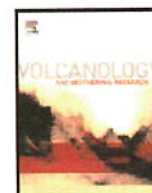




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Thermal structure beneath the Snake River Plain: Implications for the Yellowstone hotspot

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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.

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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 basalt activity (Camp and Ross, 2004). Silicic volcanism was associated with large eruptive centers, each active for 2–4 m.y., and it migrated northeastward to 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 $^{87}\text{Sr}/^{86}\text{Sr}$ and lower $^{143}\text{Nd}/^{144}\text{Nd}$,

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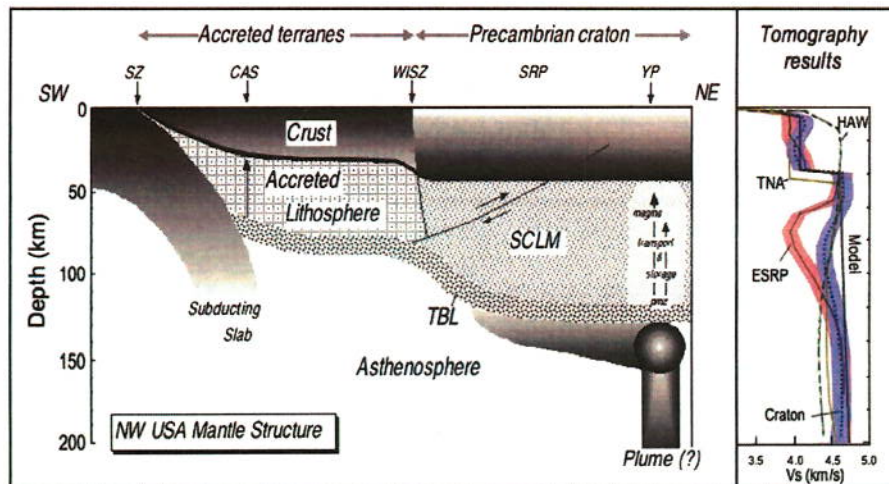


Fig. 1. Schematic lithospheric structure for NW U.S.A., with results of V_s tomography from the eastern Snake River Plain (SRP). Diagram portrays subcontinental lithospheric mantle (SCLM) beneath Precambrian cratonic North America, with locus of a postulated plume impinging beneath the Yellowstone Plateau (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 V_s tomography and estimated magma segregation depths (~60 to 100 km; this paper). Profiles show mean V_s 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).

$\delta^{11}\text{B}$, 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 (Rodgers 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 basalt–rhyolite character. However, even the earliest silicic activity can be linked to inputs of mafic magmas into the crust (e.g., Feeley and Gruner, 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-

tite is unlikely to produce much magma and then only after significant lithospheric thinning (i.e., substantially later than inception of extension). Hydrated lower lithospheric mantle also could melt in rapid response to extension, but this process is less capable of sustaining the documented large magma production rates over time in the Basin and Range province (Best and Christiansen, 1991).

Available data (e.g., Fitton et al., 1988, 1991; Kempton et al., 1991; Daly and DePaolo, 1992; Bradshaw et al., 1993) suggest that the mafic magmas changed character over time, with the earliest having lithospheric affinity (e.g., higher $^{87}\text{Sr}/^{86}\text{Sr}$, lower $^{143}\text{Nd}/^{144}\text{Nd}$, and fractionated incompatible element signatures compared to younger basalts) whereas, once extension exceeded a critical magnitude (roughly 75–100%), younger mafic magmas tend to have asthenospheric affinity (OIB-like chemistry). This transition is attributed to the notion that, once lithosphere has been critically thinned by extension, magma generation may be dominated by melting of upwelling asthenospheric mantle – i.e., the binary source model of Leeman and Harry (1993). It appears that such a transition in sources for basaltic magmas occurred by early- to mid-Miocene time in southern California, but has only migrated as far as north-central Nevada at present (Lum et al., 1989). Thus, SRPY basalts could be derived from lithospheric mantle based on the fact that their Sr–Nd–Pb isotopic compositions are distinctive from most OIB or MORB (Leeman and Manton, 1971; Doe et al., 1982; Menzies et al., 1984; Hildreth et al., 1991).

2.2. Sub-lithospheric processes (hotspot influence)

Alternatively, evidence that a mantle plume underlies the Yellowstone area warrants consideration that SRP basaltic magmas may have been derived from such a source, but acquired their unusual isotopic compositions as a consequence of interaction with old lithospheric rocks (cf. Hanan et al., 2008). If the tomographically imaged anomaly beneath Yellowstone truly is a mantle plume, is it capable of producing SRPY basaltic magmas? To address this question we consider the following issues: (1) Under what circumstances can the mantle melt to produce basaltic magmas? (2) What is the cause of the tomographic anomaly and, specifically, what do the seismic data

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 H_2O). 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% H_2O (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., $T_p \sim 1450^\circ\text{C}$ 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

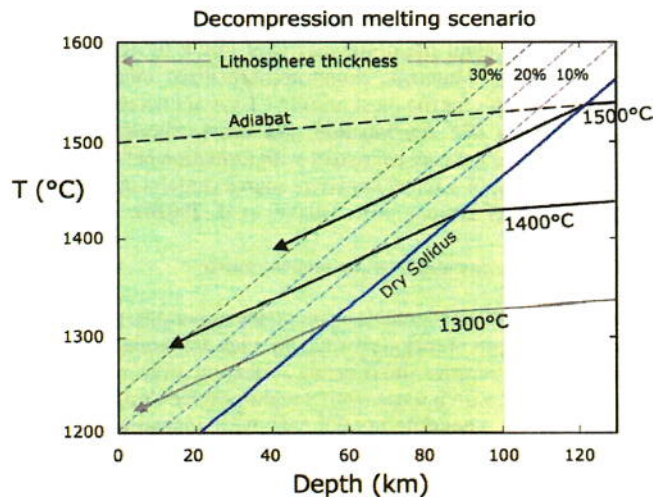


Fig. 2. Schematic diagram illustrating decompression melting scenarios for mantle material ascending along three representative adiabatic paths with potential temperatures (T_p) of 1300 °C, 1400 °C, and 1500 °C. The dry peridotite solidus is encountered at progressively greater depth (roughly at 60, 90, and 120 km, respectively) with increasing T_p . Dashed lines above the solidus are contours for degree (%) of melting. Note that heat required to produce melt causes temperature to decrease from the ideal adiabatic path once melting begins. Superposition of a viscous and chemically distinct lithospheric lid (e.g., 100 km thick shaded area) will impede ascent of upwelling asthenospheric or plume mantle. For the example shown, mantle with $T_p \leq 1430^\circ\text{C}$ will encounter the lithospheric lid before the dry solidus is reached; in these cases, no melting can occur until the lid is thinned sufficiently to allow upwelling mantle to intersect the solidus. In contrast, ascent paths with higher T_p will cross the solidus at depths below the lithospheric lid, and melting can occur. For upwelling mantle to melt beneath a cratonic lithosphere with thickness of 120 km, it must have a T_p in excess of 1500 °C, and so on. If the Wyoming craton has a thickness ≥ 150 km, T_p must exceed $\sim 1640^\circ\text{C}$ for melting to ensue.

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 $T_p < 1450^\circ\text{C}$ (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., $T_p \geq 1500^\circ\text{C}$ 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_s 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_s as a function of depth beneath Yellowstone; at depths near 115 km, V_s 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 Leshner, 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_s 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 $T_p = 1320^\circ\text{C}$ (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 ($= 1450^\circ\text{C}$, as suggested by Putirka et al. (2007), Herzberg et al. (2007), and our own work discussed in the following section), we find that V_s 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_s anomaly (Fig. 3B). In the extreme case that no melt is present, the anomaly requires a maximum excess temperature of $\sim 150^\circ\text{C}$ 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 $d\ln V_p/d\ln V_s$ (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.

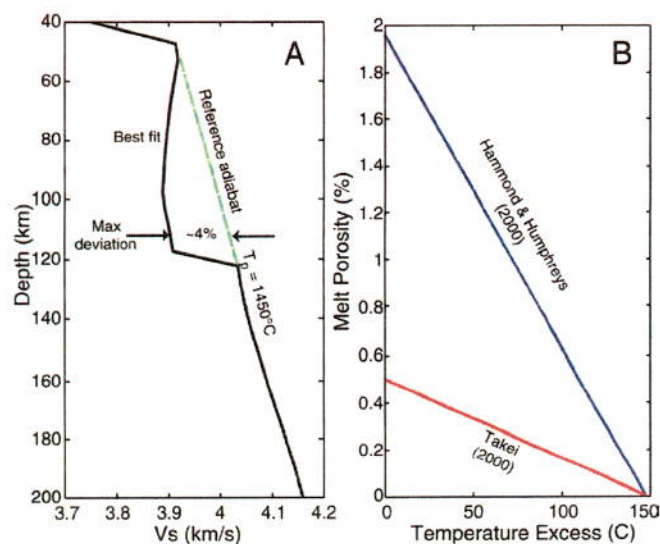


Fig. 3. Example of tomographic constraints on temperature and melt porosity for the SRPY mantle. (A) Variation in V_s as a function of depth below Yellowstone, as determined from forward modeling of the effects of temperature and grain size on fundamental mode Rayleigh wave phase velocities. Results shown are the best fitting V_s structure to the observed phase velocities, within the range of temperatures and grain sizes modeled, subject to the constraint that mantle below 120 km has $T_p = 1450$ °C (reference model) – thus the sharp V_s transition at this depth fits the data, albeit not uniquely. For comparison we show an upper mantle reference model (dashed line) with assumed $T_p = 1450$ °C. Maximum deviation (-4%) between these profiles is near 110 km depth. This velocity anomaly corresponds to either maximum excess temperature of 125 °C or melt porosity of 0.2% in the hypothetical ascending plume (see text). (B) Diagram showing trade-offs between temperature and melt porosity that can produce a -4% velocity anomaly near 110 km depth. Melt-velocity scaling follows the relations of Takei (2000) or Hammond and Humphreys (2000). Temperature-velocity scaling is calculated using the model of Faul and Jackson (2005) with activation volume (V^*) = 14 cm³/mol, activation energy = 510 kJ/mol, and frequency dependence of attenuation (with $\alpha = 0.15$ and wave period = 50 s). Note that V_s data only approximately constrain depths of melting because modeling results are non-unique. See Schutt et al. (2008) for further details on the inversion, and Schutt and Dueker (2008) for more on the forward models.

Thus, the seismic tomography results do not support the notion that the Yellowstone plume is warm enough to begin melting at depths greater than ~150 km. This, in turn, implies that the lithosphere headroom problem could be significant. Melt formation at depths of 50–100 km (cf. Fig. 3A) within an ascending plume is likely only if the lithospheric lid has been thinned significantly in the short time (a few m.y.) of plume impingement. Two lines of evidence suggest that this has not occurred. First, geochemical evidence implies that an old lithospheric mantle source is involved (Doe et al., 1982; Hildreth et al., 1991) and raises doubt that the lid has been thinned. Furthermore, geodynamic modeling suggests that impingement of a mantle plume is unlikely to result in dramatic thinning of a moving cratonic lithosphere (Burov and Guillou-Frottier, 2005; Sleep, 2006; Burov et al., 2007; Manea et al., 2009–this volume). The alternative scenario, that lithospheric mantle was melted, requires that either temperatures are substantially higher than ambient asthenospheric mantle or that the lithospheric source is relatively fertile (or hydrated). To evaluate these scenarios, petrologic and geochemical characteristics of primitive SRP basalts are used to estimate the depths and temperatures over which their parental magmas likely equilibrated with the mantle.

5. Petrologic constraints on the melting conditions

5.1. Primitive Snake River Plain basalts

SRP basalts are typically olivine $\pm$ plagioclase phyric olivine tholeiites (SROT) with rare 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 Fo₈₇) are typically in equilibrium with the host magma, although rims are commonly more Fe-rich; relatively Mg-rich (to Fo₉₁) 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) $Mg\# = 100 \times MgO / (MgO + FeO^*)$ 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 $10.0 \pm 0.7\%$ 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). V/Sc 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 notably 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 SROT 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.

Column #	1	2	3	4	5	6	7	8	9
	Bruneau Jarbidge	Bruneau Jarbidge	INEL drillhole	Mt. Bennett Hills	West of Arco	Spencer Kilgore	Spencer Kilgore	Island Park	McKinney Basalt
Sample ^a	I-524	I-755	123-505	L82-11	74-26	L73-64	L73-112	6YC-142	70-15B
SiO ₂ ^b	48.46	47.98	48.69	48.81	47.86	48.69	48.16	47.77	47.15
TiO ₂	0.98	1.33	1.38	1.18	1.07	1.21	1.32	1.14	3.03
Al ₂ O ₃	17.27	16.68	15.14	16.26	16.35	14.80	15.51	16.98	15.22
FeO	9.81	10.41	10.43	9.99	10.18	10.56	10.61	10.57	13.45
MnO	0.17	0.19	0.17	0.16	0.18	0.18	0.18	0.18	0.21
MgO	8.78	9.06	10.18	9.43	9.68	10.72	10.58	9.14	7.59
CaO	12.10	11.85	10.91	11.63	12.31	11.03	10.54	11.47	9.51
Na ₂ O	2.11	2.09	2.29	2.20	2.09	2.16	2.33	2.35	2.59
K ₂ O	0.19	0.21	0.38	0.22	0.17	0.36	0.41	0.24	0.74
P ₂ O ₅	0.14	0.19	0.43	0.12	0.11	0.28	0.35	0.16	0.50
Mg#	61.5	60.8	63.5	62.7	62.9	64.4	64.0	60.7	50.1
Ni	141	145	250	114	80	232	156	129	108
Cr	304	285	448	571	467	540	594	318	251
⁸⁷ Sr/ ⁸⁶ Sr	0.70626	0.70604	–	–	0.70668	0.70596	0.70540	0.70614	0.70679
¹⁴³ Nd/ ¹⁴⁴ Nd	0.512484	0.512508	–	–	0.512499	0.512375	0.512379	0.512460	0.512401
La	6.2	9.5	11.2	7.7	5.04	11.34	15.98	6.28	21.97
Ce	14.1	21.7	26.6	16.9	10.8	22.42	31.56	13.07	45.36
Pr	–	–	–	–	1.54	2.85	3.85	1.82	5.83
Nd	9.5	13.8	13.0	11.9	7.17	12.76	16.61	8.52	26.4
Sm	2.62	3.83	3.26	3.1	2.26	3.52	4.29	2.68	6.91
Eu	0.99	1.31	1.26	1.21	0.88	1.26	1.50	1.10	2.51
Gd	–	–	–	–	2.67	3.75	4.33	3.08	6.78
Tb	0.66	0.83	0.57	0.69	0.51	0.68	0.80	0.59	1.19
Dy	–	–	–	–	3.31	4.35	4.88	3.7	7.05
Ho	–	–	–	0.90	0.73	0.92	1.01	0.79	1.40
Er	–	–	–	–	2.19	2.65	2.94	2.29	3.90
Tm	0.42	0.47	–	0.42	0.3	0.37	0.40	0.31	0.51
Yb	2.59	3.05	2.14	2.40	1.89	2.25	2.45	1.92	3.08
Lu	0.41	0.50	0.32	0.42	0.30	0.35	0.38	0.30	0.48
Method ^c	N	N	N	N	I	I	I	I	I

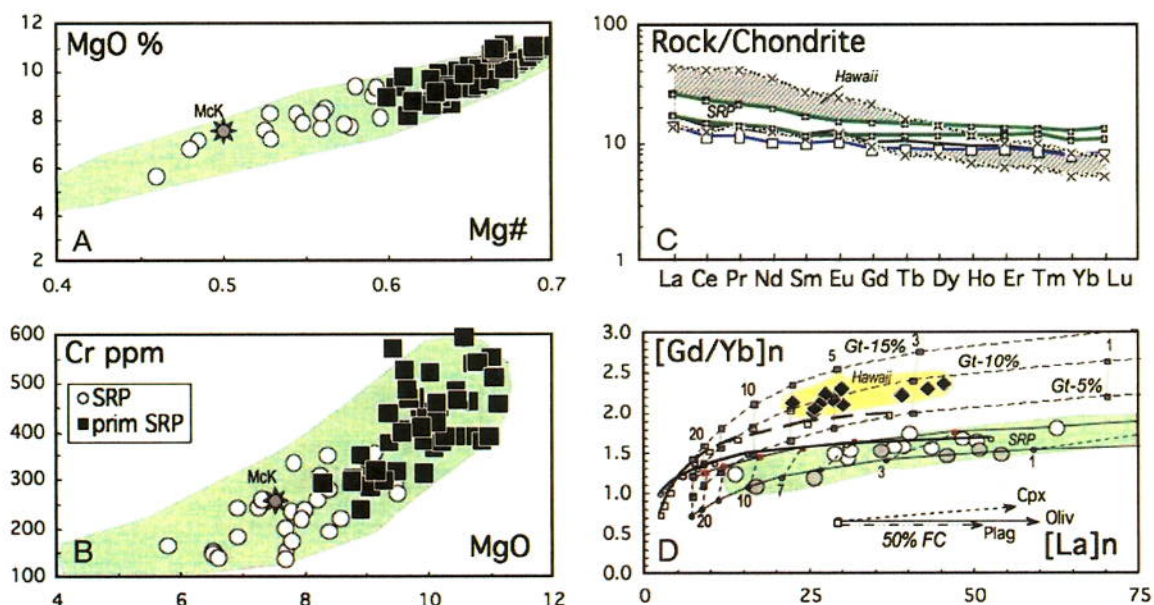
^a Areas: western SRP (column #s 1,2, and 4), central SRP (#s 5,9), eastern SRP (#s 3,6–8).^b Analyses normalized to 100% total, anhydrous with all Fe as FeO⁺; USGS wet chemistry (# 8), XRF (all others). Sr and Nd isotopes via TIMS (cf. Leeman et al., 1992).^c Rare earth elements determined by neutron 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 Kahoolawe and Kilauea tholeiites; 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 T_p of ~1450 °C (i.e., extrapolated to the surface).

Virtually the same result is obtained when the data are modeled following the approach of Putirka (2005), who considered the effects of H_2O content on olivine–melt Fe–Mg equilibria. Limited data on melt inclusions from SRP basalts suggest that magmatic water contents are unlikely to exceed 0.4% (P. Wallace, personal communication). Using Putirka's (2005) composition-dependent expression for the liquidus surface, we find that small amounts of H_2O (up to 1%) have little effect on estimated temperatures. Also, comparison between hydrous and anhydrous experiments indicates that calculated olivine temperatures are overestimated by only ~20 °C for each 1 wt.% H_2O in the magma (cf. Leeman et al., 2005). Thus, calibration uncertainty (± 25 °C) on the olivine–melt thermometer is probably more significant than effects of water on the temperature estimates. Finally, we compare our estimates with an independent approach based on thermochemical modeling of silica activity and olivine–melt–orthopyroxene equilibria. Using this approach, Leeman and Vitaliano (1976) predicted

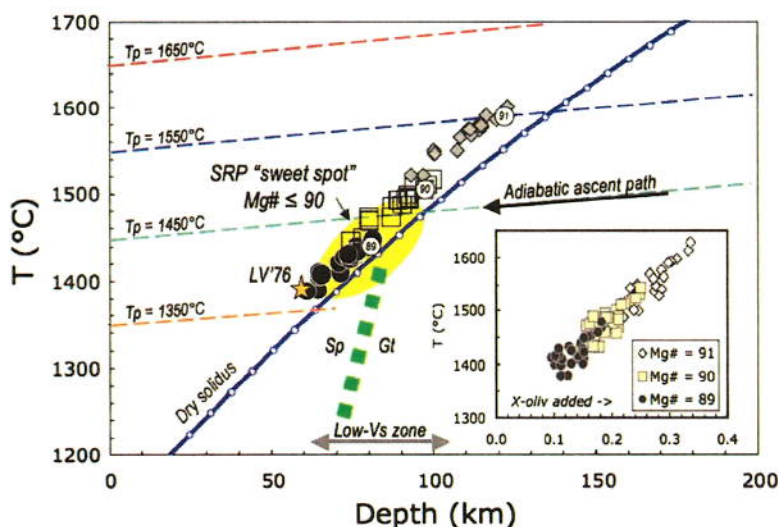


Fig. 5. Temperature–pressure systematics. Representative adiabats (dashed lines) are shown for upwelling asthenospheric mantle with a range of potential temperature (T_p) estimates. The dry peridotite solidus of Hirschmann (2000) is provided for reference. Bold arrow illustrates an ascent path for mantle with $T_p = 1450$ °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 V_s inferred from seismic tomography. Superimposed on this diagram are estimated temperatures and pressures for last equilibration of primitive SROT 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.

segregation of a relatively evolved olivine + plagioclase phyric 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 decompression 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 $X\text{-Fe}^{2+} = 0.9$ based on petrologic constraints) and that Fe–Mg partitioning ($K_d = 0.3$) between olivine and melt is constant (we allow K_d 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 maximum V_s 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 SRP melting anomaly, it is important to consider

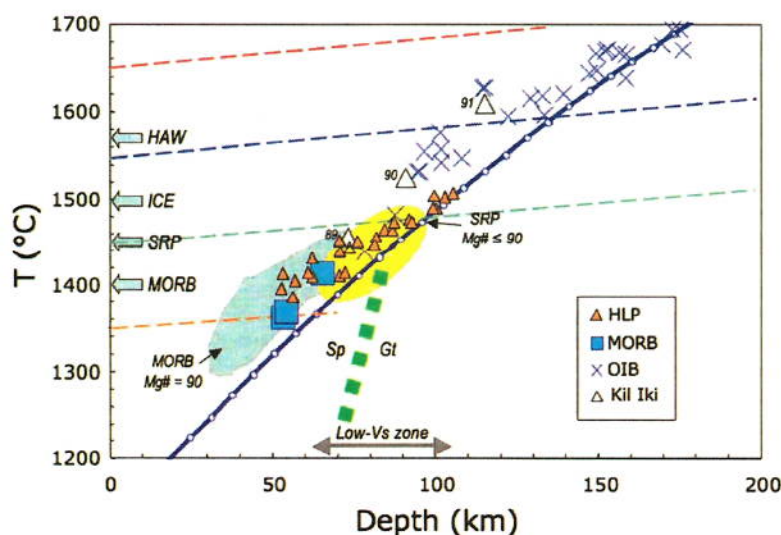


Fig. 6. Overview of P – T estimates comparing results for SRP basalts (“sweet spot” field from Fig. 5) with oceanic and other regional basalts, all calculated in a self-consistent manner. Pressure is based on the Albarède (1992) method with corrections discussed in text; temperature is based on the Sugawara (2000) olivine–melt thermometer. Primitive magma compositions for ocean island (“hotspot”) localities and average MORB estimates are taken from Herzberg and O’Hara (2002) and Putirka (2008); for these cases, P and T were calculated from the published analyses, and calculated equilibrium source Mg# typically ranged between 90 and 91. A field is also shown for several hundred MORB analyses (PETDB database) with greater than 8% MgO; compositions were adjusted by olivine addition and P and T calculated assuming source Mg# = 90. A comparable P – T estimate (upper limit) for average SRP basalts is shown by an arrow. The effect of varying source Mg# is again illustrated using data from the 1959 Kilauea Iki eruption (Hawaii). In this case, actual source composition is probably close to Mg# = 91 (Herzberg and O’Hara, 2002; Putirka et al., 2007), so the higher P – T estimates are considered most realistic. Arrows on the left hand side indicate approximate T_p values for selected localities based on extrapolation of adiabats through maximum P – T loci for each suite; for Hawaii and Iceland, these values are similar to estimates of Herzberg and O’Hara (2002). The apparent MORB adiabat is intermediate between early estimates (e.g., White and McKenzie, 1991) and those of Putirka (2005). SRP mantle T_p appears to be no more than ~50 °C higher than MORB mantle T_p (based on the warmer MORB average). P – T estimates are also shown for primitive basalts from the High Lava Plains (HLP; eastern Oregon) using data of Draper (1991) and assuming source Mg# = 90. These results suggest that mantle thermal conditions beneath the SRP and the extensional HLP are comparable to each other and only slightly warmer than the ambient sub-oceanic mantle.

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 T s 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 T_p 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–phyric 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 hercynite 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 (T_p) 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 T_p 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 T_p 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 $^3\text{He}/^4\text{He}$ (>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 (T_p) 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. T_p 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 upwelling 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 there have largely waned in lieu of asthenospheric contributions. In contrast, at the latitude of the HLP-SRP, magmatism may be in an early, more prolific stage – perhaps due to the involvement of lithospheric mantle. Regional melt productivity could also reflect lateral differences in source fertility. Modeling of segregation P – T can be reconciled with other independent constraints only if source Mg# is less than 90. Accordingly, average fertility of the subcontinental lithospheric mantle associated with the Archean Wyoming craton appears to be higher than that for other cratons (e.g., based on compositions of mantle xenoliths). Thus, lateral variation in source fertility provides a possible explanation for high melt productivity in the SRPY province.

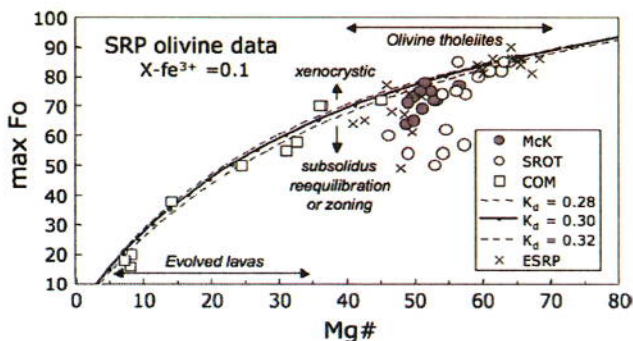
This work highlights several outstanding research questions that remain to be resolved. Foremost in discriminating between lithospheric vs. plume source alternatives would be quantitative determination of the present-day thickness of the mechanical lithospheric lid. If it is much greater than 100 km, significant magmatic contributions from a warm plume are highly unlikely. In closing, we hope that this paper provides incentive to seriously evaluate alternative ‘upper plate’ mechanisms for magma generation.

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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.

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