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High-K alkali basalts of the Western Snake River Plain: Abrupt transition from tholeiitic to mildly alkaline plume-derived basalts, Western Snake River Plain, Idaho

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ABSTRACT

Basaltic volcanism in the western Snake River Plain underwent an abrupt change circa ~700 ka to 900 ka, from low-K tholeiitic basalt and ferrobasalt to high-K transitional alkali basalt. The low-K tholeiitic basalts share major element, trace element, and isotopic characteristics with olivine tholeiites of the eastern Snake River Plain, and must have been derived by similar processes from similar sources. In contrast, the younger high-K alkali basalts share major element, trace element, and isotopic characteristics with plume-derived alkali basalts of ocean islands suites like Hawaii. We conclude that this abrupt transition reflects either or both the erosion of pre-existing mantle lithosphere in the wake of the Yellowstone–Snake River plume, or the depletion of this lithosphere in fusible components so that it no longer contributed to the overall mass flux of magma. The abruptness of the transition implies that it may have a catastrophic origin, such as lithospheric delamination caused by a Rayleigh–Taylor instability beneath the Idaho batholith.

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1. Introduction

The Neogene Snake River Plain (SRP) consists of two distinct terranes with different crustal structures, stratigraphy, and volcanic histories. The northeast-trending eastern SRP is thought to mark the track of the Yellowstone–Snake River plume from its mid-Miocene location beneath the Owyhee Plateau to its current location under Yellowstone (Smith and Braile, 1994; Pierce et al., 2002; Shervais et al., 2006). In contrast, the western SRP is a northwest-trending graben filled with up to 1.7 km of sediment (Wood and Clemens, 2002). Volcanic activity in the western SRP occurred in three stages, none of which are directly related to the track of the Yellowstone–Snake River hotspot: rhyolite volcanism associated with the opening of the western SRP graben (~10–12 Ma), basaltic volcanism that post-dates the rhyolites and underlies the thick lacustrine deposits (~7–9 Ma), and finally basaltic volcanism that largely post-dates the lacustrine deposits and is coeval with basaltic volcanism of the eastern SRP (<2.2 Ma: Amini et al., 1984; Vetter and Shervais, 1992; Shervais et al., 2002; White and Hart, 2002; Wood and Clemens, 2002; Shervais et al., 2005).

We review here the transition, circa 700–900 ka, from tholeiitic olivine basalts and ferrobasalts that form broad upland plateaus and large shield volcanoes along both the Snake River and Boise River South Fork, to mildly alkaline high-K lavas that form smaller shield volcanoes and cinder cones atop the plateaus. The abrupt change from

tholeiitic lavas with less than 0.7 wt.% K₂O to mildly alkaline lavas with 1.0–2.5 wt.% K₂O reflects a fundamental change in the source regions of these basalts that is related to the interaction of the Yellowstone–Snake River plume with pre-existing mantle lithosphere. The tholeiitic basalts and ferrobasalts have major and trace element concentrations similar to other Snake River olivine tholeiites and to Hawaiian tholeiitic basalts of the “shield forming” stage; in contrast, the high-K lavas have trace element and isotopic signatures consistent with their derivation from a mantle plume similar in composition to the mantle source of ocean island basalts, and mimic the trend seen in Hawaiian volcanoes from tholeiitic shield-building lavas to alkaline post-shield lavas.

We conclude that this transition reflects either or both the erosion of pre-existing mantle lithosphere in the wake of the Yellowstone–Snake River plume, or the depletion of this lithosphere in fusible components so that it no longer contributed to the overall mass flux of magma. The abruptness of the transition implies that it may have a catastrophic origin, such as lithospheric delamination caused by a Rayleigh–Taylor instability beneath the Idaho batholith. This model is consistent with the recent uplift of the southern portion of the Idaho batholith and the deep incision of the rivers that transect it.

2. Geologic setting

Volcanic activity in the western SRP began ~12 million years ago, coeval with extension and graben formation, with the eruption of rhyolite lavas from vents and fissures parallel to the range-front faults that bound the graben (Wood and Anderson 1981; Clemens and Wood, 1993; Bonnicksen and Godchaux, 2002). The range-front faults

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are essentially continuous along the northern margin of the SRP, separating younger sediments and basalts of the SRP from rhyolites of the Danskin Mountains and western Mount Bennett Hills, and granitic rocks of the Idaho batholith. Range-front faults are discontinuous along the southern margin of the western SRP, and basement assemblages south of these faults include rhyolites and outliers of both the Challis volcanics and Idaho batholith (Ekren et al., 1984).

Basaltic activity in the western SRP began about 9 million years ago, forming lavas that underlie sedimentary deposits of Lake Idaho and, later, local basalt horizons intercalated with these sediments (Jenks and Bonnicksen, 1989; Wood and Clemens, 2002; White and Hart, 2002; Bonnicksen and Godchaux, 2002). These basalts crop out in the central and southern portions of the SRP and have been intersected by deep drilling (Arney et al., 1982, 1984; Shervais et al., 2002) and imaged seismically (Wood, 1994).

Pleistocene and latest Pliocene basalts less than 2.2 Ma are exposed north of the Snake River near Mountain Home, Idaho (Shervais et al., 2002, 2005), farther west near Kuna–Melba, Idaho (White and Hart, 2002), and along the Boise River South Fork near Prairie, Idaho (Vetter and Shervais, 1992), some 40 km north of the Snake River (Fig. 1). In all three areas the youngest flows (<700–900 ka) form smaller shield volcanoes and cinder cones that overlie the older basalts. These younger flows are characterized by high K₂O contents and distinct isotopic compositions that set them apart from the older tholeiitic basalts (Vetter and Shervais, 1992).

3. Occurrence

The transition from low-K tholeiitic basalt to high-K alkaline basalt was first recognized by Vetter and Shervais (1992) in the Boise River

South Fork drainage near Prairie, Idaho (Fig. 1). Basalts within the Boise River South Fork drainage erupted through granites of the Idaho batholith and filled pre-existing canyons, forming hyaloclastite-pillow deltas capped with subaerial basalt flows (Howard and Shervais, 1973; Howard et al., 1982). Repeated eruptions filled the Boise River South Fork canyon and were subsequently incised to form new canyons that were filled by later eruptions, resulting in an inverted stratigraphy within the canyons in which younger flows crop out at lower elevations than earlier flows (Howard et al., 1982). Vetter and Shervais (1992) distinguished two magmatic suites: Boise River Group 1 (tholeiitic basalt) and Boise River Group 2 (younger alkaline basalts). The BRG-1 lavas were dated by Howard et al. (1982) at 1.8 to 1.9 Ma; these lavas form the largest canyon-filling plateaux that underlie Prairie; similar lavas appear to underlie the Anderson Ranch reservoir area north of Camas Prairie. The BRG-2 lavas erupted from small shield volcanoes and cinder cones that sit either on the BRG-1 basalt plateaux or directly on granite. They range in age from ~0.68 Ma to 0.20 Ma (Howard et al., 1982). Despite the potential for interaction with the granitic basement, only one flow exhibits slightly elevated $\delta^{18}\text{O}$ values (~6.4‰); most flows have $\delta^{18}\text{O}$ values below ~6.0‰, indicating that they are essentially uncontaminated with crustal material (Vetter and Shervais, 1992).

In the area around Kuna–Melba, Idaho, White and Hart, (2002) recognized three mafic magma groups, which they called M1, M2, and M3. The M1 lavas, exposed sporadically near the margins of the plain, are the oldest (7–9 Ma) and pre-date all of the Boise River Group 1 lavas. The M2 basalts (1.8 to 0.9 Ma) correspond the BRG-1 lavas of Vetter and Shervais (1992), while the M3 basalts (0.76 to 0.387 Ma) correspond to the BRG-2 lavas of Vetter and Shervais (Othberg et al., 1995; White and Hart, 2002). Many of the older M2 basalts erupted

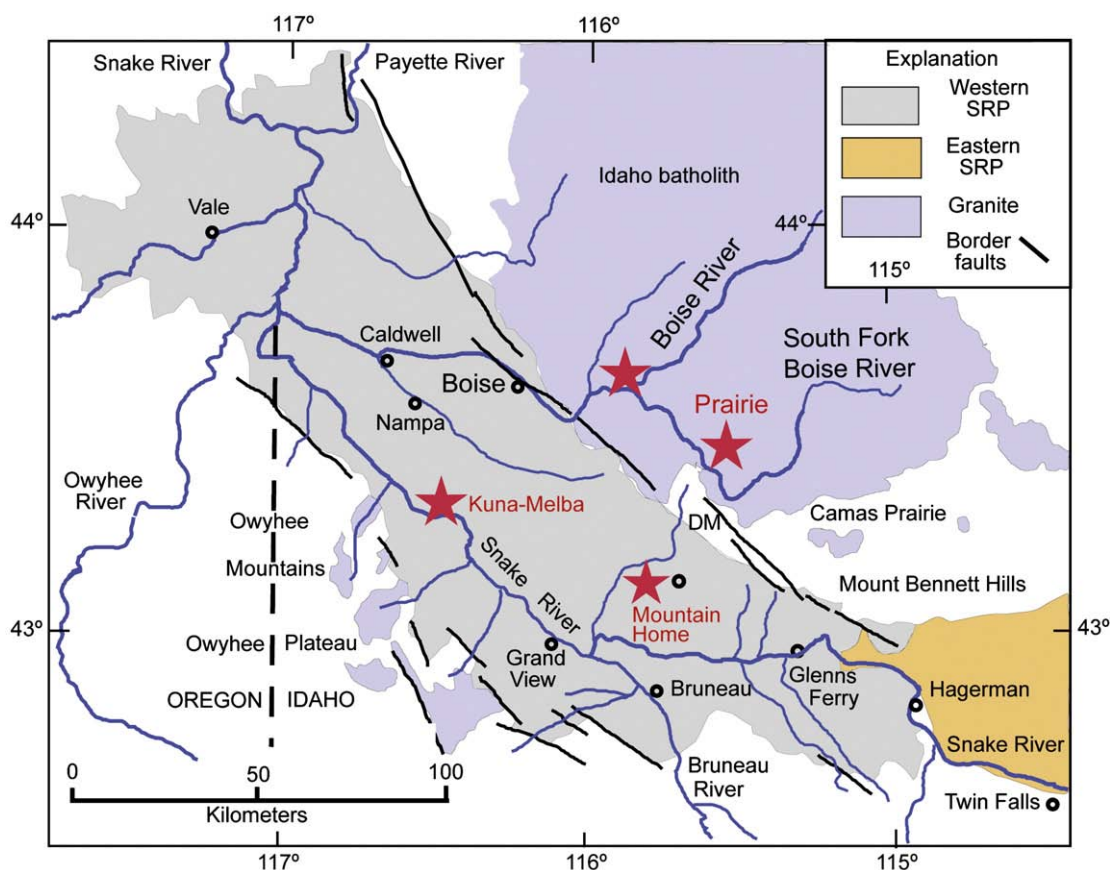


Fig. 1. Location map of southern Idaho and the western Snake River Plain. Areas with young high-K transitional alkali basalts and older low-K tholeiitic basalts shown with red stars: the Kuna–Melba area south of Boise, the Mountain Home area farther east along the Snake River, and Prairie, on the Boise River South Fork. Also shown are adjacent parts of the eastern SRP province and the Idaho batholith. Unpatterned areas include both rhyolite volcanics and Neogene sediments.

within Lake Idaho, forming hyaloclastite and pillow cones that are commonly capped by subaerial basalts (Bonnichsen and Godchaux, 2002; White and Hart, 2002). The younger M3 basalts are commonly wholly subaerial, having erupted after Lake Idaho drained around ~1.8 Ma (Wood and Clemens, 2002). We note that two of the youngest M2 basalts have high K_2O , while two of the oldest M3 basalts have low K_2O , indicating a period of transition circa ~0.8 Ma (see discussion below).

In the Mountain Home area, most of the older basalts erupted subaerially or through fluvial-deltaic deposits formed along the shore of Lake Idaho as the lake contracted (Shervais et al., 2002, 2005). The oldest tholeiitic basalts form broad shield volcanoes with subdued summit depressions, while the somewhat younger tholeiites form steep-sided shield volcanoes surrounded by lower angle basalt aprons, and commonly capped by distinct summit calderas or pit craters (Shervais et al., 2002, 2005). These basalts form extensive flow fields that cap the underlying lacustrine deposits to form the broad uplands north of the Snake River. The younger high-K basalts erupted from two cinder cones near Mountain Home (Union Buttes) and from a small shield volcano farther west (Little Joe Butte). The Union Buttes flows were confined by adjacent shield volcanoes, but one flow from Little Joe Butte flowed down the ancestral Canyon Creek drainage and entered an ancestral canyon of the Snake River near CJ Strike dam (Shervais et al., 2005). There are no radiometric dates on the Mountain Home area basalts, but we can estimate their ages from stratigraphic relationships with the Pleistocene Bruneau formation. Older pre-Lake Idaho basalts have been sampled in core from the Mountain Home Air Force Base; these correspond in age and composition to the M1 basalts of White and Hart, (2002).

4. Petrography

Tholeiitic basalts and ferrobasalts of the western SRP are commonly characterized by coarse intergranular to diktytaxitic groundmass textures, in which plagioclase laths frame vesicles or surround groundmass pyroxene and glass. Plagioclase forms large phenocrysts in the intergranular basalts, with smaller olivine phenocrysts or microphenocrysts. Fractionation of plagioclase by flotation is common in some flows, forming rafts of plagioclase-rich diktytaxitic basalt (McGee and Shervais, 1997; Shervais et al., 2005). The alkali basalts are characterized by small olivine and plagioclase phenocrysts in finer-grained, aphanitic, hyalo-ophitic, or interstitial groundmass containing sanidine and groundmass olivine without reaction rims. Some flows contain large glassy plagioclase phenocrysts up to 2 cm across, up to An_{78} in composition (e.g., Vetter and Shervais, 1992). The olivine phenocrysts are typically more magnesian in the alkali basalts (FO_{73-84} vs FO_{54-80}).

5. Geochemistry

Pleistocene and late Pliocene basalts from of the Western SRP province have been analyzed for major and trace elements by Vetter and Shervais (1992), Shervais et al. (2002), and White and Hart, (2002); representative data are presented in Table 1. These basalts can be divided into two distinct groups based on their K_2O concentrations: a generally older group of tholeiitic basalts and ferrobasalts with low K_2O (<0.7 wt.%) and a generally younger group of mildly alkaline basalts with high K_2O (~1.0 to 2.5 wt.%), as shown in Fig. 2. The older basalts here include BRG-1 of Vetter and Shervais (1992), M2 (mostly) of White and Hart, (2002), and the tholeiitic shield forming basalts of Shervais et al. (2002); we exclude from this analysis the older M1 basalts of White and Hart, (2002) and those sampled by drilling near Mountain Home. The younger alkaline basalts include here BRG-2 of Vetter and Shervais (1992), M3 (mostly) of White and Hart, (2002), and the late shield-cinder cone basalts of Shervais et al. (2002) (Table 1).

5.1. Major elements

The contrast between these two basalt suites is best seen in plots of K_2O vs MgO or $Mg\#$ ($100 \times Mg/[Mg+Fe]$), where the high K_2O concentrations of the alkali basalts place them well above the trend for the low-K tholeiitic basalts, despite their similar range in MgO concentration (Fig. 2). The high total FeO^* contents of the tholeiitic basalts and ferrobasalts results in lower $Mg\#$ s in the tholeiitic basalts than the alkali basalts, rendering any connection between the two suites by fractional crystallization or crustal assimilation unlikely (Fig. 2). The tholeiitic basalts overlap with tholeiitic basalts from the eastern SRP in K_2O – MgO and K_2O – $Mg\#$ systematics, and both overlap the field for Hawaiian tholeiitic basalts; all of these basalts exceed the K_2O content of MORB. The western SRP alkali basalts have K_2O contents in the same range as Hawaiian alkali basalts, but at higher MgO contents and $Mg\#$ s (Fig. 2).

The older basalts are chemically similar to basalts of the eastern SRP (e.g., Hughes et al., 2002; Shervais et al., 2006). All of the tholeiites display a range in MgO contents (~5–9 wt.%) that is essentially the same as mid-ocean ridge basalts (MORB) and tholeiitic basalts of the Hawaiian Islands, with high FeO^* and TiO_2 compared to MORB, and classic tholeiitic fractionation trends on MgO variation diagrams (Fig. 3). The rapid increase in FeO^* and TiO_2 with decreasing MgO , and the concomitant decrease in Al_2O_3 imply a significant component of plagioclase fractionation along with olivine. In contrast, the younger

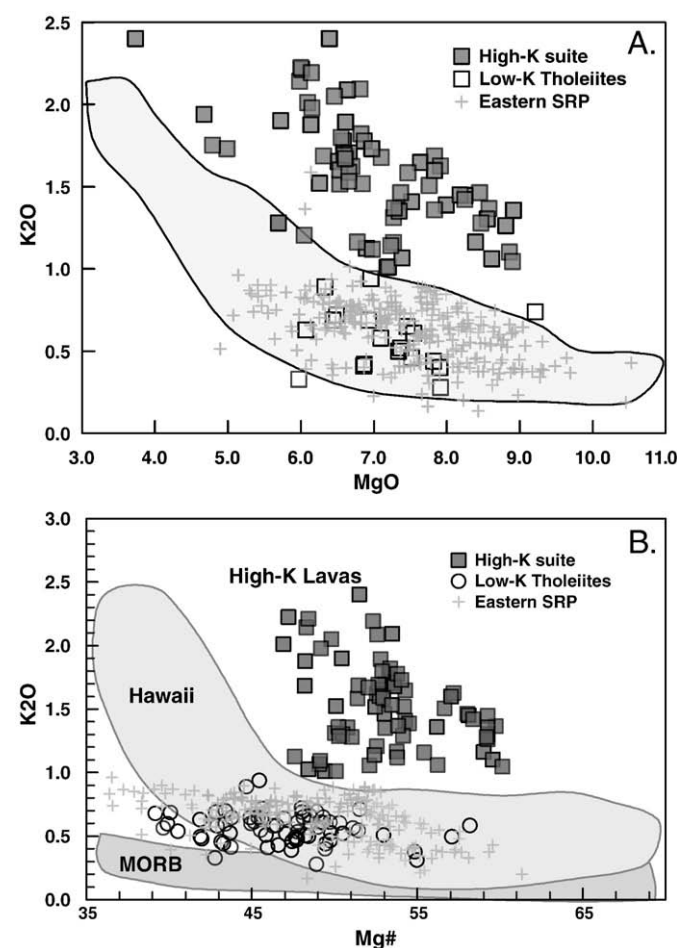


Fig. 2. K_2O plotted as function on (A) MgO wt.% and (B) $Mg\#$. Filled symbols are young high-K transitional alkali basalts, open symbols are older low-K tholeiitic basalts of the western SRP. Fields for Hawaiian basalts and MORB shown in pale grey for comparison, along with data for eastern SRP basalts (Hughes et al., 2002; Shervais et al., 2006). Note that low-K tholeiites of WSRP have same K_2O – MgO relations as basalts of ESPP.

Table 1

Selected whole rock analyses of high-K transitional alkaline basalts from the western Snake River Plain and Boise River South Fork valley

Location	96SRP-38-2	SRP 119	SRP 122	SRP 120	ML-3	DMI-18	DMK-2	DMK-4	HSP-110	HSP-111	HSP 109	SSP-76
Flow Group	Mountain Home				Kuna–Melba				Boise River South Fork			
Sample #	Union		Strike Dam		Halverson cliffs	Initial point	Kuna Butte		Fall Creek			Lava Creek
SiO ₂₀	48.24	48.40	46.92	47.62	48.13	49.25	48.71	49.93	49.57	47.80	48.71	49.08
TiO ₂	2.18	2.25	2.62	2.16	2.47	1.94	2.08	2.52	1.76	2.16	1.69	1.99
Al ₂ O ₃	15.65	15.40	14.97	15.30	16.16	16.29	16.17	16.65	17.41	16.10	17.37	16.73
FeO*	11.53	11.53	13.45	11.56	11.7	10.56	11.12	9.64	10.56	12.80	10.73	11.41
MnO	0.18	0.19	0.21	0.18	0.19	0.17	0.18	0.17	0.18	0.15	0.16	0.18
MgO	7.64	6.60	7.29	7.19	7.23	6.97	7.28	4.67	6.81	7.20	6.39	5.98
CaO	9.44	9.68	9.87	10.53	9.63	9.42	9.33	9.15	8.82	8.97	8.47	8.24
Na ₂ O	3.03	2.83	2.81	2.79	3.03	3.19	3.11	3.65	2.97	2.88	3.00	3.21
K ₂ O	1.29	1.30	1.09	1.21	1.14	1.73	1.37	1.94	2.09	1.01	2.40	2.14
P ₂ O ₅	0.56	0.69	0.68	0.61	0.55	0.48	0.55	0.83	0.37	0.31	0.45	0.43
Total	99.75	98.88	99.90	99.14	100.23	100.00	99.90	99.15	100.55	99.38	99.38	99.40
mg#	54.15	50.48	49.13	52.56	52.41	54.05	53.85	46.33	53.49	50.07	51.50	48.29
Nb	32	30	22	28	30	37	40	58	30	26	35	36
Zr	357	188	153	178	179	156	167	225	171	156	160	177
Y	45	38	36	36	33	26	28	36	27	30	23	18
Rb	8	36	27	31	19	41	30	41	51	15	68	57
Sr	364	383	314	382	379	411	426	447	477	349	473	435
Ni	65	187	127	201	73	84	80	23	78	84	72	43
Cr	133	91	108	105	138	166	152	76	112	150	100	65
V	200	103	108	102	262	224	229	257	174	230	183	191
Sc	23	59	48	46	28	24	23	24	20	23	23	13
Ba	510	812	466	516	392	636	498	810	812	519	773	751
La	11.06	13.24		11.99	20.53	21.62	22.7	34.54	20.49	16.33	20.50	19.21
Ce	26.39	30.39		28.24	43.57	42.77	45.07	67.75	45.50	36.50	44.40	41.90
Pr	3.01	3.55		3.24	5.73	5.18	5.5	8.11				
Nd	13.14	15.30		14.37	25.34	21.97	23.59	33.7	24.10	20.20	24.70	21.00
Sm	3.30	3.81		3.58	6.45	5.25	5.78	8.07	5.19	5.03	5.07	5.05
Eu	1.15	1.26		1.22	2.26	1.8	2	2.54	1.68	1.75	1.62	1.73
Gd	3.43	3.91		3.70	6.81	5.26	5.79	8				
Tb	0.55	0.64		0.61	1.09	0.84	0.93	1.31	0.63	0.76	0.63	0.71
Dy				6.49		5.02	5.57	7.97				
Ho	0.67	0.78		0.74	1.28	0.99	1.13	1.62				
Er	1.61	1.87		1.82	3.44	2.57	2.98	4.37				
Tm	0.23	0.27		0.26	0.47	0.37	0.42	0.62				
Yb	1.47	1.66		1.64	2.89	2.26	2.52	3.69	2.25	2.36	2.12	2.36
Lu	0.24	0.27		0.26	0.45	0.34	0.39	0.58	0.34	0.35	0.32	0.35
Hf	1.56	1.69		1.72	4.22	3.64	4.04	5.49	3.94	3.49	3.88	3.84
Ta	0.65	0.65		0.65	1.81	2.31	2.46	3.66	2.20	1.42	2.27	1.91
Th	2.10	2.21		2.13	1.24	1.71	1.79	2.72	1.75	1.31	1.82	1.58

Major elements in weight percent oxide, FeO* = total iron as FeO. Trace elements in ppm. Data from Vetter and Shervais (1992), White et al. (2002), Shervais et al. (2002), and Shervais and Vetter, unpublished.

alkaline basalts have a similar range in MgO (~5–9 wt.%) with lower FeO* and TiO₂, and higher K₂O, SiO₂, Al₂O₃, and Na₂O compared to the tholeiitic basalts (Fig. 3). The flat FeO* and TiO₂ trends, and the increase in alumina with decreasing MgO, all imply that olivine is the dominant fractionating phase, as shown by Vetter and Shervais (1992). As noted by Vetter and Shervais (1992), the coupling of higher silica with high alkalis is unusual and distinguishes these basalts from normal alkali basalts, which typically have lower silica contents.

5.2. Trace elements

Trace element concentrations are equally distinct. Compared to the older tholeiitic basalts, the high-K alkaline basalts are higher in Rb–Sr–Nb, and lower in Zr–Y–V (Fig. 4). Ba, Cr, and Ni have similar ranges in both groups, so it is not possible to derive the high-K alkaline basalts from the low-K tholeiitic basalts by either fractional crystallization or crustal assimilation, as shown by Vetter and Shervais (1992). The coupling of higher Nb with lower Zr and Y in the high-K alkali basalt group results in Zr/Nb ratios that are uniformly low (~5) compared to the tholeiites (~10–15; Fig. 4).

Chondrite-normalized rare earth element patterns show considerable overlap. The high-K alkali basalts extend to slightly higher REE concentrations than the associated tholeiitic basalts, but when we

consider only samples with Mg#>40 a clearer pattern emerges (Fig. 5; note that all SRP basalts are relatively high in FeO*, making Mg#s characteristically low compared to other basalt suites that have undergone similar amounts of fractionation, this is especially true for the low-K tholeiites of the western SRP). In these higher Mg# samples, the alkali basalts tend to have slightly lower overall REE concentrations (La=40–100× chondrite, Lu=10–20× chondrite) compared to the tholeiitic basalts (La=90–150× chondrite, Lu=18–28× chondrite). In addition, La/Lu ratios are slightly lower in the alkali basalts (~3× to ~5× chondrite); the low-K tholeiites extend to somewhat higher La/Lu (up to 8× chondrite). The ratios and chondrite-normalized patterns in the low-K tholeiites are essentially the same as eastern SRP basalts (e.g., Hughes et al., 2002).

Multi-element spider diagrams normalized to primitive mantle (Taylor and McLennan, 1985) are also almost identical for both the high-K alkali basalts and the low-K tholeiitic basalts (Fig. 6). The high-K alkali basalts have somewhat higher large ion lithophiles (K, Rb, Cs, Ba, Pb, Th) and Nb, but other elements (REE, Ti, Zr, Hf, Y, Sc, V) are lower in the alkali basalts compared to the tholeiitic basalts. The multi-element patterns seen in the tholeiitic basalts are essentially the same seen in basalts of the eastern SRP (Fig. 6). All groups have the positive Ba anomaly that is characteristic of mafic igneous rocks in the western United States (Fitton et al., 1991; Fig. 6).

SSP-58	HSP-7	HSP-100	HSP-49	HSP-51	HSP-76	HSP-89	HSP-188	HSP-59	HSP-64	HSP-61	HSP-33	HSP 121
Boise River South Fork												
Lava creek	Smith creek	Smith creek	SP1	SP1	SP2	SP2	SP2	SP3	SP3	SP3	Long Gulch	Rock creek
49.19	47.20	47.98	48.37	49.31	49.44	49.18	48.78	48.68	49.25	48.72	48.85	47.87
1.97	2.56	2.25	1.91	2.00	1.91	1.92	2.01	2.15	2.08	1.99	3.13	3.14
17.14	15.85	15.82	16.99	17.77	17.35	17.26	16.80	17.02	16.69	16.92	14.48	15.61
11.42	12.68	12.13	10.31	10.39	9.99	10.30	11.77	10.62	10.58	10.52	14.93	15.06
0.18	0.17	0.15	0.11	0.12	0.16	0.12	0.13	0.17	0.19	0.11	0.25	0.25
6.02	7.34	7.83	6.85	6.78	6.15	6.66	6.14	6.31	6.60	6.87	3.73	4.79
8.15	9.42	8.88	10.33	9.82	8.90	9.36	8.28	9.66	9.62	9.35	7.07	8.10
3.18	2.73	2.67	2.92	2.16	3.30	2.79	3.10	2.96	2.90	2.91	3.43	3.01
2.21	1.36	1.69	1.52	1.16	2.19	1.71	1.88	1.69	1.69	1.78	2.40	1.75
0.41	0.33	0.39	0.40	0.12	0.43	0.34	0.28	0.49	0.43	0.45	1.18	1.00
99.86	99.64	99.80	99.72	99.66	99.81	99.64	99.16	99.75	100.03	99.61	99.45	100.59
48.42	50.79	53.53	54.19	53.77	52.33	53.59	48.21	51.43	52.65	53.80	30.81	36.17
34	20	26	33	33	37	33	35	40	31	34	61	51
169	144	147	166	167	194	180	167	182	186	203	476	367
44	34	35	27	31	28	27	29	34	26	26	67	56
46	22	44	37	32	64	48	53	33	44	47	39	25
453	360	359	520	510	463	517	422	492	498	496	350	365
59	93	105	64	69	69	66	59	45	70	68	10	31
62	172	227	142	144	128	130	65	126	139	133	17	52
187	244	225	186	199	174	186	191	207	205	181	110	210
24	23	20	21	22	21	18	19	25	24	20	18	24
788	655	1187	528	620	623	602	832	630	643	618	1330	882
19.72	15.83	16.80	18.60	19.70	22.96	20.77	19.06	21.25	20.18	21.33	53.60	40.90
43.20	35.90	36.70	41.40	44.40	49.60	45.40	42.10	46.70	44.10	46.70	117.40	89.00
24.30	26.00	22.60	22.00	23.20	25.70	22.00	20.90	30.00	23.00	25.70	63.00	50.00
5.24	5.70	5.22	5.13	5.45	5.68	5.49	5.19	5.94	5.55	5.59	14.21	11.26
1.78	2.00	1.79	1.69	1.79	1.76	1.77	1.74	1.92	1.81	1.80	4.33	3.39
0.75	0.83	0.76	0.63	0.73	0.76	0.77	0.73	0.79	0.74	0.73	1.85	1.48
2.48	2.97	2.58	2.21	2.24	2.51	2.39	2.37	2.57	2.40	2.30	5.74	4.27
0.37	0.43	0.37	0.34	0.34	0.36	0.35	0.36	0.40	0.36	0.36	0.84	0.65
4.16	4.17	3.70	3.76	3.82	4.64	4.32	3.94	4.27	4.10	4.38	10.68	7.93
1.93	1.55	1.53	1.89	2.03	2.31	2.15	1.91	2.15	2.10	2.17	3.28	2.59
1.59	1.14	1.26	1.68	1.71	2.14	1.85	1.54	1.88	1.75	1.91	2.70	2.12

5.3. Isotopic data

Isotopic data are limited for all of these rocks. Vetter and Shervais (1992) present Sr, Nd, and Pb isotopic compositions for one low-K tholeiite and four high-K alkali basalts; another alkali basalt has Sr and Pb isotopic compositions but lacks Nd. White and Hart, (2002) present Sr isotopic compositions for 17 basalts, but no Nd or Pb isotope data. These data are compared to the isotopic composition of basalts from the eastern SRP, and to fields for Hawaii and MORB in Fig. 7. The single sample of low-K tholeiite for which we have Sr, Nd, and Pb isotopic data plots at the high- $^{87}\text{Sr}/^{86}\text{Sr}$ end of the SRP Sr–Nd array (Fig. 7A), and between the data arrays for Pb isotopic composition of eastern SRP basalts from the Idaho National Laboratory (INL) and central SRP Bruneau–Jarbridge (BJ) eruptive center (Fig. 7B–C), as defined by Hanan et al. (2008).

The four high-K alkalic samples plot at the low- $^{87}\text{Sr}/^{86}\text{Sr}$ end of the SRP Sr–Nd array, or between this array and the fields for oceanic basalts (Fig. 7A). In Pb isotope space, these samples form a trend between the oceanic array (MORB + Hawaii) and the Pb-isotopic composition of basalts from the Bruneau–Jarbridge eruptive center (Fig. 7B–C). As discussed by Hanan et al. (2008), the separation of the eastern SRP basalts into distinct groups in $^{206}\text{Pb}/^{204}\text{Pb}$ – $^{207}\text{Pb}/^{204}\text{Pb}$ – $^{208}\text{Pb}/^{204}\text{Pb}$ space implies that basalts erupted closer to Yellowstone interact with older, thicker lithosphere than those that erupt farther west. The trend

of the western SRP alkali basalts towards the Pb-isotope composition of the Bruneau–Jarbridge basalts suggests that both suites of basalts interacted similar age–thickness–composition lithosphere.

The Sr isotopic compositions of the high-K alkali and low-K tholeiitic suites are compared to basalts of the eastern SRP in Fig. 8, as a function of K_2O content. Here we can clearly see that the high-K alkali basalts are characterized by $^{87}\text{Sr}/^{86}\text{Sr}$ compositions ≤ 0.706 , whereas the low-K tholeiitic basalts (including those from the eastern SRP) are generally characterized by $^{87}\text{Sr}/^{86}\text{Sr}$ compositions ≥ 0.706 (Fig. 8). There are three low-K tholeiites from the western SRP with $^{87}\text{Sr}/^{86}\text{Sr}$ compositions ≤ 0.706 ; these have K_2O contents up to 0.9 wt.% and, as we shall see below, these erupted during the transition from low-K to high-K volcanism.

6. Discussion

6.1. Timing of the transition from low-K tholeiite to high-K alkaline

The transition from low-K tholeiite to high-K alkaline basalt in the western Snake River Plain occurred in the mid-Pleistocene, long after the Yellowstone hotspot passed by farther to the south. The timing of this transition is best evaluated using plots of K_2O or $^{87}\text{Sr}/^{86}\text{Sr}$ versus age of the basalt (Fig. 9). For these plots, because radiometric ages are available only for a few of the source vents (Amini et al., 1984; Howard

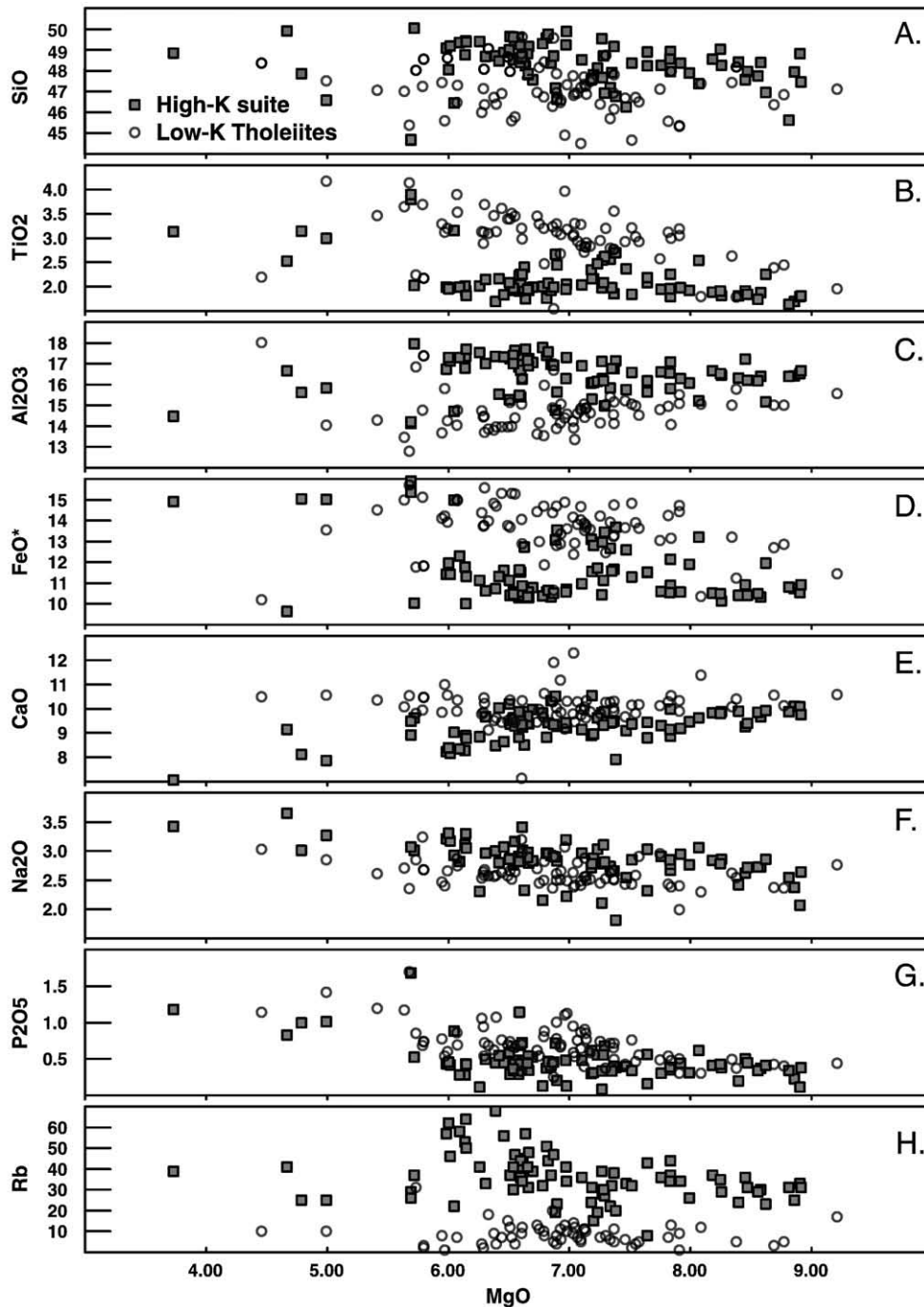


Fig. 3. MgO variation diagrams for high-K alkali basalts (filled symbols) and low-K tholeiitic basalts of WSRP: (A) SiO₂, (B) TiO₂, (C) Al₂O₃, (D) FeO*, (E) CaO, (F) Na₂O, (G) P₂O₅, (H) Rb ppm. The high-K series is higher in SiO₂, Al₂O₃, and Rb, and lower in FeO*, TiO₂, and P₂O₅, than the low-K tholeiitic series. Data from Vetter and Shervais (1992), White and Hart, (2002), Shervais et al. (2002), and unpublished.

et al., 1982; Othberg et al., 1995), we use age estimates based on the stratigraphic relations of the vents and their flows, as established by detailed geologic mapping (Howard and Shervais, 1973; Shervais et al., 2002; Bonnichsen and Godchaux, 2002; Shervais et al., 2005). We show data from the eastern SRP for comparison (Hughes et al., 2002; Shervais et al., 2006; Hanan et al., 2008).

The plot of K₂O vs age (Fig. 9A) shows that the first high-K basalts began to erupt in the western SRP around 0.9 Ma, but low-K basalts continued to erupt until around 0.7 Ma. In Fig. 9B (⁸⁷Sr/⁸⁶Sr versus age), we see that the transition to ⁸⁷Sr/⁸⁶Sr ≤ 0.706 occurred around 0.9–1.0 Ma – around the same time the first high-K lavas were

erupted. In the eastern SRP, K₂O contents have remained essentially constant for the last 2.5 Ma, although there may be a weak trend towards higher K₂O in the younger basalts (Fig. 9A). A distinct trend towards lower ⁸⁷Sr/⁸⁶Sr in the younger basalts seems evident, but there are too few data to confirm (Fig. 9B).

These data imply an abrupt onset of high-K volcanism around 0.9 Ma followed by a transition period of some 200 ka during which both low-K and high-K basalts erupted. This relatively brief time interval requires a physical mechanism for replacing one source region with another that is extremely rapid, if not catastrophic. We explore the implications of this further below.

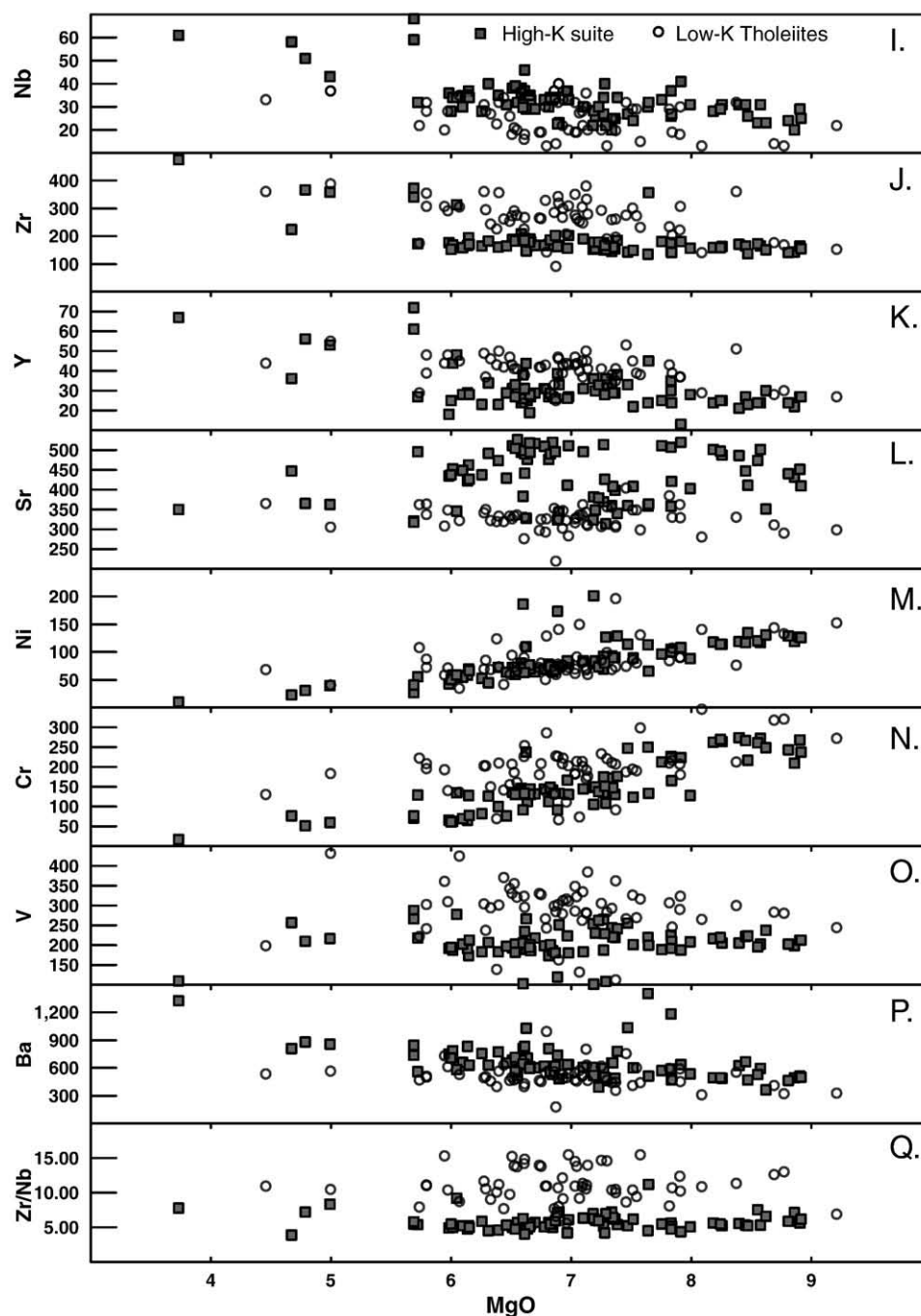


Fig. 4. MgO variation diagrams for trace elements (ppm) in high-K alkali basalts (filled symbols) and low-K tholeiitic basalts of WSRP: (I) Nb, (J) Zr, (K) Y, (L) Sr, (M) Ni, (N) Cr, (O) V, (P) Ba, (Q) Zr/Nb ratio. The high-K series is higher in Nb and Sr, and lower in Zr, Y, V, and Zr/Nb ratio, then the low-K tholeiitic series. Data as in Fig. 3.

6.2. Petrogenesis of the tholeiitic and alkali basalt suites

Petrogenetic modeling of the tholeiitic and alkali basalt suites in the western SRP show that intraflow variations in chemical composition can be accounted for by fractional crystallization of the observed phenocryst phases (olivine + plagioclase), accompanied by minor amounts of crustal assimilation in some flows (Vetter and Shervais 1992; White and Hart, 2002). Least squares mixing models show that the low-K tholeiite suite requires olivine:plagioclase ratios of ~1:2.5 to ~1:3, along with minor cryptic pyroxene fractionation that must have occurred at higher pressures within the crust (Vetter and Shervais 1992; White and Hart, 2002). In contrast, the high-K lavas require olivine:plagioclase ratios of ~1:1 to ~1:1.5 – a result

consistent with their fractionation trends on MgO variation diagrams (Fig. 3). As with the low-K tholeiite basalts, some of the alkali basalts require high pressure fractionation (circa 0.5–0.8 GPa or 15–25 km depth) of the assemblage olivine:plagioclase:clinopyroxene in the ratio 1:5:5 to account for differences between different flows from the same volcano (Vetter and Shervais 1992; White and Hart, 2002).

A few basalts show evidence for crustal assimilation (e.g., the high-K Smith Creek basalt of Prairie; low-K tholeiites of Walters Butte and Guffey Butte in the Kuna–Melba area, and the low-K Long Gulch basalt of Prairie). This evidence includes partially digested xenoliths of granite in the Smith Creek basalt, and partially melted sediments in the Guffey Butte basalt (Vetter and Shervais, 1992; White and Hart,

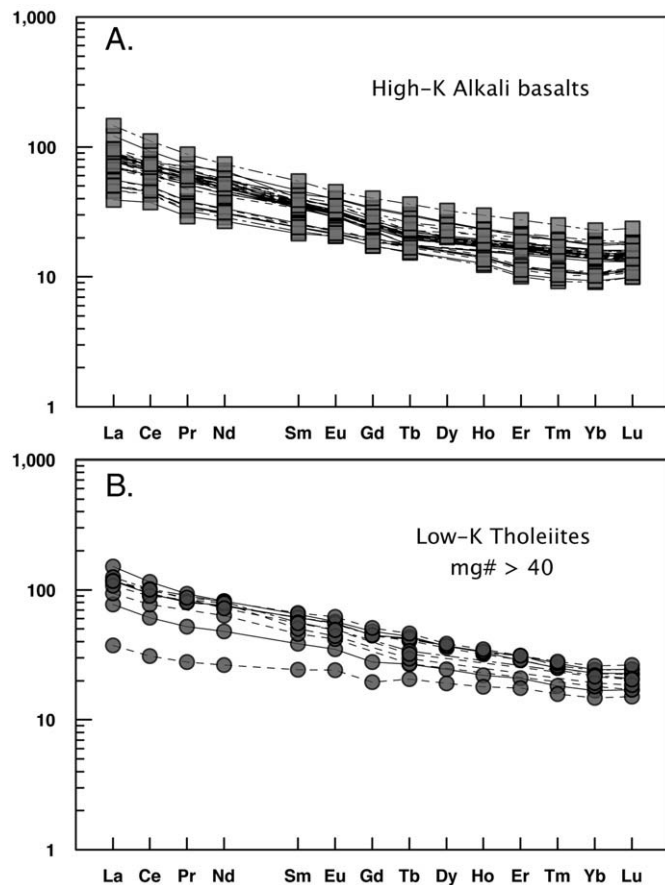


Fig. 5. Chondrite-normalized rare earth element (REE) concentrations in (A) high-K alkali basalts and (B) low-K tholeiitic basalts. Normalized to C1 chondrite compiled by Sun and McDonough 1989. Data from Vetter and Shervais (1992), White and Hart, (2002), Shervais et al. (2002), and unpublished.

2002). The Smith Creek and Long Gulch basalts of Prairie (Fig. 1) have slightly elevated $\delta^{18}\text{O}$ ratios of 6.3–6.5, consistent with limited assimilation of a crustal component such as the Idaho batholith (Vetter and Shervais, 1992). Basalts from the Kuna–Melba area with physical evidence for crustal assimilation have slightly higher silica and $\text{K}_2\text{O}/\text{P}_2\text{O}_5$ compared to unaffected basalts of the same group. Nonetheless, most basalts of both groups show little or no evidence for crustal assimilation, and oxygen isotopic data presented by Vetter and Shervais (1992) requires that crustal assimilation was negligible or absent for most of the Prairie basalts.

There are no viable mechanisms to form the high-K basalts from the low-K basalts by fractionation or by crustal assimilation: the high-K basalts have higher Mg#, MgO, and Cr than likely low-K parent basalts, and critical trace element ratios (e.g., Nb/La) are inconsistent with assimilation of a wide range of potential crustal components. In addition, the high-K basalts have lower $^{87}\text{Sr}/^{86}\text{Sr}$ ratios than the low-K basalts, rendering any relationship by crustal assimilation highly improbable. This requires that the low-K and high-K suites be derived from distinct mantle source regions with different melting histories and mantle interactions.

6.3. Melting models

In order to test various mantle melting scenarios we have constructed a series of trace element melting models using three different mantle source compositions and two different mantle phase assemblages. The three source compositions are N-MORB-source (aka depleted MORB mantle, DMM), primitive mantle, and an enriched mantle comparable to the source of EMORB (MacKenzie and O'Nions,

1995); the two mantle phase assemblages are spinel lherzolite (pressures of 1.5 to 2.0 GPa) and garnet lherzolite (pressures of 2.0 to 3.0 GPa). The results are shown on primitive mantle-normalized multi-element spider diagrams for garnet facies (Fig. 10A–C) and spinel facies (Fig. 10D–F) peridotites. The melting models were constructed using the non-modal batch melting equation and distribution coefficients from MacKenzie and O'Nions (1991, 1995). The N-MORB and primitive mantle source compositions were taken from MacKenzie and O'Nions (1995), the E-MORB source composition

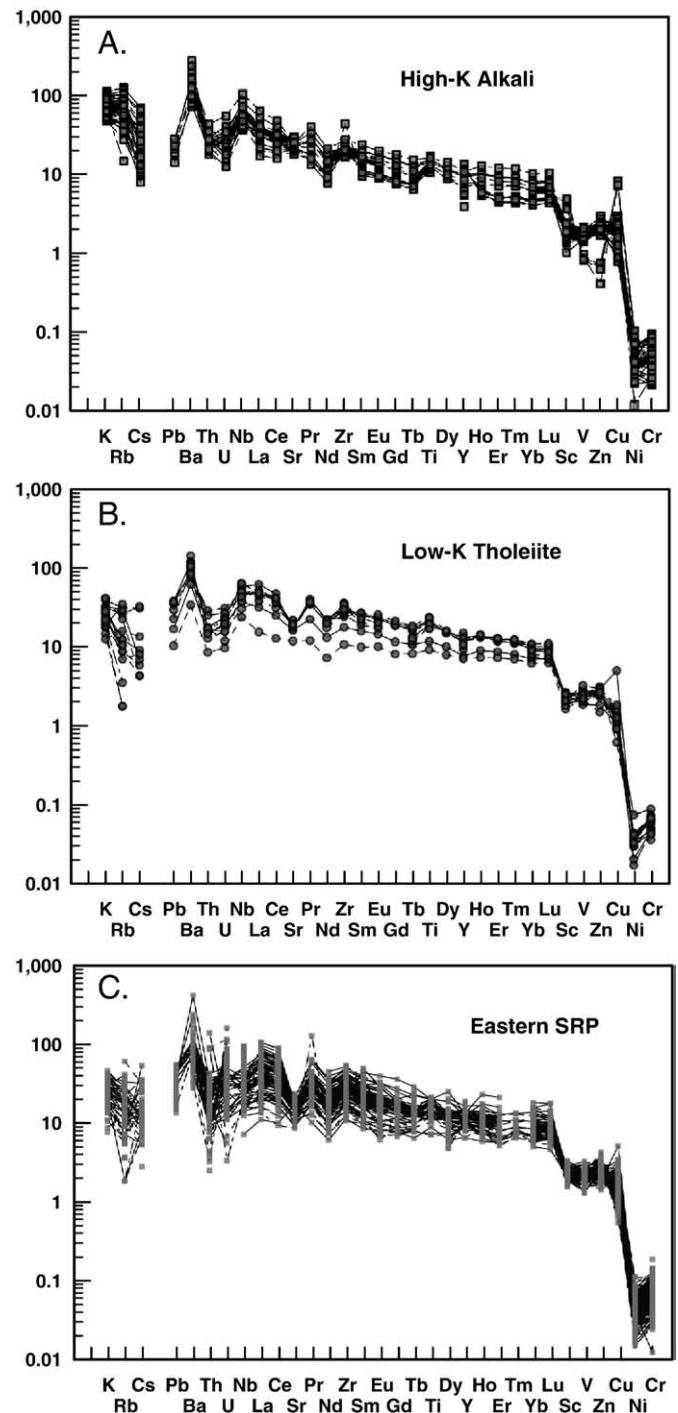


Fig. 6. Multi-element spider diagrams normalized to primitive mantle for (A) high-K alkali basalt of WSRP, (B) low-K tholeiitic basalt of WSRP, and (C) tholeiitic basalts of the eastern SRP. Primitive mantle of Taylor and McLennan 1985. Data from Vetter and Shervais (1992), White and Hart, (2002), Shervais et al. (2002, 2006), and Hughes et al. (2002).

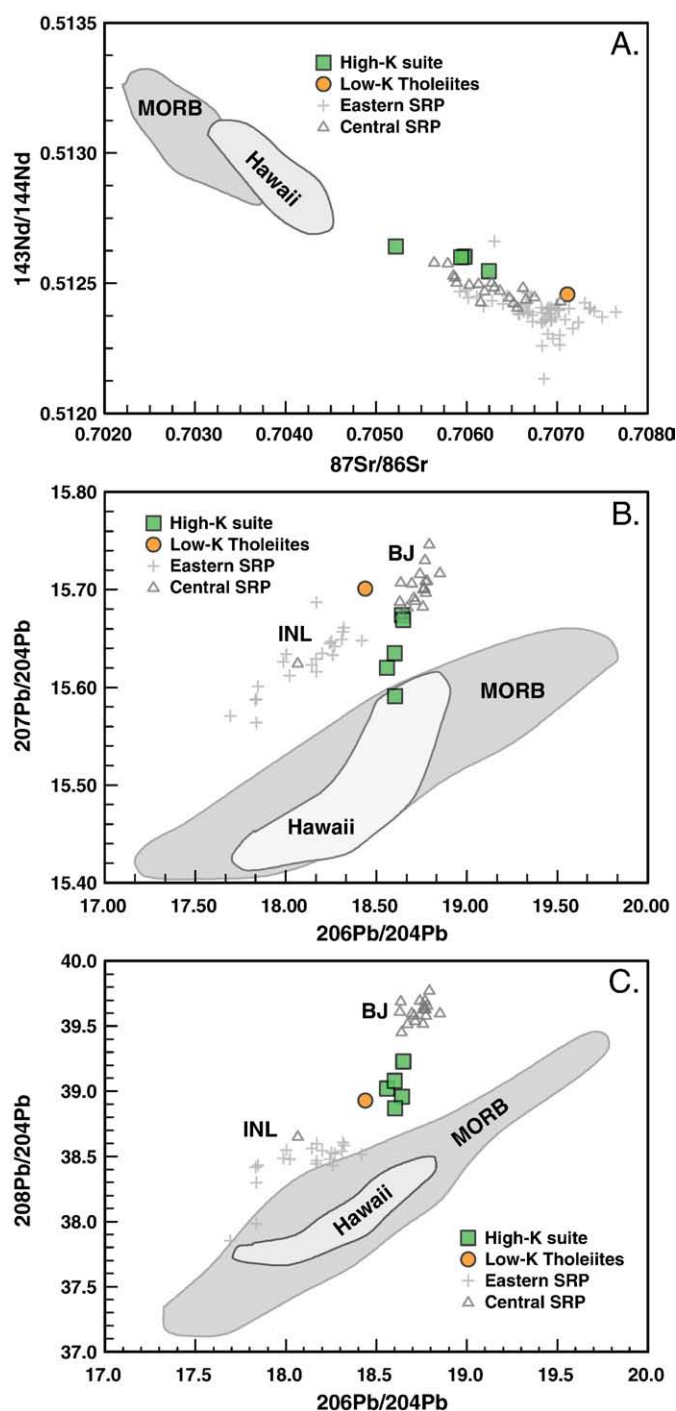


Fig. 7. Isotopic composition of low-K and high-K lavas of WSRP compared to tholeiitic basalts of the eastern SRP, ocean island basalts from Hawaii, and global MORB. (A) $^{87}\text{Sr}/^{86}\text{Sr}$ vs $^{143}\text{Nd}/^{144}\text{Nd}$, (B) $^{206}\text{Pb}/^{204}\text{Pb}$ vs $^{207}\text{Pb}/^{204}\text{Pb}$, and (C) $^{206}\text{Pb}/^{204}\text{Pb}$ vs $^{208}\text{Pb}/^{204}\text{Pb}$. The high-K basalts define trends intermediate between ESRP basalts and the fields for oceanic basalts. Data from Vetter and Shervais (1992) and Hanan et al. (2008).

was taken from Mertz et al. (2001). Non-modal batch melting yields results similar to pooled fractional melts but at slightly higher melt fractions.

None of the garnet facies models reproduces the observed magma compositions for either suite. Garnet facies melting of the N-MORB source results in model melts which show enriched LREE/HREE ratios only at very small melt fractions, but with REE patterns that cross the observed samples (Fig. 10A). Garnet facies melting of a primitive mantle source results in HREE concentrations that are too low at small melt fractions (when the LREE fit) and LREE concentrations that are

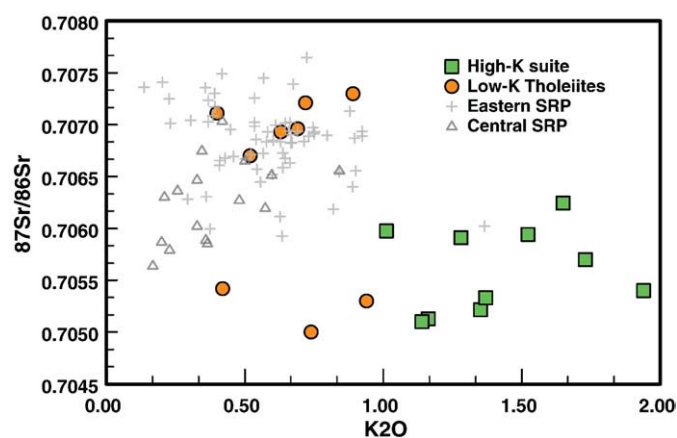


Fig. 8. Plot of K_2O vs $^{87}\text{Sr}/^{86}\text{Sr}$ for high-K alkali basalts and low-K tholeiitic basalts, compared to basalts of the central and eastern SRP. The high-K alkali basalts are characterized by low $^{87}\text{Sr}/^{86}\text{Sr} < 0.706$, whereas most low-K tholeiites have $^{87}\text{Sr}/^{86}\text{Sr} > 0.706$. A few WSRP tholeiites and central SRP basalts also have $^{87}\text{Sr}/^{86}\text{Sr} < 0.706$. Data from Vetter and Shervais (1992), White and Hart, (2002), and Hanan et al. (2008).

too low at large melt fractions (when the HREE fit; Fig. 10B). A primitive mantle source only fits at melt fractions around 3% for the high-K suite, but even here the model melts have REE slopes that cross those of the observed samples. Similar results are observed for

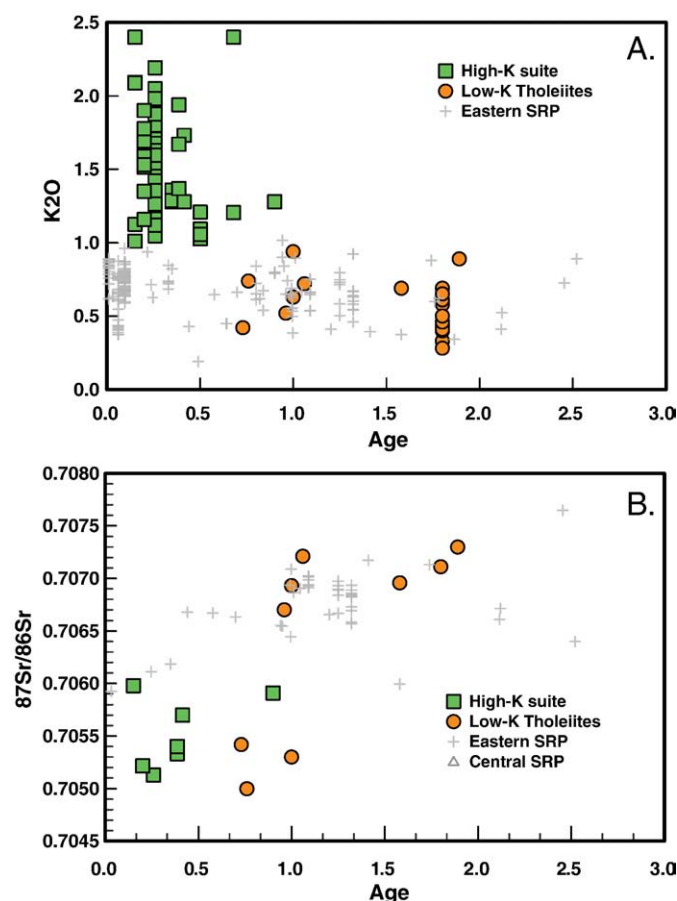


Fig. 9. Age-composition relations of western SRP basalts compared to basalts from the eastern SRP: (A) K_2O vs age and (B) $^{87}\text{Sr}/^{86}\text{Sr}$ vs age. In the western SRP transition from low-K tholeiite to high-K alkali basalt occurs ~700 ka to ~900 ka, with both types erupted during the transition; K_2O shows no change with time in the eastern SRP. The transition to low $^{87}\text{Sr}/^{86}\text{Sr}$ occurs more abruptly around 1 Ma in the WSRP, even in the low-K basalts. Data sources same as Figs. 7 and 8.

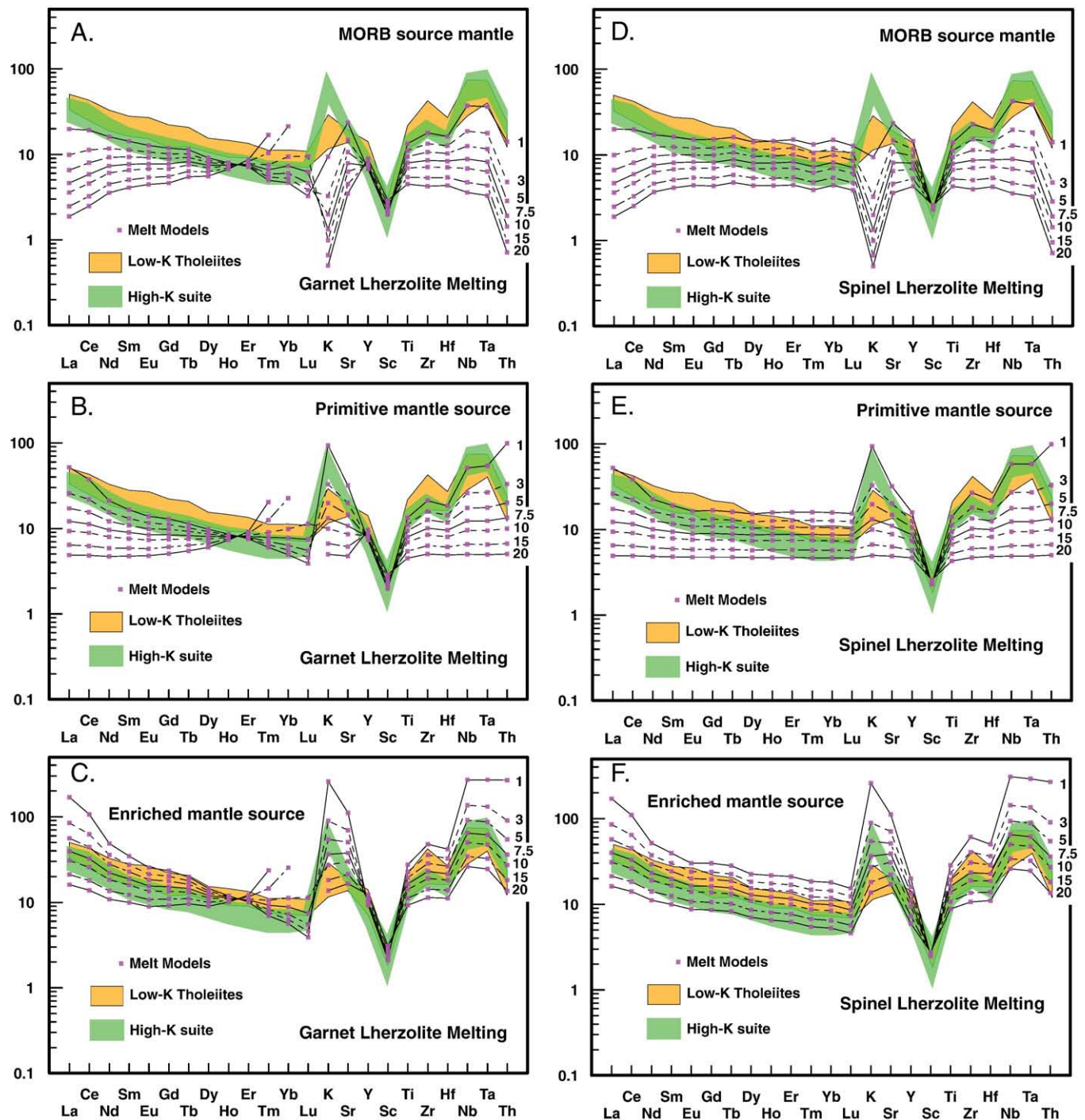


Fig. 10. Melting models for MORB-source, primitive mantle, and enriched mantle compared to WSRP basalts; all elements normalized to the primitive mantle of Mackenzie and O'Nions (1995). In all models data for low-K tholeiites represented by orange field, data for high-K basalts represented by green field. (A) MORB source mantle melting in garnet lherzolite facies; (B) primitive mantle melting in garnet lherzolite facies; (C) enriched mantle melting in garnet lherzolite facies; (D) MORB source mantle melting in spinel lherzolite facies; (E) primitive mantle melting in spinel lherzolite facies; (F) enriched mantle melting in spinel lherzolite facies. Note that only model (E) fits well for either suite.

melting of an enriched mantle source in the garnet lherzolite facies (Fig. 10C).

Results for spinel facies peridotite melting are essentially the same, as neither MORB-source nor primitive mantle source compositions are successful at reproducing the observed concentrations in either suite: models with N-MORB source composition are depleted and cross REE patterns for both suites (Fig. 10D) whereas models with primitive mantle source compositions produce melts with relatively flat patterns that also cross the REE patterns of both suites (Fig. 10E). In

addition, high field strength elements are uniformly depleted in the model melts compared to the observed melts.

Melting of an enriched mantle source in the spinel lherzolite facies reproduces that observed basalt compositions well for both suites, with melt fractions of 5% to 8% for the low-K tholeiite suite and 7% to 15% for the high-K alkali suite (Fig. 10F). The misfit between the melting models and Sr content of the basalts of both suites likely represents low pressure fractionation of plagioclase, which will quickly lower the concentration of Sr even if other elements are

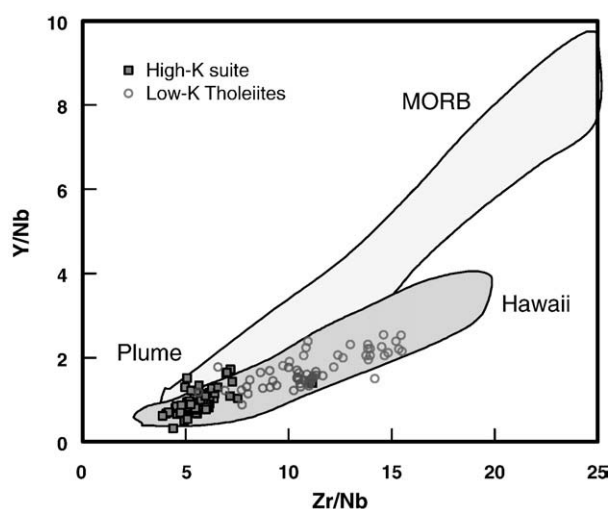


Fig. 11. Comparison of WSRP basalts to MORB and ocean island basalts of Hawaii in Zr/Nb–Y/Nb space. WSRP basalts mimic the low Y/Nb trend of Hawaiian basalts, which intersect the MORB trend at Zr/Nb ~5 and Y/Nb ~0.5; this represents a common plume component in both MORB and Hawaii. In the WSRP, the high-K basalts have the same low Zr/Nb–Y/Nb ratios as the plume component, whereas the low-K basalts lie along the mixing trend with a more depleted component (higher Zr/Nb–Y/Nb). Note that neither Hawaii nor the SRP basalts indicate mixing with a depleted MORB component. Rather, both arrays indicate mixing with a depleted component of unknown origin.

essentially unaffected. These models present a paradox, however, in that the incompatible element K has a higher concentration in basalts that apparently form by larger fractions of partial melting.

6.4. Mantle source regions

The melting models presented here require an enriched mantle composition melting at relative low pressures to form both magma series; these models cannot distinguish between the low-K and high-K suites, despite the distinct differences in their major element and isotopic compositions. How do the mantle source regions differ for these two suites? We can appraise this issue by examining a plot of Zr/Nb vs Y/Nb, which compares the high-K and low-K suites of the western SRP with ocean island basalts of Hawaii and a compilation of global MORB geochemistry (Fig. 11). In this plot, depleted components will have high Zr/Nb and Y/Nb ratios, while enriched components will have low ratios (because Nb is more incompatible than either Zr or Y).

The western SRP basalts define a trend with a wide range of Zr/Nb ratios and a more limited spread in Y/Nb ratios. This trend almost completely overlaps the trend of Hawaiian basalts, and both trends intersect the MORB trend at low Zr/Nb–Y/Nb (Fig. 11). The low Zr/Nb–Y/Nb end member in all three groups (MORB, Hawaii, WSRP) represents the deep mantle enriched Yellowstone plume component. In the western SRP, this end member composition is represented by the high-K alkali basalts (filled symbols in Fig. 11). The low-K tholeiitic basalts represent mixing with a depleted end member composition similar to that involved in the Hawaiian plume; note that this depleted end member component is *not* typical zero age Pacific-like MORB source asthenosphere. Thus models which call upon mixing between the deep-seated plume component and shallow MORB source asthenosphere are invalid. The question remains as to the physical nature of the depleted component that mixes to form the low-K tholeiite basalts; answering that question is beyond the scope of this paper.

6.5. Physical models

If both the low-K and high-K suites are derived from the same plume source mantle, how do they become distinct in their major

element, trace element, and isotopic compositions? Hanan et al. (2008) have recently proposed that basalts of the eastern SRP (low-K tholeiites similar to those in the western SRP) have been overprinted isotopically by interaction with small volumes of low-percentage fractional melts of the mantle lithosphere that underlies the continental crust. They show that the isotopic composition of central and eastern SRP basalts varies with proximity to the continental margin, implying variations in the thickness, composition, and age of the subcrustal lithosphere (Hanan et al., 2008). In this model, the older low-K tholeiites would react with the overlying subcontinental lithospheric mantle (SCLM), assimilating fractional melts with extremely high incompatible element concentrations. The isotopic composition of these fractional melts reflect the age and enrichment of the lithosphere, and will generally have high $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{207}\text{Pb}/^{204}\text{Pb}$, and low $^{143}\text{Nd}/^{144}\text{Nd}$. Because of their extreme enrichment, small volumes of these melts will dominate the isotopic composition of the mixture (Hanan et al., 2008). This model similar to previous models that have been proposed for volcanic rocks of the north-western USA in calling on progressively thickened lithosphere from west to east (e.g., Hart et al., 1997), which may also contain a subduction enrichment component (Hart et al., 1984; Hart 1985; Harry and Leeman, 1995).

The limited interaction reflected in the isotopic compositions of the high-K alkali basalts (Fig. 7) implies that interaction with the SCLM was also limited, or at least much less extensive than that experienced by the low-K tholeiites. There are two plausible explanations for this change: (1) the fusible component of the SCLM was reduced during the passage of earlier melts (low-K tholeiites) so that less melt was available for assimilation, or (2) that most of the SCLM was physically removed prior to eruption of the high-K alkali series.

A second issue involves the continued flux of plume-derived magma 10–12 Ma after passage of the Snake River–Yellowstone hotspot beneath western Idaho. Shervais and Hanan (2008) have proposed that plume-derived mantle continues to stream westward beneath the Snake River Plain along a shoaling gradient in the subcontinental lithospheric mantle. This enriched, plume-derived mantle is confined to the channel eroded into the SCLM during passage of the hotspot, and flows upward and westward along a gradient defined by the progressive thinning of the SCLM as the margin of the craton is approached (Shervais and Hanan, 2008). This active flow model is distinct from models that ascribe continued melting to flattening of stagnant plume-derived mantle against the base of the lithosphere after passage of the plume; these models limit the extent and volume of melt derived from the stagnant plume mantle, and the also limit the time frame over which this melt may be generated. The mantle material in the channel may be a hybrid mixture of MORB, plume, and SCLM components. However, the trace elements show that it is dominantly Yellowstone Plume material.

The time scale that we have established for the transition from low-K tholeiitic basalt to high-K mildly alkaline basalt is short – only some 200 ka passed between the eruption of the first high-K lavas and the last low-K lavas. This implies an equally rapid change in the SCLM. We suggest that the first explanation – reduction of the fusible component of the SCLM – would be by its nature a gradual process, not abrupt. This suggests that the second explanation – physical removal of the SCLM prior to eruption of the alkali basalts – is more likely. The abruptness of the transition implies that it may have a catastrophic origin, such as lithospheric delamination caused by a Rayleigh–Taylor instability beneath the southern Idaho. This model is consistent with the recent uplift of the southern portion of the Idaho batholith and the deep incision of the rivers that transect it, where incision rates have increased from ~0.1 mm/year to ~0.7 mm/year in the last 700 ka (Howard et al., 1982). Alternatively, this abrupt transition may represent the final removal of a highly enriched phase in the mantle lithosphere that carries most of the radiogenic isotopes, and is both highly fusible and present in small amounts modally. Testing these alternate models will require a

more detailed assessment of incision rates over a broader area to distinguish local effects from regional uplift.

7. Conclusions

Basaltic volcanism in the western SRP underwent an abrupt transition circa ~700 ka to 900 ka from low-K tholeiitic basalt volcanism to high-K alkali basalt volcanism, as documented in three separate volcanic fields in southern Idaho: Mountain Home and Kuna-Melba in the Snake River valley, and Prairie in the Boise River South Fork valley. The low-K basalts resemble tholeiitic olivine basalts of the eastern SRP and oceanic tholeiites of Hawaii, whereas the younger high-K basalts resemble alkali basalts of Hawaii. Isotopic studies show that the low-K tholeiitic basalts reacted extensively with old SCLM, while the high-K basalts are transitional to oceanic basalt compositions. We infer that the high-K basalts represent a mantle plume component that is distinct from both SCLM and from MORB-source asthenosphere.

Acknowledgements

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