Reidel, 1983, 1998; Reidel and Tolan, 1992; Swanson, et al., 1989; Takahashi et al., 1998; Ramos et al., 2005; Camp and Hanan, 2008; Wolff et al., 2008). Our focus is on the Grande Ronde Basalt, which constitutes the bulk (63%–66%) of the Columbia River Basalt Group (Camp and Hanan, 2008; Barry et al., 2010). An important feature of the Grande Ronde Basalt lavas is that they are generally aphyric or nearly phenocryst-free, that is, they essentially erupted as melts, which has prompted some authors to consider them to be near-primary melts even though they have rather evolved (e.g., low Mg/Fe ratio, discussed later herein) chemical composition (e.g., Takahashi et al., 1998; Lange, 2002).

The present study follows our earlier publication in which we presented petrographic, mineralogical, chemical, and phase equilibrium evidence to conclude that the Grande Ronde Basalt basalts were processed in shallow magma chambers prior to eruption (Durand and Sen, 2004). Significant new information and ideas concerning Columbia River Basalt Group magmatism have since emerged in the literature, which have prompted this study. Here, we attempt to simulate the pressure and temperature conditions under which these magmas ascended from their depth(s) of origin to the surface.

Grande Ronde Basalt: Problems of Petrogenesis

The majority (92.7%) of the Columbia River Basalt Group lavas, including the most voluminous Grande Ronde Basalt, erupted between 16.8 and 15.6 Ma (Jarboe et al., 2010; Reidel et al., this volume). Volcanism began in southeastern Oregon (Steens Basalts) and rapidly moved toward northeastern Oregon, where the Imnaha and Grande Ronde Basalt Formation lavas erupted through the Chief Joseph dike swarm (Fig. 1). Smaller and episodic eruptions continued from the northern end of the Chief Joseph dike swarm, giving rise to the Wanapum and Saddle Mountains Formations. Much farther to the southeast, rhyolites were erupting 16.5 Ma onward from the Yellowstone hotspot (e.g., Camp and Hanan, 2008).

Most authors accept the hypothesis that the Yellowstone plume was responsible for the generation of the Columbia River Basalt Group (Duncan, and Richards, 1991; Richards et al., 1989; Campbell and Griffiths, 1990). The location of the Yellowstone plume ~17 m.y. ago and its subsequent track have been established through plate reconstructions by Engebretson et al. (1984) and Pierce and Morgan (1992); they place the plume near the Columbia River Basalt Group eruptive center in southeastern Oregon at that time. The plume or hotspot track can be traced from the site of the initial eruptions, which produced a rhyolitic

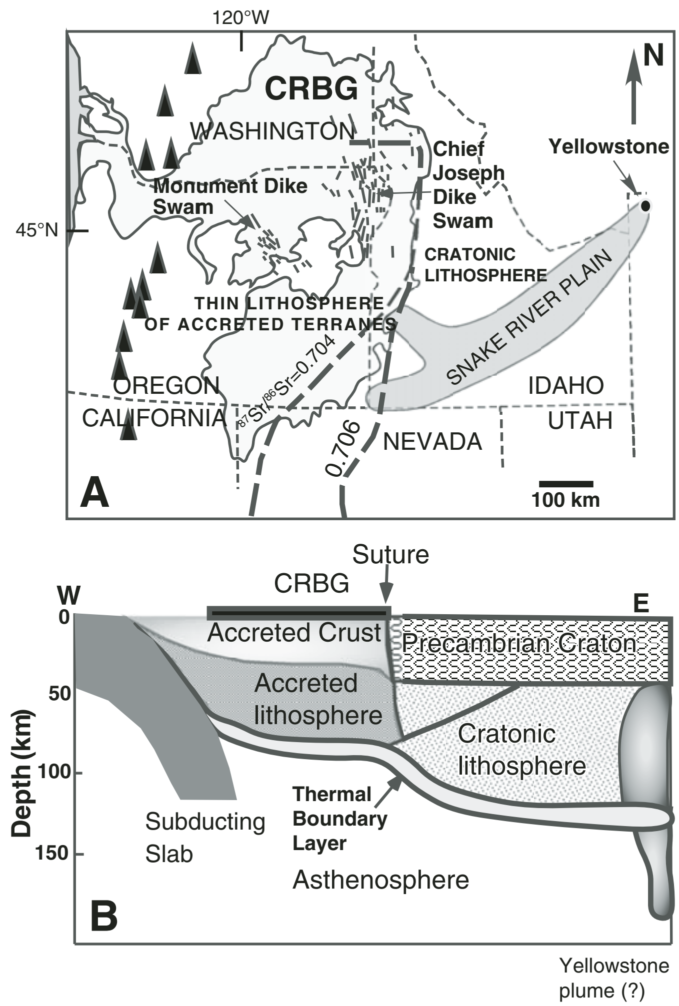

Figure 1. (A) Map showing distribution of the Columbia River Basalt Group (CRBG; slight gray), feeder dike swarms related to the Columbia River Basalt Group, Cascades volcanoes (dark triangles), the Snake River Plain, and the Yellowstone hotspot. The dashed line defined by $^{87}Sr/^{86}Sr$ = 0.7060 is generally thought to represent a suture zone along which the accreted lithosphere to the west is "glued" to the cratonic lithosphere to the east. (B) A schematic cross section (after Leeman et al., 2004) that depicts the complicated nature of the lithosphere at the present time: accreted and thinner lithosphere to the west and thicker, significantly older (Precambrian) lithosphere to the east of the suture zone described in Figure 1A. Active magmatism due to the subduction of the Juan de Fuca plate at the western margin has given rise to the Cascades volcanoes. At the eastern end, Yellowstone plume activity has given rise to the formation of the Snake River Plain volcanic track.

field from ca. 16.5 Ma onward, through the silicic eastern Snake River Plain and ending in Wyoming at the Yellowstone caldera, where the plume is located at the present time.

Takahashi et al. (1998) presented a complex plume model that calls for the generation of Columbia River Basalt Group magmas from a plume head composed of blobs of eclogite embedded in a peridotite matrix. The appeal of this model is that it provides for a mechanism to generate large volumes of mafic magma that has the appearance of being chemically evolved, such as the Grande Ronde Basalt, at relatively low temperature. Some other authors have proposed models that do not involve a plume. One such hypothesis is that the Columbia River Basalt Group magmas formed due to backarc spreading behind the Cascade arc and at the northern end of the Basin and Range Province (Hart and Carlson, 1987; Carlson and Hart, 1988; Swanson et al., 1989; Smith, 1992). Hooper et al. (2007) suggested that the voluminous production of magmas in a very short time cannot be accomplished in a nonplume environment, and the excess heat supply over a short time is more consistent with a plume source.

More complex models of magma generation via plume melting and plume-lithosphere interaction have appeared recently in the literature (Hales et al., 2005; Camp and Hanan, 2008). Hales et al. (2005) and Camp and Hanan (2008) proposed a model of lithospheric delamination triggered by the emplacement of a plume, and, according to Camp and Hanan (2008), Columbia River Basalt Group magmas represent a progressive sequence of delamination and "chewing up" of the subcontinental lithosphere by an impacting plume head. In this model, the older and more magnesian Imnaha magmas were directly generated by plume melting, whereas the Grande Ronde Basalt magmas were produced from a mixture of partially melted crust, lithosphere, and the plume itself.

Finally, Wolff et al. (2008) reexamined all the isotopic systematics and proposed that Columbia River Basalt Group magmas were generated by mixing of melts derived from a plume, surrounding upper mantle, and crust. Most importantly, these authors suggested that the Grande Ronde Basalt gained its crustal isotopic signal via mixing between plume-derived melts and crustal melts that occurred in a central magma chamber system located ~15–30 km below the Chief Joseph dike swarm.

Although Hooper et al. (2007) and Camp and Hanan (2008) agreed with the idea of a crustal component, they suggested that the rapid eruption of Grande Ronde Basalt lavas goes against the magma chamber idea, and they favored incorporation of the crustal signal directly by large-scale melting of eclogitic crust at ~70 km.

In this study, we try to retrace the evolutionary path of Grande Ronde Basalt lavas and attempt to answer the following important questions using model simulations:

1. Did the magmas rise straight up from 70 km to the surface, or did they actually spend some time in shallow (or deep) crustal magma chambers?
2. What was the nature of the source rock(s)—peridotite, eclogite, or a mixture of both?
3. At what depth did the melting occur, and under what conditions? What was the temperature range of melting?
4. Was H_2O involved, and if so, to what extent?

Constraints from Petrography, Petrology, and Geochemistry of the Grande Ronde Basalt

Phenocrysts

Depth(s) of phenocryst formation. As stated before, the Grande Ronde Basalt lavas are aphyric to weakly porphyritic, with less than 5% phenocrysts (e.g., Caprarelli and Reidel, 2004; Durand and Sen, 2004; S. Durand, 2006). The phenocrysts are of plagioclase, olivine, augite, and pigeonite—a typical assemblage that forms by low-pressure (shallow crust) crystallization of basaltic magma (Durand and Sen, 2004; Caprarelli and Reidel, 2004). Durand and Sen (2004) presented strong textural evidence, such as resorption of some plagioclase and augite phenocrysts, development of augite rims around resorbed pigeonite phenocrysts, and reverse and normal zoning in plagioclase in the same lava, in support of their conclusion that magma mixing played a major role in defining the major-element composition of the erupted lavas. Observing that Grande Ronde Basalt lavas define a linear trend that closely follows the 0.001 GPa pseudocotectic olivine (ol) + plagioclase (pl) + augite (aug) + liquid (liq), Durand and Sen (2004) concluded that such mixing took place in shallow magma chambers. This conclusion is generally consistent with the conclusion of Ramos et al. (2005), who estimated the depth of formation of the phenocrysts to correlate to ~0.5 GPa using MELTS software.

Caprarelli and Reidel (2004) performed geothermobarometric calculations on the augite phenocrysts of the Grande Ronde Basalt and found that they formed over a pressure range from ~0.01 GPa (i.e., near the surface) to ~0.65 GPa (~18 km). Hence, they suggested that the magmas traveled through the crust without stalling anywhere within the shallow crust. Because this conclusion is at odds with ours, and yet both studies were essentially based on mineral chemistry, we decided to evaluate their conclusion. Although Caprarelli and Reidel (2004) showed many pressure-temperature (*P-T*) data in their Figure 8, they presented a few augite analyses in their Table 2, representing the range of pressures. We calculated aug/melt (whole rock) Mg/Fe partitioning for two samples (GR014-b and BLS30) for which they obtained the higher pressures of 0.6 and 0.4 GPa, respectively. From their Tables 1 (whole-rock composition) and 3 (mineral composition), we calculated a K_d value (0.54) that is very different from the equilibrium partitioning value of 0.24 (±0.02). Therefore, we suggest that their calculated pressures do not represent host magma–augite equilibrium pressures. This invalidates their calculated polybaric path for magma ascent (figure 8 *in* Caprarelli and Reidel, 2004). In passing, we note that Caprarelli and Reidel's (2004) barometric calculations on *equilibrium* (as deciphered from K_d consideration; Durand and Sen, 2004) augite crystals consistently give <0.3 GPa (i.e., less than 9 km).

Durand (2006) was careful in selecting only equilibrium plagioclase crystals in her pressure calculation based on plagioclase-melt barometry. She found that the plagioclase crystals formed within a very narrow pressure range (0.2–0.3 GPa, 6–9 km). In sum, there is little doubt that the *equilibrium* plagioclase and augite crystals formed at low pressure (0.2–0.3 GPa), which supports the low-pressure mixing and equilibration interpretation (discussed earlier); however, we also note that disequilibrium crystals of augite and the An_{96} plagioclase crystals exist (Durand and Sen, 2004), and these likely formed at greater depths. We have no way of estimating their depth(s) of formation.

Even though the lavas and disequilibrium phenocrysts support magma mixing and formation of the equilibrium phenocrysts at 0.2–0.3 GPa, they do not provide any constraints on the size(s) or shape(s) of the magma chamber(s) in which they formed. However, as discussed here, the phenocrysts provide some constraints on magma residence times in such shallow chambers.

Time scales of phenocryst growth, and eruption rates. In this study, we carried out a simple "back-of-the-envelope" calculation of the age (t) of a plagioclase phenocryst from growth rate (G) and size (s), where $t = s/G$. We recognize that plagioclase growth rate (and morphology) under near-surface conditions (as lava or in shallow conduits) can be influenced by the degree of magma undercooling and melt/crystal ratio, and the growth rate may not even be constant (hence, our use of the average growth rate). We use average growth rate here because (1) the crystal morphologies do not exhibit any evidence of strong undercooling, and (2) we used a range of rates that represent the extremes of published values—two of these are from natural "systems," 0.99×10^{-11} cm/s from a Hawaiian lava lake (Cashman and Marsh, 1988) and 3.03×10^{-10} cm/s from a volcano in Kamchatka (Izbekov et al., 2002). The high value of 5.37×10^{-10} cm/s was derived from laboratory experiments by Burkhard (2005). Our use of 5.37×10^{-10} cm/s at the high end and 0.99×10^{-11} cm/s at the low end should cover average natural growth rate of plagioclase crystals at shallow depths. Using the higher rate for G, we obtain an age of 30 yr for a 5 mm crystal that Ramos et al. (2005) analyzed. The same crystal gives an age of 160 yr if we use a G of 0.99×10^{-11} cm/s. It is our view that the experimental G is less reliable than those estimated from natural systems, and we consider the 160 yr value for the crystal's age to be more reasonable.

What significance may be given to such calculated plagioclase growth time? Could this be a minimum or maximum time of residence of the magma in the shallow chamber or conduit? The presence of disequilibrium plagioclase crystals (presumably of higher-pressure origin) as well as equilibrium crystals can be taken to suggest that the magmas that were input into the shallow chamber (or intrusive system) were already saturated with plagioclase. In such a case, the particular plagioclase crystal in consideration could have already been present in the magma, and therefore the estimated crystal age would be greater than the actual residence time of the host magma in the shallow chamber. On the other hand, if the said crystal formed postemplacement in a shallow conduit/chamber, then crystal age would be less than the magma's residence time in the chamber. All these caveats notwithstanding, we suggest that the 160 yr time is a reasonable average time of magma residence in the pre-eruptive chamber(s) because it is based on a growth rate estimated from natural systems. This is a relatively short time and suggests that a shallow intrusive network, such as a feeder dike-sill complex, is perhaps more appropriate than large, kilometer-sized (say, Skaergaard-like or even mid-ocean-ridge basalt [MORB]–like) magma chambers.

The Grande Ronde Basalt is composed of at least ~110 lava flows, which have very few (but unknown) hiatuses between eruptions, as represented by thin soil horizons (e.g., Barry et al., 2010). If we ignore such noneruptive intervals and extend our average magma residence time of 160 yr to the ~110 flows of the entire Grande Ronde Basalt, then we estimate a cumulative eruption duration of 17,600 yr for the Grande Ronde Basalt. This, of course, assumes a discontinuous process in which each batch of magma comes in, grows plagioclase crystals, and erupts en masse. Based on our understanding of modern volcanic systems, it is more likely that the process is relatively continuous: a magma chamber (or, in the Grande Ronde Basalt case, we prefer a conduit network) undergoes episodic emplacement of new batches of magma, which squeezes out a portion of the extant magma while mixing with the remainder of the magma that was already in the chamber. Also, some Grande Ronde Basalt flows are thinner than others, implying that the residence times varied, and hence an average residence time is more appropriate. In passing, we note that both Barry et al. (2010) and Jarboe et al. (2010), despite differences in absolute ages, estimated that the entire volume (~150,000 km^3; Reidel and Tolan, this volume) of Grande Ronde Basalt basalts erupted within 420,000–570,000 yr. We note that this would be the maximum duration time because this is limited by the resolving power of the Ar isotopic system (Barry et al., 2010). If we compare our calculated duration time (17,600 yr) with the maximum time of 420,000 ± 18,000 yr for the entire Grande Ronde Basalt (lavas + hiatuses) as determined by Barry et al. (2010), then we find that the total of all the noneruptive intervals between lava eruptions would be equal to 402,400 (±18,000) yr. Therefore, the average interval between eruptions is of the order of 402,400 ÷ 110 = 3658 (±1636) yr, which is consistent with Barry et al.'s (2010) estimation of less than 4000 yr for the average eruption hiatus.

Taking 17,600 yr to be the cumulative eruption duration of the Grande Ronde Basalt, we obtain an average active lava production rate of 8.5 km^3/yr. It is worthwhile to compare this rate with that of the largest known historic lava eruption, Laki 1783 (Iceland), which produced 14.73 km^3 of basalt in 8 mo, yielding an eruption rate of 22 km^3/yr (Guilbaud et al., 2007, and references therein). Thus, our estimated lava production rate for the Grande Ronde Basalt is significantly less than the largest basalt eruption in modern history.

Pre-eruption H_2O content of the Grande Ronde Basalt lavas. In our previous study, we concluded that the erupted Grande Ronde Basalt lavas were all well degassed (Durand and Sen, 2004). Using Putirka's (2005) plagioclase hygrometer (algorithm given in Appendix), we calculate that the dissolved H_2O contents were on the order of 0–0.3 wt% in the pre-eruptive magmas from which the plagioclase crystals had formed. This finding substantiates our earlier conclusion (Durand and Sen, 2004). However, Durand and Sen (2004) also noted that there exist rare high-An (An_{96}) plagioclase phenocrysts in Grande Ronde Basalt lavas, which are best explained as having crystallized from hydrous magmas at greater depths. However, it is difficult to provide a quantitative assessment of (1) the amount of dissolved water that was present in the magmas that produced such crystals, (2) the fraction of Grande Ronde Basalt magmas that started out as significantly H_2O-rich magmas, or (3) whether all Grande Ronde Basalt magmas were initially significantly hydrous and degassed during ascent. With regard to item 1, we nonetheless made an attempt to estimate the H_2O content of Grande Ronde Basalt magmas that could have formed the An_{96} crystals at 0.2 GPa pressure using a plagioclase hygrometer calculator given by R. Lange (Lange et al., 2009). We assumed (1) major-element composition of the whole rock as the melt in equilibrium with An_{96} plagioclase and (2) a temperature of 1100 °C, and we obtained an estimate of 4.5% dissolved H_2O in the magma that would have formed the An_{96} plagioclase crystals. We feel that these assumptions are not too far-fetched based on our modeling work (presented later).

Chemical Constraints on Magma Contamination

The Grande Ronde Basalt is impressively uniform in its major- and trace-element composition (Figs. 2 and 3). The lavas are chemically evolved and have low MgO and relatively high SiO_2 contents; thus, the lavas do not appear to be what we typically think of as primary magmas derived from peridotite mantle; in fact, they are basaltic andesites. Isotopic evidence suggests that at least some of the evolved character, such as high silica, is due to contamination with old continental crust (discussed later; Carlson et al., 1981; Carlson, 1984). Note that we have plotted the majority of the Grande Ronde Basalt flows that contain roughly 54% SiO_2, and ignored those with higher SiO_2 (which goes up to 57%), since our purpose was to obtain constraints on the less-differentiated and volumetrically more significant lavas.

Among all the Columbia River Basalt Group formations, the Imnaha is thought to represent largely the plume-derived magmas because of its high $^3He/^4He$ ($R/R_A = 11.4 \pm 0.7$; Dodson et al., 1997), and also, all the other Columbia River Basalt Group formations radiate away from the relatively primitive Imnaha on Pb-Sr-Nd isotope plots (Wolff et al., 2008). Wolff et al. (2008) pointed out that the Grande Ronde Basalt compositional spread can only be explained by shallow-level crustal contamination. In an earlier study, Carlson et al. (1981) used the Sr-Nd isotope variation exhibited by the Columbia River Basalt Group to suggest that the Grande Ronde Basalt lavas are a result of contamination of Imnaha magma by 10% to 20% of a crustal component.

Crustal melting (contaminant) requires heat, which must come from the basalt magma, and this loss of heat would cause magma to crystallize. The question is how much magma mass would be crystallized if it loses heat needed for 10%–20% crustal

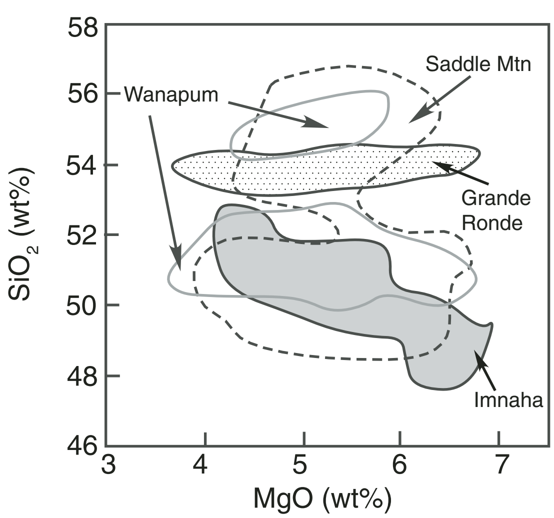

Figure 2. MgO vs. SiO_2 variation in the lavas of various formations of the Columbia River Basalt Group. Data source: GEOROC database: http://georoc.mpch-mainz.gwdg.de/georoc/. Relative to the other formations, the most voluminous Grande Ronde Basalt has spectacularly constant and relatively high levels of SiO_2 over a limited range of MgO. Note that we have excluded all Grande Ronde Basalt lavas that have >54.5 wt% SiO_2.

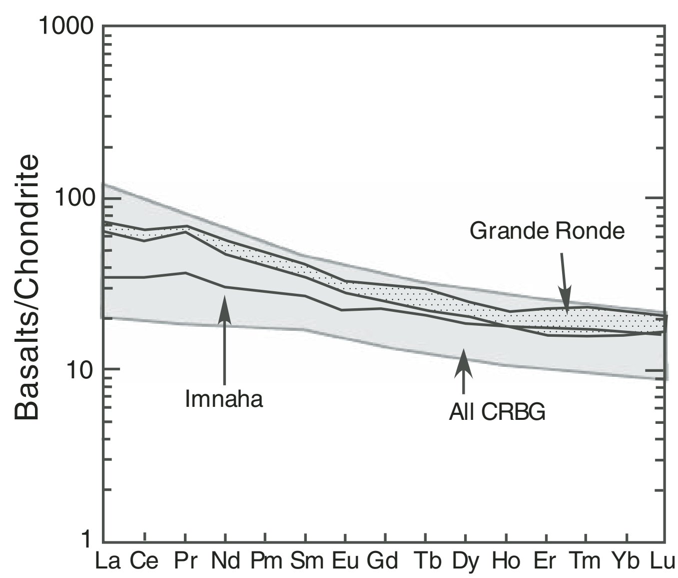

Figure 3. The Grande Ronde Basalt shows very limited variation in chondrite-normalized rare earth element (REE) abundances relative to the other Columbia River Basalt Group units, once again suggesting some spectacularly efficient process that controlled the REE composition. That process is unlikely to have involved any role of garnet, as garnet is known to fractionate middle to heavy REEs, which is not shown by the Grande Ronde Basalt. The light (L) REE enrichment of the Grande Ronde Basalt is suggestive of LREE-enriched source rocks.

melting (Fig. 4A). Reiners et al. (1995) calculated that at 0.1 GPa pressure (~3 km deep), basaltic host magma can assimilate 2–2.7 times the mass of contaminant melt (granitic) as long as olivine is the only liquidus phase. However, fractional crystallization of the hybrid magma would severely slow down the assimilation process when pyroxene and plagioclase join the crystalline assemblage. Figure 4B is modified from Reiners et al.'s (1995) Figure 5, in which the authors plotted calculated contamination trends vis-à-vis magma mass crystallized due to the incorporation of 2% granitic contaminant melt at 400 °C and 800 °C. The dashed line for the wall-rock temperature at 600 °C is simply an interpolation between the 400 °C and 800 °C curves. Each curve has a break in slope that is marked by the change from olivine + melt (steeper slope) to olivine + plagioclase + augite + melt transition. We have included a field for the Grande Ronde Basalt in Figure 4B based on the Nd-isotopic variation shown by the Grande Ronde Basalt and Carlson et al.'s (1981) mixing calculations requiring the production of Grande Ronde Basalt magmas by mixing 8%–18% granitic contaminant in the Imnaha basalt (discussed earlier). Although the Grande Ronde Basalt field is an estimate, it is a reasonable approximation because it is constrained from the observed isotopic variation and calculated extent of contamination. Furthermore, it is consistent with the observation that the Grande Ronde Basalt lavas with ε_{Nd} <+2 carry all three phenocryst phases. The significance of this figure is that the erupted Grande Ronde Basalt lavas could be explained as contaminated Imnaha-like parental melts that lost 8%–18% crystals in shallow crustal chambers due to assimilation–fractional crystallization processes, as pointed out by earlier authors (Carlson et al., 1981; Wolff et al., 2008).

Rare Earth Element Geochemistry Constraints on Depth of Magma Generation

The lack of middle-to-heavy rare earth element (REE) fractionation (Fig. 3) suggests that garnet was not present in the partial melting residue related to the production of Grande Ronde Basalt magmas, which limits the pressure of melting to less than ~3 GPa (e.g., Robinson and Wood, 1998; Wolff et al., 2008; Camp and Hanan, 2008, and references therein). Therefore, garnet peridotite or eclogite can only be called upon as source materials if they were melted to a large degree such that garnet was no longer a residual phase. Wolff et al. (2008) also pointed out that if previously subducted eclogites were involved (as suggested by Takahashi et al., 1998), then they should have been depleted in large ion lithophile elements (LILEs), and magmas produced from them should have also been depleted in LILEs. However, Grande Ronde Basalt lavas are all enriched in LILEs. The lack of residual garnet, based on middle to heavy REEs, and the light (L) REE–enriched nature of Grande Ronde Basalt weaken the eclogite melting hypothesis of Takahashi and others. However, LREE-enriched eclogites do exist, which are found in orogenic areas (Jacob, 2004). These could certainly be source material for the Grande Ronde Basalt, and, in such a case, the lack of residual garnet can be attributed to the high degree of melting of the eclogite source.

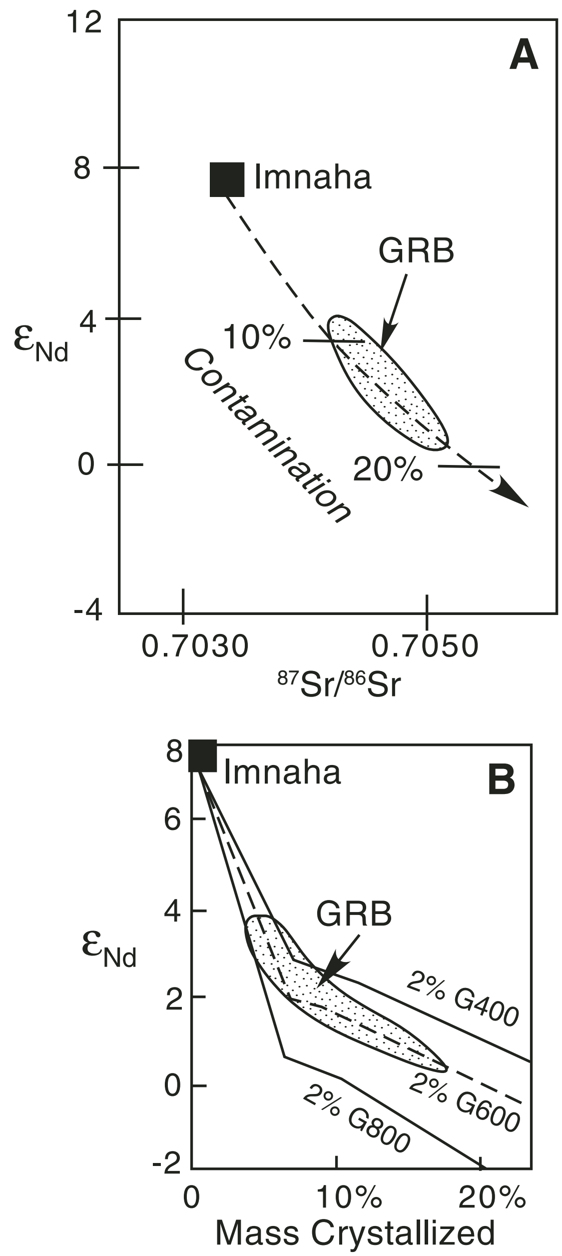

Figure 4. (A) This figure is a simplified version of Figure 5 in Carlson et al. (1981) and shows the Imnaha basalts as uncontaminated mantle melts on the Nd-Sr isotope diagram and that the Grande Ronde Basalt (GRB, dotted field) plots on a mixing line defined by progressive crustal contamination of Imnaha-like magma. The calculated percentages of crustal contamination are shown as tick marks on the mixing line (Carlson et al., 1981). (B) This figure shows how much mass % Imnaha-like magma would crystallize if it were to generate Grande Ronde Basalt magma by 2% melting of the continental crust (granitic melt) and mixing with such melts at various temperatures. The curves are from Reiners et al. (1995). See text for further discussion.

Geochemical Modeling

We used the software COMAGMAT (Ariskin et al., 1993) to simulate the physico-chemical conditions (i.e., P, T, and X_{H2O}) appropriate for fractionation, contamination, and mixing that could explain the "liquid line of descent" (LLD), i.e., the major-element chemical variations in a suite of lavas from a given geographic area. Taking this one step further, we have attempted to deduce the suitable primary compositions for such a LLD and the *P-T* conditions of magma generation using both inverse and forward modeling techniques (Fig. 5; explained in the next section). This exercise demonstrated to us that even though the melts underwent efficient shallow mixing along the olivine + plagioclase + augite + pigeonite (pig) pseudocotectic (Durand and Sen, 2004), their compositions could only be explained by LLDs derived from a select few parent magmas under specific physical conditions of crystallization.

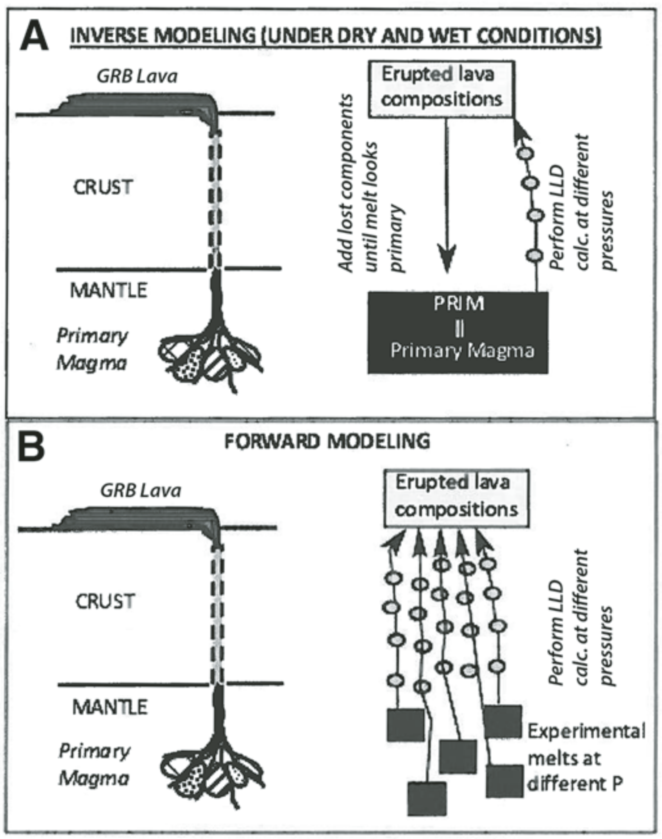

Figure 5. This diagram explains the inverse and forward modeling approaches used here. In forward modeling, partial melts generated experimentally at different pressures and p_{H2O} from peridotite in previous studies are fractionally crystallized at different pressures in the crust using COMAGMAT software. Such fractionation paths are compared with erupted lava compositions in order to evaluate which paths are more plausible than others. In inverse modeling, a "back calculation" is performed in which the erupted lava compositions are made to become reasonable mantle-equilibrated primary magmas by adding back the chemical components that were presumably "lost" from the magma during its rise from the mantle through the continental crust. PRIM is such a primary magma calculated for the Grande Ronde Basalt basalts. See text for further explanation. GRB—Grande Ronde Basalt; LLD—liquid line of descent.

Although this paper is not meant to be a critique of thermodynamic rigor of published simulation software like COMAGMAT, we wish to point out that such software packages are based on semi-empirical algorithms that are not strictly rigorous in a thermodynamic sense, and hence they have their limitations. They are built in a general sense on the basis of free-energy minimization methods, and their success is evaluated on how closely they can reproduce crystallization behavior of natural magmas at various pressures in laboratory experiments.

Before proceeding further, it is useful to be clear about terminology: We use the terms "primary" magma, "primitive" magma, and "parental" magma in the same sense as detailed in *Basaltic Volcanism on Terrestrial Planets* (BVTP, 1981). A primary magma is one that has equilibrated with the source rock, whether it is a peridotite or an eclogite. It is often thought that tiny fractions of partial melt are generated within the source materials (peridotite, eclogite) over a wide volume (i.e., *P-T* space), and that these melts accumulate into larger bodies that equilibrate with the upper mantle before ascending toward the crust. We consider such tiny melt fractions and the accumulated melt volume that equilibrated last with the source mantle volume to be primary magma. As a rule of thumb, basaltic magma with Mg/(Mg + Fe^{2+}) ratio of ≥0.69 is a potential candidate for peridotite-derived primary magma. However, if the source rock is an eclogite, which is a metamorphosed basalt, the Mg/(Mg + Fe^{2+}) ratio of a magma derived from it can be as low as 0.45.

In a suite of spatially related erupted lavas of varying compositions derived ultimately from a peridotite source, the one with the highest Mg/(Mg + Fe*) ratio is considered to be the parental magma because such magma could potentially differentiate in subsurface conduits and generate the other less magnesian lavas. A primitive magma, on the other hand, is any lava with a fairly high Mg/(Mg + Fe*). Thus, a primary magma is both primitive and parental to a suite of lavas, but the inverse is not necessarily true.

Forward Modeling

In the forward modeling approach, we chose starting magma compositions to be experimentally generated partial melts of peridotite and eclogite sources at a variety of *P-T* conditions. Using COMAGMAT, we modeled the compositions of these magmas from their P and T of the last mantle equilibration to the surface, while crystallizing in magma conduits at a single pressure or multiple pressures (Fig. 5). The goal of such exercise was to see if we could derive the tightly grouped Grande Ronde Basalt liquid compositions from any experimentally produced melt by modeling the crystallization of the parent melt under any set of conditions.

As with all modeling efforts, there exist certain caveats: (1) COMAGMAT has an upper pressure limit of 2.0 GPa (Ariskin et al., 1993). (2) The high-pressure (>1 GPa) crystallization studies of basaltic or andesitic magmas are few relative to 0.001 GPa crystallization, adding further uncertainty in terms of polybaric crystallization in these simulations. Temperature, melt fraction, and phase compositions sometimes do not exactly duplicate experimental conditions; however, the overall trends match in most cases (e.g., Ariskin et al., 1993; Asimow and Longhi, 2004; Hirschmann et al., 1998, 1999). (3) Published literature on the Grande Ronde Basalt and the Columbia River Basalt Group as a whole shows that there is no strong constraint on the starting eclogite or peridotite compositions. Therefore, for forward modeling efforts on eclogite-derived magmas, our simulations utilized laboratory-generated melts that formed from the starting composition CRB72-31 (Takahashi et al., 1998; Table 1). For forward modeling of peridotite-derived magmas, we used the experimental melt compositions from starting peridotite composition KLB-1 (Hirose and Kushiro, 1993; Table 1). Notwithstanding these difficulties, this exercise revealed surprisingly tight constraints on P, T, and X_{H2O} conditions for the primary Grande Ronde Basalt magma, as described further in the following.

Inverse Modeling

This approach requires extrapolation of the lava compositions to a more primitive magma composition by correcting for all material losses (crystal fractionation) and gains (magma mixing and wall-rock contaminations). In a numerical sense, losses are easier to tackle than gains. In the Grande Ronde Basalt, all lavas are ol + aug + pig + pl saturated (Durand and Sen, 2004), and the maximum MgO content is 6.5%, indicating that even the most primitive Grande Ronde Basalt lavas have too low MgO to be mantle peridotite-equilibrated primary magma. Because we have few constraints on any eclogitic source materials, it is unreasonable to use the inverse approach for reproducing LLDs from an eclogite source. Instead, we focused on obtaining compositions of peridotite-equilibrated primary magma(s) that could have generated the Grande Ronde Basalt. There are problems with this approach in the case of Grande Ronde Basalt, because it assumes that contamination and magma mixing did not significantly alter the LLD, such that one can use equilibrium-based simulations to back track the primary/parental magma composition(s). Crustal contamination can profoundly change isotope ratios (e.g., $^{87}Sr/^{86}Sr$) and trace-element composition of mafic magmas, but the effect is much less on major elements, which are largely controlled by crystal-melt equilibrium processes (Gangopadhyay et al., 2005). In the inverse modeling technique, success or failure of the calculated parental/primary magmas to reproduce the observed LLD could be taken as a measure of the impact of contamination on the LLD.

The impact of magma mixing, which seems to be a major mechanism in the case of Grande Ronde Basalt, is subtler because the mixing occurs (1) between melts at different stages of a LLD at a single pressure or (2) between melts that have evolved to different stages at different pressures but mix in a common chamber. In case 1, one would obtain a linear trend that does not exactly fall on the pseudocotectic in a phase diagram (say, olivine-augite-plagioclase) but will fall very close to an isobaric pseudocotectic. In case 2, the mixing patterns would be more scattered, cutting across pseudocotectics at different pressures. In the case of Grande Ronde Basalt, Durand and Sen (2004) showed that the magma mixing occurred at shallow depth under near isobaric conditions (i.e., scenario 1).

One added problem is that the erupted Grande Ronde Basalt lavas were saturated with olivine, plagioclase, and augite. Peridotite-derived melts typically have only olivine on their liquidus during their ascent, and other phases appear as they cool during ascent. Therefore, in order to obtain olivine-saturated parental melt composition, the lava compositions needed to be extrapolated to the point where the calculated parental melt was saturated only with olivine. In order to do this, we examined a large data set on experimental crystallization of tholeiitic melts at low pressure and noticed that they become olivine-saturated at ~8 wt% MgO (Herzberg and O'Hara, 2002). We used a linear regression through the Grande Ronde Basalt data to 8 wt% MgO, and selected a composition with 8% MgO to be the parental magma to Grande Ronde Basalt basalts. This procedure assumes that the magmas went from olivine + melt to olivine + plagioclase + augite + melt during its evolution without an intermediate period when an ol + pl assemblage would have crystallized. This method is also not entirely desirable in view of contamination and magma mixing. Notwithstanding these drawbacks and in the absence of better alternatives, there really is no better way to go about it due to the lack of constraints on early crystallization. We followed this procedure anyway because our objective was to eliminate implausible scenarios. Furthermore, one can also check whether the inferred parent magma (8% MgO) can reproduce the observed LLD under shallow crystallization or mixing conditions. We have done this using COMAGMAT and noticed that even slight differences in the starting melt composition can result in calculated LLDs that do not go through the observed data set. Therefore, we proceeded with inverse approach keeping in mind the inherent difficulties of interpretation.

TABLE 1. STARTING EXPERIMENTAL MELTS (HIROSE AND KUSHIRO, 1993)

P (GPa)	T (°C)	F	SiO_2 (wt%)	TiO_2 (wt%)	Al_2O_3 (wt%)	FeO* (wt%)	MnO (wt%)	MgO (wt%)	CaO (wt%)	Na_2O (wt%)	K_2O (wt%)
1	1250	0.07	51.32	1.09	19.09	6.38	0.23	8.14	8.85	4.60	0.27
1.5	1300	0.06	50.71	1.04	19.31	6.37	0.14	8.31	7.75	5.47	0.73
2	1375	0.14	47.47	0.75	15.53	8.51	0.18	13.94	11.11	2.22	0.08
3	1500	0.17	45.67	0.99	14.33	9.59	0.17	16.73	10.64	1.80	0.07

Note: F—melting degree.

The extrapolated 8 wt% MgO parental magma (PAM, Table 2) is close to a composition that could form by fractionation from a mantle-equilibrated primary magma. To this postulated olivine-saturated melt, we then added olivine in 2 wt% incremental steps, while maintaining a constant K_d $(FeO/MgO)^{Ol/liq}$ of 0.3, until a reasonable primary magma composition was obtained (PRIM; Table 2) that would have equilibrated with a typical mantle olivine (Fo_{89}). We cannot rely too much on the inverse approach in the case of Grande Ronde Basalt at higher pressure because of all the reasons stated already. Instead, we limit our calculations to the shallow crust (0.2–0.6 GPa) in order to evaluate the sensitivity of LLDs to pressure from a plausible bulk melt composition.

RESULTS AND DISCUSSION

Forward Models

Eclogite-Derived Melts Crystallized at 0.2 GPa

Takahashi et al. (1998) suggested that Grande Ronde Basalt magmas can be produced as 30%–50% batch melts from an eclogitic source at ~2 GPa and 1300–1350 °C. Here, we used only the maximum magnesian melt generated at 2 GPa from a single eclogitic bulk composition (CRB72-31; Takahashi et al., 1998) and crystallized them at 0.2 GPa (Fig. 6). We should point out that maximum MgO in Takahashi et al.'s high-pressure experimental melts was 5.7 wt%. None of those experimental melts has enough MgO to cover the more magnesian part (MgO > 6%) of the Grande Ronde Basalt spectrum, and therefore they cannot be suitable parents of Grande Ronde Basalt magmas. Even though CRB72-31 is not an appropriate starting material for the more magnesian lavas of the Grande Ronde Basalt, some lessons can be learned from our simulation experiments: For example, a very distinct type of LLD is shown by eclogite (at least CRB72-31)–derived melts compared to peridotite-derived melts (discussed later). During shallow crystallization of eclogite-derived melts, plagioclase appears on the liquidus first, causing an initial increase in MgO as plagioclase fractionates. As the residual melt evolves, olivine or augite appears, and then the LLD makes a switch toward decreasing MgO. Although we do not show examples of other eclogite-derived melts, they all exhibit the same characteristic trend, i.e., they have the characteristic "hook" that forms as a result of switch over from plagioclase to a mafic phase on the liquidus. Such "hooks" may be erased when multiple batches mix extensively in a shallow chamber. In the case of Grande Ronde Basalt, persistent presence of resorbed (disequilibrium) plagioclase crystals in the lavas may be relicts of such early liquidus plagioclase crystals.

TABLE 2. STARTING COMPOSITIONS FOR INVERSE MODELING

(wt%)	8 wt% MgO composition (parental magma)	Olivine addition composition (primary magma)
SiO_2	52.0	51.16
TiO_2	0.82	0.76
Al_2O_3	17.5	16.16
FeO*	7.5	7.86
MgO	8.0	11.07
CaO	11.5	10.62
Na_2O	2.57	2.37
Total	100.0	100.0

Melts from a Peridotite Source Crystallized at Low Pressure

We used experimental partial melts from KLB-1 generated at 1–3 GPa by low to moderate degrees (<17%) of melting (F) (Table 1). To simulate low-pressure LLD, each melt was "brought up" to the shallow-levels. LLDs for both equilibrium and fractional crystallization were then calculated for each melt composition over a pressure range (3–0.2 GPa) using COMAGMAT. The resultant LLDs were then compared with Grande Ronde Basalt whole rock data (GEOROC database: http://georoc.mpch-mainz.gwdg.de/georoc/).

Figure 7 shows major oxide-oxide comparison between Grande Ronde Basalt whole-rock data and LLDs calculated for the different experimental partial melts using COMAGMAT. It is

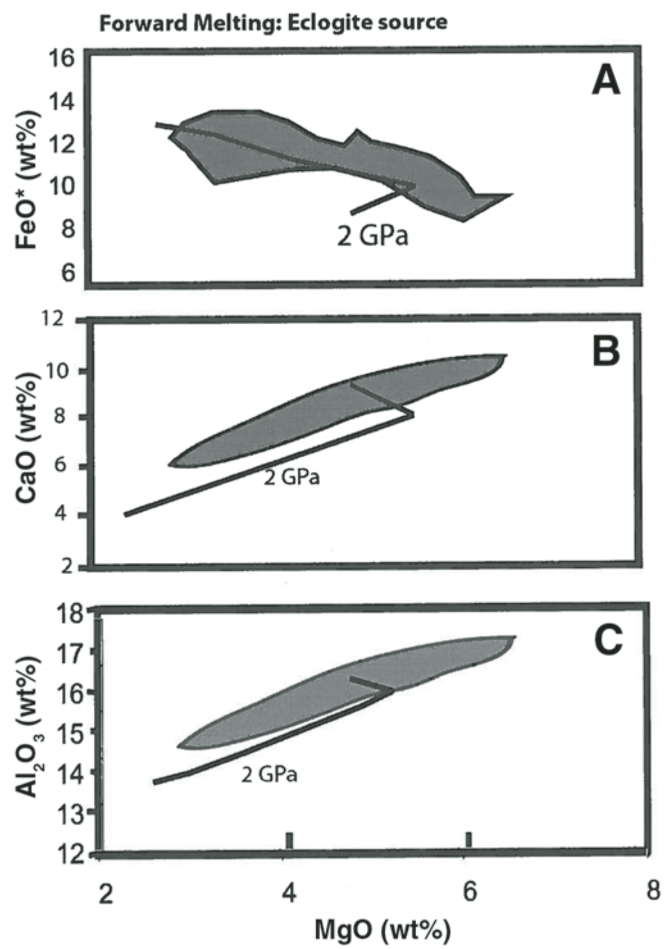

Figure 6. (A–C) Comparison among calculated 0.2 GPa liquid lines of descent (LLDs) for a magma formed by partial melting at 2 GPa from an eclogite starting material (Takahashi et al., 1998). Grande Ronde Basalt represented by shading.

clear from this figure that the melts generated at 2 and 3 GPa and fractionated at low pressure (0.001–0.2 GPa) do not at all resemble the Grande Ronde Basalt data. Partial melts derived at low degrees of melting and at 1.5 GPa from KLB-1 peridotite, when brought up to low pressure (0.2 GPa) and allowed to fractionate, generate a LLD that fits the Grande Ronde Basalt data the best.

Durand and Sen (2004) presented textural and chemical evidence to suggest that Grande Ronde Basalt magmas were well degassed by the time they reached shallow crust. Hygrometry in the present study suggests 0–0.3 wt% dissolved H_2O in the pre-eruptive magma from which the equilibrium plagioclase phenocrysts formed. The effect of such low amounts of dissolved H_2O on LLD at low pressure is not significant, and therefore we do not consider it further. Durand and Sen (2004) also stated that some of the magmas were originally hydrous based on the occurrence of rare An_{96} plagioclase; however, the H_2O content of the parent/primary magma remains unconstrained, and therefore there is no way of knowing the extent to which such magmas degassed on the way up.

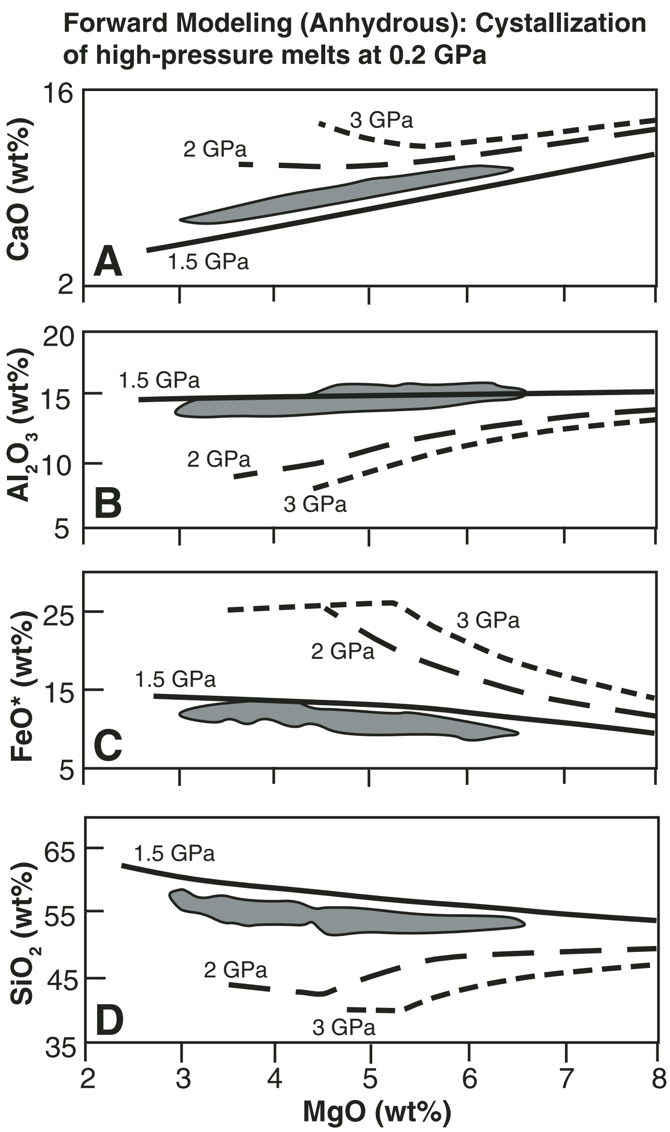

Figure 7. (A–D) Comparison of composition of Grande Ronde Basalt (represented by shading) with liquid lines of descent calculated at 0.2 GPa for different starting magmas generated by melting a peridotite KLB-1 at 1.5, 2, and 3 GPa (Hirose and Kushiro, 1993). Compositions of the starting magmas in these forward models are given in Table 1. (A) CaO vs. MgO. (B) Al_2O_3 vs. MgO plot. (C) FeO* vs. MgO. (D) SiO_2 vs. MgO plot.

Inverse Modeling

Crystallization of Primary Magma

We carried out simulations of isobaric fractional crystallization at 0.2, 0.4, and 0.6 GPa under volatile-free conditions using the COMAGMAT software. The resultant LLDs are best compared in terms of the CaO/Al_2O_3 ratio as a function of MgO variation (Fig. 8). This is because this diagram shows a clear separation of the LLDs at different pressures due mainly to appearance of plagioclase at different pressures. There are three parts to each curve shown in Figure 8: The flat part represents fractionation of olivine alone, because the CaO/Al_2O_3 ratio is unaffected by olivine separation, which also decreases MgO content of the residual melt. Fractionation of plagioclase causes an increase in CaO/Al_2O_3 in the liquid at ~8.7 wt% MgO. Once augite (±pigeonite) joins the fractionating assemblage olivine + plagioclase, CaO/Al_2O_3 decreases sharply. The effect of pressure of crystallization on CaO/Al_2O_3 trends is obvious in Figure 8. The coincidence of the 0.2 GPa LLD with Grande Ronde Basalt field (Fig. 8) suggests that our choice of primary magma is perhaps reasonable as parent magma and supports our earlier conclusion of shallow mixing and fractionation. However, the resultant low-pressure LLD is only slightly inconsistent with our original assumption

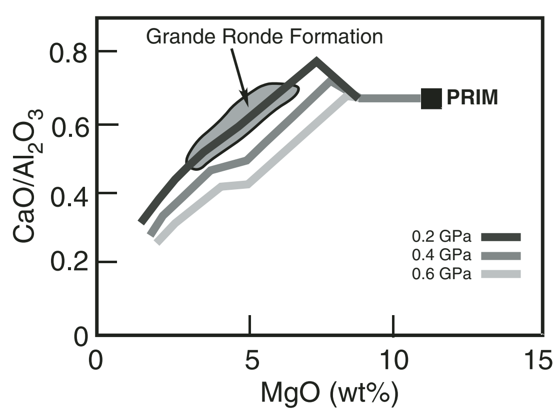

Figure 8. CaO/Al_2O_3 vs. MgO plot of liquid lines of descent (LLDs) at three different pressures from a parental (PRIM) magma composition. Lower-pressure (0.2 GPa) LLD fits the Grande Ronde Basalt lava compositions, whereas LLDs of >0.2 GPa completely miss the Grande Ronde Basalt field. This rules out a deep fractionation origin for Grande Ronde Basalt in magma chambers at the base of the crust (~1.3 GPa), under volatile-free conditions.

in creating the primary magma, that is, we assumed that the ol + aug + pl assemblage is stable on the liquidus at MgO < 8%, and above 8% MgO, olivine is stable. Figure 8 shows that this assemblage starts at 7.3% MgO, and there is a small interval of 7.3%–8.7% where only the ol + plag assemblage is stable. Given all the assumptions made in our modeling, we consider this to be an excellent result.

SUMMARY AND DISCUSSION

The main findings in this paper are summarized as follows:

1. Fractionation and magma mixing occurred at 0.2 GPa. Textural evidence, mineral barometry, phase relationships, and simulation modeling of LLDs all point to the fact that Grande Ronde Basalt lavas underwent significant mixing and fractionation at relatively low pressure (0.2 GPa, ~6 km). Ramos et al. (2005) also saw evidence of shallow-level fractionation of Columbia River Basalt Group lavas in their study of Sr isotope disequilibrium. Most recently, Wolff et al. (2008) proposed that Columbia River Basalt Group lavas were derived from a single centralized crustal storage system where magma rose to the surface through an extensive network of dikes. Our conclusion is consistent with these two studies and not consistent with the conclusions of many other authors who suggest that magmas rose straight from deeper crust or mantle, bypassing any "crustal filter" by stalling in a shallow crustal magma chamber (e.g., Lange, 2002; Hooper et al., 2007).

Hooper et al. (2007) cited the absence of large collapse structures and large dense bodies (representing fossil magma chamber cumulates) in opposing the shallow magma chamber(s) hypothesis. We contend that our work cannot impose any size constraint on such magma chambers. However, given the short magma storage time and rapid eruption rate that we calculated, one would not expect a large magma body to be present in the shallow crust; instead, it is likely to be a network of dikes and sills. In such case, gravity collapse structures would not be expected; on the contrary, a regional uplift caused by mass compensation would be more consistent with such scenario (Bruce D. Marsh, 2000, personal commun.).

2. Grande Ronde Basalt lavas were degassed. The rare presence of high-An (An_{96}) plagioclase phenocrysts points to the possibility of ~4.5 wt% dissolved water in some of the more primitive magmas of the Grande Ronde Basalt. Plagioclase hygrometry on the equilibrium crystals utilizing Putirka's (2005) algorithm suggests that the vast majority of the Grande Ronde Basalt lavas had largely degassed and contained less than 0.3 wt% dissolved water. This is also consistent with our simulations.

3. We calculated the average magma residence time to be 160 yr and an average active lava production rate of 8.6 km^3/yr for the Grande Ronde Basalt, i.e., ignoring the intervals between eruptions. There are some important assumptions made in arriving at these values, which the reader is urged to take into consideration.

4. In the context of magma generation and eruption, it is important to mention constraints from geophysics. Based on a seismic tomographic study, Hales et al. (2005) discovered a high-velocity (denser) body at 70–150 km (~2.2–5 GPa) below the Wallowa Mountains. They interpreted this body to be residue of melting that generated the Columbia River Basalt Group. While the lower limit of this pressure range is consistent with magma production in the spinel peridotite stability field, the 3–5 Pa pressure range would put magma generation well within the garnet peridotite stability field (e.g., Sen, 2001). We noted before that the flat middle-to-heavy REE pattern shown by the Grande Ronde Basalt is not consistent with residual garnet. Alternatively, unusually high temperatures (≥1500 °C; Robinson and Wood, 1998) would be needed to completely melt garnet from a garnet peridotite source, which would be required to explain the REE behavior of Grande Ronde Basalt. Of course, eclogite melting would bypass such temperature requirement, because eclogite can melt at temperatures almost 150 °C lower than peridotite, and, also, eclogite is stable at 1.5 GPa (e.g., Peterman and Hirschmann, 2003). Presence of water in the source can further lower the solidus. Our simulations suggest that partial melts generated at 1.5 GPa (i.e., from spinel peridotite) and fractionating at 0.2 GPa best explain the Grande Ronde Basalt data. The consistently flat middle-to-heavy REE patterns shown by Grande Ronde Basalt lavas are also consistent with a spinel peridotite source rock. However, we cannot rule out an important role played by deep crustal LREE-enriched eclogites as source materials; in such a case, the melting degree would have to have been high enough so that garnet was not a residual phase.

Petrotectonic Model

These conclusions/observations allow us to develop a reasonable petrotectonic model for the generation of the Grande Ronde Basalt (Fig. 9). For melting to have occurred at 1.5 GPa, the lithosphere beneath Columbia Plateau had to be greatly thinned via delamination, allowing hot asthenosphere, possibly hydrated by the fluids released from the subducting Farallon plate, to rise to shallow depths (~45–50 km). It has been reported that the Columbia Plateau is underlain by a thin lithosphere consisting of accreted oceanic terrain of Mesozoic origin (Vallier, 1995; Hales et al., 2005; O'Driscoll, 2007). Hales et al. (2005) suggested that thinning of the lithosphere was caused by the continued backarc extension created by subduction to the west. In the delamination process, large chunks of the eclogitic lower crust may have detached and became heavily involved in the magma generation process (see also Camp and Hanan, 2008).

Leeman et al. (2004) suggested that the temperature beneath the Cascade arc today is warmer than in normal arc settings due to the subduction of a younger slab (Farallon plate). If similar thermal conditions existed 16–17 Ma, then that could have significantly aided the process of generation of large volumes of magma beneath the plateau. The combination of detachment of the lower lithosphere and addition of fluids to the asthenosphere may have created an unusually hot and wet situation to exist in the asthenosphere reaching close to the Moho, resulting in sudden production of melts that became the Grande Ronde Basalt

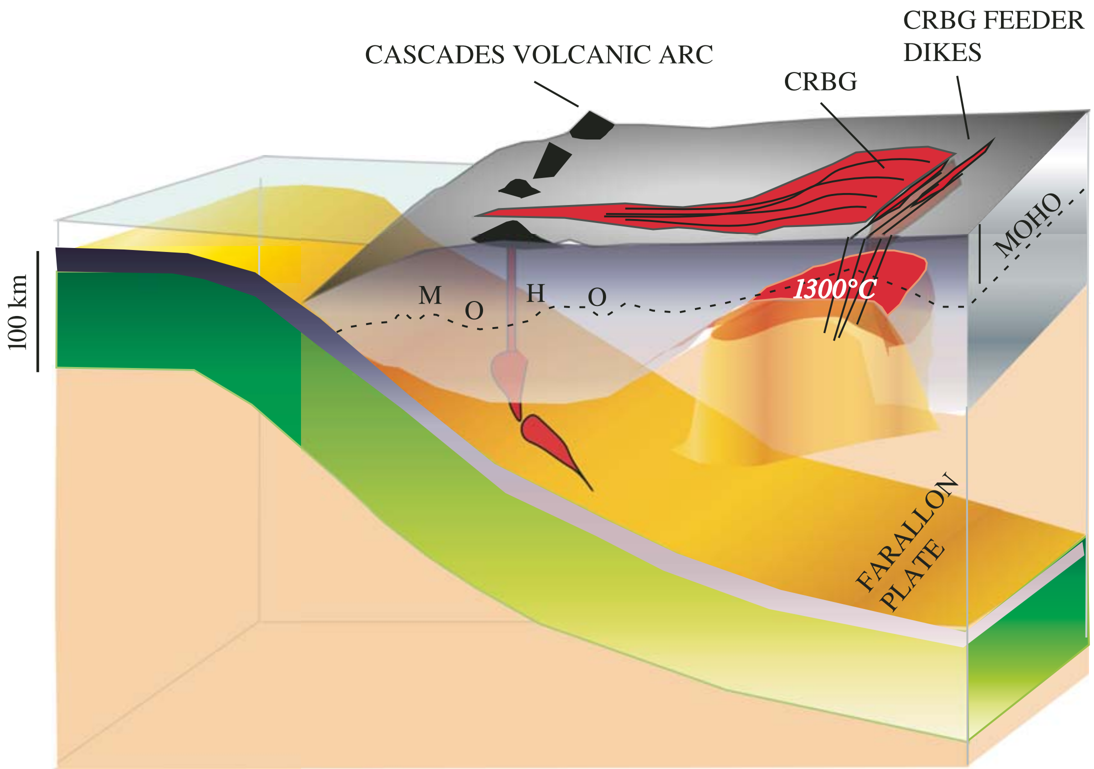

Figure 9. Diagram depicting the conditions that led to magma segregation below the Columbia River Basalt Group (CRBG) plateau. As the Farallon plate was being subducted under the North America plate, the release of volatiles caused melting to occur above the Farallon plate. In addition, the extensional environment in this area caused thinning of the lithosphere. As a consequence, decompressional melting occurred. It is possible that a combination of two or more processes combined to form the large volumes of lava that erupted to produce the flood basalt province.

primary magmas. This model is presented schematically in Figure 9. These magmas were generated or last equilibrated with largely peridotitic material at a shallow depth (45 km). Magmas were briefly stored and modified in a series of interconnected sills and dikes before erupting. Such magmas escaped through feeder dikes, which formed during the backarc extension event.

The model we have described here does not a priori require the presence of a very hot plume from the deeper mantle; however, as a thermal source, such a plume would help the model in accelerating the magma production rate. A recent numerical modeling study by Liu and Stegman (2012) suggests that the development of a slab tear in the Farallon plate was necessary for allowing hotter, deeper material to rapidly rise through the Farallon plate. Their model requires an active role by the subducted Farallon plate as source material for the magmas. As pointed out earlier, this is not consistent with the geochemistry of the Grande Ronde Basalt lavas (Wolff et al., 2008).

ACKNOWLEDGMENTS

Peter Hooper introduced us to the Columbia River Basalt Group, encouraged us along the way, and shared samples from his own collection from the type section of the Grande Ronde Basalt. We will miss him and humbly dedicate this paper to his memory. We thank Steve Reidel for his help and encouragement over the years. We are grateful to Scott Burroughs, Kirsten Nicolaysen, and an anonymous reviewer for their excellent reviews. The final version of the paper is much improved from the original manuscript as a result of the reviewers' comments and the excellent editorial suggestions made by John Wolff. Rodriguez is grateful to Florida International University's Graduate School for a Dissertation Year Fellowship, and to the Department of Earth & Environment for assistantships.

APPENDIX: THERMOMETRY AND BAROMETRY

We calculated eruption conditions using the plagioclase phenocryst data obtained during this study and whole-rock data from Hooper and Hawkesworth (1993). *P-T* conditions were calculated for plagioclase crystals that were in equilibrium with appropriate liquid compositions (whole-rock data were assumed to be the same as the liquid in this case, since most of the Grande Ronde Basalt is aphyric and fine grained).

Plagioclase-liquid thermobarometers of Putirka (2005) were used to estimate the *P-T* of crystallization and the water contents of the Grande Ronde Basalt lavas. Although the Sugawara (2001) and Ghiorso (1994) thermometers provide more accurate temperature estimates than earlier calibrations, these do not provide a means for calculating pressure from plagioclase-liquid equilibrium (Putirka, 2005). To remedy this shortcoming, Putirka developed a plagioclase-liquid thermometer and barometer. Putirka's (2005) thermometers (below) yield an error that is 30%–40% lower than Sugawara (2001) and Ghiorso (1994) models:

Putirka's Plagioclase-liquid thermometer based on his model A:

$$104/T\ (\mathrm{K}) = 68.8 - 0.86\ln\{An^{P}/(Ca^{liq}[Al^{liq}]^2(Si^{liq})^2\} + 179(Al^{liq}) - 113(Al/[Al + Si])^{liq} - 7.92(AbAn^{P}) - 6.13 \times 10^{-2}(P\ [\mathrm{kbar}]) - 91.6(CaAl^{liq}) - 155(Si^{liq}) + 110.3(Si^{liq})^2 - 149(Al^{liq})^2,$$

where Pl and liq denote plagioclase and liquid phases, respectively, and Ca^{liq}, Al^{liq}, and Si^{liq} represent the abundances of these elements in the liquid.

Temperatures calculated using Putirka's model A range from 1089 °C to 1127 °C, which are 50–100°C lower than those calculated by Caprarelli and Reidel (2005) using Putirka's clinopyroxene-liquid geothermometer (1120–1222 °C).

We also calculated pressure using Putirka's (2005) model C:

$$P\ (\mathrm{kbar}) = -42.2 + 4.94 \times 10^{-2}(T\ [\mathrm{K}]) + 1.16 \times 10^{-2}T\ (\mathrm{K})\ \ln(Ab^{Pl}Al^{liq}Ca^{liq}/[An^{Pl}Na^{liq}Si^{liq}]) - 382.3(Si^{liq})^2 + 514.2(Si^{liq})^3 - 19.6\ln(Ab^{Pl}) - 139.8(Ca^{liq}) + 287.2(Na^{liq}) + 163.9(K^{liq}),$$

where Pl and liq denote plagioclase and liquid phases, respectively, and Ca^{liq}, Al^{liq}, and Si^{liq} represent the abundances of these elements in the liquid.

We obtain pressures of 0.3–0.6 GPa from the plagioclase-liquid barometer.

H_2O contents of the magmas were derived by using the temperatures calculated above and Putirka's model H:

$$H_2O\ (\mathrm{wt\%}) = 24.757 - 2.26 \times 10^{-3}(\mathrm{T}\ [\mathrm{K}])\ \ln\{An^{Pl}/(Ca^{liq}[Al^{liq}]^2Si^{liq})^2\} - 3.847(Ab^{Pl}) + 1.927\{An^{Pl}/(Ca^{liq}/[Ca^{liq} + Na^{liq}])\},$$

where Pl and liq denote plagioclase and liquid phases, respectively, and Ca^{liq}, Al^{liq}, and Si^{liq} represent the abundances of these elements in the liquid.

The results of the H_2O calculations showed that water contents were very low, reaching the limits of the model, and some results were negative. We obtained dissolved H_2O contents ranging from 0 to 0.3 wt% H_2O, which indicate either that the Grande Ronde Basalt magmas were very dry to begin with, or that, even if the original magmas were wet, they degassed significantly during ascent.

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