

Explosive silicic volcanism of the Yellowstone hotspot: The ash fall tuff record

Michael E. Perkins*

Barbara P. Nash

Department of Geology and Geophysics, University of Utah, Salt Lake City, Utah 84112-0111, USA

ABSTRACT

Unaltered silicic ash fall tuffs are abundant in Neogene sedimentary basins of the western U.S. and constitute an important record of explosive silicic volcanism in this region. In particular, ash fall tuffs from silicic volcanic centers along the Yellowstone hotspot track are common in these basins and provide a detailed record of explosive volcanism along the hotspot track. The available hotspot ash fall tuff record commences at ca. 16 Ma, shortly after the initiation of hotspot silicic volcanism at ca. 16.5 Ma, and continues through the most recent explosive eruptions in the late Pleistocene. Post-16 Ma hotspot silicic volcanism has been dominated by eruption of metaluminous ash flow tuffs and rhyolites, and ash fall tuffs produced by these eruptions dominate the Yellowstone hotspot ash fall tuff record. Evaluation of a well-dated composite sequence of 142 of these tuffs reveals systematic variation in magma composition, magma temperature, eruption frequency, and, possibly, volumetric discharge as the hotspot migrated eastward from the western edge of the North America craton to its current location in the Yellowstone Plateau.

On the basis of these variations, three primary stages of metaluminous rhyolite magmatism (M stages) are recognized: M1 (16.0–15.2 Ma), M2 (15.2–7.5 Ma), and M3 (7.5–0 Ma). Each of these stages is marked by distinctive magma compositions, eruption frequencies, and magma temperature ranges and trends, with an overall decline in average eruption frequency and magma temperature from stage to stage. The partitioning of explosive hotspot volcanism into stages likely reflects variation in the style

and intensity of the interaction between the mantle anomaly powering the hotspot magmatic systems and a spatially and temporally heterogeneous lithosphere along the hotspot track. Although the ash fall tuff record does not provide definitive constraints on the nature of these variable interactions, correlations between changes in eruption frequency, ash fall tuff discharge, and magma temperature point to variable input of mantle basalt into hotspot crustal magmatic systems as a first order control on intensity of explosive volcanism. As future studies reveal more about the processes controlling the variations in hotspot silicic volcanism, a fuller understanding of both the nature of silicic magmatism and the nature of the Yellowstone hotspot should emerge.

Keywords: ash fall tuff, composition, eruption frequency, explosive volcanism, volumetric discharge, Yellowstone hotspot.

INTRODUCTION

The Yellowstone hotspot is named after the Yellowstone Plateau volcanic field of the western U.S. Cordillera. This volcanic field has been the focus of vigorous explosive and effusive silicic volcanism and subordinate basaltic volcanism since ca. 2.5 Ma (Christiansen, 1984) and is currently characterized by a regime of extraordinarily high heat flow and associated, world renowned geysers and hot springs. The thermal anomaly that drives the volcanism and geothermal activity is commonly termed the Yellowstone hotspot (Morgan, 1972; Smith et al., 1977), whereas the plateau itself and associated older silicic volcanic centers to the southwest (Fig. 1) define the volcanic track of Yellowstone hotspot (Pierce and Morgan, 1992).

Geophysical, geological, and geochemical

observations in the Yellowstone Plateau region support the hypothesis, first suggested by Morgan (1972), that the volcanism and tectonism in this region is driven by upwelling of a deep seated mantle plume (Smith et al., 1977; Craig et al., 1978; Leeman, 1982; Anders et al., 1989; Pierce and Morgan, 1992; Craig, 1993; Smith and Braile, 1994). Further, the northeastward progression of silicic volcanism along the hotspot track, antiparallel to the motion of the North America plate (Fig. 1), is consistent with the hypothesis that the older silicic volcanic fields along this track mark past positions of this plume relative to the North America plate (Smith and Sbar, 1974; Armstrong et al., 1975; Rodgers et al., 1990; Pierce and Morgan, 1992; Smith and Braile, 1994). However, as emphasized by Leeman (1982), the nature of the origin and nature of the mantle anomaly driving the volcanism is uncertain. Thus, the origin of the Yellowstone hotspot remains a topic of active discussion with current research focusing on possible “nonplume” origins for the hotspot (Humphreys et al., 2000).

Although the nature of the mantle phenomenon driving Yellowstone hotspot track volcanism remains uncertain, the volcanism itself is of considerable interest. In particular, the variation in the intensity of silicic volcanism over the lifetime of the hotspot provides an opportunity to explore the response of silicic magmatism to variable inputs. In this investigation we monitor volcanic output from hotspot sources by using a detailed record of associated silicic ash fall tuffs preserved in Neogene sedimentary deposits of the northern Basin and Range province and the central Great Plains. The tuffs examined in this record document temporal variation in magma composition, magma temperature, and eruption frequency as the hotspot transgressed eastward from the western edge of

*E mail: mperk@mines.utah.edu.

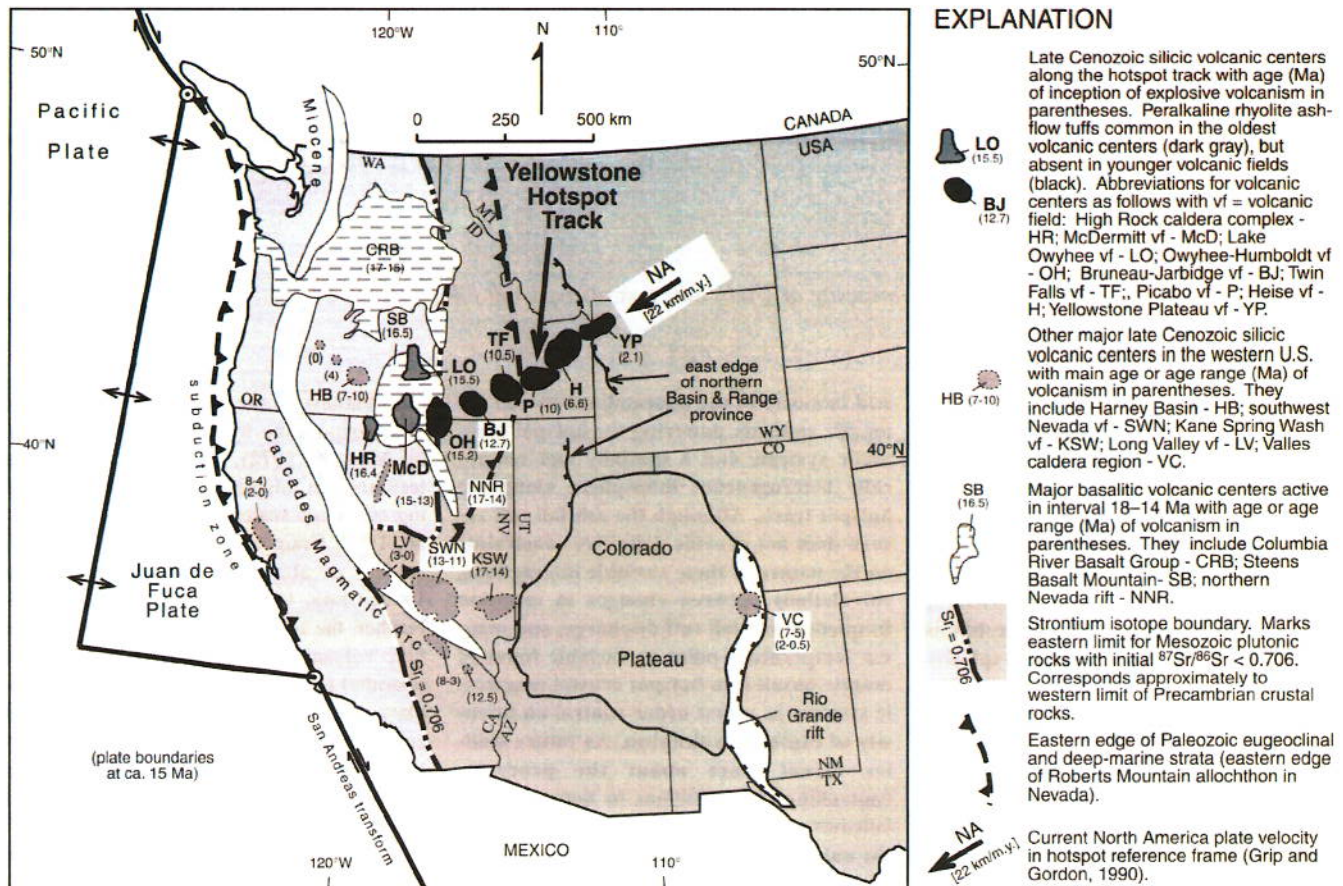


Figure 1. Location and tectonic setting of the Yellowstone hotspot track. Modified from Dickinson (1997).

the North America craton to its present position deep within the craton. Correlations between this variable volcanism, the heterogeneous lithosphere along the hotspot track, and variable hotspot migration rate offer insights into processes modulating hotspot volcanism.

YELLOWSTONE HOTSPOT TRACK VOLCANISM

Silicic volcanism along the Yellowstone hotspot track began at ca. 16.6 Ma along the Nevada–Oregon border as the voluminous outpouring of basalt in the Columbia Plateau and associated basaltic volcanic centers to the south was near a peak and as partly coeval rhyolitic volcanic centers were developing to the south (Fig. 1). The initiation of widespread bimodal basalt rhyolite volcanism in the region at ca. 17–15 Ma marks a fundamental change from the pattern of intense calc alkalic volcanism that dominated the region from ca. 45 to 17 Ma (Best et al., 1989).

The volcanic centers that mark the hotspot

track have been the focus of vigorous explosive and effusive silicic volcanism. The track itself is best delineated by silicic ash flow tuffs, as it is these tuffs that show a well focused space time transgression (Fig. 2; named ash flow tuffs, along with identifying initials, are listed in Table 1). Rhyolite flows also young eastward, but with a more diffuse space time pattern. In addition, basalt flows are common along the hotspot track but generally postdate the silicic volcanism (Fig. 2). These differing space time patterns of hotspot track volcanism suggest that the explosive volcanism marks the hotspot core and that more peripheral hotspot processes produce the spatially associated silicic and basaltic flows.

Silicic volcanic rocks along the hotspot track are typically Fe rich, Ca poor, high-temperature rhyolites ($\geq 800^\circ\text{C}$) with a sparse phenocryst assemblage dominated by quartz, feldspar, and generally anhydrous mafic minerals (Leeman, 1982; Conrad, 1984; Honjo et al., 1992). These rhyolites range in composition from mildly peralkaline to peraluminous with the peralkaline rhyolites reported only

from volcanic centers west of the initial Sr ratio (Sr_i) = 0.706 line (Figs. 1, 2). Although there is a general uniformity to the hotspot rhyolites (Leeman, 1982), compilation of analyses indicates some temporal variation in their composition. This variation includes a transition from predominantly peralkaline rhyolites at ~16 Ma along with marked temporal fluctuations the Fe content of the metaluminous rhyolites (Fig. 3). Furthermore, among the metaluminous silicic volcanic rocks there is a trend toward less evolved, more Fe rich rocks in the interval ca. 15–9 Ma followed by a decline to less Fe rich rocks after 8 Ma. Associated variation in the frequency of ash flow tuffs is also suggested with a very high eruption rate of ~32 tuffs/m.y. prior to 16 Ma followed by markedly lower rates for younger ash flow tuffs of ~4 tuffs/m.y. in the interval 15–9 Ma and <1 tuff/m.y. after 9 Ma (Fig. 3). As discussed in this paper, similar variations in the magma composition and the eruption frequency are captured in detail in the record of the hotspot ash fall tuffs.

EXPLOSIVE SILICIC VOLCANISM OF THE YELLOWSTONE HOTSPOT

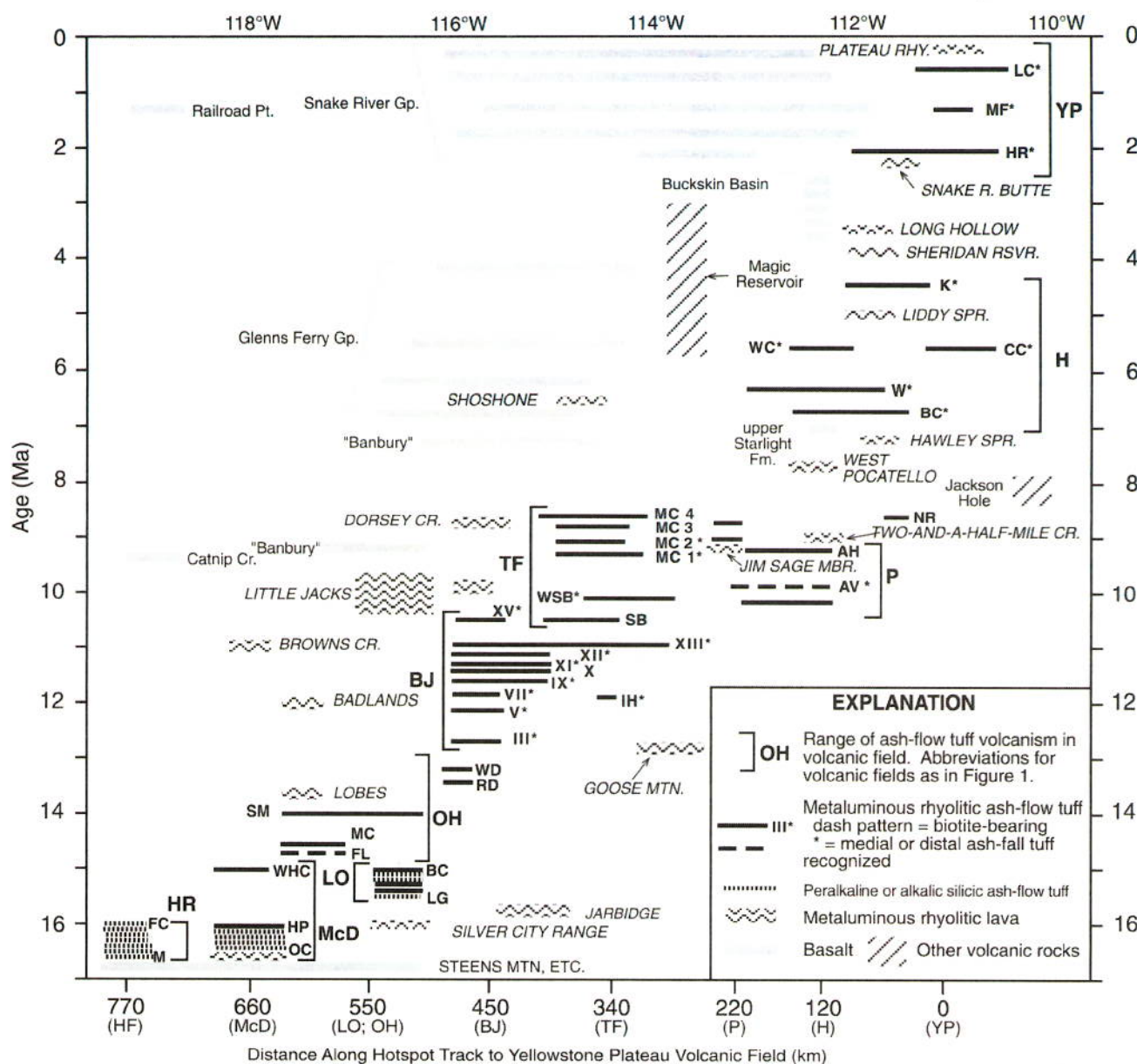


Figure 2. Space time distribution of Yellowstone hotspot track volcanic rocks. Ash flow tuffs in the interval 14–4 Ma are those exposed along the southern margin, but temporally equivalent tuffs are recognized along the northern margin as well. Named ash flow tuffs shown here, along with identifying initials, are listed in Table 1; fields abbreviated along x axis are explained in Figure 1. Basalts mainly from Armstrong et al. (1975) and Hart and Carlson (1985). Age for Steens Mountain Basalt from Swisher (1992). See Table DRI for details on silicic volcanic rocks.

METHODS AND CONVENTIONS

The Yellowstone hotspot ash fall tuff record is a synthesis of chemical analyses, stratigraphic investigations, and geochronologic studies of late Cenozoic silicic ash fall tuffs conducted at the University of Utah (Smith and Nash, 1976; Williams, 1994; Perkins et al., 1995a, 1995b, 1998; Perkins, 1998a, 1998b) and supplemented by results of other

studies, particularly those of Izett and Wilcox (1982) and Moutoux (1995).

The University of Utah tephra database includes information on the glass shard composition, stratigraphy, and age of ~2400 samples of late Neogene ash fall tuffs from localities in the intermontane west and Great Plains. Glass compositional data include electron microprobe analyses of individual glass shards for most of the samples and X-ray fluo-

rescence analyses of bulk glass shard separates of ~1100 samples. These analyses were done at the University of Utah following methods discussed in Perkins et al. (1995b, 1998). Glass shard analyses for a suite of chronologically and chemically representative gray metaluminous hotspot ash fall tuffs are given in Table 2. In addition, analyses of Fe_2O_3 , K_2O , Cl , Sr , Rb , Zr , and the element ratio Mn/Fe are given for all 146 hotspot tuffs

TABLE 1. SILICIC ASH FLOW TUFFS AND LAVAS SHOWN IN FIGURE 2

Unit	Type	Initials	Age* (Ma)	Field†	Unit	Type	Initials	Age* (Ma)	Field†
Lava Creek Tuff	Af	LC	0.60	YP	Cougar Point Tuff V	Af	V	12.07	BJ
Mesa Falls Tuff	Af	MF	1.30	YP	Cougar Point Tuff III	Af	III	12.67	BJ
Huckleberry Ridge Tuff	Af	HR	2.06	YP	Whiskey Draw rhyolite	Af?	WD	ca. 13?	BJ?
Snake River Butte Rhyolite	L		2.10	YP	Rattlesnake Draw tuff	Af	RD	ca. 13?	BJ?
Rhyolite of Long Hollow	L		3.50	H?	Jarbridge Rhyolite	L		15.80	
Rhyolite of Sheridan Reservoir	L		4.00		Tuff of Little Jacks	?		ca. 10	
Tuff of Kilgore	Af	K	4.45	H	Tuff of Browns Creek	?		11.00	
Rhyolite of Lidy Spring	Af		5.40		Tuff of the Badlands	?		12.00	
Connant Creek Tuff	Af	CC	5.58	H	Lower Lobes Tuff	?		13.80	
Tuff of Wolverine Creek	Af	WC	5.58	H	Upper Lobes Tuff	?		13.90	
Walcott Tuff	Af	W	6.20	H	Swisher Mtn. Tuff	Af?	SM	13.80	OH
Tuff of Blacktail Creek	Af	BC	6.62	H	Mink Creek	Af	MC	N.A.	OH?
Rhyolite of Hawley Spring	L		7.20		Tuff of Flint Creek	Af	FL	N.A.	OH?
Tuff of Newby Ranch	Af	NR	8.60	H?	Rhyolite of Silver City Rng.	L		ca. 16	
Rhyolite of west Pocatello	L		7.90		Tuff of Birch Creek	Af	BC	15.00	LO
Rhyolite of 2-and-1/2-Mile Creek	L		9.10		Tuff of Crowley Lake	Af		15.00	LO
Arbon Valley Tuff	Af	AV	10.20	P	Wild Cat Tuff	Af		15.00	LO
Tuff of McMullen Creek 4	Af	MC4	8.60	TF	Dinner Creek Tuff	Af		15.30	LO
Tuff of McMullen Creek 3	Af	MC3	N.A.	TF	Tuff of Spring Creek	Af		15.40	LO
Tuff of McMullen Creek 2	Af	MC2	N.A.	TF	Leslie Gulch Tuff	Af	LG	15.50	LO
Tuff of Appendicitis Hill	Af	AH	9.23	P?	Tuff of Whitehorse Creek	Af		15.00	McD
Tuff of McMullen Creek 1	Af	MC1	9.16	TF	Tuff of Hoppin Peaks	Af		16.90	McD
Tuff of Wooden Shoe Butte	Af	WSB	10.13	TF	Tuff of Long Ridge	Af		16.30	McD
Tuff of Steer Basin	Af	SB	10.50	TF	Tuff of Double H	Af		N.A.	McD
Rhyolite in Goose Mountain area	L		12.80		Tuff of Trout Creek Mountains	Af		16.20	McD
Dorsey Creek Rhyolite	L		8.11		Tuff of Oregon Canyon	Af		16.44	McD
Sheep Creek Rhyolite	L		9.88		Tuff of Fly Canyon	Af	FC	16.08	HR
Cougar Point Tuff XV	Af	XV	10.50	BJ	Soldier Meadow Tuff	Af		16.12	HR
Cougar Point Tuff XIII	Af	XIII	10.94	BJ	Tuff of Trough Mountain	Af		N.A.	HR
Cougar Point Tuff XII	Af	XII	11.18	BJ	Tuff of Big Mountain	Af		N.A.	HR
Cougar Point Tuff XI	Af	XI	11.31	BJ	Tuff of Indian Spring	Af		N.A.	HR
Cougar Point Tuff X	Af	X	11.45	BJ	Summit Lake Tuff	Af		16.36	HR
Cougar Point Tuff IX	Af	IX	11.59	BJ	Idaho Canyon Tuff	Af		16.48	HR
Cougar Point Tuff VII	Af	VII	11.90	BJ	Tuff of Craine Creek	Af		16.10	HR
Tuff of Ibex Hollow	Af	IH	11.93	BJ?	Tuff of the Mesa	Af	M	16.60	HR

Note: See Data Repository Table DR1 for additional information. Af—ash flow tuff; L—lava. Queried where uncertain.

*N.A.—no reported age dates but units placed in reported stratigraphic order in a given volcanic field.

†Initials are the same as in Figure 2.

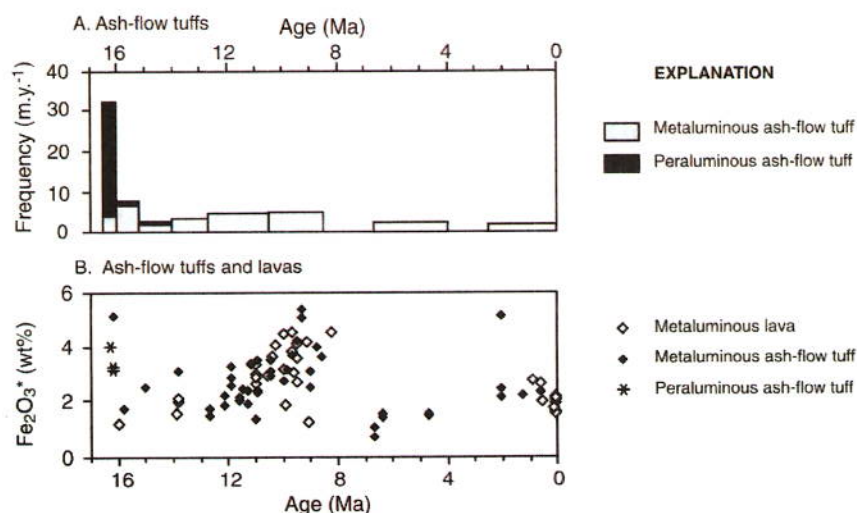


Figure 3. Variation in the frequency and Fe_2O_3^* content of silicic volcanic rocks along the hotspot track. (A) Frequency of ash flow tuffs shown in Figure 2. (B) Variation of whole rock $\text{Fe}_2\text{O}_3^* = \text{total Fe as } \text{Fe}_2\text{O}_3$ with age; analyses from Bonnichsen (1982), Ekren et al. (1984), Hildreth et al. (1991), Honjo (1990), Kellogg and Marvin (1988), and Kellogg et al. (1994).

in Data Repository Table DR3.¹ The temporal variations in these six elements and the ratio Mn/Fe are shown in figures that follow.

The sequence of ash fall tuffs assembled for this study comes from 15 stratigraphic sections in the Basin and Range (14 sections) and Rio Grande rift (1 section), 5 research drill holes (Bonneville cores) and 5 petroleum exploration wells (Salt Lake wells) in and around the Great Salt Lake in Utah, and a suite of samples from upper Cenozoic strata of the Great Plains (Fig. 4). The composite sequence of hotspot ash fall tuffs identified in these localities is shown in Figure 5, and ages and other information for all these ash beds is included in Table DR2. Correlation between localities is based on both glass shard composition and relative stratigraphic sequence of the tuffs by using procedures of Perkins et al. (1995b, 1998). Generally, tuffs from several

¹GSA Data Repository item 2002039, tables listing Yellowstone hotspot silicic ash flow tuffs, lavas, and ash fall tuffs with information on the age and composition of these units, is available on the Web at <http://www.geosociety.org/pubs/ft2002.htm>. Requests may also be sent to editing@geosociety.org.

EXPLOSIVE SILICIC VOLCANISM OF THE YELLOWSTONE HOTSPOT

TABLE 2. GLASS-SHARD COMPOSITION FOR REPRESENTATIVE HOTSPOT GRAY METALUMINOUS ASH FALL TUFFS

Sample ID:	Oc	dee93	rir93	mor94	rv88	rv88	rv88	epb92	tc90	tc89	tc89	tc89	tc89	tc90	sv93	buf94	buf94	buf94	buf94	Std.
Unit: [†]	LCB	HR	SAN	W	BC	F	RV	un	WSB	XIII	XI	V	III	RR	un	PV1	VV8	un	VV1	dev.*
Age (Ma): [‡]	0.60	2.06	5.1	6.2	6.62	7.49	7.9	9.3	10.13	10.94	11.3	12.07	12.67	13.7	14.1	15.17	15.4	15.5	15.6	
Type:	SRP	SRP	SRP	SRP	SRP	SRP	SRP	SRP	SRP	SRP	SRP	SRP	SRP	SRP	SRP	SRP	BC	BC	BC	
Stage:	M3	M3	M3	M3	M3	M3	M2	M2	M2	M2	M2	M2	M2	M2	M2	M2	M1	M1	M1	
Source: [‡]	YP	YP	H	H	H	H?	P?	TF	TF	BJ	BJ	BJ	BJ	OH?	OH?	OH?	OV?	OV?	OV?	
Major and minor oxides (wt%): electron microprobe analyses																				
SiO ₂	74.2	73.3	73.6	71.7	73.1	73.1	72.9	70.6	72.7	72.9	72.8	73.3	73.2	71.9	70.1	71.9	72.5	67.5	69.9	0.50
TiO ₂	0.13	0.11	0.17	0.18	0.17	0.19	0.18	0.34	0.27	0.19	0.24	0.20	0.15	0.16	0.20	0.21	0.25	0.41	0.35	0.01
Al ₂ O ₃	11.7	11.7	11.7	11.4	11.3	11.3	11.2	11.4	11.4	11.5	11.2	11.5	11.5	11.3	11.3	12.1	11.5	11.9	11.7	0.20
Fe ₂ O ₃ **	1.65	1.70	1.50	1.17	1.23	1.29	1.82	2.02	2.03	2.08	1.81	1.88	1.47	1.37	1.81	2.23	2.95	5.3	4.18	0.02
MnO	0.04	0.04	0.04	0.03	0.04	0.02	0.04	0.03	0.03	0.03	0.03	0.03	0.03	0.03	0.02	0.03	0.05	0.13	0.09	0.005
MgO	0.02	0.02	0.08	0.09	0.07	0.08	0.04	0.14	0.08	0.03	0.07	0.06	0.06	0.06	0.06	0.10	0.03	0.09	0.11	0.003
CaO	0.52	0.57	0.57	0.45	0.45	0.44	0.51	0.73	0.64	0.63	0.54	0.59	0.52	0.53	0.56	0.68	0.74	1.58	1.21	0.01
BaO	0.01	0.02	0.07	0.10	0.05	0.03	0.04	0.14	0.06	0.02	0.07	0.07	0.00	0.02	0.04	0.02	0.11	0.12	0.12	0.01
Na ₂ O	2.8	3.4	2.8	2.6	2.7	2.5	2.3	1.8	2.4	2.4	2.8	2.3	1.8	1.8	1.7	1.8	2.5	3.0	2.7	0.20
K ₂ O	5.1	4.9	3.5	5.7	5.8	6.0	5.7	6.8	6.0	5.6	5.2	6.2	6.8	6.6	6.9	5.0	5.8	4.6	4.8	0.10
Cl	0.13	0.14	0.09	0.11	0.13	0.08	0.03	0.03	0.03	0.04	0.03	0.05	0.07	0.08	0.06	0.04	0.03	0.03	0.02	0.002
F	0.12	0.20	0.12	0.15	0.16	0.18	0.12	0.14	0.10	0.18	0.10	0.24	0.30	0.22	0.17	0.14	0.17	0.14	0.12	0.01
H ₂ O ^{††}	4.3	3.9	7.0	6.4	5.1	4.7	5.4	6.5	4.4	4.8	4.1	5.3	6.1	5.3	6.1	4.8	5.1	5.7	5.4	0.70
-OH ^{††}	0.10	0.13	0.08	0.10	0.11	0.10	0.07	0.08	0.07	0.10	0.07	0.13	0.15	0.12	0.10	0.08	0.09	0.10	0.09	0.03
Sum	100.7	99.9	101.2	100.0	100.2	99.8	100.3	100.6	100.1	100.3	99.0	101.6	101.9	99.3	99.0	99.0	101.7	100.5	100.7	0.70
Trace elements (ppmw): analyses by wavelength-dispersive X-ray fluorescence spectrometry																				
Ti	682	649	1004	1171	997	1138	1068	1871	1452	1152	1340	1182	951	865	1140	1519	1421	2750	2032	25
Mn	242	247	216	250	238	166	257	200	201	199	174	186	180	132	171	228	394	813	651	4
Zn	72	91	40	42	37	30	52	30	46	70	36	56	60	34	41	45	98	108	88	4
Rb	180	186	156	159	193	198	181	176	179	186	201	197	229	241	186	188	153	111	121	3
Sr	4	8	16	22	20	16	19	42	36	26	26	25	9	9	20	24	24	95	88	2
Zr	226	231	273	245	200	234	319	413	424	356	334	290	221	192	294	432	568	530	465	8
Nb	60	57	50	42	38	41	56	42	44	41	40	43	50	40	29	39	45	32	28	1
Ba	127	232	662	827	400	329	480	983	804	676	592	506	125	112	365	642	845	1540	1396	15
Ce	150	161	143	135	139	140	167	138	157	168	135	162	173	163	145	146	152	161	135	9
Th	23	29	23	27	27	30	27	26	27	27	33	30	33	37	28	24	17	17	17	2

*Estimated analytical precision (1 σ).

†un—unnamed; see Table 3 for key to named ash beds.

‡See Data Repository Table DR2 for data sources for age dates.

§Sources as in Figure 1; queried where speculative.

**Total Fe as Fe₂O₃.

††Estimated on basis of stoichiometry.

different sources are present in each locality, and all such tuffs were used to establish correlations between localities. However, only hotspot tuffs are included in Figure 5.

The ages of the ash fall tuffs have been determined variously by direct isotopic age dating, correlation to isotopically dated ash fall or ash flow tuffs, or by linear interpolation between dated horizons in measured sections (Table DR2). Methods of age estimation by interpolation are those of Perkins et al. (1998). In general the ages of hotspot tuffs are known to ± 0.3 m.y. (1 σ), though age estimates of a few tuffs are no better than ± 0.75 m.y.

In figures and text to follow, ash fall tuffs in particular time intervals are assigned to the source in the volcanic field that is most active during that interval. This convention is, in part, a convenience, but it also reflects the general shift in source with time along the hotspot track. The correlation of intervals and sources assumed in this study are as follows: (1) 16.0–15.2 Ma, Orevada volcanic province; (2) 15.2–12.7 Ma, Owyhee Humboldt volcanic field; (3) 12.67–10.5 Ma, Bruneau-Jarbridge volcanic field; (4) 10.5–8.5 Ma, Twin Falls

and, possibly, Picabo volcanic fields; (5) 8.5–7.0 Ma, Picabo volcanic field?—few hotspot ash flow tuffs are reported in this interval so the particular source(s) of hotspot tuffs in this interval is uncertain; (6) 7.0–3.5 Ma, Heise volcanic field; (7) 2.5–0 Ma, Yellowstone Plateau volcanic field. The interval 4.0–2.5 Ma represents a lull in explosive volcanism. Only one ash fall tuff is recognized in this interval, an unnamed ash fall tuff near the base of the Glens Ferry Formation (Fig. 5). The source of this tuff is unknown. These age ranges generally mark the initiation and cessation of ash flow tuff volcanism in a particular volcanic field and a change in the compositional character of the ash fall tuffs. However, the linkage between age range and source is only a useful first approximation, and we do not claim that all tuffs within these intervals have a single source.

CHARACTERISTICS OF HOTSPOT ASH FALL TUFFS

The ash fall tuff record commences at ca. 16 Ma as the early peralkaline rhyolitic erup-

tions were ending at the west end of the hotspot track (Figs. 2, 3). Thus, the record consists primarily of the metaluminous tuffs, and it is these tuffs that are the focus of this study. The composite record of hotspot tuffs consists of 142 typical metaluminous tuffs as well as 3 peralkaline tuffs and 1 unusual metaluminous tuff that we correlate with the 10.3 Ma Arbon Valley tuff (AV in Figs. 2 and 5) of the Picabo volcanic field.

The 142 typical post-16 Ma hotspot tuffs are metaluminous tuffs lying within the compositional range of the gray metaluminous tuffs of Perkins et al. (1998). These tuffs generally consist of $\geq 95\%$ glass shards along with minor quartz and feldspar, trace amounts of pyroxene and iron titanium oxides, and, only rarely, hydrous mafic minerals. Thus, the tuffs are mineralogically similar to the metaluminous to peraluminous rhyolite ash flow tuffs and lavas along the hotspot track.

As shown by representative analyses (Table 2), the gray metaluminous hotspot tuffs have glass shard compositions similar to A type granites with high SiO₂, alkalis, Fe₂O₃/MgO, Zr, Nb, and Ce and low CaO and Sr compared

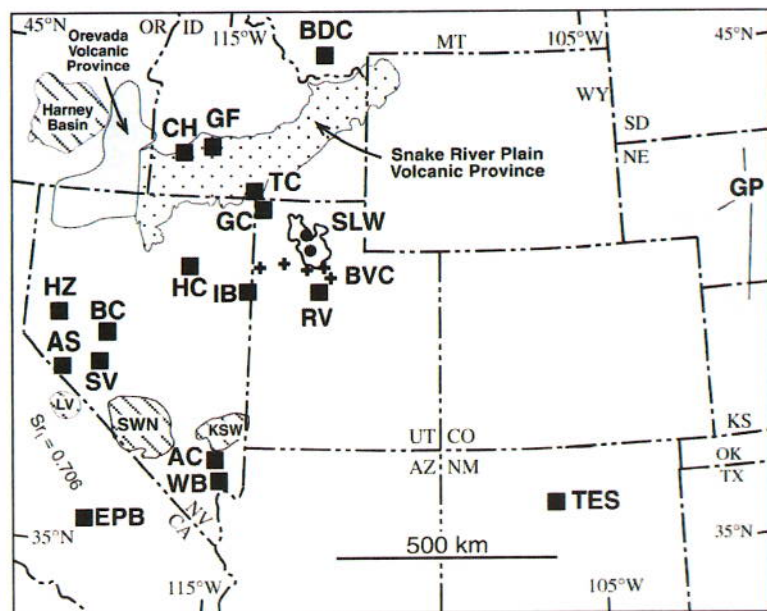


Figure 4. Location of stratigraphic sections. Surface stratigraphic sections (solid squares) include the following: AS—Aldrich Station, AC—Arrow Canyon, BC—Buffalo Canyon, CH—Chalk Hills, EPB—El Paso basin, GC—Grouse Creek, HZ—Hazen quarries, HC—Huntington Creek, IB—Ibapah badlands, RV—Rush Valley, SV—Stewart Valley, TC—Trapper Creek, WB—White Basin from Perkins et al. (1998); BDC—Blacktail Deer Creek and TES—Tesuque sections sampled for this study; and GF—Glenns Ferry section from Izett (1981) and Williams (1994; also 1994, written commun.). Other stratigraphic information includes ash beds in Bonneville basin cores (BVC; marked by crosses) from Williams (1994), ash beds sampled in cuttings from Great Salt Lake wells (SLW; marked by dots) from Moutoux (1995), and ash beds in the Great Plains (GP) from Izett and Wilcox (1982) and Perkins et al. (1995b; this study). Major silicic volcanic centers shown for reference: Harney Basin, KSW—Kane Springs Wash caldera complex, LV—Long Valley volcanic field, SWN—southwestern Nevada volcanic field, and the Orevada and Snake River Plain volcanic provinces along the hotspot track. Age ranges for these volcanic centers are shown in Figure 2.

to orogenic granites (Whalen et al., 1987) and are comparable in composition to the hotspot track rhyolites (Leeman, 1982). The three peralkaline hotspot tuffs in the post-16 Ma record (Fig. 5) resemble the gray metaluminous tuffs in some respects, but they have significantly lower concentrations of CaO (<0.20 wt%), Sr (<5 ppmw [parts per million by weight]), and Ba (<15 ppmw). These lower concentrations clearly distinguish the peralkaline tuffs from the metaluminous hotspot tuffs with similar Fe_2O_3 contents (1.60–2.9 wt%). One unusual metaluminous hotspot tuff, the biotite bearing Arbon Valley ash fall tuff, has a glass shard composition outside the compositional range of gray metaluminous hotspot tuffs with low Fe_2O_3 (1.1 wt%), high $\text{Al}_2\text{O}_3/\text{Fe}_2\text{O}_3$ (11.4), and high $\text{MnO}/\text{Fe}_2\text{O}_3$ (0.08).

The gray metaluminous hotspot tuffs fall into two compositionally and stratigraphically

distinct groups termed BC type (Buffalo Canyon type) and SRP type (Snake River Plain type) by Perkins et al. (1998). The earlier of these two groups, the BC type tuffs, are similar in many respects to the SRP type tuffs but are distinguished from them on the basis of generally higher contents of Fe, Ba, Mn, Sr, Zn, and Zr and lower contents of Rb and Th (Table 2). The least Fe rich of the BC type tuffs approaches the composition of SRP type tuffs, so the boundary between these two types of tuffs is somewhat arbitrary. In this study, the boundary is placed at $\text{Mn}/\text{Rb} = 2.2$; BC type tuffs have $\text{Mn}/\text{Rb} \geq 2.2$ (Mn and Rb contents as ppmw). By this criterion, there is neither a compositional nor stratigraphic overlap of BC type and SRP type tuffs (Fig. 5).

SRP type ash fall tuffs are linked to hotspot sources by a number of observations. First, and particularly compelling, is the compositional and chronological correlation of a num-

ber of SRP type ash fall tuffs to proximal ash flow tuffs along the hotspot track (Izett and Wilcox, 1982; Perkins et al., 1995b, 1998). Nineteen such correlations are now established for SRP type tuffs in the age range 0.60–12.7 Ma (Table 3). Second, as discussed shortly, SRP type ash fall tuffs show a distinct temporal variation in composition, and this variation parallels that seen in metaluminous hotspot ash flow tuffs and lavas (Fig. 3). Because it is implausible that other sources would show a similar temporal variation in composition, we have confidence that, with possible rare exception, all the SRP type tuffs come from hotspot sources. Finally, grain size, frequency, and thickness of SRP type ash fall tuffs decrease away from hotspot sources. The grain size decrease is shown in Figure 6 and frequency and thickness decreases are discussed later in this paper.

Linkage of the BC type ash fall tuffs to hotspot sources is more indirect, but several observations place the general source region for these tuffs at the western end of the hotspot track rhyolite. First, Nd isotope analyses of representative BC type tuffs show ϵ_{Nd} values that are consistently higher (all > -3.7) than those of SRP tuffs (all < -4.5) and within the range expected for hotspot rhyolites with sources between longitudes 117.5° and 120°W (Nash et al., 1997; Leeman et al., 1992). Second, as shown subsequently, there is a temporal gradation in the chemical composition of glass shards of BC type tuffs toward that of SRP type tuffs such that the youngest BC type tuffs approach the composition of SRP type tuffs. This chemical gradation, in conjunction with Nd isotope characteristics, indicates that the youngest BC type tuffs had sources just west of the Snake River Plain province, i.e., in the general region of the Orevada volcanic province. Third, there is a decline in grain size of BC type tuffs with distance from the Orevada province (Fig. 6). Finally, as with SRP type tuffs, BC type tuffs show rough decreases in both frequency and thickness with distance from the western end of the hotspot track rhyolite. All of these observations strongly support the Orevada volcanic province as a general source for BC type tuffs.

TEMPORAL VARIATIONS

The 142 gray metaluminous ash fall tuffs record well-defined temporal variations in magma composition, magma temperature, and eruption frequency. The pattern of these variations indicates three principal stages of post-16 Ma explosive metaluminous volcanism

EXPLOSIVE SILICIC VOLCANISM OF THE YELLOWSTONE HOTSPOT

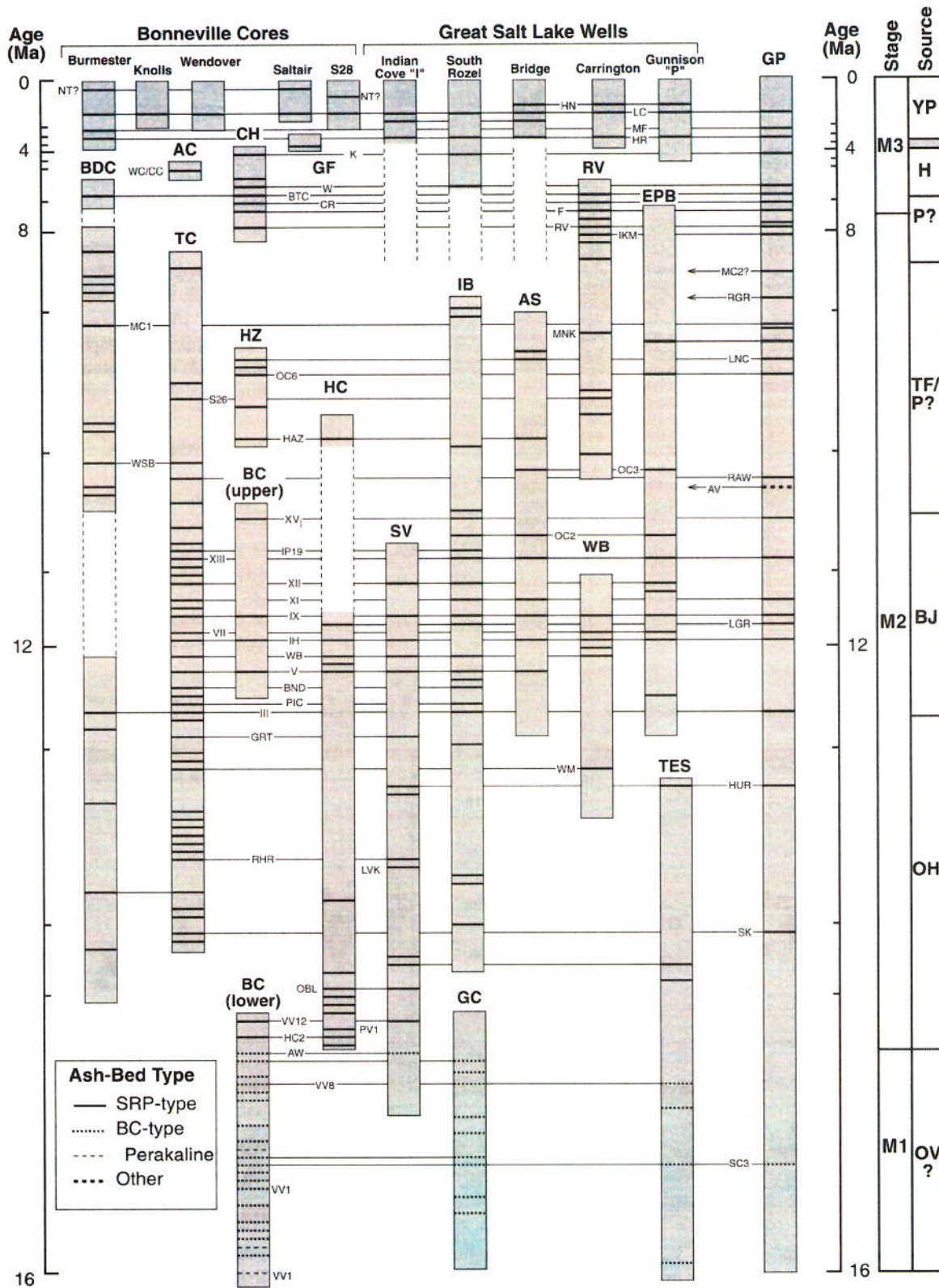


Figure 5. Correlation chart. Composite sequence of ash fall tuffs from hotspot sources from sections, core holes, and wells shown in Figure 4 and a preliminary composite sequence for the Great Plains (GP). Initialed ash beds referred to named ash beds. See Figure 2 and Table 1 for keys for these initials.

TABLE 3. NAMED ASH FALL TUFFS SHOWN IN FIGURE 5

Ash bed*	Initials	Age (Ma)	Ash bed	Initials	Age (Ma)
Natural Trap?	NT?	0.1	Ibex Peak 19	IP19	10.74
Natural Trap?	NT?	0.15	Cougar Point XIII	XIII	10.94
Hebgen Narrows	HN	0.39	Cougar Point XII	XII	11.19
Lava Creek B	LC	0.60	Cougar Point XI	XI	11.31
Indian Cove	IC	0.70	Cougar Point IX	IX	11.59
Mesa Falls	MF	1.30	Logan Ranch	LGR	11.79
Huckleberry Ridge	HR	2.06	Cougar Point VII	VII	11.80
Kilgore	K	4.45	Ibex Hollow	IH	11.93
Santee	SAN	5.10	Cougar Point V	V	12.07
Conant Creek–Wolverine Creek	CC-WC	5.58	White Basin	WB	12.09
Walcott	W	6.20	Banded	BND	12.14
Blacktail Creek	BTC	6.62	Pictograph	PIC	12.34
Cub River	CR	7.02	Cougar Point III	III	12.67
Faust	F	7.49	Grant Claim	GRT	12.96
Rush Valley	RV	7.90	Worthington Mine	WM	13.25
Inkom	IKM	8.30	Hurlbut	HUR	13.50
Reagar	RGR	8.80	Rawhide Ranch	RH	13.73
McMullen Creek 2	MC2	9.16	Lovelock	LVK	13.78
McMullen Creek 1	MC1	9.16	Shimek	SK	14.04
Mink Creek	MNK	9.24	Obliterator	OBL	14.93
Loneran Creek	LNC	9.41	Virgin Valley 12	VV12	15.18
Opal Canyon 6	OC6	9.52	Paradise Valley 1	PV1	15.17
Section 26	S26	9.70	Huntington Creek 2	HC2	15.21
Hazen	HAZ	9.81	Huntington Creek 1	HC1	15.21
Wooden Shoe Butte	WSB	10.13	Antonne Wash	AW	15.24
Opal Canyon 3	OC3	10.19	Virgin Valley 8	VV8	15.42
Arbon Valley	AV	10.20	Sheep Creek #3	SC3	15.55
Rawlins	RAW	10.25	Virgin Valley 1	VV1	15.60
Cougar Point XV]	XV]	10.45	Peacock	PEA	16.00
Opal Canyon 2	OC2	10.54	-	-	-

Note: See Data Repository Table DR2 for additional information.

*Ash fall tuffs correlated to ash flow tuffs along the hotspot track are underlined.

along the hotspot track rhyolite. The initial stage, stage M1 (metaluminous stage 1), was in progress at ca. 16 Ma and ended by 15.2 Ma. Stage M2 follows M1 without interruption and was active from 15.2 to 7.5 Ma. Stage M3 follows M2 without interruption at ca. 7.5 Ma and continues into the late Pleistocene with the explosive volcanism in the Yellowstone Plateau volcanic field. A pre-M1 stage, stage P, is hypothesized. Ash fall tuffs from this stage should be predominately peralkaline tuffs generated during the vigorous peralkaline volcanism in the Orevada province in the interval 16.5–16.0 Ma (Figs. 2, 3). However, we do not yet have stratigraphic sections covering this interval, so the nature of the hotspot ash fall tuff record during stage P is speculative.

Magma Composition

Variations in the glass shard composition of the tuffs record temporal changes in metaluminous rhyolite magmas generated along the hotspot track rhyolite. These variations include changes in the general character of magmas from stage to stage as well as systematic temporal variations in content of many com-

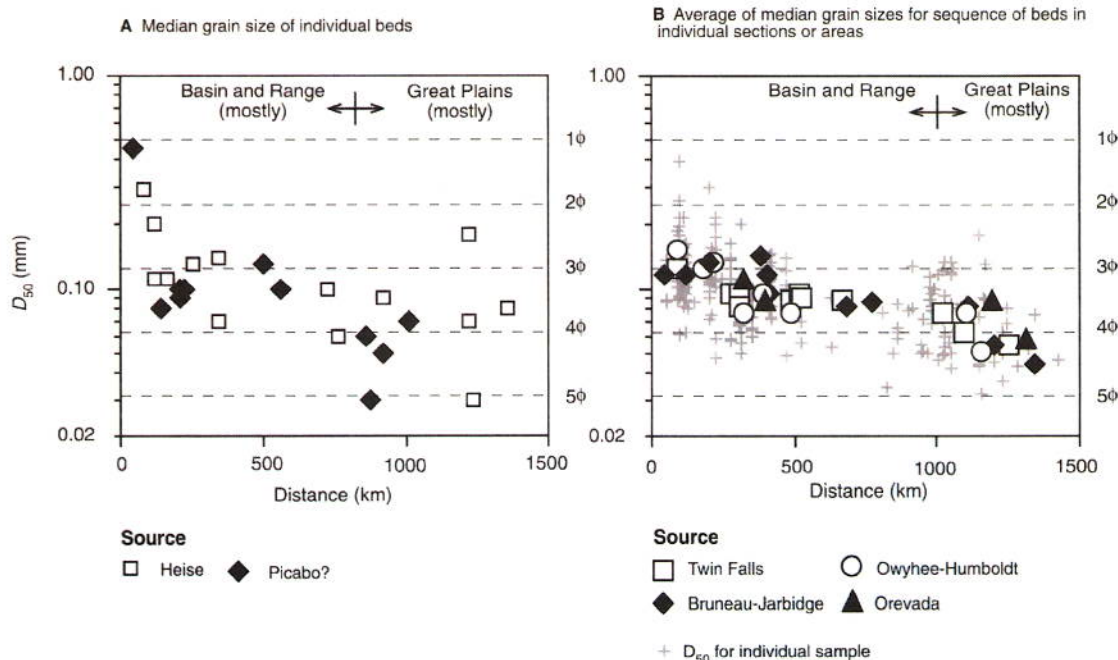


Figure 6. Approximate median grain size vs. distance from source for gray metaluminous hotspot tuffs. (A) Tuffs in 8.5–4.5 Ma age range. There are only one or two ash beds present in any one section, so the median size of individual beds is plotted. (B) Tuffs in 16.0–8.5 Ma age range. Two or more beds are present in typical sections covering this age range, so the average of the medians is plotted. Note grain size analyses are lower limits because mild crushing was needed to disaggregate these typically friable tuffs. This crushing likely reduced sample grain size. However, the median grain size of these samples lies in the size range estimated by hand lens inspection, so grain size reduction was $<1\phi$.

EXPLOSIVE SILICIC VOLCANISM OF THE YELLOWSTONE HOTSPOT

Ash-fall tuff glass shards

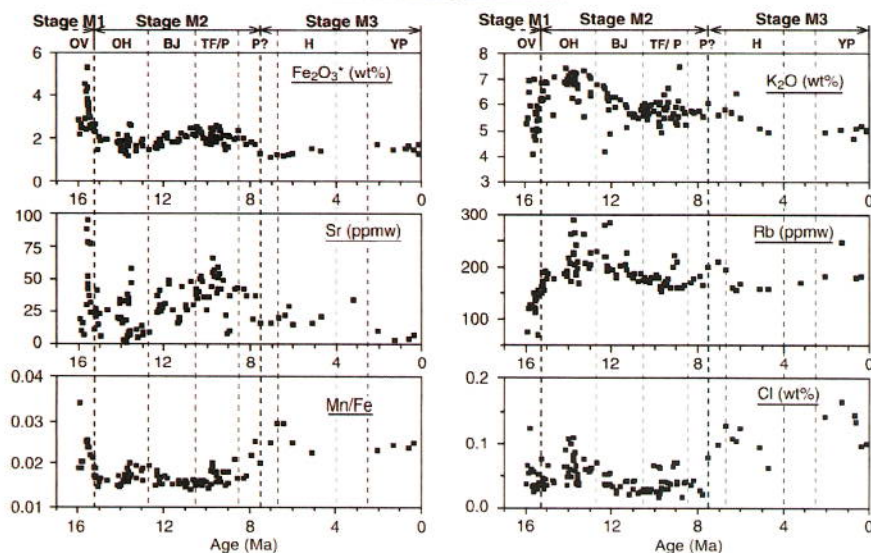


Figure 7. Temporal variation in the compositions of glass shards from hotspot metaluminous rhyolite ash fall tuffs. Each data point is an analysis for a separate ash bed. Fe_2O_3^* (= total Fe as Fe_2O_3), K_2O , and Cl contents are from electron microprobe analyses and are the averages for ~20 shards per tuff. Sr and Rb contents and the Mn/Fe ratios from X-ray fluorescence analyses. See Table DR3 for ages and analyses of individual tuffs. Volcanic fields: OV—Orevida volcanic province, OH—Owyhee-Humboldt volcanic field, BJ—Bruneau-Jarbridge volcanic field, TW/P—Twin Falls or Picabo volcanic fields, P?—possibly Picabo volcanic field, H—Heise volcanic field, YP—Yellowstone Plateau volcanic field.

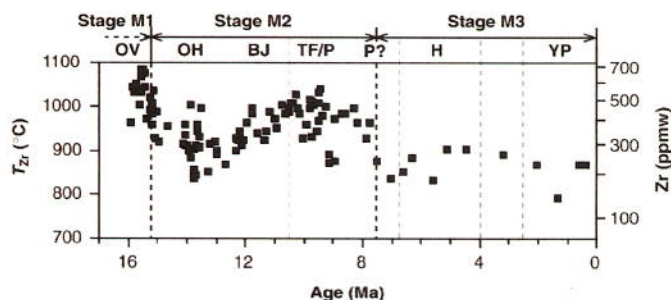


Figure 8. Temporal variation in magma temperature of metaluminous ash fall tuffs from hotspot track sources. Magma temperatures estimated from Zr content (X_{Zr}) in glass shards (ppmw) using relationship $10\,000/T = -1.02 + 1.27 \times \ln(500\,000/X_{\text{Zr}})$, where temperature, T , is in Kelvins. This relationship is from Watson and Harrison (1983) with constants from calibration of Perkins et al. (1995a). For volcanic field abbreviations, see Figure 7.

patible and incompatible elements during stages M1 and M2 (Fig. 7).

The BC type tuffs were erupted in stage M1. Temporal variation in the composition of these tuffs includes increases in the concentration of incompatible elements such as Rb, a decrease in the Mn/Fe, and, in the later half of stage M1, a marked decline in concentration of compatible elements such as Fe and

Ba (Fig. 7). The youngest M1 tuffs approach the composition of the oldest SRP type tuffs of stage M2. Thus, there is no pronounced break in magma composition at the M1/M2 boundary. As discussed previously, sources for the BC type tuffs of stage M1 were most likely within the Orevida volcanic province.

Stage M2 began at ca. 15.2 Ma with a shift from BC type tuffs to SRP type tuffs (Fig. 5).

During M2 there was an overall trend toward eruption of progressively less evolved tuffs. This trend is expressed by the increase of compatible element concentrations and associated decline in incompatible element concentrations (Fig. 7). Sources for M2 tuffs include the Owyhee-Humboldt, Bruneau-Jarbridge, and Twin Falls volcanic fields (Perkins et al., 1995a). Some younger M2 tuffs may be from sources in the Picabo volcanic field. At the end of M2 there is a decline in compatible element concentrations and an increase in Mn/Fe values toward those of typical M3 tuffs.

Eruption of SRP type tuffs continued through stage M3. For most elements the glass shard compositional range of stage M3 tuffs is coincident with that of the more evolved of M2 tuffs. However, M3 tuffs are distinguished by higher Cl concentrations, higher Mn/Fe values, and, for a given Zr concentration, lower Rb concentrations. The sources of the 6.7 Ma and younger M3 tuffs are mainly in the Heise and Yellowstone Plateau volcanic fields as shown by direct correlation of most of these tuffs to proximal ash flow tuffs in these volcanic fields (Fig. 2; Table 3). Sources for the oldest M3 tuffs are uncertain, but may include both the Heise and Picabo volcanic fields. Unlike stages M1 and M2, pronounced temporal trends in composition are lacking in M3 tuffs, but tuffs from the Heise volcanic field are generally less evolved than those from the Yellowstone Plateau volcanic field (i.e., higher Ba and lower Rb and Cl; Fig. 7).

Magma Temperature

The temporal variation in magma temperature (Fig. 8) is based on the zircon saturation thermometer of Watson and Harrison (1983) adjusted to fit mineral pair temperature estimates of metaluminous rhyolite tuffs and lavas (Perkins et al., 1995a). Zircon is a common trace constituent in the metaluminous tuffs, and the decline of Zr with fractionation (Perkins, 1998b) is consistent with coprecipitation of trace amounts of zircon. So, as required by the Watson and Harrison model, the metaluminous hotspot magmas were likely saturated in zircon. Although these temperature estimates lack the precision of those derived from mineral pair equilibria, they are in accordance with previously reported mineral pair temperatures for hotspot metaluminous tuffs (Izett, 1981; Hildreth, 1981; Honjo et al., 1992; Manley, 1996). Temperatures peaked in the middle of stage M1 (at >1000 °C) and again near the end of stage M2 (at ~1000 °C). They declined to 800–900 °C by the start of

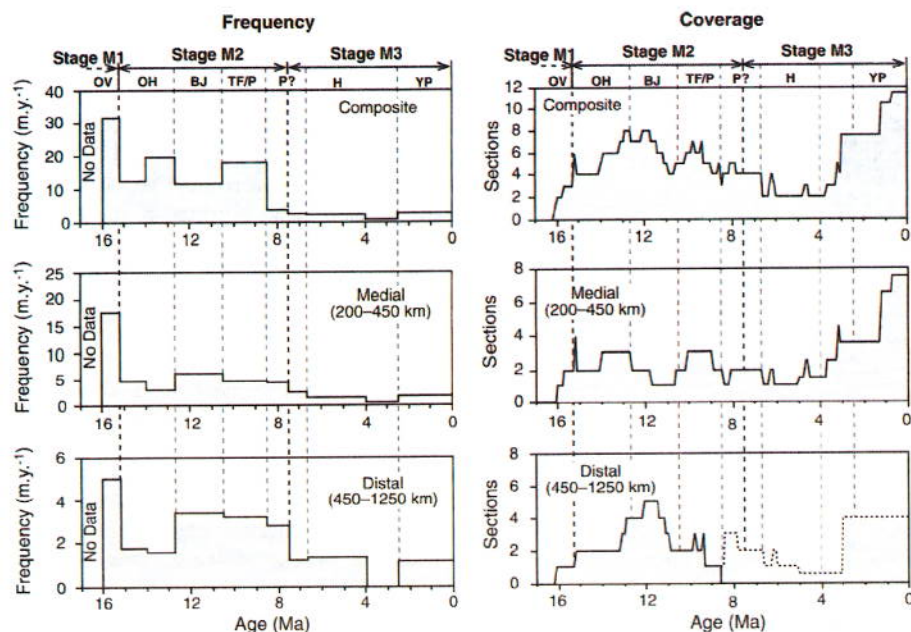


Figure 9. Temporal variation in frequency of gray metaluminous ash fall tuffs. Frequency is the number of tuffs in a particular time interval divided by the interval duration. For the composite record, the frequency is the total number of tuffs in the interval divided by the interval duration. For the medial and distal records, the frequency is a relative frequency, i.e., the average of frequencies in individual sections covering the interval. Interval breaks for the frequencies are at 16, 15.2, 14, 12.7, 10.5, 8.5, 7.5, 6.7, 4, 2.5, and 0 Ma. This coverage graph shows the number of sections that sample a particular time and provides an estimate of the variation in relative coverage with time. The solid line in the coverage graphs indicates the number of sections or wells that sample each 0.1 m.y. interval, and a dashed line (distal sections only) is a qualitative estimate of coverage. The estimated precision for relative frequency in the medial and distal records is $\sigma = n^{1/2}$, where n is the average number of sections sampling the interval. For volcanic field abbreviations, see Figure 7.

stage M3 and persisted at these relatively low values throughout stage M3.

Eruption Frequency

The gray metaluminous ash fall tuffs show pronounced variation in the frequency of eruption with relatively few tuffs erupted since ca. 8 Ma and a great many more erupted from 16 to 8 Ma (Fig. 5). Here, three subsets of the ash fall record are examined, the composite, medial (200–450 km), and distal (450–1300 km) records. Each of these records reflects different facets of hotspot explosive volcanism, and each has particular strengths and biases. In aggregate, these records show three intervals of differing eruption frequency that are coincident with stages M1, M2, and M3.

Composite Record

The composite frequency record for the gray metaluminous tuffs is shown in (Fig. 9). Temporal variation in eruption frequency is

clearly suggested by the composite record that shows very high frequency from 16 to 15.2 Ma (32 tuffs/m.y.), high frequency from 15.2–8.5 Ma (~10–20 tuffs/m.y.), and low frequency since 8.5 Ma (~2.5 tuffs/m.y.), i.e., roughly coincident with the M1, M2, and M3 compositional intervals. The composite record provides the best estimate of total number of explosive eruptions. Thus, it is a proxy for proximal eruption frequency. However, this record has biases that need to be considered when assessing its significance.

For example, different time periods are sampled by varying numbers of stratigraphic sections, and these sections are located at differing distances and directions from sources. The periods 15–7 Ma and 3–0 Ma are sampled most frequently; less dense sampling has occurred for other periods (Fig. 9). Because no single section or boring preserves all tuffs, as is clearly evident in Figure 5, more will appear in the more frequently sampled time intervals. If the completeness of record for any interval

could be estimated, then the composite frequency record could be adjusted to yield an unbiased record. However, estimation of sampling completeness is complex and cannot yet be made with full confidence.

In addition, correlation difficulties within some groups of tuffs introduce further bias. In particular, the tuffs in the 10.5–8.5 Ma interval have proved difficult to correlate between sections with confidence. The reasons for this are not fully understood but include compositional similarity among these tuffs and stratigraphic variation in the composition of glass shards within individual tuffs. A conservative approach to correlation is used in this study, namely, “when in doubt, don’t correlate.” Hence, the number of 10.5–8.5 Ma tuffs is likely overestimated in the composite record.

Coverage bias is reduced by using subsets of the record with more uniform coverage than the composite record. Correlation bias is reduced by using average frequencies for individual sections. Two distance-based subsets of the record are examined: The medial record (200–450 km from source) and the distal record (450–1400 km from source); each subset was assembled by using the average frequency of tuffs in individual sections or areas over particular time periods.

Medial Record

The medial record (Fig. 9) provides the best representation of variation in relative eruption frequency. The reasons for this statement are several. First, most of the metaluminous ash fall tuffs, 107 out of 142, are present in medial sections. Second, these sections are mostly south of the hotspot track rhyolite (Fig. 4) and mostly consist of similar fine grained fluvial deposits with some lacustrine intervals. Finally, coverage is relatively uniform with 1–3 sections covering any time interval from 16 to 4 Ma (Fig. 9). Thus, the medial record provides a relatively unbiased sampling in the sense that (1) time intervals are sampled at similar distances and directions from sources and with similar number of sections, and (2) the tuffs themselves occur in similar sedimentary environments that likely have similar tuff preservation efficiencies.

The pattern of frequency variation observed in the composite record is repeated in the medial record. The main differences between the two records occur in stage M2, i.e., the frequency peaks at 14–12.7 Ma and 10.5–8.5 Ma in the composite record are not duplicated in the medial record. These differences probably reflect biases in the composite record. The 14–12.7 Ma peak in the composite record reflects the high frequency of 14–12.7 Ma tuffs in the

proximal Trapper Creek section (Fig. 5). Apparently, many tuffs erupted in this interval are not widely distributed. The frequency peak at 10.5–8.5 Ma in the composite record is thought to reflect the correlation difficulties already discussed, and the result is an overestimation of the number of individual tuffs for this interval in this record.

Distal Record

The distal record (Fig. 9) is an amalgamation of several data sources and has a variable consistency. For the interval 16–8.5 Ma, individual sections are used to estimate average frequency per section. For the interval <8.5 Ma, such sections are not available. Instead, results of regional studies in the northern Great Plains (Boellstorff, 1978; Izett and Wilcox, 1982; Perkins et al., 1995b; Perkins, 1998a), along with our information from the Basin and Range, are used to estimate composite distal frequencies for time periods since 8.5 Ma.

Although the distal record is less comprehensive and more variable in quality than the other records, there is, nonetheless, a strong consistency between this record and the medial and composite records. Both stages M1 and M2 are well defined. The transition between these stages is sampled in individual stratigraphic sections (Fig. 5), so the M1/M2 contrast likely reflects actual differences in relative frequency rather than sampling bias. If anything, the contrast in distal frequency between stages M1 and M2 is muted by distance bias, i.e., the M1 record is from the very distal (~1300 km) Tesuque Formation section in the Rio Grande rift, whereas the younger than 13.5 Ma part of the M2 record is from less distal (450–900 km) sections in the Basin and Range. Comparisons of the distal M2–M3 transition are less direct because the distal M2 frequencies are average section frequencies, whereas the distal M3 frequencies are composite frequencies from combined records in the Great Plains and Basin and Range. Composite frequencies are, in general, higher than average section frequencies. So it is likely that the distal M3 frequencies are exaggerated relative to distal M2 frequencies. Hence, the distal stage M2–M3 frequency contrast is likely muted by this bias.

Distance is a strong filter of the ash fall tuff record. On average, only the thicker, more widespread ash layers are likely to be preserved in regions far from source. Thus, the distal record preferentially preserves large magnitude explosive eruptions. This record indicates that such eruptions have occurred re-

peatedly over the past 16 Ma but with generally declining frequency.

Comparison with the Record of Ash Flow Tuffs

Comparisons of compositional and frequency trends between ash fall tuffs and proximal ash flow tuffs reveal both similarities and differences. Characteristics of both stages M2 and M3 trends are apparent in the temporal variation in the frequency of occurrence of ash flows and their Fe_2O_3 contents (Fig. 3). In particular, frequency patterns for ash flow tuffs are similar to those of the medial ash fall tuff record: an average of 4–5 tuffs/m.y. during the M2 stage, an average of ~1 tuff/m.y. during M3, and a peak in tuff frequency for the brief M1 interval. Similarly, the trends in whole-rock Fe_2O_3 contents in the ash flow tuffs parallel the trends in glass shard Fe_2O_3 contents in ash fall tuffs; the distinct rise in Fe_2O_3 content during M2 is evident in both records followed by a sharp decline to low Fe_2O_3 contents during M3. Differences in the records include the greater number of eruptions recorded in the ash fall tuff record than in the ash flow tuff record (142 vs. ~70) and the possible absence of ash flow tuffs with M1 compositions.

The difference in the number of recorded M2 and M3 eruptions in the two records is thought to be mainly one of exposure bias. It is likely that only the most widespread of ash flow tuffs are commonly preserved in the M2 and M3 volcanic fields, with less extensive ash flow tuffs concealed beneath the widespread tuffs and the typically extensive cover of younger lava flows and sedimentary strata. If one compares the record of exposed M2 and M3 ash flow tuffs with that of the distal M2 and M3 ash fall tuffs, then the number of ash flow and ash fall tuffs (~70 vs. 52) and their relative frequencies are roughly equivalent. Recalling that only the larger magnitude explosive eruptions are likely preserved in the distal sections, there is fair agreement between the number of exposed ash flow tuffs and distal ash fall tuffs. The ash fall tuffs that are known only from the proximal record are likely associated with either the eruption of smaller magnitude ash flow tuffs or the less explosive eruption of silicic lava flows.

The apparent absence of either ash flow tuffs or lavas with the compositional signature of the M1 tuffs is striking (Fig. 3). One might argue from this that the M1 tuffs are not from hotspot sources, but we regard the indirect evidence for hotspot sources as compelling. We consider the apparent absence of M1 ash flow

tuffs to reflect an information gap more than an actual absence. Much of the region between the Bruneau-Jarvis volcanic field on the east and the McDermitt volcanic field on the west (Figs. 1, 2) has extensive cover of post-hotspot basalt and lacks detailed studies of the stratigraphy, geochemistry, and geochronology of exposed ash flow tuffs and lavas. We anticipate that the source(s) of the M1 ash fall tuffs will be identified in future studies.

Discharge

The gray metaluminous ash fall tuffs are typically thick, widespread, and abundant throughout the intermontane west and northern Great Plains. This fact points to exceptionally high rates of discharge of atmospherically transported hotspot tephra. Although the quantitative estimation of discharge rates is complex and not undertaken in this study, some first-order estimates of relative volumetric discharge are feasible. These estimates, based on comparisons with reference ash beds, indicate that typical hotspot ash fall tuffs have volumes of 250–600 km^3 and that the average ash fall discharge was on the order of 600 $\text{km}^3/\text{m.y.}$ during stage M3 though earlier stages had higher discharges.

The thicknesses of older hotspot ash fall tuffs and three reference ash beds are shown in Figure 10. The reference ash beds are the 0.6 Ma Lava Creek B ash bed, the 2.06 Ma Huckleberry Ridge ash bed, and the 0.76 Ma Bishop ash bed. These widespread reference ash beds were generated during large volume ash flow tuff eruptions. The first two were generated during the eruptions of the >1000 km^3 Ma Lava Creek Tuff and >2400 km^3 Huckleberry Ridge Tuff from Yellowstone Plateau sources (Christiansen, 1984; Izett and Wilcox, 1982). The third ash bed was generated during the eruption of the ~500 km^3 Bishop Tuff from Long Valley caldera in southeastern California (Bailey et al., 1976; Izett et al., 1988). These ash beds are selected because their distribution areas and thicknesses are similar to that of the older hotspot tuffs.

The volume of the reference ash beds is considerable but not well determined. Estimates available for some large magnitude Quaternary eruptions suggest that ash fall tuff volume is ≥ 0.5 times ash flow tuff volume (Izett et al., 1988). Thus, the volume of the widespread (~3 $\times 10^6$ km^3) Bishop tuff is likely ≥ 250 km^3 , whereas the volumes of the more widespread Lava Creek B and Huckleberry Ridge ash beds (~4 $\times 10^6$ km^3) are likely >500 km^3 . As discussed in the next section,

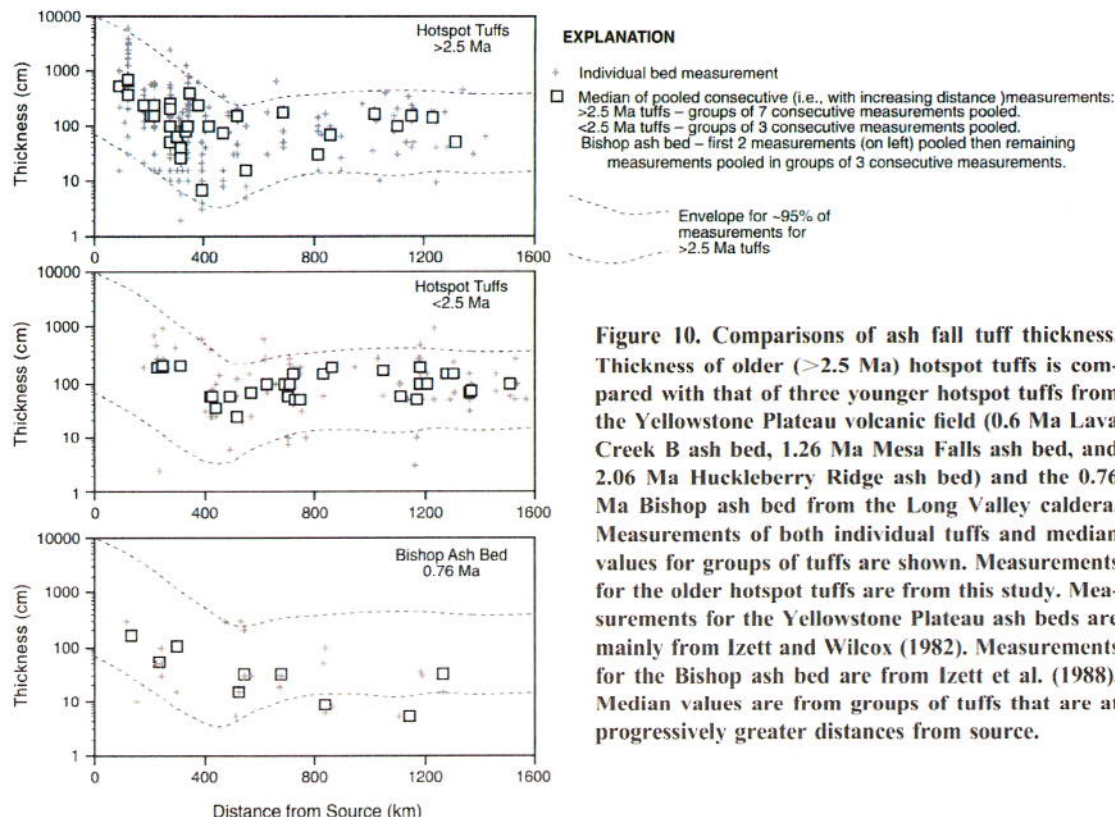


Figure 10. Comparisons of ash fall tuff thickness. Thickness of older (>2.5 Ma) hotspot tuffs is compared with that of three younger hotspot tuffs from the Yellowstone Plateau volcanic field (0.6 Ma Lava Creek B ash bed, 1.26 Ma Mesa Falls ash bed, and 2.06 Ma Huckleberry Ridge ash bed) and the 0.76 Ma Bishop ash bed from the Long Valley caldera. Measurements of both individual tuffs and median values for groups of tuffs are shown. Measurements for the older hotspot tuffs are from this study. Measurements for the Yellowstone Plateau ash beds are mainly from Izett and Wilcox (1982). Measurements for the Bishop ash bed are from Izett et al. (1988). Median values are from groups of tuffs that are at progressively greater distances from source.

we use the lower bound estimates for the average volume of widespread hotspot tuffs.

The thicknesses of both the older hotspot tuffs and the reference ash beds show considerable scatter at any particular distance interval from the source (Fig. 10). This scatter reflects fluvial and eolian reworking and, for the source groupings of hotspot tuffs, variation in primary ash fall thickness for different eruptions and different transport directions. Given the similar environments in which the older hotspot tuffs and reference ash beds are preserved, i.e., fluvial deposits of the Basin and Range and Great Plains, the impact on preserved thickness due to reworking should be the similar, on average, for both groups. So scaling of the older tuffs relative to the reference ash beds is justified.

Both the median thickness and the general range of ash bed thickness are roughly comparable between the older hotspot tuffs and the Yellowstone Plateau ash beds and 3–4 times greater than those of the Bishop ash bed. So our first-order estimates are that the more widespread of the older hotspot ash fall tuffs (i.e., those present in distal sections) have volumes in excess of the Bishop ash bed and that the most widespread have volumes similar to the Lava Creek B and Huckleberry Ridge ash

beds. We thus estimate the average volume of the most widespread of distal tuffs at 500 km³. These are the distal tuffs identified in the Basin and Range as well as in either the Great Plains or Rio Grande rift. For less widely distributed distal tuffs, we use an average volume of 250 km³. This approach is a conservative weighting of the data for several reasons: First, the knowledge of the distal distribution of the hotspot tuffs, particularly in the intervals 16.0–12.7 and 8.5–6.7 Ma, is incomplete, and as additional distal tuffs are identified, both the number of tuffs and the proportion of very widespread distal tuffs will increase. Second, the considerable volume of tuffs found only in proximal or medial sections is ignored. With these caveats in mind, our first-order estimates of ash fall tuff discharge are given in Table 4. We claim no great accuracy for these estimates, but we regard them as plausible both as estimates of relative variation in discharge and as conservative first order approximations of actual discharge.

DISCUSSION

The hotspot ash fall tuffs provide an unparalleled, long-term record of explosive volcanism in one of the world's premier active si-

lic volcanic systems. As shown by this record, there have been important changes in the volcanic expression of hotspot silicic magmatism since 16 Ma. These changes include temporal variation in magma composition and temperature, eruption frequency, and, likely, discharge. When the processes and interactions controlling such changes are better understood, a fuller understanding of both the nature of silicic magmatism and the nature of the Yellowstone hotspot should emerge.

One striking feature of the temporal variation in the hotspot tuffs is the roughly synchronous variation in magma temperature with eruption frequency (Figs. 8, 9). This variation suggests that a principal process controlling explosive volcanism is variable thermal input into hotspot silicic magmatic systems, with intervals of higher magma temperature and eruption frequency corresponding to intervals of greater thermal input. That this may indeed be the case is reinforced by our estimates of discharge, with intervals of higher discharge also corresponding to those of higher magma temperature and eruption frequency.

Because the primary source of heat driving hotspot magmatic systems is mantle derived basalt (Hildreth et al., 1991), we suggest that

EXPLOSIVE SILICIC VOLCANISM OF THE YELLOWSTONE HOTSPOT

TABLE 4. ESTIMATED DISTAL ASH FALL TUFF DISCHARGE

Age range (Ma)	Source	Distal tuffs*	Frequency (m.y. ⁻¹)	Average tuff volume (km ³)†	Discharge (km ³ /m.y.)‡
0.0–2.5	YP	3 (2)	1.2	425	500 [750]
2.5–4.0	N.D.	0	0.0	N.D.	N.D.
4.0–6.7	H	4 (2)	1.5	375	550 [650]
6.7–7.5	H?	2	2.5	250	625
7.5–8.5	P?	4	4.0	250	1000
8.5–10.5	TF	13 (2)	6.5	300	1880
10.5–12.7	BJ	14 (6)	6.5	350	2300
12.7–14.0	OH	4 (1)	3.0	325	950
14.0–15.24	OH	4	3.2	250	800
15.2–16.0	OV	4 (1)	5.3	325	1650

Note: N.D.—not determined.

*Total number of tuffs and number of tuffs identified in the Basin and Range and either the Great Plains or Rio Grande rift given in parentheses. Includes all tuffs shown in Figure 4. Further information on several other distal Basin and Range sections with Heise volcanic field tuffs incorporated into the distal tally.

†Weighted average. Volume of distal tuffs that are identified only in the Basin and Range, or Rio Grande rift, or Great Plains is estimated at 250 km³, whereas that of the more widespread tuffs is estimated at 500 km³.

‡Values in brackets are estimates on the basis of half of the reported discharge from major ash-flows. Plausible volume estimates are only available for ash-flow tuffs in the Yellowstone Plateau and Heise volcanic fields.

variable input of basalt from the mantle is a first-order process controlling explosive hotspot volcanism. Factors that may influence the input of mantle basalt into silicic magma-generating systems are likely numerous and may have varied over the lifetime of the hotspot track rhyolite. Several that may be important are suggested by temporal and spatial correlations between hotspot explosive volcanism and changes in both the crust and mantle along the hotspot track as well as changes in the hotspot migration rate.

Since its inception, the Yellowstone hotspot has traversed a variety of geologic terranes (Fig. 1). The first-order boundary traversed by the hotspot is the interface between accreted Paleozoic and Mesozoic oceanic terranes lying west of the $S_r = 0.706$ line and the Precambrian craton of North America to the east. As discussed earlier, this boundary was being traversed during stage M1, a stage in which mantle input into hotspot silicic magmatic systems appears to have been high. We suggest that the productivity of basalt within the mantle anomaly driving the hotspot was higher in the oceanic mantle to the west of the 0.706 line than it was within continental mantle immediately to the west of the 0.706 line. Greater productivity is suggested by the voluminous outpouring of the Steens Mountain Basalt and Columbia River Basalt Group at the initiation of hotspot volcanism (Zoback et al., 1994). Such voluminous precursory basalts (relative to silicic volcanism) are largely absent east of the 0.706 line (Fig. 2), suggesting exceptional mantle productivity west of this lithospheric boundary. It is also worth repeating that the M1/M2 transition appears quite abrupt. This abruptness may reflect both the contrast in basalt productivity across the 0.706 line and the

sharp character of this lithospheric boundary (Fleck and Criss, 1985; Leeman et al., 1992).

East of the 0.706 line, the isotopic characteristics of hotspot track basalts appear uniform (Leeman et al., 1992), suggesting a uniform mantle source for these basalts. So the processes modulating mantle input into M2 and M3 magmatic systems may lie within the crust or along the crust/mantle interface rather than within the mantle. One controlling process may be the proportion of mantle basalt that underplates the crust. If this proportion is high, then the only mantle input into the crustal magmatic systems will be heat diffused into the crust. Such a process is thermally inefficient and capable of producing only a limited volume of melt within the crust (Pedersen et al., 1998). In contrast, if a high proportion of mantle basalt is intruded into the crust, then widespread fusion of crustal protolith is feasible (Bergantz, 1995). The variations observed during stages M2 and M3 could be controlled by the degree of basaltic underplating; intervals of lower magma temperatures and eruption frequency would correspond to greater underplating and the associated lower efficiency of thermal input into the crustal magmatic systems.

The degree of underplating may be influenced by the character of the crust. There are correlations between crustal character and hotspot volcanism. As mentioned previously, the M1/M2 transition marks the transit of the hotspot onto the leading western edge of the craton. Hence, variation in amount of underplating on either side of this boundary, as well as mantle productivity, may contribute to the M1/M2 contrast. During M2, the hotspot track was within the northern part of the Paleozoic Roberts Mountain allochthon with its "eugeo-

clinal" suite of basalt and deep water marine strata (Fig. 1). By the start of M3, the hotspot was east of the Roberts Mountain allochthon and within crust of the miogeoclinal part of the Paleozoic geocline along the western margin of the craton. Although we cannot offer specific mechanisms that could control the degree of crustal underplating between these two contrasting Paleozoic terranes, it is feasible that differences within deeper crustal levels of these terranes contribute to the M2/M3 contrasts.

Finally, we want to explore the possible implications of the variation in the apparent rate of hotspot migration to hotspot volcanism. The average migration rate during M3 has been ~25 km/m.y. (Fig. 2). However, the rate during M2 and M1 was ~50 km/m.y., twice the M3 rate. This earlier higher rate could have been due to a higher drift rate for the North America plate, a higher rate of extension along the hotspot track, a higher rate of hotspot migration relative to deep mantle, or some combination of these. We suggest that the important factor governing production of silicic magmas is the relative rate of motion between the mantle thermal anomaly and the crust. The intensity of this anomaly may be sufficient to melt a fixed fraction of the crust, regardless of migration rate. Thus, the amount of silicic melt generated could increase as the rate at which fertile crust passes over the mantle anomaly increases, given a fixed ratio of underplated to intruded basalt. Furthermore, if the increased migration rate is, in part, due to extension, such extension could have enhanced intrusion of basalt into the crust and facilitated additional melting. Thus, we propose that some of the M2/M3 contrasts, particularly the apparent higher discharge during M2, reflects the higher migration rate prior to ca. 8 Ma.

CONCLUSIONS

Late Cenozoic silicic volcanic centers along the Yellowstone hotspot track have generated numerous, widespread, compositionally distinctive gray metaluminous rhyolite ash fall tuffs. These tuffs offer an unparalleled record of explosive volcanic activity along the hotspot track. Evaluation of a comprehensive record of 142 of these tuffs reveals systematic temporal variation in magma composition and temperature, eruption frequency, and, possibly, magma discharge rates from hotspot sources since 16 Ma. These variations show that explosive activity along the Yellowstone hotspot track proceeded in stages, with each stage marked by distinctive magma composi-

tions, eruption frequencies, and magma temperature ranges and trends. The partitioning of explosive volcanism into stages likely reflects variation in the style and intensity of the interaction between the mantle anomaly powering the hotspot magmatic systems and a spatially and temporally heterogeneous lithosphere along the hotspot track. Although the ash fall tuff record does not provide definitive constraints on the nature of these variable interactions, correlations between changes in eruption frequency, ash fall tuff discharge and magma temperature point to variable input of mantle derived basalt into hotspot crustal magmatic systems as a first-order control on intensity of explosive volcanism. As future studies reveal more about the processes controlling the variations in hotspot silicic volcanism, a fuller understanding of both the nature of silicic magmatism and the nature of the Yellowstone hotspot should emerge.

ACKNOWLEDGMENTS

We thank Jay Quade, University of Arizona, for samples of ash fall tuffs from the Tesuque Formation in the Rio Grande rift and for showing us the outcrops where these samples were collected. We thank William Fritz, Georgia State University, for a tour of Ruby basin, Montana, and for showing us some of the key exposures of ash fall tuffs in this basin. We thank Robert F. Diffendal Jr., Michael R. Voorhies, and Bruce Bailey, University of Nebraska, for providing samples of ash fall tuffs from Nebraska. We also thank Francis H. Brown and Robert B. Smith, University of Utah, for valuable discussions on the tephrochronology, the volcanism, and the nature of the Yellowstone hotspot. Finally, we thank Eric H. Christiansen and Anita L. Grunder for thoughtful reviews of the original manuscript. This investigation was supported by National Science Foundation grant EAR 9316289 (to B.P. Nash) and EAR 9204700 (to F.H. Brown and B.P. Nash).

REFERENCES CITED

- Anders, M.H., Geissman, J.W., Piety, L., and Sullivan, T., 1989, Parabolic distribution of circum eastern Snake River Plain seismicity and latest Quaternary faulting: Migratory pattern and association with the Yellowstone hotspot: *Journal of Geophysical Research*, v. 94, p. 1589–1621.
- Armstrong, R.L., Leeman, W.P., and Malde, H.E., 1975, K Ar dating, Quaternary and Neogene volcanic rocks of the Snake River Plain, Idaho: *American Journal of Science*, v. 264, p. 225–251.
- Bailey, R.A., Dalrymple, G.B., and Lanphere, M.A., 1976, Volcanism, structure, and geochronology of Long Valley caldera, Mono County, California: *Journal of Geophysical Research*, v. 81, p. 725–744.
- Bergantz, G.W., 1995, Changing techniques and paradigms for the evaluation of magmatic processes: *Journal of Geophysical Research*, v. 100, p. 17603–17613.
- Best, M.G., Christiansen, E.H., Deino, A.L., Gromme, C.S., McKee, E.H., and Noble, D.C., 1989, Excursion 3A: Eocene through Miocene volcanism in the Great Basin of the western United States, in Chapin, C.E., and Zidek, J., eds., *Field excursions to volcanic terranes in the western United States*, Volume II: Cascades and Intermountain West: New Mexico Bureau of Mines and Mineral Resources Memoir 47, p. 91–133.
- Boellstorff, J.D., 1978, Chronology of some late Cenozoic deposits from the central United States and the ice ages: *Nebraska Academy of Sciences Transactions*, v. 6, p. 35–49.
- Bonnichsen, B., 1982, Chemical composition of the Cougar Point Tuff and rhyolite lava flows from the Bruneau-Jurbidge eruptive center, Owyhee County, Idaho: Idaho Bureau of Mines and Geology Technical Report 82-1, p. 1–22.
- Christiansen, R.L., 1984, Yellowstone magmatic evolution: Its bearing on understanding large volume explosive volcanism, in *Geophysics Study Committee, Explosive volcanism: Inception, evolution and hazards*: Washington, D.C., National Academy Press, National Research Council Studies in Geophysics, p. 84–95.
- Conrad, W.K., 1984, The mineralogy and petrology of compositionally zoned ash flow tuffs, and related silicic volcanic rocks, from the McDermitt caldera complex, Nevada Oregon: *Journal of Geophysical Research*, v. 89, p. 8639–8644.
- Craig, H., 1993, Yellowstone hotspot, a continental mantle plume [abs.]: *Eos (Transactions, American Geophysical Union)*, v. 74, p. 602.
- Craig, H., Lupton, J.E., Welhan, J.A., and Poreda, R., 1978, Helium isotope ratios in Yellowstone and Lassen Park volcanic gasses: *Geophysical Research Letters*, v. 5, p. 897–900.
- Dickinson, W.R., 1997, Tectonic implications of Cenozoic volcanism in coastal California: *Geological Society of America Bulletin*, v. 109, p. 936–954.
- Ekren, E.B., McIntyre, D.H., and Bennett, E.H., 1984, High temperature, large volume, lavalike ash flow tuffs without calderas in southwestern Idaho: U.S. Geological Survey Professional Paper 1272, 76 p.
- Fleck, R.J., and Criss, R.E., 1985, Strontium and oxygen isotopic variations in Mesozoic and Tertiary plutons of central Idaho: Contributions to Mineralogy and Petrology, v. 90, p. 291–308.
- Gripp, A.E., and Gordon, R.G., 1990, Current plate velocities related to hotspots incorporating the NUVEL 1 global plate motion model: *Geophysical Research Letters*, v. 17, p. 1109–1112.
- Hart, W.K., and Carlson, R.W., 1985, Distribution and geochronology of Steens Mountain-type basalts from the northwestern Great Basin: *Isotopes West*, no. 43, p. 5–10.
- Hildreth, W., 1981, Gradients in silicic magma chambers: Implications for lithospheric magmatism: *Journal of Geophysical Research*, v. 86, p. 10153–10192.
- Hildreth, W., Halliday, A.N., and Christiansen, R.L., 1991, Isotopic and chemical evidence concerning the genesis and contamination of basaltic and rhyolitic magma beneath the Yellowstone Plateau volcanic field: *Journal of Petrology*, v. 32, p. 63–138.
- Honjo, N., 1990, Geology and stratigraphy of the Mount Bennett Hills, and the origin of west central Snake River Plain rhyolites [Ph.D. thesis]: Houston, Texas, Rice University, 259 p.
- Honjo, N., Bonnichsen, B., Leeman, W.P., and Storer, J.C.J., 1992, Mineralogy and geothermometry of high temperature rhyolites from the central and western Snake River Plain: *Bulletin of Volcanology*, v. 54, p. 220–237.
- Humphreys, E.D., Dueker, K.G., Schutt, D.L., and Smith, R.B., 2000, Beneath Yellowstone: Evaluating plume and nonplume models using teleseismic images of the upper mantle: *GSA Today*, v. 10, p. 1–7.
- Izett, G.A., 1981, Volcanic ash beds: Recorders of upper Cenozoic silicic pyroclastic volcanism in the western United States: *Journal of Geophysical Research*, v. 86, p. 10200–10222.
- Izett, G.A., and Wilcox, R.A., 1982, Map showing localities and inferred distribution of the Huckleberry Ridge, Mesa Falls, and Lava Creek ash beds (Pearlette family ash beds) of Pliocene and Pleistocene age in the western United States and Canada: U.S. Geological Survey Miscellaneous Map 1-1325, 1 sheet, scale 1:4 000 000.
- Izett, G.A., Obradovich, J.D., and Melnert, H.H., 1988, The Bishop ash bed (middle Pleistocene) and some older (Pliocene and Pleistocene) chemically and mineralogically similar ash beds in California, Nevada, and Utah: U.S. Geological Survey Bulletin 1675, p. 1–37.
- Kellogg, K.S., and Marvin, R.F., 1988, New potassium argon ages, geochemistry, and tectonic setting of upper Cenozoic volcanic rocks near Blackfoot, Idaho: U.S. Geological Survey Bulletin 1806, 18 p.
- Kellogg, K.S., Harlan, S.S., Melnert, H.H., Snee, L.W., Pierce, K.L., Hackett, W.R., and Rodgers, D.W., 1994, Major 10.2 Ma rhyolitic volcanism in the eastern Snake River Plain, Idaho—Isotopic age and stratigraphic setting of the Arbon Valley Tuff Member of the Starlight Formation: U.S. Geological Survey Bulletin 2091, 18 p.
- Leeman, W.P., 1982, Development of the Snake River Plain Yellowstone Plateau province, Idaho and Wyoming: An overview and petrologic model, in Bonnichsen, B., and Breckenridge, R.M., eds., *Cenozoic geology of Idaho*: Idaho Bureau of Mines and Geology Bulletin 26, p. 155–157.
- Leeman, W.P., Oldow, J.J., and Hart, W.K., 1992, Lithosphere scale thrusting in the western U.S. Cordillera as constrained by Sr and Nd isotopic transitions in Neogene volcanic rocks: *Geology*, v. 20, p. 63–66.
- Manley, C.R., 1996, Physical volcanology of a voluminous rhyolite lava flow: The Badlands lava, Owyhee Plateau, southwestern Idaho: *Journal of Volcanology and Geothermal Research*, v. 71, p. 129–153.
- Morgan, W.J., 1972, Deep mantle convection plumes and plate motions: *American Association of Petroleum Geologists Bulletin*, v. 56, p. 203–213.
- Moutoux, T.E., 1995, Palynological and tephra correlations among deep wells in the modern Great Salt Lake, Utah, U.S.A.—Implications for Neogene through Pleistocene climatic reconstructions [Master's thesis]: Tucson, University of Arizona, 30 p.
- Nash, W.P., Perkins, M.E., Christensen, J.N., Lee, D. C., and Halliday, A.N., 1997, Isotopic tracking of the Yellowstone hotspot: Temporal variation in neodymium and hafnium: *Geological Society of America Abstracts with Programs*, v. 29, no. 6, p. A 70.
- Pedersen, T., Heeremans, M., and van der Beek, P., 1998, Models of crustal anatexis in volcanic rifts: Applications to southern Finland and the Oslo Graben: *Geophysical Journal International*, v. 132, p. 239–255.
- Perkins, M.E., 1998a, Miocene ash beds and Miocene mammals in the intermontane west and Great Plains, USA [abs.]: *Journal of Vertebrate Paleontology*, v. 18, p. 70A.
- Perkins, M.E., 1998b, Tephrochronologic and volcanologic studies of silicic fallout tuffs in Miocene basins of the northern Basin and Range province [Ph.D. thesis]: Salt Lake City, University of Utah, 207 p.
- Perkins, M.E., Diffendal, R.F., Jr., and Voorhies, M.R., 1995a, Tephrochronology of the Ash Hollow Formation (Ogallala Group)—Northern Great Plains [abs.]: *Geological Society of America Abstracts with Programs*, v. 27, no. 3, p. 79.
- Perkins, M.E., Nash, W.P., Brown, F.H., and Fleck, R.J., 1995b, Fallout tuffs of Trapper Creek Idaho—A record of Miocene explosive volcanism in the Snake River Plains volcanic province: *Geological Society of America Bulletin*, v. 107, p. 1484–1506.
- Perkins, M.E., Brown, F.H., Nash, W.P., McIntosh, W., and Williams, S.K., 1998, Sequence, age, and source of silicic fallout tuffs in middle to late Miocene basins of the northern Basin and Range province: *Geological Society of America Bulletin*, v. 110, p. 344–360.
- Pierce, K.L., and Morgan, L.A., 1992, The track of the Yellowstone hot spot: Volcanism, faulting, and uplift, in Link, P.K., Kuntz, M.A., and Platt, L.B., eds., *Regional geology of eastern and western Wyoming*: Geological Society of America Memoir 179, p. 1–53.
- Rodgers, D.W., Hackett, W.R., and Ore, H.T., 1990, Extension of the Yellowstone Plateau, eastern Snake River Plain, and Owyhee Plateau: *Geology*, v. 18, p. 1138–1141.
- Smith, R.B., and Braile, L.W., 1994, The Yellowstone hotspot: *Journal of Volcanology and Geothermal Research*, v. 61, p. 121–187.
- Smith, R.P., and Nash, W.P., 1976, Chemical correlation of volcanic ash deposits in the Salt Lake Group, Utah,

EXPLOSIVE SILICIC VOLCANISM OF THE YELLOWSTONE HOTSPOT

- Idaho and Nevada: *Journal of Sedimentary Petrology*, v. 46, p. 930–939.
- Smith, R.B., and Shar, M.L., 1974, Contemporary tectonics and seismicity of the western United States, with emphasis on the Intermountain seismic belt: *Geological Society of America Bulletin*, v. 85, p. 1205–1218.
- Smith, R.B., Shuey, R.T., Pelton, J.R., and Bailey, J.P., 1977, Yellowstone hot spot: contemporary tectonics and crustal properties from earthquakes and aeromagnetic data: *Journal of Geophysical Research*, v. 82, p. 3665–3676.
- Swisher, C.C., III, 1992, $^{40}\text{Ar}/^{39}\text{Ar}$ dating and its application to the calibration of the North American land mammal ages [Ph.D. thesis]: Berkeley, University of California, 239 p.
- Watson, E.B., and Harrison, T.M., 1983, Zircon saturation revisited: Temperature and composition effects in a variety of crustal magma types: *Earth and Planetary Science Letters*, v. 64, p. 295–304.
- Whalen, J.B., Currie, K.L., and Chappell, B.W., 1987, A type granites: Geochemical characteristics, discrimination and petrogenesis: *Contributions to Mineral and Petrology*, v. 95, p. 407–419.
- Williams, S.K., 1994, late Cenozoic tephrostratigraphy of deep sediment cores from the Bonneville basin, north-west Utah: *Geological Society of America Bulletin*, v. 106, p. 1517–1530.
- Zoback, M.L., McKee, E.H., Blakely, R.J., and Thompson, G.A., 1994, The northern Nevada rift: Regional tectono magmatic relations and middle Miocene stress direction: *Geological Society of America Bulletin*, v. 106, p. 371–382.

MANUSCRIPT RECEIVED BY THE SOCIETY AUGUST 31, 2000

REVISED MANUSCRIPT RECEIVED JUNE 27, 2001

MANUSCRIPT ACCEPTED SEPTEMBER 24, 2001

Printed in the USA

