Thermal structure beneath the Snake River Plain: Implications for the Yellowstone hotspot

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Abstract

Basaltic magmatism associated with the Yellowstone hotspot has been widely attributed to upwelling of a mantle plume, yet the temporal and spatial distribution of these magmas and their compositional characteristics are distinctive from oceanic hotspot magmatism. Fundamental questions concern the influence of continental cratonic lithosphere in producing the differences, and the extent to which upper plate processes contribute to magma production. To better understand scenarios of melt generation, P-T conditions are estimated for segregation of primitive Snake River Plain (SRP) basalts from the mantle. Combined with analysis of trace element and seismic constraints, we conclude from this that (1) melt production was concentrated at depths between roughly 70–100 km, (2) mantle temperature was only slightly higher than ambient conditions with a maximum potential temperature of 1450 °C, and (3) the mantle source was relatively fertile (Mg# < 90). These results suggest that the seismically imaged plume below Yellowstone is significantly cooler than upwellings beneath Hawaii, Iceland and many other oceanic “hotspots”. Our findings, in combination with other geochemical and geodynamic considerations, are permissive of magma generation within the ancient lithospheric mantle keel associated with the Wyoming craton. Plume contributions, while not excluded, involve physical and geochemical implications that suggest they are subordinate.

1. Introduction

The Snake River Plain-Yellowstone (SRPY) province is a region of diachronous, bimodal rhyolite–basalt magmatism that has been attributed to migration of North America over the Yellowstone hotspot (e.g., Armstrong et al., 1975; Pierce and Morgan, 1992; Smith and Braile, 1994). A variety of other observations appear to support this hypothesis (Pierce et al., 2002). Perhaps the most compelling evidence comes from seismic tomography that defines a sub-vertical low-velocity pipe beneath Yellowstone that extends into the mantle to a depth of 500 km (Fee and Dueker, 2004; Yuan and Dueker, 2005; Waite et al., 2006). This feature connects to a low-velocity region between 60 and 120 km depth under the eastern Snake River Plain (Schutt and Humphreys, 2004; Yuan and Dueker, 2005; Waite et al., 2006; Schutt et al., 2008). Together, these images suggest that a plume of hot mantle is ascending from the transition zone to beneath Yellowstone Park, where it is swept to the SW by the North America Plate flow field as shown in Fig. 1.

The earliest volcanism directly associated with this phenomenon was manifest as eruptions of voluminous silicic ignimbrites and lavas in southwestern Idaho and adjacent areas of Oregon and Nevada around 16 Ma, shortly following onset of the Columbia River flood basalts (Camp and Ross, 2004). Silicic volcanism was associated with large eruptive centers, each active for 2–4 m.y., and it migrated northeasterward toward Yellowstone over time. Although it is inferred that this magmatism was fundamentally driven by injections of voluminous mafic magma into the crust (Hildreth et al., 1991; Bonnichsen et al., 2008), eruptions of basalt were delayed until waning stages of silicic activity, after which volcanism in any given area was dominantly basaltic. At Yellowstone, this transition has only just begun, whereas further west basaltic volcanism has persisted intermittently to Quaternary time across much of the SRP.

The pattern of SRPY volcanic activity differs conspicuously from that associated with oceanic island hotspots, for which magmatism is almost entirely basaltic and defines simpler time-transgressive patterns. The preponderance of silicic volcanism in southern Idaho is attributed to the fact that the region is underlain by cratonic crust, whereas roughly west of the Idaho border the basement comprises an amalgamation of accreted oceanic terranes with relatively juvenile crust (Fig. 1). This difference in basement type is clearly reflected in gravity and magnetic signatures for these areas as well as in the compositions of both basaltic and rhyolitic magmas, with those to the east having systematically higher 87Sr/86Sr and lower 143Nd/144Nd.
magma formation and storage coincides with anomalously low seismic velocities (relevant in the context of the northern migration of the Mendocino triple junction. The tectonic setting over the SW US and propagated northward over the Western Idaho Suture Zone (WISZ); mantle thrust fault signifies shortening related to a late Cretaceous accretionary ‘event’ (cf. Leeman et al., 1992). Beneath YP a thick zone of magma formation and storage coincides with anomalously low seismic velocities (reflected in V̇s tomography and estimated magma segregation depths ~60 to 100 km; this paper). Profiles show mean V̇s structure, derived from surface wave phase velocities (Schutt et al., 2008), for the hotspot track (red, ESRP) and the Wyoming Craton (blue, Craton). One sigma model error is shown by the width of the bands. Reference model is shown as a black line (Model). For comparison, velocity variations with depth are shown for Hawaii (HAW; Priestley and Tilmann, 1999) and Tectonic North America (TNA; Grand and Helmberger, 1984).

Fig. 1. Schematic lithospheric structure for NW U.S.A., with results of V̇s tomography from the eastern Snake River Plain (SRP). Diagram portrays subcontinental lithospheric mantle (SCLM) beneath the Precambrian cratonic North America, with locus of a postulated plume impinging beneath the Yellowstone Plateau (YP); an intervening thermal boundary layer (TBL) marks the base of the lithosphere (thickness is schematic). The Cascadia arc (CAS) and subduction zone (SZ) and Phanerozoic accreted terranes are juxtaposed outboard the Western Idaho Suture Zone (WISZ); mantle thrust fault signifies shortening related to a late Cretaceous accretionary ‘event’ (cf. Leeman et al., 1992). Beneath YP a thick zone of magma formation and storage coincides with anomalously low seismic velocities (reflected in V̇s tomography and estimated magma segregation depths ~60 to 100 km; this paper). Profiles show mean V̇s structure, derived from surface wave phase velocities (Schutt et al., 2008), for the hotspot track (red, ESRP) and the Wyoming Craton (blue, Craton). One sigma model error is shown by the width of the bands. Reference model is shown as a black line (Model). For comparison, velocity variations with depth are shown for Hawaii (HAW; Priestley and Tilmann, 1999) and Tectonic North America (TNA; Grand and Helmberger, 1984).

δ¹³B, and B/Rb (Leeman et al., 1992; Savov et al. 2009-this volume). If SRP magmatism is plume-related, the presence of cratonic lithosphere appears to have profoundly influenced its surface expression. Alternatively, magmatism could be related to Basin and Range style extension, possibly enhanced by sub-lithospheric processes. To better constrain the origin of the basaltic magmas, we have evaluated the thermal structure of their mantle source and how this may relate to contrasting tectonic scenarios.

2. Tectonic processes

2.1. Lithospheric processes (extensional deformation)

The SRP is superimposed on the northern reaches of the Basin and Range province of western North America. Thus, it is relevant to review possible influences of the underlying extensional processes. Considerable evidence (cf. Snyder et al., 1976; Oldow et al., 1989; McQuarrie and Wernicke, 2005) indicates that extensional deformation began in the southwestern US and propagated northward over time following northward migration of the Mendocino triple junction. Extensional strain exceeds 100% (since early Oligocene time) at the latitude of Las Vegas and is on the order of at least 15–20% (since mid-Miocene time) at the latitude of the SRP (Rogers et al., 2002). Moreover, associated magmatism was synchronous with or closely followed onset of extension as it propagated laterally (Gans et al., 1989; Armstrong and Ward, 1991; Axen et al., 1993). Initial mid-Tertiary magmatism was predominantly characterized by eruptions of voluminous silicic magmas, and in Neogene time evolved to bimodal basalts – rhyolite character. However, even the earliest silicic activity can be linked to inputs of mafic magmas into the crust (e.g., Feeley and Grunder, 1991).

Harry and Leeman (1995) considered the effects of extension on mantle melting and proposed that (for the southern and central Basin and Range province) early synextensional magmatism was driven by basalt production in the continental lithospheric mantle owing to decompression melting of entrained fertile (i.e., mafic) domains that were close to their solidus temperatures. This mechanism was suggested because depleted or refractory lithospheric mantle perido-
tell us about mantle temperature? (3) What information can be extracted directly from the basaltic magmas concerning the composition and temperature of their source(s)?

3. The melting process

An upwelling mantle plume can melt only if ascent brings it across its solidus curve (Fig. 2). For a specific mantle solidus, the depth at which decompression melting can begin is determined by the potential temperature \( T_p \), or adiabat curve, for the ascending material (cf. McKenzie and Bickle, 1988). The solidus also may be affected by variations in composition, and is particularly sensitive to content of volatile components (especially \( \text{H}_2\text{O} \)). Because we are concerned with a deep mantle upwelling, it is reasonable to assume that the plume is similar to its sub-oceanic counterparts and therefore likely to consist of relatively ‘dry’ peridotite (although some Hawaiian and other OIB magmas may contain up to ca. 0.5\% \( \text{H}_2\text{O} \) (Hauri, 2002), implying that their sources could contain traces of water; see further discussion below). A more critical factor in this case is the structure of the lithosphere. If ascending mantle follows an adiabat representative of sub-oceanic mantle beneath mid-ocean ridges (e.g., \( T_p = 1450 \) °C or lower; Stein and Stein, 1992; Putirka, 2005; Herzberg et al., 2007), the upwelling material must ascend to depths shallower than ~100 km before it crosses the dry peridotite solidus and melting can proceed. If the plume consists of more fertile material or contains a small amount of water, the solidus could be shifted toward lower temperatures and melting could ensue at greater depths. These options potentially can be distinguished if we know the temperatures of the magmas that are produced. For now, we consider the end member case in which the mantle is dry.

Given that much of southern Idaho and nearby areas are underlain by the Archean Wyoming craton, it is expected that the lithosphere prior to Miocene time could have been more than 150 km thick based on seismic tomography and thermal models (Dueker et al., 2001; Goes and van der Lee, 2002; Artimieva, 2006; Priestley and McKenzie, 2006). If the lithospheric (mechanical or rheologic) lid today exceeds roughly 100 km in thickness, then for decompression paths with \( T_p < 1450 \) °C (i.e., ‘normal mantle’), there would be insufficient headroom to generate any magma for the scenario under consideration. It is worth noting that accreted oceanic terranes outboard of the Archean craton are likely to have sufficiently thin lithosphere, such that (as in ocean basins) there would be little impediment to melting of upwelling plume material. To promote melting of dry mantle at greater depths, so as to minimize the headroom problem, requires that the upwelling material follows a much hotter adiabat (e.g., \( T_p \geq -1500 \) °C if the lid is 120 km thick; see Fig. 2).

4. Seismic constraints on mantle temperature and the melt distribution

Considerable effort has been expended to determine the regional seismic velocity structure in order to constrain distributions of mantle temperature and melting domains. Using seismic tomography, a number of workers (e.g., Humphreys et al., 2000; Schutt and Humphreys, 2004; Waite et al., 2006; Schutt et al., 2008) have established the presence of a significant low-\( V_p \) anomaly. The top of the anomaly is constrained to depths between 50 and 60 km, whereas the bottom is less precisely defined between 120 and 200 km beneath the eastern SRP and Yellowstone. The maximum regional amplitude of this anomaly is about 11% (comparing mantle velocities at depths near 80 km beneath Yellowstone with those beneath the Wyoming craton). Fig. 3A shows a model for variation in \( V_p \) as a function of depth beneath Yellowstone. At depths near 115 km, \( V_p \) shows a maximum anomaly of about -4% relative to velocity calculated for mantle with a potential temperature of 1450 °C. This anomaly could reflect elevated temperature, presence of partial melt, decreased grain size, changes in composition or combinations of these factors (Minster and Anderson, 1981; Karato, 1993; Goes et al., 1997; Lee, 2003; Faul and Jackson, 2005; Schutt and Lesher, 2006). Since grain size and temperature tend to be positively correlated (e.g., Solomatov, 2001; Hall and Parmentier, 2003) it is improbable that the low velocities reflect decreased grain sizes. Rather, it is more likely that low-\( V_p \) anomalies result from increased temperature and/or melting.

Schutt and Dueker (2008) modeled the effects of temperature, melting, and grain size on the fundamental mode Rayleigh wave phase velocity for mantle under the eastern Snake River Plain. Assuming that average mantle has a \( T_p = 1320 \) °C (as assumed in some early studies; e.g., White and McKenzie, 1995), an excess temperature of 55–200 °C was found to best reproduce the velocity anomaly. For comparison, following an identical approach and assuming a higher \( T_p (=1450 \) °C, as suggested by Putirka et al. (2007), Herzberg et al. (2007), and our own work discussed in the following section), we find that \( V_p \) structure can be matched by a plume excess temperature of 125 °C with melt porosity of 0.2% (Fig. 3A). Because this solution is not unique, we also consider possible trade-offs between melt porosity and temperature that could produce the observed \( V_p \) anomaly (Fig. 3B). In the extreme case that no melt is present, the anomaly requires a maximum excess temperature of ~150 °C relative to nearby ambient mantle. If ascribed entirely to the presence of melt (i.e., no excess temperature), the anomaly could be explained by melt porosities between 0.5 and 1.95%, depending on the melt-velocity scaling used (e.g., Hammond and Humphreys, 2000; Takei, 2000). The upper limit on this range may be unrealistic if maximum melt porosity for the upper mantle is on the order of 1% (Renner et al., 2003). However, values of \( \text{dln} V_p / \text{dln} V_s \) (Waite et al., 2006) and geologic intuition suggest that there is at least some melt present. So, realistically, both factors likely contribute to the velocity anomaly below Yellowstone. It is safe to say that the mantle thermal anomaly is unlikely to exceed 150 °C, but it could be significantly smaller.
velocity scaling is calculated using the model of Faul and Jackson (2005) with activation

5. Petrologic constraints on the melting conditions

5.1. Primitive Snake River Plain basalts

SRP basalts are typically olivine ± plagioclase phytic olivine tholeiites (SROT) with low \( K_2O \), high \( Al_2O_3 \) variants (Leeman and Manton, 1971; Leeman and Vitaliano, 1976; Leeman, 1976, 1982a; Geist et al., 2001; Hughes et al., 2002a; Shervais et al., 2006). Low pressure liquidus temperatures near 1200 °C have been estimated for many SROT (Leeman and Vitaliano, 1976; Geist et al., 2001; Hughes, 2005), and range up to 1300 °C for the most primitive ones. Olivine compositions (up to \( Fop_2 \)) are typically in equilibrium with the host magma, although rims are commonly more Fe-rich; relatively Mg-rich (to \( Fom_1 \)) xenocrystic olivine was found in one sample (see Appendix for details). Overall, SRP basalts are variably fractionated owing to storage in crustal level magma reservoirs prior to eruption, and in some cases they have experienced contamination by crustal materials (Leeman et al., 1976; Leeman, 1982b). Analyses of several hundred lavas (from the above references and our unpublished data) were screened to identify the most primitive samples using the criteria that (1) Mg# \( \geq 1=100 \cdot MgO/(MgO + FeO^*) \) on a molar basis with all iron as FeO* exceeds a value of 60 and (2) olivine is the only phenocryst or is clearly the dominant one. The first condition ensures that the bulk composition is not far removed from equilibrium with realistic mantle compositions. The second criterion screens out magmas whose ascent was interrupted by residence in shallow crustal reservoirs where crystallization of plagioclase or interaction with wall rocks could significantly modify their composition. Some 40 samples meet these criteria, with an average of 10.0 ± 0.7% MgO; all have high Ni and Cr contents, averaging 157 and 426 ppm respectively (with maxima of 265 and 594 ppm). These samples represent the most primitive lavas erupted in the SRP (Table 1; Fig. 4A, B).

The systemic absence of clinopyroxene as a phenocryst (coupled with high Cr/Ni and Cr content in the most primitive lavas; cf. Leeman, 1976) also suggests that the parental magmas did not experience significant crystallization at pressures greater than 8–10 kbar, where clinopyroxene is the primary liquidus phase (Thompson, 1975). VSc averages 6.6 ± 0.8, which overlaps with MORB values (Lee et al., 2005) and is consistent with redox conditions near the QFM buffer as also inferred from mineral chemistry (Leeman and Vitaliano, 1976). Under these conditions about 10% of the magmatic iron is ferric (Kress and Carmichael, 1991). Moreover, chondrite-normalized rare earth element (REE) profiles for the most primitive lavas are not flat with little light/heavy REE fractionation and nearly chondritic Gd/Yb (Fig. 4C, D). These data are consistent with formation of SROT basalts by partial melting of a spinel lherzolite source containing little or no residual garnet (cf. Leeman, 1976; Hughes et al., 2002b).

5.2. Geochemical constraints on the magma source

Because the REE data provide a potentially important independent constraint on depth of magma formation, they are modeled in some detail (Fig. 4D). Primitive SROTs define a coherent array in chondrite-normalized La vs. Gd/Yb space, with a slight positive slope. The range in \([La]_n \) (14–55; chondrite-normalized) with increasing \([Gd/Yb]_n \) (1.1–1.7) is difficult to reconcile with simple fractional crystallization because the slope of the array exceeds that produced by removal of possible phenocryst minerals (olivine, plagioclase, or even clinopyroxene), and the range of La enrichment is equivalent to more than 90% closed system crystallization. Thus, the data are more plausibly modeled in terms of varied degrees of melting of primitive and depleted mantle sources. For a range of model conditions, the data are consistent with melting of spinel lherzolite sources; estimated degree of melting varies between roughly 1 and 10% assuming a ‘primitive mantle’ source (Sun and McDonough, 1989), or 0.1–5% if a ‘depleted mantle’ source (Salters and Stracke, 2004) is assumed. Very small amounts of garnet in the source cannot be precluded, but model sources with more than a few percent garnet result in significant enhancement of \([Gd/Yb]_n \), and would be easily detected (cf. Fig. 4D for representative Hawaiian lavas). White and McKenzie (1995) arrived at a similar conclusion based on dynamic melting models to simulate REE data (Lum et al., 1989) for western SRP basalts.
Table 1
Representative primitive SRP basalts and evolved McKinney Basalt.

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a Areas: western SRP (column #s 1, 2, and 4), central SRP (#s 5, 9), eastern SRP (#s 3, 6–8).
b Analyses normalized to 100% total, anhydrous with all Fe as FeO*. USGS wet chemistry (# 8), XRF (all others). Sr and Nd isotopes via TIMS (cf. Leeman et al., 1992).
c Rare earth elements determined by neutron activation (N) or ICPMS (I); Cr and Ni by XRF and ICP-OES.

Fig. 4. Compositional variations for SRP basalts. (A, B) Primitive basalts (filled squares; cf. Table 1 for representative analyses) with high Mg#, MgO, Cr, and Ni (not shown) are used for calculations discussed in this paper; circles and shaded field represent several hundred variably evolved lavas from sources cited in the text; McK (McKinney Basalt) is discussed in text. (C) Representative chondrite-normalized REE patterns for primitive SRP basalts (Table 1; cols. 1, 2, 5, 8) and representative Hawaiian basalts (shaded field outlines data for Kaho'olawe and Kilauea tholeites; Leeman et al., 1994; unpublished). Relatively flat REE profiles for primitive SRP basalts are consistent with little or no residual garnet in sources for the parental magmas. (D) Chondrite-normalized La vs. Gd/Yb for SRP basalts (data for western and eastern SRP are distinguished as filled and open circles, respectively) and selected Hawaiian basalts (filled diamonds). Representative equilibrium melting models are shown for spinel lherzolite (solid curves; shaded field) and garnet lherzolite (dashed curves) sources. Heavy lines indicate models assuming depleted mantle composition, and lighter curves show models assuming primitive mantle composition (see text). Fractional crystallization (FC) vectors show effect of 50% removal of olivine (Oliv), plagioclase (Plag), or clinopyroxene (Cpx); net effect of crystallization of various combinations of these phases is to drive residual liquids horizontally to the right.
5.3. Estimated P–T conditions for magma segregation

Knowledge of the temperature and depth of magma production is critical to resolving the underlying mantle thermal structure. We address this problem by following the approach used for Cascades arc basalts by Leeman et al. (2005). The erupted lavas have Mg#s (60–69) that are too low to have been in equilibrium with reasonable mantle compositions. It is assumed that Mg#s of actual mantle melts were reduced primarily by olivine fractionation during magma ascent. Accordingly, the composition of each primitive lava was adjusted in small steps by incremental addition of calculated equilibrium olivine until a magmatic Mg# appropriate for equilibration with mantle peridotite was obtained; at that point, the recalculated liquid composition was assumed to represent the parental magma that last equilibrated with the mantle source. Because the exact composition of the mantle source is unknown, solutions were obtained for an arbitrary range of mantle Mg#s (88–91). Compositions of representative parental liquids and estimates of T, P, and depth are given in the Appendix tables. It should be noted that this approach approximates an average composition of pooled magma just as it separates from its source. It does not capture finer complexities of the melting process prior to melt segregation.

Each of the resulting primitive magma compositions so obtained was then used in combination with the Sugawara (2000) olivine thermometer and the Albarède (1992) barometer to calculate the temperature and pressure conditions at which the postulated primitive melts could have last equilibrated with peridotitic mantle. A correction was applied to the pressures using the empirical fit of calculated vs. nominal pressures for experimental melts reported subsequent to 1992 (cf. Leeman et al., 2005); uncertainties for T and P are on the order of ±30 °C and ±3 kbar, respectively, based on replication of laboratory P–T conditions for mantle melting experiments. Conceptually, these calculations define a range of segregation T and P (or depth) conditions at which specific magmas could have last equilibrated with a mantle mineral assemblage. The loci of these points closely parallel Hirschmann’s (2000) dry peridotite solidus (cf. Fig. 5).

For specific samples, the required amount of olivine addition ranges between ~10 and 30%, increasing (for a given sample) with assumed mantle Mg# (cf. inset in Fig. 5). Because erupted compositions of some SRP lavas are close to Fe–Mg equilibrium with mantle of Mg# = 88, more fertile (lower Mg#) sources are considered unrealistic. If REE compositions of SRP basalts preclude significant involvement of residual garnet in the magma sources (i.e., a spinel lherzolite source is appropriate), then melting is restricted to depths no greater than ~85 km (which approximates the minimal stability pressure for garnet in the mantle; Robinson and Wood, 1998). Calculated segregation depths in this range are obtained only for source compositions with Mg# ≤ 90 (Fig. 5). In detail, all of the primitive basalts considered give depth estimates below 90 km if source Mg# = 89; about half exceed this depth if source Mg# = 90, and all exceed this depth if source Mg# = 91. With source composition constrained to Mg# < 90, conditions of last equilibration between primitive SRP basalts and the mantle are bracketed as follows: T = 1380–1500 °C, depth = 60–100 km. The upper limit is consistent with mantle ascent along an adiabat with a maximum Tp of ~1450 °C (i.e., extrapolated to the surface).

Fig. 5. Temperature–pressure systematics. Representative adiabats (dashed lines) are shown for upwelling asthenospheric mantle with a range of potential temperature (Tm) estimates. The dry peridotite solidus of Hirschmann (2000) is provided for reference. Bold arrow illustrates an ascent path for mantle with Tm = 1450 °C (i.e., the ideal temperature of this material if it reached the surface with no losses of heat). This material would not undergo melting until it reached a depth of 100 km (i.e., where the adiabat crosses the solidus). The sub-solidus transition from spinel (Sp)-bearing to garnet (Gt)-bearing peridotite intersects the dry solidus at a depth near 85 km (Robinson and Wood, 1998). Thus, for this example (and for warmer mantle upwellings), melting would initiate in the garnet peridotite stability field. Double-headed arrow corresponds to depths of anomalously low Vp inferred from seismic tomography. Superimposed on this diagram are estimated temperatures and pressures for last equilibration of primitive SRP magmas with peridotitic mantle assuming different Mg#s for the source ranging between 89 and 91 (see text for details). White circles illustrate how calculated P–T varies for a specific sample (6YC-142; Table 1, col. 8) with this range of source compositions. Solid star indicates thermodynamic estimate for segregation P–T (Leeman and Vitaliano, 1976). The “sweet spot” field indicates segregation P–T conditions that best satisfy all constraints discussed in this paper. Inset shows variation in calculated segregation temperatures vs. weight fraction of equilibrium olivine added to attain Fe–Mg equilibrium for respective peridotite sources.
A relatively evolved olivine + plagioclase phryic SROT (McKinney Basalt) from the mantle at ~1390 °C and 60 km. Although this analysis was based on the composition of a rapidly quenched pillow lava glass (Table 1, col. 9) with no attempt to back-calculate to the primitive liquid composition, the result is in close agreement with segregation P–T estimates in this paper (Fig. 5).

The range in P–T segregation conditions obtained using multiple samples and specific mantle source Mg# deserves comment. In part, it reflects an inherent systemic correlation between assumed source Mg#, calculated melt MgO content, and estimated temperature. The errors in P and T cited above capture the calibration uncertainties. Propagation of analytical errors is generally smaller based on estimates using multiple analyses/samples of the same basalt unit (cf. Leeman et al., 2005). For example, for a specific source Mg#, multiple samples from Kilauea Iki (with MgO between 7 and 14%) give P and T estimates that agree with standard deviations less than 10 °C and 1 kbar, respectively (see discussion below).

The spectrum of predicted P–T conditions could reflect real variations in source composition. That is, if the approach worked perfectly, slightly changing the assumed source composition for a given sample will produce an array of results like that seen in Fig. 5 (e.g., shown by the shift of points for one sample as assumed source composition is varied). However, as long as variation of Mg# is small within local melting domains, relative differences in P–T estimates for different samples are likely to be meaningful. For example, Mg# could vary systematically as a consequence of melting degree. For decomposition melting in a column with increasing degree of melting toward the top, calculated P–T arrays might partly reflect polybaric melt extraction. Alternatively, if our calculated P–T arrays largely reflect statistical noise, then the maximum P–T estimated for a given source composition is likely to provide a reasonable upper limit value for segregation conditions.

Finally, we note that our preferred segregation depth for SRP basalts is shallower than that inferred (depth ~140 km) by Wang et al. (2002) who use a different approach based primarily on the concentrations of FeO+ and Na2O in estimated primitive melts. The difference partly results from the Wang et al. assumptions that all iron is ferrous (we assume X–Fe2+ = 0.9 based on petrologic constraints) and that Fe–Mg partitioning (Kg = 0.3) between olivine and melt is constant (we allow Kg to vary with melt composition; cf. Leeman et al., 2005). As discussed above, segregation of SRP magmas from depths greater than ~100 km is inconsistent with both their observed REE profiles and the seismic constraints on depths of melting.

6. Comparative mantle thermometry

We present three independent constraints on the depth of melt segregation/formation beneath the Snake River Plain province, as well as estimates of mantle temperature and bulk composition (Mg#). These estimates converge in a mutually consistent manner implying that primitive SROT basalts segregated from the mantle at maximum P–T conditions averaging near 1500 °C and 2.8 GPa (or ~90–100 km; Fig. 5). These conditions closely correspond with the depth of the Vg anomaly (Fig. 3A), and support the notion that the anomaly reflects the presence of melt. The REE systematics for SROT basalts suggest that melting occurred largely within spinel lherzolite facies (i.e., shallow) mantle, although small contributions of melt generated from garnet peridotite cannot be ruled out. If segregated magmas are pooled accumulations of polybaric melts, the REE data could be interpreted in terms of a melting column that extends to depths slightly below the spinel stability field. However, the bulk of the magma must be generated from essentially garnet-free mantle to explain the observed REE profiles. Calculated segregation P–T conditions are consistent with these constraints for all samples only if source Mg# ≤ 89. Thus, it appears that the SRP mantle source must be relatively fertile (lower Mg#) compared to typical estimates for subcratonic lithospheric mantle based on compositions of mantle xenoliths (Walter, 2003).

To better appreciate the implications of predicted segregation conditions for the SRPY melting anomaly, it is important to consider...
these results in the context of estimates for other melting regimes. This is not a straightforward task for several reasons. First, there is considerable latitude in estimates for hotspot and MORB parental magma compositions, reflecting different approaches to the problem of determining mantle thermal structure (cf. Herzberg et al., 2007; Putirka et al., 2007). Second, there is an inherent inverse relation between the fertility of basalt sources and the P–T conditions attending melting and, without independent constraints, it is difficult to specify the fertility of the sources. To make comparisons as direct as possible, we have estimated P–T conditions for primitive ocean island (OIB) and mid-ocean ridge (MORB) basalt compositions proposed by Herzberg and O’Hara (2002) and Putirka (2008). Most of these postulated parental magmas have calculated equilibrium source compositions with Mg# between 90 and 91; this range is consistent with maximum forsterite content in olivines from primitive lavas from each suite. The same thermometer and barometer used for SRP samples was applied to these compositions and the results are shown in Fig. 6. We also show a field of P–T conditions obtained from several hundred MORB analyses with MgO–8% (from PETDB); here we have assumed a source with Mg# = 90, which is consistent with other MORB estimates. Our P–T solutions for MORB suggest maximum segregation is above 1400 °C and depths largely within the spinel peridotite stability field, which is consistent with the characteristic flat HREE profiles for MORB. Although higher source Mg# is conceivable for some MORB, this would yield higher P–T solutions and result in a garnet signature in their REE profiles. These results are consistent with the findings of Wang et al. (2002) who also suggest an ambient upper mantle Tp between 1400 and 1450 °C. P–T estimates for postulated OIB parental liquids are more widely scattered, but generally higher than our preferred range for SRP basalts. Exceptions are Putirka’s liquids from the Galapagos and Jan Mayen, which yield P–T values overlapping those of the SRP. Part of the overall scatter may relate to difficulties in defining the parental liquid compositions. In many OIBs, pyroxene is a significant liquidus phase along with olivine, and its omission in the back-calculation could introduce large uncertainty in the melt composition. To partly circumvent this problem, data for olivine–pychic Kilauea Iki lavas (Hawaii; Leeman and Scheidegger, 1977) were processed for comparison with other OIB estimates. Although the starting samples ranged widely in MgO content (7–14%), the back-calculation resulted in similar parental compositions for each specific source Mg#. Based on petrologic arguments (Herzberg et al., 2007; Putirka et al., 2007), the source Mg# for Kilauea is plausibly near 91, resulting in preferred segregation T near 1610 ± 10 °C (one sigma, 5 samples) and P near 3.5 ± 0.1 GPa (or 115 ± 4 km depth); these statistics only reflect the reproducibility among the five samples used and actual uncertainties could be several times larger. However, this result is similar to P–T estimates of Herzberg and O’Hara (2002) for Hawaiian basalts, and suggests that the Hawaiian magma source is both warmer and more refractory than that for the SRP. Also, this result is qualitatively consistent with a significant garnet signature in REE profiles for Kilauea and other Hawaiian basalts (Fig. 3C, D). The effect of varying source Mg# in this case is shown in Fig. 6. Lower source Mg#s (e.g. 89–90) for Kilauea will bring segregation P–T values closer to the preferred range for SRP basalts, but would predict melting in the spinel lherzolite stability field, and is inconsistent with the REE data. To facilitate comparisons of P–T conditions, adiabats are projected through the likely maximum P–T conditions for selected lava suites, and the temperature intercepts (Tp) highlighted in Fig. 6. These values are considered lower limits, particularly where calculated P–T values lie above the dry solidus. From this comparison, it appears that the SRP mantle Tp is at most ~50 °C higher than that for the warmest MORB average, which we consider representative of ambient asthenospheric mantle. In contrast, most OIB sources appear to be considerably warmer. Mantle Tp values for Iceland and Hawaii (Kilauea) are roughly 60° and 120 °C warmer than our preferred upper limit for SRP. Finally, it is of interest to contrast the SRP with nearby areas of the tectonically active western United States. First, we note that SRP potential temperatures are comparable to estimates for the southern and central Basin and Range province (Wang et al., 2002). Second, we show (Fig. 6) estimated P–T conditions for primitive basalts from the High Lava Plains (HLP) of eastern Oregon using the data of Draper (1991), assuming that source Mg# = 90, and following the same approach used for the SRP, Kilauea, and MORB. This area is located due west of the SRP and constitutes the northernmost part of the Basin and Range extensional province. We find extensive overlap in P–T segregation conditions between HLP, SRP, and the upper end of the MORB field. This result implies that mantle thermal conditions below the SRP and the adjacent parts of the Basin and Range province are more typical of regional lithospheric extension than upwelling of an unusually warm mantle plume. Since a plume-like feature is imaged beneath Yellowstone, we conclude that it must have a relatively weak thermal signature compared to other plume-related or hotspot volcanoes (Watson and McKenzie, 1991; Thompson and Gibson, 2000; Herzberg and O’Hara, 2002; Putirka, 2005). Based on our earlier discussion, the velocity anomaly associated with this feature must be attributed solely to thermal effects as melting is unlikely to occur at depths greater than ~100 km. However, if SRP basaltic magmas actually form by melting lithospheric mantle, they provide no direct constraints on thermal conditions within the imaged plume. 7. Conclusions and implications for magma genesis Here, we summarize the principal conclusions derived from this study and briefly comment on their implications regarding genesis of SRPY basaltic magmas.

1. A first order result is that maximum melt segregation P–T conditions, geochemical arguments, and seismic tomography for the SRPY province preclude melting at depths much greater than 100 km. In this case, the geochemical results provide a unique constraint in that it is difficult to define precisely the lower limit of the tomographic anomaly using seismic data alone.

2. At most, there appear to be only weak thermal contributions from sublithospheric sources, and the tomographically imaged Yellowstone plume is unlikely to have sufficient thermal strength to erode a thick lithospheric plate (cf. Manea et al., 2009-this volume). Isotopic data for SRP basalts indeed suggest that the subcontinental lithospheric mantle has remained intact and is involved in magma generation or modification. Elevated 3He/4He (>13 Ra) in Yellowstone fluids (Craig et al., 1978) and SRP basalts (Graham et al., 2009-this volume) can be reconciled with this scenario if He is selectively transported from upwelling deep mantle to the lower lithosphere (e.g., in a discrete fluid/gas phase).

3. The locus of estimated segregation P–T conditions is proximal to (or slightly above) the dry peridotite solidus. This result is inconsistent with melting of hydrated mantle to produce SRP basalts. If melting involved portions of lithospheric mantle that had been previously hydrated by fluids derived from the subducting Farallon slab (e.g., during Laramide ‘low angle subduction’; cf., Humphreys et al., 2003), we would expect to see magmatic temperatures lower than the dry peridotite solidus.

4. The potential temperature (Tp) inferred for the SRPY mantle is unlikely to exceed ~1450 °C. This is the maximum value that is consistent with all available constraints. Similar values are also inferred for mantle beneath the HLP, other parts of the Basin and Range province, and portions of the mid-ocean ridges. Tp of this magnitude is representative of asthenospheric mantle, but could also be associated with thermally equilibrated lithospheric mantle. This result suggests that a mantle plume is not necessarily the driving force behind SRPY magmatism. Rather, thermal conditions beneath the SRP
may be more representative of regional lithospheric extension (cf. Hernlund et al., 2008). Heating of the lower lithosphere by upwellings asthenosphere or a deeper mantle plume could contribute to or enhance mantle melting beneath Yellowstone (as reflected in the He isotopic data), but may not be required to promote melting.

5. An important dilemma remains concerning the relative melt productivity, which today appears to be greater in the SRP than in the adjacent Basin and Range province. In part this may be an artifact of the diachronous nature of extension and magmatism in the western US. That is, activity to the south was prolific in the past, and lithospheric contributions have largely waned in lieu of asthenospheric contributions. In contrast, at the latitude of the HLP-SRP, magmatism may be in an early, more proli

Acknowledgements

Leeman and Schutt acknowledge support from the National Science Foundation for time spent in preparation of this paper. Hughes acknowledges the NASA Idaho Space Grant Consortium and the EPSCoR program for supporting laboratory analyses and student research on the Snake River Plain. We thank Mike McCurry, Cin-Ty Lee, Claude Herzberg, and Keith Putirka for discussions about this work. Finally, we thank Keith Putirka, Chris Hawkesworth and an anonymous reviewer for helpful suggestions to improve the paper.

Appendix A

Appendix Fig. A1. Plot of Mg# for host SRP lavas and maximum forsterite (Fo) content of their olivine phenocrysts. Data sources are provided in the Appendix. Curves are shown for equilibrium loci of olivine and host liquid compositions as dictated by experimentally determined Fe–Mg partitioning.

Appendix B. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.jvolgeores.2009.01.034.

References


