

Lead isotopes by LA-MC-ICPMS: Tracking the emergence of mantle signatures in an evolving silicic magma system

J.I. Simon ^{a,*}, M.R. Reid ^b, E.D. Young ^{a,c}

^a Department of Earth and Space Sciences, University of California at Los Angeles, 595 Charles E. Young Drive East, 2676 Geology Building, Los Angeles, CA 90095, USA

^b Department of Geology, Northern Arizona University, Flagstaff, AZ 86011, USA

^c Institute of Geophysics and Planetary Physics, University of California at Los Angeles, 595 Charles E. Young Drive East, 2676 Geology Building, Los Angeles, CA 90095, USA

Received 5 July 2006; accepted in revised form 30 January 2007; available online 3 February 2007

Abstract

At Long Valley (LV) model Sr isotope phenocryst ages and absolute U–Pb zircon ages from precaldera Glass Mountain (GM) and caldera-related Bishop Tuff (BT) rhyolites show that these crystals track >1 Myr of evolution of a voluminous rhyolite magmatic system. In detail, strong disparities between the different age populations complicate ideas for a unified model for rhyolite generation, differentiation, and storage. To better elucidate the age discrepancies a new *in situ* Pb isotope technique has been developed to measure the compositions of 113 individual LV feldspars (mainly sanidine) and their host glasses by UV laser ablation MC-ICPMS. Given sufficient signal the accuracy and precision of this technique approaches that of double-spike thermal ionization mass spectrometry. The utility of our technique for many geologic materials is, however, limited to determining Pb isotope ratios that include ²⁰⁶Pb, ²⁰⁷Pb, and ²⁰⁸Pb, but exclude ²⁰⁴Pb. New zircon ²³⁸U–²⁰⁶Pb crystallization ages were also obtained for two older Glass Mountain domes.

A >1.5‰ difference between the Pb isotope compositions of feldspars from older (1.7–2.2 Ma) precaldera Glass Mountain (GM) rhyolites and younger LV rhyolites, including the BT, is found. The Pb isotope data for feldspars and their host glasses lie along a regional trend line between young basalts and evolved crust compositions, spanning ~15% of that isotopic difference, and show a secular change towards increasing mantle contribution. Most feldspars have Pb isotope compositions that are similar to their host glasses and, as such, there persists an apparent >100 k.y. difference between Sr model feldspar ages and zircon ages for some GM rhyolites. Collectively, the feldspars define a Sr–Pb isotope mixing curve. Evidence for mixing complicates the interpretation that the Sr isotope data solely reflect radiogenic ingrowth. Where isotopically heterogeneous feldspar populations occur, there is greater uncertainty about the veracity of the Sr model ages. Specifically, we find no Pb isotope evidence that BT feldspars grew from older GM-like magmas.

The distinct Pb isotope signatures for individual rhyolites and their feldspars support evidence based on zircon dating that LV volcanism did not erupt from a single long-lived magma chamber but rather tapped a number of different magmas. Moreover, contrary to the conventional model of gradual build-up prior to cataclysmic eruption, secular changes in the U–Pb age constraints on magma residence times and the magmas' distinct Pb isotopic compositions suggest that, at Long Valley, eruptive volumes increase with shorter magma residence time and correlate with greater mantle input. Evidently, the plumbing and therefore activity at Long Valley was influenced by the evolving interaction between source and crustal magma system.

© 2007 Elsevier Ltd. All rights reserved.

* Corresponding author. Present address: Department of Earth and Planetary Science, University of California at Berkeley, Center for Isotope Geochemistry and Berkeley Geochronology Center, 483 McCone Hall, Berkeley, CA 94720, USA. Fax: +1 510 642 9520.

E-mail address: simon@eps.berkeley.edu (J.I. Simon).

1. INTRODUCTION

Cataclysmic eruptions of silicic magma are among the most awe-inspiring natural phenomena found in the geologic record, in terms of size, power, and potential hazard. Caldera-related eruptions occur in hours to days, are hundreds to thousands of cubic kilometers in volume, and modify the landscape hundreds of kilometers from their source. One must ask then, how are the large thermal and mass requirements needed to generate large eruptible volumes of silicic magma obtained? Based on the repose intervals between eruptions of this magnitude, the magmas responsible for them could accumulate gradually in the shallow crust over timescales that may be in excess of 1 Myr (Smith, 1979; Spera and Crisp, 1981; Shaw, 1985). Is long-term melt storage in the shallow crust required or could voluminous silicic magmas reflect rapid melt influx from greater depths? Accurate isotopic tracer and geochronologic constraints can provide insights into both the source(s) and tempo of rhyolite magmagenesis.

At Long Valley, California, Sr model ages and Ar apparent ages well in excess of eruption ages have been used as evidence that a long-lived (>1 Myr) silicic magma chamber was responsible for the climatic eruption of the Bishop Tuff rhyolite (e.g., Christensen and DePaolo, 1993; van den Bogaard and Schirnick, 1995; Davies and Halliday, 1998). In contrast, U–Pb accessory phase dating (Reid and Coath, 2000; Simon and Reid, 2005) reveal crystal populations whose absolute ages imply accumulation of the voluminous Bishop Tuff magma in <200 k.y. and, if anything, imply a much more rapid accumulation rate than for the small volume eruptions that preceded it. The more gradual versus more punctuated magma accumulation rates required by the model and absolute ages, respectively, imply important differences in the mass and heat fluxes associated with the generation, differentiation, and storage of voluminous rhyolites and emphasize the need to reconcile the magmatic age differences.

Modest secular variations in Nd, Pb, and O isotopes have been documented for the Long Valley system (e.g., Halliday et al., 1989; Davies et al., 1994; Davies and Halliday, 1998; Bindeman and Valley, 2002). If initial isotopic heterogeneities also apply at the mineral scale, they could account for some of the discrepancy between the model and the absolute crystal ages. In order to track the overall Pb isotopic evolution of the magma system and its crystal “cargo”, we performed Pb isotope ratio measurements of glass and feldspars from rhyolites of the Long Valley magma center by ultraviolet (UV) laser ablation combined with multiple collector inductively coupled plasma-source mass spectrometry (MC-ICPMS). Our results for older (~1.7–2.2 Ma) and younger (0.9–1.2 Ma) precaldern Glass Mountain rhyolites and for the 0.76 Ma caldera-related Bishop Tuff reveal distinct if sometimes overlapping Pb isotope populations. Some feldspar–glass pairs and feldspar populations are more isotopically heterogeneous than can be explained by radiogenic ingrowth and provide evidence for variable amounts of crystal–melt exchange and/or melt modification prior to eruption. Collectively, the Pb isotope data obtained on feldspars and glasses comprise a linear trend between regional crustal (Cousens, 1996) and mantle (Heumann and

Davies, 1997) end-member signatures which, when coupled with Sr isotope data, is interpreted as a mixing array. Thus it cannot be assumed that minerals and host glasses initially had the same isotopic composition and model Sr isochron ages have to be evaluated on a case-by-case basis (e.g., Kalsbeek, 1992; Wendt, 1993; Davidson et al., 2005).

The distinct feldspar populations observed for individual Long Valley rhyolites support evidence from their zircon populations that the Glass Mountain and Bishop Tuff eruptions reflect a number of transient and distinct magma bodies (Hildreth, 2004; Simon and Reid, 2005). Sources for the rhyolites become increasingly mantle-dominated over time. When considered in conjunction with the decreasing absolute pre-eruption crystallization ages and increasing eruption volumes, these secular changes lead to a new picture for the Long Valley magma system, one in which mantle-dominated sources are increasingly tapped and crustal interaction and/or transfer times are reduced as the system evolves.

2. EXPERIMENTAL APPROACH

2.1. Mineral and obsidian samples from Glass Mountain and Bishop Tuff rhyolites

The glass, feldspar, and zircon separates studied here are mostly from the samples of Reid and Coath (2000), Schmitt and Simon (2004), and Simon and Reid (2005), and have been described therein. Two additional Bishop Tuff pumice samples have been included (Fig. 1).

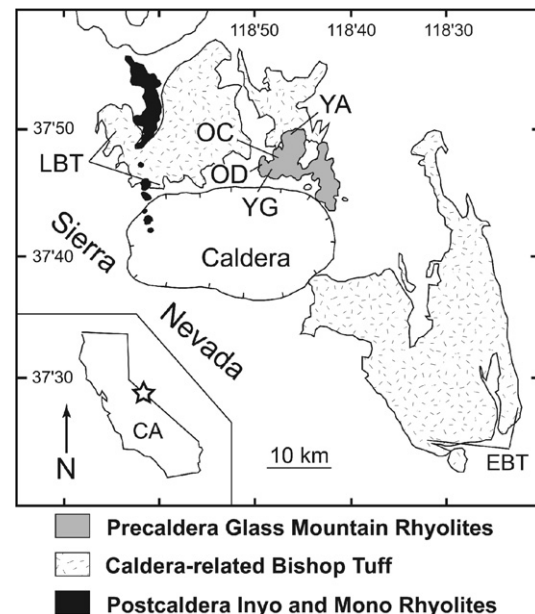


Fig. 1. Map of Long Valley showing sample locations, modified after Metz and Mahood (1985) and Wilson and Hildreth (1997). OC, OD, YG, and YA indicate the sample locations for the Glass Mountain rhyolites within their respective domes. EBT and LBT reflect the sample locations for the “early” and “late” Bishop Tuff samples (see text and references therein for details).

JS03LV20-16/17 is an “early” (EBT) erupted pumice from a fall deposit ~2 m below the distal portion of the Ig1Eb (Chidago) ignimbrite lobe (Wilson and Hildreth, 1997) from the same locality as Reid and Coath (2000). JS03LV18 is a “late” (LBT) erupted pumice from ~10 m above the base of the Ig2NWa (Mono) ignimbrite lobe where it has been exposed in a road cut near the Crestview highway maintenance station (locality 208 of Wilson and Hildreth, 1997). Feldspar and glass were cleaned by ultrasonication and rinsed with deionized water, but otherwise untreated. Lead isotope analyses come from euhedral to subhedral feldspar crystals that range in size between ~730 to 2750 μm . Conventional optical oil immersion methods were used to distinguish among albite, sanidine, and anorthoclase feldspars. The euhedral to subhedral zircons used for dating are typically ~100–200 μm in length.

2.2. Pb isotope measurements of feldspar and obsidian by laser ablation

Pb isotope measurements of plagioclase and sanidine feldspar crystals and their rhyolitic host glasses were obtained by ultraviolet laser ablation combined with multiple collector inductively coupled plasma-source mass spectrometry (MC-ICPMS). Target materials were ablated by a 213-nm UV laser built by NewWave Research™. The laser was operated at a fluence of 22–30 J/cm² with a pulse repetition rate of 4–6 Hz. Analyses were performed by scanning the laser across the sample. Prior to acquisition of data a series of broad (300 μm) weak laser pulses was used to “clean” the target surface. The laser scans produced ~25 μm deep, ≤ 800 μm long, and 100–150 μm wide tracks with width and length dimensions depending on the intensity of the Pb signal of each sample.

A sample-standard comparison approach was used for instrumental mass bias correction rather than Tl-doping (cf. Willigers et al., 2002; Mathez and Waight, 2003; Gagnevin et al., 2005). The measured Pb isotope ratios of the standards and rhyolitic materials were compared to those of a NBS981 standard solution. Signal intensities for the solution standards were matched to within ~10% of the solid standards. The intensity of the NIST glass standard NBS612 was ~1 V and most closely matched the intensities of the measured rhyolitic materials (sample signals ranged from ≥ 0.4 to ~1.8 V). A complete description of our laser ablation system design and analytical protocol has been included in the Appendix A.

2.3. U–Pb isotope analyses of zircon by ion microprobe

The CAMECA ims 1270 ion microprobe at UCLA was used to measure zircon ²³⁸U–²⁰⁶Pb crystallization ages, employing established methods (e.g., Quidelleur et al., 1997; Dalrymple et al., 1999). Individual zircon ages were ²⁰⁷Pb-corrected using a typical whole rock initial common ²⁰⁷Pb/²⁰⁶Pb composition of 0.818 (this study; modern surface Pb contamination is minimal). The ²⁰⁷Pb-corrected ages were also adjusted for initial U–Th disequilibrium (Schärer, 1984). See Simon and Reid (2005) for further details.

3. RESULTS

3.1. LA-MC-ICPMS measurement precision and accuracy

The precision and accuracy for our sample-standard comparison technique can be seen in Appendix A (Fig. 1A) and Table 1. Given sufficient Pb ion signals (²⁰⁸Pb ~2 V) measured values for NIST glass standards that represent a range of Pb isotope compositions can yield an accuracy and precision for Pb isotopes approaching that obtained with the double-spike TIMS (e.g., Woodhead and Hergt, 2001) and MC-ICPMS (Baker et al., 2004) methods. For many geologic materials the utility of our technique is limited to determining Pb isotope ratios comprised of the ²⁰⁶Pb, ²⁰⁷Pb, and ²⁰⁸Pb nuclides because of insufficient ²⁰⁴Pb ion yields. Our sample-standard comparison results are considerably better than LA-MC-ICPMS results obtained using Tl as a means for correcting for instrumental mass fractionation (Willigers et al., 2002; Mathez and Waight, 2003; Gagnevin et al., 2005) and for ²⁰⁷Pb/²⁰⁶Pb and ²⁰⁸Pb/²⁰⁶Pb ratios are a dramatic improvement over conventional TIMS analyses. For example, the external reproducibility of NIST glass NBS612 (~40 ppm Pb) is better than $\pm 0.2\text{‰}$ 2SD for linearized delta values $\delta^{207}\text{Pb}/^{206}\text{Pb}'$ and $\delta^{208}\text{Pb}/^{206}\text{Pb}'$.

3.2. Intra- and inter-sample Pb isotope heterogeneity among Long Valley rhyolites

Lead isotope data for 113 individual feldspar (sanidine, plagioclase, $\pm$ anorthoclase) crystals from three older Glass Mountain rhyolites, two younger Glass Mountain rhyolites, and four compositionally distinct Bishop Tuff pumice fragments are reported in Table 2. Most of the analyzed feldspars, with the salient exception of dome YA feldspars, are sanidine. Host glasses of the Glass Mountain samples were also measured. The data are expressed in the linear delta notation (δ') reported as a deviation from a representative local mafic composition (Cousens, 1996) where $\delta^x\text{Pb}/^{206}\text{Pb}' = 10^3 \ln((^x\text{Pb}/^{206}\text{Pb})_{\text{sample}}/(^x\text{Pb}/^{206}\text{Pb})_{\text{LVref}})$ and x represents either 207 or 208. On this scale the value for “LVref” (²⁰⁷Pb/²⁰⁶Pb = 0.8174 and ²⁰⁸Pb/²⁰⁶Pb = 2.0282) is 0.0‰. Glass measurements are age-corrected to initial ²⁰⁷Pb/²⁰⁶Pb and ²⁰⁸Pb/²⁰⁶Pb values based on the U and Th concentrations determined by ion microprobe (Schmitt and Simon, 2004) and by assuming Pb concentrations of 30 ($\pm 10\%$) ppm for younger Glass Mountain rhyolites and 45 ($\pm 10\%$) ppm for older Glass Mountain rhyolites (Table 2). Some confidence that our estimate of the Pb concentration is correct can be gained from the fact that beam intensities on the NIST612 standard (Pb content nominally 50 ppm, but more realistically ~40 ppm), are typically within ~25% of the glasses. No age-correction is applied to the feldspars. They contain relatively low U and Th concentrations and therefore corrections to both $\delta^{208}\text{Pb}/^{206}\text{Pb}'$ and $\delta^{207}\text{Pb}/^{206}\text{Pb}'$ values are only ~0.02‰ per Myr, or less.

The initial Pb isotope compositions of feldspars and glasses from the Long Valley rhyolites range from $\delta^{207}\text{Pb}/^{206}\text{Pb}'$ ~0.0‰ to >1.5‰ and from $\delta^{208}\text{Pb}/^{206}\text{Pb}'$ ~0.5‰ to 2.5‰ (Fig. 2). In general the isotopic compositions of the rhyolitic host glasses mirror the differences observed in the feldspar

Table 1

Primary and secondary Pb isotope standards^a

	²⁰⁶ Pb/ ²⁰⁴ Pb	2σ _m	²⁰⁷ Pb/ ²⁰⁴ Pb	2σ _m	²⁰⁸ Pb/ ²⁰⁴ Pb	2σ _m	²⁰⁷ Pb/ ²⁰⁶ Pb	2σ _m	²⁰⁸ Pb/ ²⁰⁶ Pb	2σ _m	δ ²⁰⁶ Pb/ ²⁰⁴ Pb'	2σ _m	δ ²⁰⁷ Pb/ ²⁰⁴ Pb'	2σ _m	δ ²⁰⁸ Pb/ ²⁰⁴ Pb'	2σ _m	δ ²⁰⁷ Pb/ ²⁰⁶ Pb'	2σ _m	δ ²⁰⁸ Pb/ ²⁰⁶ Pb'	2σ _m
<i>Glass 610</i>																				
LA-MC-ICPMS ^b	17.0497	0.0006	15.5096	0.0006	36.9762	0.0013	0.90967	0.00001	2.16875	0.00002	6.40	0.04	0.77	0.04	6.78	0.04	−5.63	0.01	0.39	0.01
w/Tl corr. ^b	17.0172	0.0005	15.4739	0.0005	36.8687	0.0014	0.90931	0.00001	2.16655	0.00003	4.50	0.03	−1.53	0.03	3.87	0.04	−6.02	0.01	−0.62	0.01
w/Tl corr. ^c	17.038	0.012	15.492	0.011	36.919	0.031	0.9093	0.0001	2.1670	0.0005	5.72	0.70	−0.36	0.71	5.25	0.84	−6.02	0.11	−0.42	0.23
w/Tl corr. ^d	17.041	0.011	15.498	0.011	36.924	0.032	0.90945	<i>0.0009</i>	2.16677	<i>0.0023</i>	5.89	0.65	0.03	0.71	5.39	0.87	−5.85	<i>0.94</i>	−0.52	<i>1.08</i>
LA-ICPMS ^e							0.9115	0.0048									−3.61	5.25		
TIMS ^f	17.049	0.012	15.506	0.01	36.989	0.024	0.9095	<i>0.0009</i>	2.1696	<i>0.0021</i>	6.36	0.70	0.54	0.64	7.16	0.65	−5.80	<i>0.94</i>	0.78	<i>0.96</i>
DS-TIMS ^g	17.0509	0.0018	15.5156	0.0010	36.9756	0.0026	0.9097	<i>0.0001</i>	2.1692	<i>0.0003</i>	6.47	0.11	1.16	0.06	6.79	0.07	−5.54	<i>0.11</i>	0.58	<i>0.13</i>
TIMS ^h	17.0640	0.0070					0.9098	0.0002	2.1710	0.001	7.24	0.41					−5.47	0.20	1.43	0.46
DS-MC-ICPMS ^j	17.0513	0.0009	15.5126	0.0009	36.9867	0.0022	0.90976	0.00002	2.16924	0.00004	6.50	0.05	0.97	0.06	7.18	0.06	−5.53	0.02	0.73	0.02
<i>Glass 612</i>																				
LA-MC-ICPMS ⁱ	17.091	0.002	15.505	0.002	36.992	0.005	0.90724	0.00002	2.16450	0.00003	8.85	0.14	0.49	0.13	7.20	0.14	−8.30	0.03	−1.57	0.01
DS-TIMS ^g	17.098	0.003	15.517	0.004	37.001	0.009	0.9073	<i>0.0003</i>	2.1647	<i>0.0006</i>	9.23	0.15	1.23	0.23	7.47	0.25	−8.21	<i>0.24</i>	−1.50	<i>0.29</i>
DS-MC-ICPMS ^j	17.098	0.001	15.514	0.001	37.016	0.003	0.90735	0.00002	2.16494	0.00004	9.25	0.06	1.03	0.06	7.97	0.08	−8.18	0.02	−1.25	0.02
<i>Glass 614</i>																				
LA-MC-ICPMS ^a							0.8699	0.0001	2.0992	0.0002							−50.32	0.11	−32.18	0.09
DS-TIMS ^g							0.8710	<i>0.0008</i>	2.1014	<i>0.0020</i>							−47.90	<i>0.32</i>	−30.66	<i>0.46</i>
DS-MC-ICPMS ^j							0.8709	0.0002	2.1015	0.0003							−49.17	0.00	−30.97	0.14
<i>NBS981</i>																				
MC-ICPMS ^k	16.9414	0.0013	15.4976	0.0011	36.7261	0.0028	0.91479	0.00001	2.16788	0.00002	0.03	0.08	0.00	0.07	0.00	0.08	−0.02	0.01	−0.01	0.01
DS-MC-ICPMS ^j	16.9402	0.0002	15.4952	0.0002	36.7176	0.0004	0.914703	0.000004	2.16748	0.00001	−0.04	0.01	−0.15	0.01	−0.12	0.01	−0.11	0.004	−0.08	0.01

Italicized font, estimated uncertainties calculated from reported ²⁰⁶Pb/²⁰⁴Pb, ²⁰⁷Pb/²⁰⁴Pb, and ²⁰⁸Pb/²⁰⁴Pb errors.

Individual glass analyses reported in Appendix B.

^a Linearized delta values (δ') = 1000 * ln (R_{spl}/R_{NBS981}). NBS981 values from Eisele et al. (2003).^b This study (within session reproducibility, *n* = 10). Measured ²⁰⁵Tl/²⁰³Tl value (=2.42409 ± 0.00003) for mass bias correction was explored, but not used (see text).^c Mathez and Waight (2003).^d Willigers et al. (2002).^e Hirata and Nesbitt (1995).^f Walder et al. (1993).^g Woodhead and Hergt (2001), *n* = 4, normalized to Eisele et al. (2003).^h Machado and Gauthier (1996).ⁱ This study (multiple session reproducibility, *n* = 14). ²⁰⁸Pb intensity for 612 glass was 0.83–1.3 V.^j Baker et al. (2004) (610, *n* = 5; 612, *n* = 5, 614, *n* = 5; NBS981, *n* = 119; normalized to Eisele et al. (2003).^k This study (zero enrichments from all sessions, *n* = 33). Individual analyses reported in Appendix C.

Table 2
Laser ablation Pb isotope measurements of feldspars and glasses

Rhyolite ^a	Age ^b ka	Material	<i>n</i>	²⁰⁸ Pb intensity (V)	$\delta^{207}\text{Pb}/^{206}\text{Pb}'_{\text{LVref}}$ measured	2 σ_m	$\delta^{207}\text{Pb}/^{206}\text{Pb}'_{\text{LVref}}$ age-corrected ^c	2 σ_m	$\delta^{208}\text{Pb}/^{206}\text{Pb}'_{\text{LVref}}$ measured	2 σ_m	$\delta^{208}\text{Pb}/^{206}\text{Pb}'_{\text{LVref}}$ age-corrected ^c	2 σ_m
<i>Bishop Tuff</i>												
Early (MR99LV58)	760	Sanidine wt. mean	11	0.71	0.12	0.04	—	—	0.85	0.03	—	—
		Plagioclase wt. mean	4	0.45	0.12	0.06	—	—	0.97	0.08	—	—
Early (JS03LV20-16/17)	760	Sanidine wt. mean	6	0.77	0.04	0.05	—	—	0.81	0.04	—	—
		Plagioclase	1	0.54	−0.06	0.18	—	—	0.73	0.19	—	—
Late (MR99LV51)	760	Sanidine wt. mean	9	0.97	0.17	0.03	—	—	0.77	0.02	—	—
Late (JS03LV18)	760	Sanidine wt. mean	10	0.73	0.12	0.03	—	—	0.69	0.03	—	—
<i>Younger Glass Mountain rhyolites</i>												
Dome YA (JS01LV04)	870	Plagioclase wt. mean	9	0.53	0.34	0.05	—	—	1.20	0.04	—	—
		Sanidine	1	0.46	0.34	0.19	—	—	1.07	0.15	—	—
		Glass	1	0.88	0.01	0.11	0.10	0.11	1.05	0.09	1.09	0.09
Dome YG (MR00LV62)	900	Sanidine wt. mean	12	0.93	0.20	0.03	—	—	0.97	0.03	—	—
		Glass wt. mean	3	0.77	0.25	0.04	0.32	0.04	1.12	0.05	1.16	0.06
<i>Older Glass Mountain rhyolites</i>												
Dome OD (MR00LV63)	1686	Sanidine wt. mean	10	1.49	1.02	0.03	—	—	1.89	0.02	—	—
		Plagioclase wt. mean	3	0.82	1.37	0.14	—	—	2.43	0.15	—	—
		Glass	1	1.22	0.83	0.09	1.06	0.11	1.94	0.12	2.09	0.13
Dome OL (JS01LV05)	1867	Sanidine wt. mean	8	1.36	1.38	0.02	—	—	2.44	0.02	—	—
		Plagioclase wt. mean	2	0.77	1.40	0.37	—	—	2.50	0.34	—	—
		Glass	1	1.27	1.26	0.06	1.43	0.07	2.52	0.08	2.65	0.09
Dome OC-I (MR00LV60)	1990	Sanidine wt. mean	10	1.19	1.25	0.03	—	—	2.27	0.02	—	—
		Glass	1	1.03	0.58	0.23	0.99	0.26	1.88	0.22	2.17	0.24
Dome OC-II (MR00LV61)	1990	Sanidine wt. mean	12	1.31	1.24	0.02	—	—	2.28	0.02	—	—
		Glass	1	1.21	0.95	0.07	1.22	0.10	2.23	0.06	2.41	0.09

Results are expressed in the linearized delta notation $\delta^x\text{Pb}/^x\text{Pb}' = 10^3 \ln (^x\text{Pb}/^x\text{Pb}_{\text{spl}} / ^x\text{Pb}/^x\text{Pb}_{\text{LVref}})$. LVref is a lead isotope composition representative of a primitive Long Valley reservoir based on mafic lavas from [Cousens, 1996](#) (e.g., $^{207}\text{Pb}/^{206}\text{Pb} = 0.817945$; $^{208}\text{Pb}/^{206}\text{Pb} = 2.02869$).

^a Nomenclature of “Early” and “Late” is used for reference, but also indicates pre-eruption temperature differences ([Hildreth, 1977](#)); Glass Mountain rhyolite domes after [Metz \(1987\)](#).

^b Ar/Ar ages of sanidine for Bishop Tuff from [van den Bogaard and Schirnick \(1995\)](#); Ar/Ar and K–Ar ages of host glass for younger Glass Mountain rhyolites from [Renne et al. \(2006\)](#) and [Metz and Mahood \(1985\)](#), respectively; Ar/Ar ages of sanidine for older Glass Mountain rhyolites from [Davies et al. \(1994\)](#).

^c Corrected values defined by Ar eruption ages; SIMS data for U and Th from [Schmitt and Simon \(2004\)](#); assumes nominal Pb glass abundances of 30 ± 5 ppm for younger Glass Mountain and 45 ± 5 ppm for older Glass Mountain, respectively (see text).

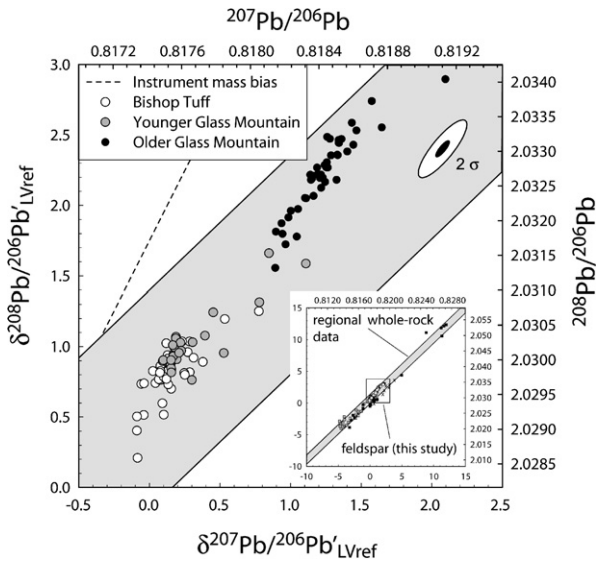


Fig. 2. Three isotope representation of Pb measurements of Long Valley feldspars. Distinct Pb isotope compositions were found among Glass Mountain and Bishop Tuff rhyolites. Inset shows that studied feldspar lie along a regional Pb isotopic trend that includes high and low $\delta^{208}\text{Pb}/\delta^{206}\text{Pb}'$ evolved end-member (e.g., old crustal granites, Cousens, 1996) and $\delta^{208}\text{Pb}/\delta^{206}\text{Pb}'$ believed to reflect mantle-like isotopic compositions (i.e., young basalts and the resurgent dome, Heumann and Davies, 1997) (see text). Older GM rhyolites trend towards higher $\delta^{208}\text{Pb}/\delta^{206}\text{Pb}'$ compositions, whereas younger GM rhyolites yield intermediate Pb compositions that overlap with the even lower Pb compositions measured in the BT feldspar population. The 2σ error ellipses show internal precision (filled) and external reproducibility (open) for NBS612 glass, respectively.

populations that they host. Some of the data spread along an instrument mass fractionation line that can be ascribed to residual inter-element matrix effects (Fig. 2). This spread is within our reported uncertainty. The principal component of the data taken as a whole lies along a trend defined by the Pb isotopic composition of whole rock samples from the Long Valley region (Fig. 2, inset) and is clearly distinct from the theoretical mass fractionation line. This and the good match to existing Pb isotope data for rhyolite whole rock values (Halliday et al., 1989; Davies et al., 1994) support the veracity of our measurements.

There is a broad increase in the $\delta^{208}\text{Pb}/\delta^{206}\text{Pb}'$ of the feldspars and glasses with increasing eruption age, with feldspars and glasses from older Glass Mountain rhyolites (~1.7–2.2 Ma eruption ages) having the highest $\delta^{208}\text{Pb}/\delta^{206}\text{Pb}'$ compositions. A similar difference between the isotopic compositions of older Glass Mountain rhyolites compared to the younger ones is also suggested by whole rock Pb isotope data for many of the same rhyolites (Halliday et al., 1989) but the scatter in those data is much greater than the scatter in the results reported here. The isotopic compositions of younger Glass Mountain (0.9–1.2 Ma eruption ages) feldspars and glasses overlap considerably those of the Bishop Tuff feldspars (0.76 Ma) but a few approach values obtained for older Glass Mountain feldspars. This $\geq 1.1\text{‰}$ difference in $\delta^{207}\text{Pb}/\delta^{206}\text{Pb}'$ of the feldspars is significantly greater than can be explained by radiogenic ingrowth in the time interval

between eruption of the oldest Glass Mountain rhyolite to eruption of the Bishop Tuff.

The scatter in sanidine Pb isotopic compositions for individual rhyolites is generally within our analytical reproducibility ($\pm 0.2\text{‰}$ 2SD). Intra-rhyolite heterogeneity is observed in domes OD and YA (see Fig. 3). Dome OD exhibits isotopic differences between and heterogeneity within its sanidine and plagioclase populations, while dome YA exhibits heterogeneity within its plagioclase population. Less significant intra-sample heterogeneity may exist in a second younger Glass Mountain rhyolite, dome YG, and parts of the caldera-related Bishop Tuff (see Fig. 3). These heterogeneities require contributions from isotopically distinct sources.

3.3. U–Pb zircon age constraints on crystallization of the older Glass Mountain rhyolites

It is important to know the affinity of the feldspar crystals to each other and their host magmas in order to interpret the Pb isotope heterogeneity they reveal. Although we cannot date the feldspars themselves, we can use the absolute crystallization ages of the zircon to provide insights into the relative proportions of xenocrystic and cognate crystals. The zircon ages reported here for precaldera Glass Mountain rhyolite domes OC and OL (Metz and Mahood, 1985) extend the results for zircon age populations determined previously for other Long Valley rhyolites (Reid and Coath, 2000; Simon and Reid, 2005). The weighted mean zircon crystallization ages and associated values for the mean square of the weighted deviates (MSWD) (York, 1969; Mahon, 1996) of the dome OC and OL zircon populations are summarized in Table 3; individual ages, Th/U ratios, and U concentrations for zircons analyzed in this study can be found in the Appendices E and F. Because the MSWD of small datasets that are normally distributed around a single mean value may deviate from the expected value of unity, we also provide the MSWD values at the upper 95%-confidence interval on a single age distribution as defined by the degrees of freedom for each rhyolite. Both of the zircon populations yield MSWD values that exceed that expected for single ages and, as such, crystallization of each zircon population appears to have been protracted (≥ 50 k.y.). Estimates for model pre-eruption ages and for the duration of crystallization are also given in Table 3, following the approach of Simon and Reid (2005). The uncertainties for weighted mean, end-member, and individual zircon ages are reported as 1 standard error (σ_m) throughout the text, unless stated otherwise.

Nearly all zircons from domes OC and OL have U concentrations (>4000 ppm) like those that characterize zircon from another older Glass Mountain rhyolite, dome OD, and distinctly higher than those of zircon from the younger Glass Mountain and Bishop Tuff rhyolites (Simon and Reid, 2005). Based on the maximum and minimum ages calculated from the age distributions, zircon crystallization in domes OC and OL likely occurred sometime between 2431 ± 36 and 1991 ± 27 ka (Table 3). The mean pre-eruption age of the zircons is ~ 310 k.y. older than the reported 1990 ± 6 ka Ar/Ar eruption age for dome OC (Davies et al., 1994). Alternatively, if the minimum zircon age of 2137 ± 49 ka for dome OC better estimates the time of

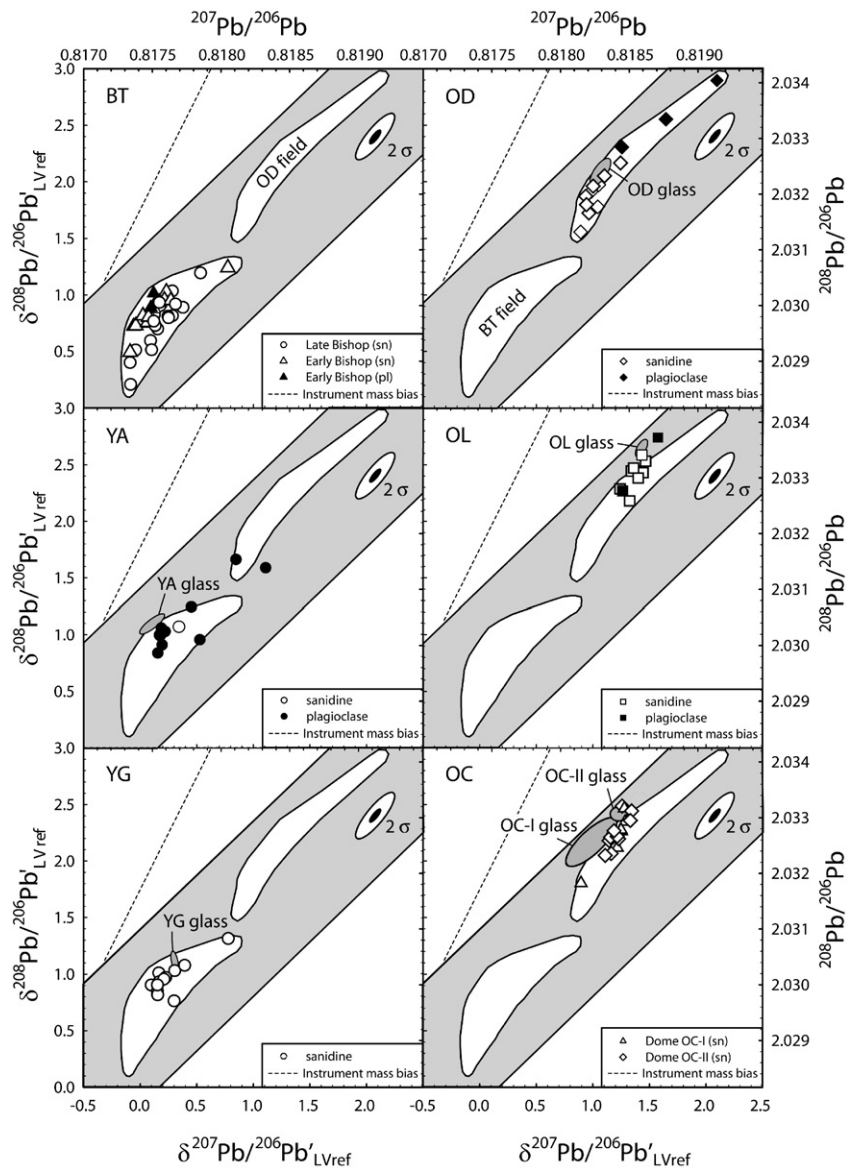


Fig. 3. Three isotope representation of Pb isotope variations within individual rhyolites with columns organized from youngest (upper left) to oldest (lower right) eruption ages. Fields for feldspar from dome OD (upper right field) and Bishop Tuff (lower left field) are shown in each panel for reference. $2\sigma_m$ error ellipses for Glass Mountain host glasses are shown for reference. The orientation of dome YG glass ellipse reflects the combined uncertainty of three replicate analyses. Unlabeled error ellipse in each panel comes from repeated analyses of NBS612 glass and shows representative 2σ internal precision (filled) and external reproducibility (open), respectively, for individual analyses.

eruption, the mean pre-eruption age is ~ 165 k.y. older than eruption. Zircon from dome OL likely crystallized no more than 350 k.y. before the reported Ar/Ar eruption age of 1866 ± 7 ka (Davies et al., 1994). The mean pre-eruption age predates that eruption age by 285 k.y. and the minimum zircon age estimate by ~ 160 k.y.

4. DISCUSSION

4.1. Lead isotope evolution of magmas in the Long Valley region

Hildreth (2004) suggests that silicic magmatism in Long Valley is localized by the concentrations of basaltic magmas

generated in response to extensional unloading and decompression melting of the upper mantle. Mantle-derived contributions to large silicic magma centers have been reported for a number of localities in western North America, including Long Valley (e.g., Hildreth, 1981; Lipman, 1988; Johnson, 1991). Pb isotope data for mafic to intermediate volcanic rocks regionally associated with Long Valley caldera (Ormrod, 1988; Nielsen et al., 1991; Cousens, 1996; Reid and Ramos, 1996) are initially similar to, but then trend along a linear array to lower values than the relatively high $\delta^{x}\text{Pb}/^{206}\text{Pb}'$ of local Mesozoic granitoids (Fig. 4). The earliest lavas (~ 3.6 to 2.2 Ma) are trachytic and reflect significant amounts of assimilated Sierran granitoids (up to 30%, Cousens, 1996). Basalts that erupted ~ 1.0 to 0.8 Ma are

Table 3
Summary of zircon crystallization ages of precaldera Glass Mountain Domes OC and OL^a

Sample	Mean age (ka)	Analyses (<i>n</i>)	Grains	MSWD ^b	Eruption age ^c	Mean pre-eruption age (ka) ^d (<i>n</i> , MSWD)	Limits on age range (ka) ^e		Sample location
							Min (<i>n</i> , MSWD)	Max (<i>n</i> , MSWD)	
<i>Glass Mountain Rhyolites</i>									
Dome OC-I (MR00LV60)									
Interiors	2232 ± 29	17	10	5.6 (1.8)	1990 ± 6	2303 ± 24 (<i>n</i> = 12, 2.0)	2137 ± 49 (<i>n</i> = 5, 4.1) ^f	2431 ± 36 (<i>n</i> = 4, 0.15)	37°47.756' lat. 118°47.491' long.
Dome OL (JS01LV05) ^g									
Interiors	2099 ± 33	11	11	4.0 (2.1)	1866 ± 7	2151 ± 34 (<i>n</i> = 7, 2.4)	1991 ± 27 (<i>n</i> = 4, 0.19)	2212 ± 25 (<i>n</i> = 4, 0.85)	37°45.796' lat. 118°41.187' long.
Near surface	2081 ± 55	7	3	4.4 (2.6)					

Individual zircon dates are included in Appendices E and F.

^a All uncertainties are reported as $1\sigma_m$.

^b Mean square of the weighted deviates (MSWD) for the zircon age population. MSWD value expected for a single age population at the upper 95%-conf. interval (Mahon, 1996) is provided in parentheses for comparison.

^c Ar/Ar eruption ages (ka).

^d Model mean pre-eruption age obtained by sequential omission of all young ages until the MSWD value for the remaining population is consistent with a single age (i.e., just within the upper 95%-conf. interval for the relevant number of analyses).

^e Minimum and maximum end-member ages obtained by pooling the youngest and oldest ages, respectively, until their associated MSWD values are high enough to be consistent with a single age (i.e., as defined by the lower 95%-conf. interval for the relevant number of analyses).

^f If analysis GMII-r5g3sl-pol is excluded the calculated minimum age for the next four youngest measurements of zircon in LV60 is 2173 ± 20 ka (MSWD = 0.27).

^g Xenocryst GMVI-r1g15, possibly from “older GM-like” precursor, that has a surface age of 2348 ± 45 and mean interior age of 2603 ± 60 ka (MSWD = 1.1) is excluded from model mean pre-eruption age calculation.

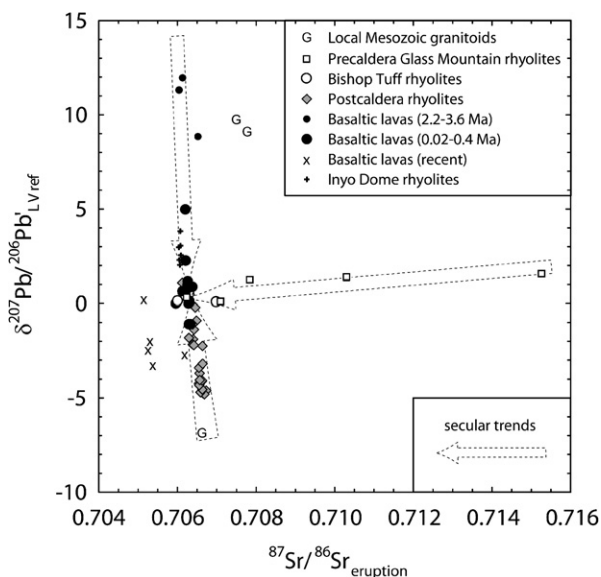


Fig. 4. Secular eruption age-corrected Sr and Pb isotope trends within mafic and rhyolitic volcanic rocks from the Long Valley region, California (Sr data of Halliday et al., 1989; Davies and Halliday, 1998; Heumann and Davies, 1997; Heumann et al., 2002; Simon and Reid, 2005. Pb data of Chen and Tilton, 1991; Cousens, 1996; Halliday et al., 1989; Heumann and Davies, 1997; Heumann et al., 2002; Nielsen et al., 1991; Ormerod, 1988; Reid and Ramos, 1996; and this study). The intersection of trends defined by the mafic lavas and the pre- and postcaldera rhyolites support the idea of a common source component (“LVref”, see text). Black symbols, indicate mafic lavas; white, precaldera and Bishop Tuff rhyolites, and grey, postcaldera rhyolites; individual groups defined in the figure).

chemically and isotopically similar to other lavas in the western Basin and Range Province and thought to have an enriched lithospheric mantle source. Younger lavas (<0.5 Ma) may represent increasing interaction with mafic lower crust or a second less enriched mantle source (Cousens, 1996). The observed trend in the mafic to intermediate volcanic rocks may thus represent rapid damping of a local Mesozoic crustal signature on magmatism; at the same time, this isotopic heterogeneity in the mafic-intermediate rocks precludes identification of a unique mantle end-member for the Long Valley rhyolites.

The Pb isotope data for the rhyolite glasses reported in this study lie along the same trend as other volcanic rocks as well as Mesozoic granitoids from the Long Valley region (see inset Fig. 2). The $\sim 2.0\%$ range in $\delta^{207}\text{Pb}/^{206}\text{Pb}'$ of the rhyolites spans only $\sim 15\%$ of the isotopic variation exhibited by the mafic to intermediate volcanic rocks. The trend towards decreasing Pb isotope compositions for the precaldera and caldera-related rhyolites broadly mimics that of the more mafic magmas and could be explained solely by derivation from heterogeneous mantle sources.

When the secular decrease in $\delta^{207}\text{Pb}/^{206}\text{Pb}'$ towards the predominant Pb isotope signatures of the basalts is coupled with evidence for a secular increase in ϵ_{Nd} (Davies et al., 1994; Davies and Halliday, 1998; Halliday et al., 1989) and decreases in $\delta^{18}\text{O}$ (Bindeman and Valley, 2002), it seems likely that the precaldera and caldera rhyolites at Long Valley reflect a tem-

poral trend towards proportionately greater mantle contributions to the melt. A general composition for this source component may be identified from the convergence of Pb and Sr isotopic trends defined by the Long Valley mafic lavas with those of the precaldera and postcaldera rhyolites, shown in Fig. 4. Either the immediate source areas for precaldera and caldera-related Long Valley magmatism (e.g., lower crust) becomes progressively mantle-dominated or mantle-derived magmas assimilate lesser amounts of older crustal material during their ascent. A relatively refractory crust might be the expected result after repeated interactions of magma during passage through the crust but, given the geographical changes in magmatic loci associated with Long Valley volcanism, a more likely explanation for these changes would seem to be related to an increasing flux of mantle-derived magma.

Lead isotope data for the postcaldera rhyolites overlaps and considerably extends the range observed in the precaldera and caldera-forming rhyolites. Heumann and Davies (1997) attribute a secular variation in the Pb isotope compositions of these rhyolites, towards the predominant Pb isotope composition of the basalts, to the addition of more primitive magma in that system. Collectively these data suggest that the precaldera and postcaldera rhyolites evolved towards an end-member common to both systems and require both high and low $\delta^{207}\text{Pb}/^{206}\text{Pb}'$ crustal end-members. Possible crustal end-members for the Pb isotope compositions exist among eastern Sierran granitoids (Chen and Tilton, 1991; Cousens, 1996), as shown in Fig. 4.

4.2. Implications of open system processes recorded in Glass Mountain and Bishop Tuff rhyolites

Eruption-age Sr isotope signatures of Glass Mountain and Bishop Tuff whole rocks, glasses, and feldspars are heterogeneous ($^{87}\text{Sr}/^{86}\text{Sr} \sim 0.706$ to >0.714). In an important series of papers, Halliday et al. (1989), Christensen and DePaolo (1993), Davies et al. (1994), Christensen and Halliday (1996), and Davies and Halliday (1998) recognized that *in situ* radiogenic decay within the lifespan of the Long Valley magmatic system can wholly account for the Sr isotope heterogeneity. Pseudo-isochrons defined by eruption-age Sr isotope characteristics of Bishop Tuff minerals and glasses (Christensen and Halliday, 1996) along with isotopic affinities between Bishop Tuff crystals and Glass Mountain rhyolites (Davies and Halliday, 1998) could indicate differentiation of Bishop Tuff parental magmas ~ 300 k.y. to up to ~ 1300 k.y. before eruption; the minerals' radiogenic isotope signatures could be produced by crystallization in the episodes between magma differentiation and eruption. Alternatively, the initial Sr isotope heterogeneity between the glasses and crystals could signal the presence of “foreign” crystals as documented elsewhere (e.g., Davidson and Tepley, 1997; Ramos and Reid, 2005).

The results presented here are the first to convincingly show internal heterogeneity in not only Sr, but Pb isotope characteristics of Long Valley feldspar populations as well (Fig. 3). Unlike the Sr isotope variations, the Pb isotope differences cannot be explained by aging (cf. Wolff et al., 1999 and Wolff and Ramos, 2003). At least some of the feldspars

could be xenocrysts, or so-called “antecrysts” (e.g., Charlier et al., 2004)—crystals related to the system but not to the magma in which they are hosted.

The zircon age populations presented here for the Glass Mountain domes OC and OL show no evidence for inheritance of zircons older than the Long Valley magmatic system, in keeping with previous results for other Glass Mountain rhyolites (Simon and Reid, 2005). The observation that zircon crystallization ages range to hundreds of k.y. before eruption generally supports the inference that at least some of the eruption-aged Sr isotope heterogeneity could reflect the onset of differentiation and crystallization well before eruption. For dome OD (Simon and Reid, 2005), there is a good match between the feldspar Sr model ages and the zircon ages, both of which are hundreds of k.y. older than the eruption age. For domes OC and OL, this is not the case (Table 3). The near eruption ~ 2035 to 1968 ka feldspar Sr model ages reported for dome OC (Davies et al., 1994) are significantly younger than the 2431 ± 36 to 2120 ± 65 ka zircon ages for dome OC. Similarly, the ~ 1908 to 1877 ka feldspar Sr model ages for dome OL are younger than its 2212 ± 25 to 1991 ± 27 ka zircon ages. Although the onset of zircon and feldspar crystallization could occur at different times in a silicic magma, when coupled with the Pb isotopic heterogeneity of the feldspars, these age differences could be the result of open system processes (e.g., mixing/contamination, crystal–melt exchange, etc.).

4.2.1. Open system processes and older Glass Mountain magma chronology

Evidence from Pb isotopes for open system processes and from the zircon crystallization ages for pre-eruption crystal residence begs the question of whether an exotic contaminant or radiogenic ingrowth more strongly influenced the eruption-age Sr isotope characteristics of the older Glass Mountain rhyolites. It is particularly notable, therefore, that the sanidines have Pb and Sr isotope compositions which when paired together define a possible mixing curve (Fig. 5). Even at magmatic temperatures Pb and Sr diffusion in feldspar is relatively slow (Cherniak and Watson, 1994; Giletti and Casserly, 1994). Thus the relative homogeneity of the sanidine Pb isotope signatures in each sample implies that, if magma mixing and/or contamination were responsible for the apparent mixing trend, the different mixtures must have been produced prior to most sanidine growth.

Model curves with $[Pb]/[Sr]_2:[Pb]/[Sr]_1 \sim 20$ defined by the two end-members fit the sanidine data. End-member 1 is isotopically like the Bishop Tuff rhyolite as well as many mafic magmas erupted from Long Valley (“LVref” value). The higher eruption-aged $^{87}Sr/^{86}Sr$ of end-member 2 is in concert with our inference from Nd, O, and Pb isotopes that the older Glass Mountain rhyolites contain a greater fraction of crustal input (Section 4.1). That end-member, if it were a melt or silicic rock, would have to be much more compositionally evolved given its significantly higher Pb/Sr (Pb is more incompatible than Sr during rhyolite differentiation). It could be youthful intrusions with Glass Mountain-like compositions (i.e., highly fractionated, high Pb/Sr and Rb/Sr) that were affected by crustal contamination and, because they predate Glass Mountain, radiogenic ingrowth. Alterna-

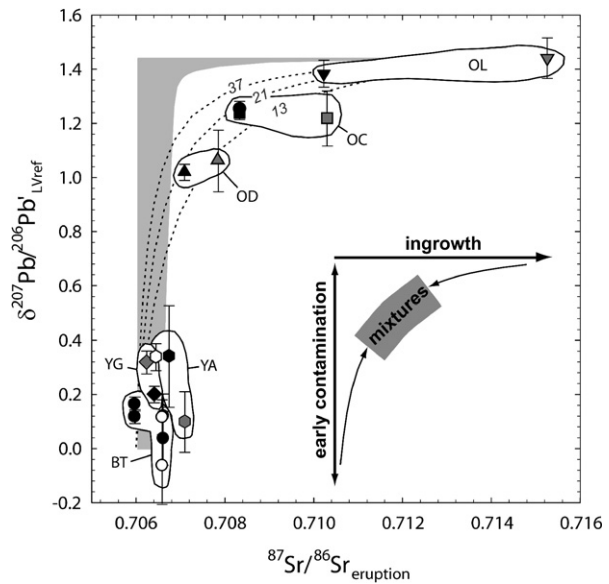


Fig. 5. Eruption age-corrected Sr versus Pb isotope diagram of weighted mean Pb and the bulk Sr isotope compositions of feldspars and their associated host glasses (Sr data of Halliday et al., 1989; Davies and Halliday, 1998). Black symbols indicate sanidine; white, plagioclase; and grey, glass. The plagioclase from OD and OL domes are heterogeneous, lie above the mixing trend, and host glass from OC-II are omitted for clarity. Individual rhyolites are distinguished by: pentagon, dome YA; diamond, dome YG; up triangle, dome OD; down triangle, dome OL; and circle, dome OC-I. Error bars are $2\sigma_m$. Curves are hypothetical two-component mixing trends between a radiogenic evolved end-member (e.g., older Glass Mountain-like) and an unradiogenic low-silica rhyolite end-member (i.e., with the isotopic composition of the predominant mafic lavas in Long Valley, from Cousens, 1996). Numbers superimposed on mixing hyperbola are the $[Pb]/[Sr]_2:[Pb]/[Sr]_1$ values used to calculate each curve, respectively. Shaded region delineates compositional space obtainable by a process whereby mixing occurs between distinct Pb isotopic sources prior to Sr ingrowth. The $[Rb]/[Sr]$ ratio used in this calculation spans three orders of magnitude (~ 8 to 600) as would be expected for the variably differentiated magmas. The concave right side of the shaded region represents the maximum $^{87}Sr/^{86}Sr$ values possible given realistic Rb and Sr concentrations and 300 k.y. of ^{87}Rb decay. Cartoon illustrates possible isotopic evolution pathways (e.g., early source mixing/contamination and mixing recorded by feldspar). In some rhyolites evidence for later mixing producing isotopic disequilibrium between feldspar and their host glasses also exists (see text).

tively, it could be mineral phases in the country rocks that are highly radiogenic with respect to Sr isotopes and are selectively assimilated (e.g., Duffield and Ruiz, 1998).

For the glasses, there is some latitude in their Pb isotope compositions on account of uncertainty in corrections for Pb isotope ingrowth since eruption. Assuming, as concluded above, that the Pb isotope uncertainties of the glasses are realistically delimited by the error bars shown in Fig. 5, the Pb isotopes measured in the older Glass Mountain glasses match those of the sanidine they host but their $^{87}Sr/^{86}Sr$ is higher. This similarity in Pb isotopes is evidence that the sanidines grew from their host melts. Accordingly, the difference in Sr isotopes between sanidine and glass would reflect radiogenic ingrowth and therefore sanidine crystallization 40 k.y. (domes OL and OC) to 280 k.y. (dome OD) before eruption,

as previously inferred (Davies et al., 1994). As such, there persists a difference between the feldspar and zircon ages for domes OL and OC. Given that the absolute value of the Sr model ages largely depends on the $^{87}\text{Sr}/^{86}\text{Sr}$ of the host glass and rhyolitic glass becomes relatively Pb-rich (tens of ppm) and Sr-poor with increasing differentiation, the possibility remains that the Sr isotope signatures of the glasses were more susceptible to late-stage contamination than the Pb isotopes.

The isotopic compositions of plagioclase in the older Glass Mountain rhyolites tend to fall to higher $\delta^{207}\text{Pb}/^{206}\text{Pb}$ and lower $^{87}\text{Sr}/^{86}\text{Sr}$ than the associated sanidines. Relatively few plagioclase grains were analyzed and there is no overlap between their Pb isotope compositions, making comparisons to the generally uniform Sr isotope ratios obtained on plagioclase from the same domes suspect. Even so, the Sr and Pb isotope signatures of plagioclase will be biased towards different stages in an evolving magmatic system: Sr, being much more compatible than Pb in plagioclase, will be biased towards the earlier stages of magma evolution. Coupled to this is the possibility that plagioclase may have started to crystallize before sanidine and therefore its Pb isotope composition may record earlier events in melt evolution. Alternatively, given the magnitude of Pb isotope heterogeneity delimited by the plagioclase, portions or all of the grains may be antecrysts. Irrespective of its origin, the Pb isotopic heterogeneity of plagioclase implies open system behavior in the evolution of the older Glass Mountain rhyolites and greater uncertainty about the veracity of the plagioclase model ages.

4.2.2. Open system processes and younger Glass Mountain and Bishop Tuff magma chronology

The Bishop Tuff and younger Glass Mountain rhyolites exhibit relatively modest Sr isotope heterogeneity and collectively lie at (and could be used to define) one end of the Pb–Sr isotope mixing array defined by the older Glass Mountain rhyolites (Fig. 5). Unlike the older Glass Mountain rhyolites, dome YG host glass has higher (“more contaminated”) $\delta^{207}\text{Pb}/^{206}\text{Pb}$ than has the sanidine it contains, reflecting contamination/mixing that occurred after the feldspar began to grow. Similarly, but in the opposite sense, the host glass of dome YA has lower (“less contaminated”) $\delta^{207}\text{Pb}/^{206}\text{Pb}$ and it also has higher $^{87}\text{Sr}/^{86}\text{Sr}$ than has its feldspars. A slightly older eruption age based on the zircon crystallization ages rather than a K–Ar glass age (Simon and Reid, 2005) would diminish these differences only slightly. Simon and Reid (2005) reconciled differences in the YA zircon crystallization ages and Sr isotope characteristics by inferring that the feldspars may have grown from a melt significantly different from the one in which they erupted. The Pb isotope data support this scenario and further require that the late-added melt have a less contaminated, more mantle-like Pb signature.

For the Bishop Tuff, sanidines erupted later in the pyroclastic sequence range from only 0.70595 to 0.70603 in Sr isotope composition (Christensen and DePaolo, 1993; Davies and Halliday, 1998). When cast in terms of model ages, this limited range amplifies to crystallization ages in excess of 400 k.y. older than eruption. The scatter in late Bishop Tuff sanidine Pb isotope compositions, while small, ranges up to three times that of analytical uncertainty and could signify

that the subtle variation in Sr isotopes reflects an isotopically heterogeneous system rather a protracted interval of crystallization. If, as for the young Glass Mountain rhyolites, the glass is differentially more affected by open system processes, the sanidine-glass model ages could overestimate both the range and absolute value of the crystallization ages. This source of uncertainty could account for the ~150 k.y. difference between the mean feldspar model age (~1000 ka) and mean zircon crystallization age (~850 ka) (Simon and Reid, 2005). The extent of chemical relaxation of Sr and Ba heterogeneity in zoned Bishop Tuff sanidine also supports a duration of no more than a few hundred k.y. for magma accumulation (Morgan and Blake, 2006). Pb isotope analyses of Bishop Tuff glasses would have enabled us to more rigorously assess the effect of open system processes, but were not possible using laser ablation because of the porous texture of the glass.

Relative to late Bishop Tuff sanidines, feldspars erupted early in the sequence are characterized by more heterogeneous Sr isotopes, with $^{87}\text{Sr}/^{86}\text{Sr}$ in individual feldspars ranging to >0.713 (Davies and Halliday, 1998). This is in marked contrast with the strong clustering in Pb isotopes that we find for the same phases (Fig. 3). Assuming that the Pb isotope analyses obtained here included the uncommon sanidines with highly radiogenic Sr like those analyzed by Davies and Halliday (1998), both the Pb and Nd isotope signatures of those feldspars would be consistent with crystallization from Bishop Tuff-like melts (i.e., LVref). Accordingly, these feldspars could have crystallized from melts similar to the Bishop Tuff but after a long interval (>1 Myr) of radiogenic ^{87}Sr ingrowth in the melt, i.e., the melts would have long residence times. As described elsewhere (Simon and Reid, 2005) these melts could have been admixed into the main body of the Bishop Tuff with negligible isotopic effects. Alternatively, the feldspars with high $^{87}\text{Sr}/^{86}\text{Sr}$ could be antecrysts from earlier intrusions. The significance of these crystals for magma residence time is ambiguous, in part, because some early erupted feldspars yield negative Sr model ages when coupled with their host melt. Additionally, because no comparably “old” (>1 Ma) crystallization ages have been found in the nearly 50 early Bishop Tuff zircons that have been analyzed (Mesozoic xenocrysts aside), alternative explanations warrant exploration. As judged by their lower than average Sr contents (Davis and Halliday, 1998), the sanidines characterized by high $^{87}\text{Sr}/^{86}\text{Sr}$ must have crystallized from extremely evolved melts. Given the high Pb/Sr and Nd/Sr that characterize evolved melts, the Sr isotope composition of the melts would have been more responsive to contamination than Pb and Nd. Alternatively, given their low Ba contents (Dunbar and Hervig, 1992), the sanidines with low Sr contents may also have relatively low Pb contents. If so, it is possible that these sanidines were systematically excluded from Pb isotope analysis because of inadequate Pb ion yield. The implications of our Pb isotope data for the relative roles of decay versus contamination in the early Bishop Tuff could be further assessed by paired Sr–Pb analyses on individual feldspars or, more indirectly, by determining whether any of the sanidines we analyzed had the low Sr contents (<2 ppm) that characterize Bishop Tuff sanidines with high $^{87}\text{Sr}/^{86}\text{Sr}$.

4.2.3. Whole rock Rb–Sr pseudo-isochrons in the context of mixing/contamination

We envision that the Long Valley rhyolites originated by mixing/contamination involving less evolved, recently mantle-derived melts or sources and more evolved crustal sources, especially previously emplaced Long Valley magma bodies. At possible variance with this is the interpretation that the near eruption-aged Rb–Sr pseudo-isochrons require closed system evolution (i.e., no magma mixing or assimilation of crust), especially because the low Sr concentrations of the rhyolites make magmatic Rb–Sr systematics susceptible to disturbances (Halliday et al., 1989). This restriction only holds if open system processes involve an exotic end-member. Open system processes could have taken place and still preserve near eruption-aged Rb–Sr pseudo-isochrons if mixing (*sensu lato*) occurred mainly between materials represented by the more radiogenic rhyolites (Schmitt and Simon, 2004; Simon and Reid, 2005) and an end-member with an isotopic composition represented by the late Bishop Tuff.

4.3. Long Valley rhyolites reflect accelerated magma production rates rather than long storage

Large silicic magma systems are characterized by long periods of relative calm between caldera-forming eruptions. The volcanic quiescence has been interpreted as accompanying the progressive buildup of magma volumes before major eruptive activity (Smith, 1979; Spera and Crisp, 1981; Shaw, 1985). Rates of magma chamber growth leading to the production and eruption of voluminous highly evolved magmas is a key constraint for understanding hazards related to volcanic eruptions (Simkin, 1993) and mechanisms of crustal differentiation and growth (Clemens, 1998).

The increase in relative eruption sizes at Long Valley up through eruption of the Bishop Tuff (Fig. 6) has been seen by many as the classic example of fractional tapping of a growing magma reservoir. Here we point out that a gradual (>1 Myr) buildup of a long-lived magma chamber likely never occurred at Long Valley: even allowing for the possibility that the zircons are antecrysts as well as phenocrysts (e.g., Charlier et al., 2004; Bacon and Lowenstern, 2005), the zircon ages delimit the magma residence times of individual rhyolites to ≤ 0.3 Myr (this study; Reid and Coath, 2000; Simon and Reid, 2005). Moreover, eruption volumes are inversely correlated with these pre-eruption ages (Fig. 6). This anticorrelation is contrary to the standard model of steady magma buildup (Smith, 1979; Spera and Crisp, 1981; Shaw, 1985). Given the greater crustal signature of the older Glass Mountain rhyolites, differentially older zircons might be expected if zircon cores were inherited. Although Mesozoic xenocrysts have been identified, no Mesozoic cores have been found in the >120 Long Valley zircons studied so far (Reid and Coath, 2000; Simon and Reid, 2005; and this study). Rather, the longer pre-eruption intervals spanned by zircons for the older rhyolites may represent longer magma storage/crustal transfer times and/or greater salvaging of zircon from youthful co-genetic intrusions (e.g., Schmitt and Simon, 2004). An evolving magma system, likely due to an accelerated rate of mantle-derived influx, can explain the relationships between pre-eruption crystallization

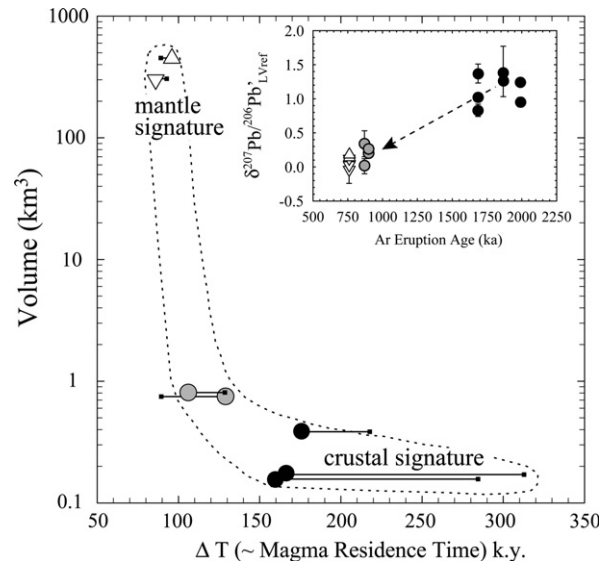


Fig. 6. Magma volume versus mean residence timescale for precaldera Glass Mountain and caldera-related Bishop Tuff rhyolites. The volume of erupted rhyolite accelerates through time. Magma residence times are based on zircon dating (this study; Reid and Coath, 2000; and Simon and Reid, 2005) and Ar eruption ages (Metz and Mahood, 1985; Davies et al., 1994; van den Bogaard and Schirnack, 1995; and Renne et al., 2006). Age spans are calculated two ways: as the difference between mean pre-eruption and the youngest zircon ages (see text for details) and as the difference between mean pre-eruption zircon ages and Ar eruption ages. The timescales based on zircon ages are shown as symbols (open up triangle is LBT, open down triangle is EBT, grey circles are younger Glass Mountain rhyolites, and black circles are older Glass Mountain rhyolites). Each timescale based on zircon and Ar eruption dating is shown as a square and connected, by a horizontal line, to each rhyolite, respectively. Eruption ages are, with one exception (dome YG), from Ar/Ar dating. Volume (km^3) estimates for Glass Mountain rhyolites are based on areal measurements and dome thicknesses (approximated by contour intervals) from map of Metz and Bailey (1993). Volumes for the studied Long Valley eruptions have been calculated after Hildreth and Mahood (1986) and Metz (1987). The approximately equal proportions of fall versus flow deposits of the >750 km^3 Bishop Tuff (Hildreth and Mahood, 1986) come from Hildreth (1977) and references therein. Inset shows shift of Pb isotope composition (for crystals and host glasses) versus eruption age.

time periods, isotopic affinities, and eruptive volumes by limiting the degree to which ascending melt interacts with resident crustal materials.

5. CONCLUSIONS

The accurate and precise *in situ* Pb isotope measurements of Long Valley rhyolitic materials afforded by laser ablation MC-ICPMS demonstrate the utility of this new approach to investigating crystal–melt dynamics. The evidence for secular Pb isotope evolution of the precaldera and caldera-related Long Valley rhyolites obtained here shows that the Rb–Sr isotope characteristics of these rhyolites cannot be treated in strictly closed-system terms. Moreover, the Pb isotope heterogeneity

within individual rhyolites implies that some magmas experienced open system processes and therefore that Rb–Sr isotope ages should be used with caution. The distinct Pb isotope characteristics of the various feldspar populations support evidence from the zircon age populations for derivation of the Glass Mountain and Bishop Tuff rhyolites from a number of transient and distinct magma bodies (Hildreth, 2004; Simon and Reid, 2005). Collectively, feldspar and zircon data imply that at Long Valley the rhyolite magma production rate increases before the climatic eruption while the duration and degree of silicic magmas interaction with the crust decrease.

ACKNOWLEDGMENTS

We thank Charlie Bacon for his thoughtful and constructive contributions to this study. We are grateful to Joel Baker, Jon Davidson, and an anonymous reviewer for their input. We also thank Yuri Amelin for his constructive input and editorial handling. Funding was provided by NSF Grant EAR-003601 to M.R.R. J.I.S. acknowledges use of the Berkeley Geochronology Center facilities. The LA-MC-ICPMS and the UCLA ion microprobe are partially supported by grants from the NSF Instrumentation and Facilities Program.

APPENDIX A. LA-MC-ICPMS BY SAMPLE-STANDARD COMPARISON

A.1. Analytical design and protocol

A 213-nm wavelength UV Nd-YAG laser is used to vaporize the target sample and standard materials. Analysis of feldspar and glass produce $\sim 25\ \mu\text{m}$ deep, $\leq 800\ \mu\text{m}$ long, and $100\text{--}150\ \mu\text{m}$ wide ablation tracks. Typical sample traverse rates are $\sim 10\ \mu\text{m/s}$. The laser is operated at a fluence of $22\text{--}30\ \text{J/cm}^2$ with a pulse repetition rate of 4–6 Hz. The relatively short widely spaced pulses of UV photons minimize thermal effects at the sample surface and thereby potential isotope fractionation at the sample surface. The beam delivery system is built by New Wave ResearchTM. Real-time digital images of the target materials can be viewed (and captured) with a microscope mounted video camera. A small (cylindrical inner diameter = $1.27\ \text{cm}$) Teflon sample chamber built at UCLA is used to reduce dead volume for laser ablation. Solids are pressed into sticky putty so that dead volume due to interstitial sample spacing is further minimized. Experiments are run so that the upper surface of sample and standard materials are at the same image plane of the laser beam. Ablated materials caught up in a micro-plasma above the focal point of ablation and desolvated solutions are directed into the plasma torch of the MC-ICPMS instrument (analogous to approach of Young et al., 2002). The laser chamber is flushed of ablated material by a He gas flow of $0.32\ \text{l/min}$. Helium is directed across the ablation plume above the focal and sample surface. Ablation products entrained in He are carried from the sample chamber where they are mixed with Ar ($0.76\ \text{l/min}$) gas exiting a desolvating nebulizer (CETAC AridusTM) before being introduced into the plasma source (see schematic in Appendix Fig. 1A). Proper adjustment

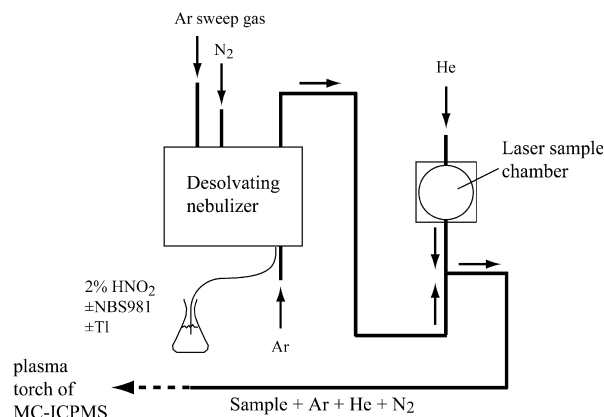


Fig. 1A. Schematic illustration of laser ablation system at UCLA. Ablation products are entrained in He gas and carried from the sample chamber where they are mixed with Ar gas exiting a desolvating nebulizer before being introduced into the plasma source.

of the sample gas flow rates is a key component to efficient transport of sampled material to the plasma source and to minimizing isotope fractionation in the plasma due to pressure differences between the sampler and skimmer cones. High gas pressures in between the cones (e.g., caused by excess addition of He to the plasma) exacerbates isotope fractionation effects related to elemental shifts in the plasma. The Ar sweep gas in the desolvating nebulizer includes a small amount of N_2 ($0.02\text{--}0.04\ \text{l/min}$) gas. We find that bleeding a small amount of N_2 into the plasma stabilizes and increases the sensitivity for Pb $\sim 2\text{--}3$ times. Flow of the Ar sweep gas entering the desolvating nebulizer is tuned daily to maximize the Pb ion signal by setting the flow rate to between 4.5 and $5.5\ \text{l/min}$. In order to maintain constant gas pressures we use the desolvating nebulizer at all times (i.e., we co-aspirated ultra clean DI water while ablation measurements are run). Solutions doped with Tl for mass bias correction are also co-aspirated with ablated materials (occasionally used during the course of technique development, but not for any of the reported sample measurements). Samples are compared to standard solutions of NBS981 that are introduced to the plasma source by passing through the desolvating nebulizer. This sample-standard scheme eliminates the need for instrumental mass fractionation corrections.

A.2. Mass spectrometry

Isotope measurements are measured using a Thermo Electron MAT MC-ICPMS instrument (the NeptuneTM). This double focusing instrument can produce the flat-top peak shapes ideal for precise isotopic ratio measurements. Seven Faraday collectors are spaced to collect ^{202}Hg , ^{203}Tl , ^{204}Pb , ^{205}Tl , ^{206}Pb , ^{207}Pb , and ^{208}Pb ion beams simultaneously. A mass resolving power (width of focal plane/width of focused ion beam) of ~ 2160 is used. All sample and standard analyses are corrected for on-peak blank. A ^{204}Hg interference correction is also applied to all analyses

($^{202}\text{Hg}/^{204}\text{Hg}$ ratio assumed to be 4.350). We find no significant difference between Hg correct and uncorrected results (the $^{20x}\text{Pb}/^{204}\text{Pb}$ ratios where $x = 206, 207$ or 208 had more significant issues related to variable backgrounds). The samples and standards are measured for 10 cycles with integration times of ~ 8 s per cycle. Each block of cycles is preceded by a ~ 160 s interval that includes a pre-analysis blank measurement and a ~ 5 s incubation period during which time the ablated ion signal becomes more uniform.

A.3. Calibration and precision of Pb isotopes for LA-MC-ICPMS by sample-standard comparison

Calibration of Pb isotope measurements is performed by alternating the measurement of ablated samples with measurements of NBS981 solution standards. The measurement of NIST glass standard NBS612 and feldspar from all of the Long Valley rhyolites were reproduced in multiple sessions over several weeks. The variability of NBS612 represents our best estimate of our external reproducibility (Table 1)

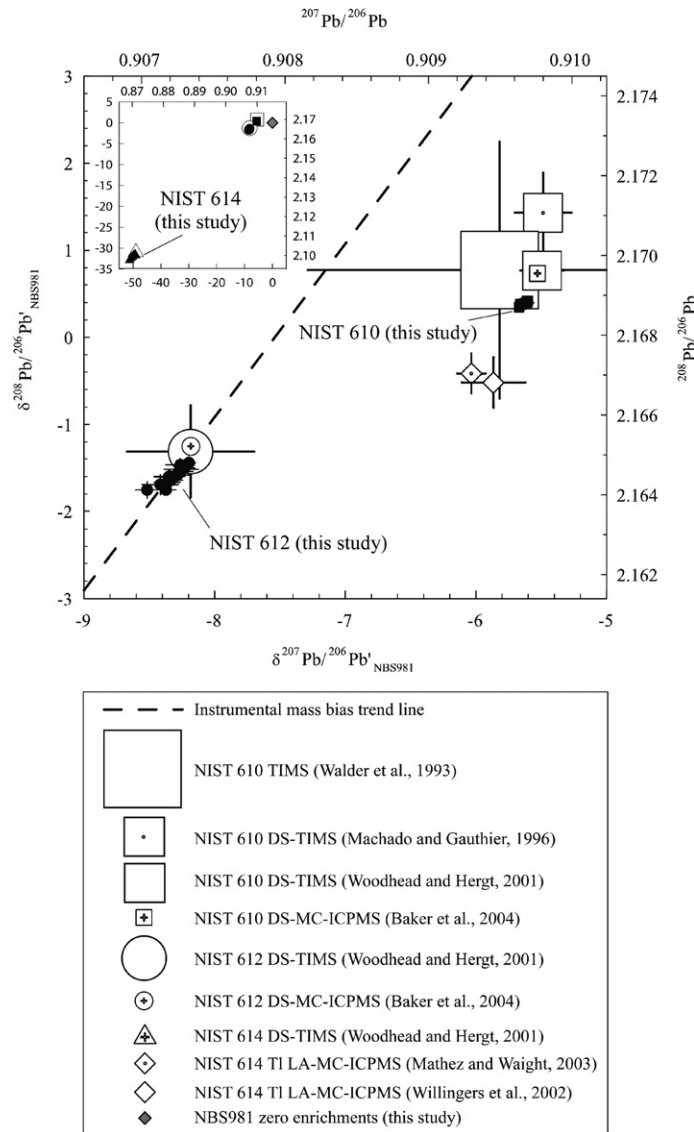


Fig. 1B. Three isotope representation of Pb isotope measurements of NIST reference materials NBS610, 612, and 614 glasses and NBS981 (in solution) by laser ablation and solution MC-ICPMS, respectively. Measurements are made by sample-standard comparison (standard, NBS981) and reported as linearized ‰ deviations from NBS981 values of Eisele et al. (2003) ($\delta^x\text{Pb}/^x\text{Pb}' = 10^3 \ln[(^x\text{Pb}/^x\text{Pb}_{\text{spl}})/(^x\text{Pb}/^x\text{Pb}_{\text{NBS981}})]$). Ion beam (i.e., concentration) of NBS981 solution used for comparison is matched to the ion signals of the NIST glasses (NBS612 has a Pb concentration similar to the studied rhyolitic materials). Repeated measurements are compared to existing TI-doped laser ablation MC-ICPMS, conventional TIMS, DS-TIMS, and DS-MC-ICPMS studies. Inset shows accuracy of Pb isotope standards over a large Pb isotope compositional range. Symbols are defined in figure and reported uncertainties are $2\sigma_m$ (equal to the symbol size if not seen).

because it has a Pb abundance that is similar to the rhyolite materials (cf. Gagnevin et al., 2005). Reproducibly over longer time periods is less well-documented because the sample measurements were made fairly rapidly (~month) and the LA-MC-ICPMS facility at UCLA is primarily used for measurement of lighter isotope systems. Zero enrichments comparing pairs of NBS981 measurements (i.e., where solutions of NBS981 are used for both sample and standard) are made daily before any sample measurements are made, throughout the day interspersed between samples, and at the end of the session (these can be seen in Appendix C). In order to maximize precision and accuracy, we restricted data collection to ^{208}Pb intensities of ~ 400 mV or greater. Even with this restriction, ratios involving ^{204}Pb are anomalously low and increase as Pb signals increase until ^{208}Pb intensities reach ~ 2 V. We suspect that this is due to a variable background with ^{204}Pb abundances of common Pb composition ($\sim$ NBS981). Reliable $^{20x}\text{Pb}/^{204}\text{Pb}$ ratios (where x represents either 206, 207 or 208) can be made for measurements that have ^{208}Pb intensities of ≥ 3 V, such as the NBS610 glass (Table 1). Rhyolitic materials typically have Pb abundances that are an order of magnitude lower and for this reason, useful Pb isotope ratios involving ^{204}Pb are absent for the studied rhyolites. The data are measured as differences from the NBS981 standard. We then express the composition of the samples in terms of the linear delta notation (δ) reported as a deviation from a representative local mafic composition (Cousens, 1996) where $\delta^x\text{Pb}/^{206}\text{Pb} = 10^3 \ln((^x\text{Pb}/^{206}\text{Pb})_{\text{sample}}/(^x\text{Pb}/^{206}\text{Pb})_{\text{LVref}})$ and x represents either 207 or 208. On this scale the value for “LVref” ($^{207}\text{Pb}/^{206}\text{Pb} = 0.8174$ and $^{208}\text{Pb}/^{206}\text{Pb} = 2.0282$) is 0.0‰.

Laser ablation liberates all of the elements found in the target material. Undesirable shifts in Pb isotope ratios can result from isobaric interferences and potential interelemental matrix effects. Despite our improvements in analytical precision no significant interference has been observed (cf. Mathez and Waight, 2003). Isotope fractionation due to interelemental interactions (the so-called “matrix effect”) obeys well-behaved theoretical mass-dependent laws. It can be readily identified because the shifts in isotopic composition lie along characteristic trends with slopes that range between those expected for equilibrium (the high temperature limit) and kinetic behavior.

Equilibrium fractionation law $\alpha_{208/206} = (\alpha_{207/206})^\beta$, where $\beta = \frac{(\frac{1}{206} - \frac{1}{208})}{(\frac{1}{206} - \frac{1}{207})}$. The exponent $\beta = 1.990$ comes from consideration of the reduced masses.

Kinetic fractionation law $\alpha_{208/206} = (\alpha_{207/206})^\beta$, where $\beta = \frac{\ln(\frac{206}{208})}{\ln(\frac{206}{207})}$. The exponent $\beta = 1.995$ comes from consideration of the masses in motion related to the kinetic process.

If the $^{208}\text{Pb}/^{206}\text{Pb}$ ratio is plotted on the ordinate and $^{207}\text{Pb}/^{206}\text{Pb}$ ratio is plotted on the abscissa (i.e., a three-isotope plot) equilibrium and kinetic laws follow trends with slopes of 1.990 and 1.995, respectively. For our purposes these are indistinguishable and we use a generalized instrumental fractionation trend with a slope of 2.0 for reference in all figures in which the three major Pb isotopes are presented. To some degree the enrichment and depletion effects of individual matrix elements tend to cancel each other (i.e., the difference in the total multi-element “matrix effect” of one sample to the next that both have relevant “matrix” produce only minor shifts, $\leq 0.5\text{‰}$ in all Pb isotope ratios). The best way to determine whether accurate measurements are possible is by running empirical tests.

The NIST glass certified reference materials used for internal standardization have bulk CNAS compositions that have been doped with Pb and 60 other trace elements to nominal concentrations of 500, 50, and 1 ppm (Woodhead and Hergt, 2001). As a result of the doping and dilution processes the original support matrix composition relative to the Pb concentrations change considerably, and non-linearly, between the different glasses (Pb:matrix elements vary by ~ 2 orders of magnitude). The excellent match of our results to the Pb isotope compositions of all three NIST glass standards attest to the applicability of our approach in the face of large changes between Pb concentration and host chemical composition (Fig. 1A and Appendix B). Additional empirical evidence that the reported Pb isotope measurements are robust come from the fact that: (1) we get a general correspondence between feldspar and their rhyolite host glasses (despite their relatively distinct major element compositions), (2) there is no systematic isotopic difference between plagioclase and sanidine measurements for individual rhyolites, and (3) we get the “right” answer for the glass and feldspar as compared to existing TIMS measurements of Long Valley materials (e.g., Halliday et al., 1989; Davies et al., 1994) Appendix Fig. 1B.

APPENDIX B

Individual laser ablation NIST glass standard Pb isotope measurements

	²⁰⁸ Pb intensities	$\delta^{206}\text{Pb}/^{204}\text{Pb}'$	σ_m	$\delta^{207}\text{Pb}/^{204}\text{Pb}'$	σ_m	$\delta^{208}\text{Pb}/^{204}\text{Pb}'$	σ_m	$\delta^{207}\text{Pb}/^{206}\text{Pb}'$	σ_m	$\delta^{208}\text{Pb}/^{206}\text{Pb}'$	σ_m
Glass 610											
	6.26	6.24	0.07	0.63	0.07	6.65	0.07	-5.61	0.02	0.42	0.03
	6.21	6.38	0.07	0.76	0.05	6.77	0.05	-5.62	0.03	0.39	0.02
	6.40	6.47	0.04	0.86	0.05	6.86	0.05	-5.61	0.02	0.40	0.02
	6.28	6.24	0.11	0.59	0.10	6.61	0.10	-5.65	0.02	0.37	0.02
	6.61	6.40	0.06	0.78	0.06	6.78	0.06	-5.62	0.02	0.38	0.02
	6.54	6.36	0.06	0.70	0.06	6.74	0.05	-5.66	0.02	0.38	0.01
	6.39	6.44	0.05	0.80	0.04	6.82	0.05	-5.64	0.02	0.38	0.01
	6.59	6.49	0.05	0.89	0.04	6.91	0.04	-5.60	0.01	0.42	0.01
	6.81	6.22	0.10	0.59	0.09	6.62	0.09	-5.62	0.02	0.40	0.01
	6.83	6.28	0.08	0.61	0.07	6.62	0.07	-5.67	0.02	0.34	0.02
<i>n</i> = 10	Wt. average	6.40	0.04	0.77	0.04	6.78	0.04	-5.63	0.01	0.39	0.01
Glass 612											
	1.01	10.22	0.24	1.86	0.21	8.55	0.23	-8.36	0.05	-1.67	0.05
	1.00	9.57	0.30	1.37	0.29	8.13	0.30	-8.20	0.03	-1.44	0.03
	0.94	8.52	0.30	0.25	0.28	7.05	0.29	-8.26	0.05	-1.47	0.04
	1.26	9.00	0.17	0.30	0.18	6.67	0.19	-8.70	0.05	-2.32	0.05
	1.24	9.14	0.21	0.77	0.22	7.38	0.21	-8.37	0.04	-1.76	0.03
	1.20	9.42	0.19	1.09	0.20	7.78	0.19	-8.34	0.05	-1.64	0.03
	1.01	9.00	0.25	0.74	0.24	7.46	0.24	-8.27	0.05	-1.54	0.01
	1.02	8.58	0.33	0.35	0.33	7.07	0.32	-8.23	0.04	-1.50	0.04
	0.95	8.16	0.19	-0.12	0.19	6.58	0.18	-8.28	0.05	-1.58	0.04
	0.95	8.10	0.30	-0.15	0.27	6.59	0.29	-8.26	0.07	-1.52	0.02
	1.11	9.70	0.22	1.28	0.21	8.01	0.22	-8.42	0.07	-1.69	0.06
	0.91	7.71	0.30	-0.81	0.30	5.95	0.30	-8.52	0.04	-1.76	0.05
	0.83	6.85	0.29	-1.51	0.25	5.24	0.27	-8.35	0.06	-1.61	0.04
	0.89	8.37	0.30	0.03	0.27	6.77	0.29	-8.35	0.06	-1.60	0.04
<i>n</i> = 14	Wt. average	8.85	0.14	0.49	0.13	7.20	0.14	-8.30	0.03	-1.57	0.01
Glass 614											
	0.09	13.57	1.96	-35.56	1.91	-18.11	1.95	-49.13	0.21	-31.69	0.09
	0.09	13.59	1.42	-35.64	1.44	-17.92	1.45	-49.23	0.22	-31.51	0.16
	0.08	7.92	1.54	-41.35	1.58	-23.83	1.58	-49.27	0.20	-31.75	0.17
	0.08	6.18	2.34	-43.54	2.37	-25.59	2.36	-49.73	0.17	-31.77	0.18
	0.08	12.32	2.21	-36.81	2.30	-19.38	2.32	-49.13	0.20	-31.70	0.18
	0.08	14.77	1.22	-35.03	1.29	-17.00	1.36	-49.80	0.24	-31.77	0.21
	0.08	9.04	2.78	-41.71	2.73	-23.32	2.77	-50.74	0.25	-32.35	0.21
	0.08	6.50	1.93	-43.00	1.96	-25.24	1.97	-49.49	0.28	-31.73	0.20
	0.08	10.92	1.58	-39.28	1.67	-21.13	1.59	-50.20	0.24	-32.06	0.28
	0.08	20.10	2.31	-31.08	2.34	-12.83	2.35	-51.18	0.08	-32.93	0.07
<i>n</i> = 10	Wt. average	11.90	2.70	-37.80	2.60	-20.10	2.70	-50.32	0.11	-32.18	0.09

Results are expressed in the linearized delta notation $\delta^x\text{Pb}/^x\text{Pb}' = 10^3 \ln(^x\text{Pb}/^x\text{Pb}_{\text{spl}}/^x\text{Pb}_{\text{NBS981}})$.

NBS981 values from Eisele et al. (2003) ($^{207}\text{Pb}/^{206}\text{Pb} = 0.817945$; $^{208}\text{Pb}/^{206}\text{Pb} = 2.02869$).

APPENDIX C

Individual solution NBS981 Pb isotope measurements (“zero enrichments”)

	²⁰⁸ Pb intensities	$\delta^{206}\text{Pb}/^{204}\text{Pb}'$	σ_m	$\delta^{207}\text{Pb}/^{204}\text{Pb}'$	σ_m	$\delta^{208}\text{Pb}/^{204}\text{Pb}'$	σ_m	$\delta^{207}\text{Pb}/^{206}\text{Pb}'$	σ_m	$\delta^{208}\text{Pb}/^{206}\text{Pb}'$	σ_m
	0.66	0.05	0.35	0.02	0.35	0.12	0.35	-0.03	0.08	0.07	0.06
	0.68	-0.20	0.38	-0.32	0.39	-0.25	0.40	-0.12	0.06	-0.05	0.04
	0.69	0.21	0.29	0.23	0.30	0.28	0.32	0.02	0.06	0.08	0.04
	0.70	-0.36	0.38	-0.44	0.37	-0.40	0.39	-0.08	0.06	-0.05	0.05
	0.70	0.46	0.25	0.52	0.27	0.46	0.28	0.06	0.04	0.00	0.07
	0.70	-0.87	0.42	-0.81	0.45	-0.89	0.43	0.05	0.06	-0.02	0.04
	1.25	-0.38	0.34	-0.37	0.35	-0.39	0.35	0.00	0.03	-0.01	0.03
	1.24	-0.05	0.16	-0.07	0.16	-0.11	0.16	-0.03	0.05	-0.07	0.04

(continued on next page)

Appendix C (continued)

²⁰⁸ Pb intensities	$\delta^{206}\text{Pb}/^{204}\text{Pb}'$	σ_m	$\delta^{207}\text{Pb}/^{204}\text{Pb}'$	σ_m	$\delta^{208}\text{Pb}/^{204}\text{Pb}'$	σ_m	$\delta^{207}\text{Pb}/^{206}\text{Pb}'$	σ_m	$\delta^{208}\text{Pb}/^{206}\text{Pb}'$	σ_m
1.29	0.26	0.13	0.29	0.14	0.25	0.14	0.03	0.04	−0.02	0.03
1.29	−0.41	0.13	−0.56	0.12	−0.54	0.13	−0.15	0.04	−0.13	0.04
1.36	−0.11	0.15	−0.06	0.12	−0.09	0.14	0.05	0.03	0.02	0.02
1.24	−0.04	0.19	−0.13	0.19	−0.08	0.19	−0.10	0.03	−0.04	0.03
1.25	0.30	0.19	0.33	0.19	0.29	0.19	0.03	0.02	−0.01	0.03
1.26	−0.35	0.26	−0.35	0.28	−0.34	0.26	0.00	0.05	0.01	0.02
1.26	0.22	0.25	0.20	0.26	0.23	0.25	−0.02	0.04	0.01	0.04
1.23	0.11	0.24	0.07	0.24	0.08	0.25	−0.04	0.04	−0.03	0.02
1.16	0.29	0.34	0.25	0.34	0.26	0.33	−0.04	0.04	−0.03	0.04
1.11	0.02	0.29	−0.01	0.30	0.02	0.31	−0.04	0.03	0.00	0.03
1.14	0.35	0.22	0.37	0.20	0.36	0.22	0.03	0.04	0.01	0.03
1.60	−0.04	0.24	−0.04	0.22	−0.05	0.25	0.00	0.03	−0.02	0.05
1.64	−0.21	0.20	−0.16	0.20	−0.16	0.20	0.06	0.04	0.06	0.02
1.59	0.16	0.19	0.13	0.18	0.09	0.19	−0.03	0.04	−0.07	0.03
1.64	0.18	0.18	0.14	0.17	0.26	0.17	−0.04	0.04	0.07	0.03
1.02	−0.02	0.17	−0.12	0.17	−0.11	0.17	−0.10	0.02	−0.09	0.04
1.05	0.38	0.25	0.32	0.24	0.31	0.25	−0.05	0.06	−0.07	0.04
1.05	−0.04	0.24	−0.07	0.20	−0.13	0.21	−0.03	0.07	−0.09	0.04
1.05	0.16	0.24	0.19	0.23	0.17	0.23	0.02	0.04	0.01	0.03
1.34	0.15	0.16	0.10	0.15	0.05	0.16	−0.05	0.03	−0.10	0.05
1.36	0.25	0.26	0.19	0.27	0.24	0.27	−0.06	0.03	−0.01	0.05
1.33	−0.06	0.25	−0.08	0.22	−0.10	0.23	−0.03	0.04	−0.04	0.03
1.35	−0.07	0.20	−0.04	0.22	−0.07	0.20	0.03	0.04	−0.01	0.03
0.29	0.44	0.54	0.40	0.50	0.37	0.51	−0.04	0.14	−0.07	0.06
0.30	0.47	0.63	0.49	0.66	0.44	0.63	0.02	0.05	−0.03	0.05
<i>n</i> = 33	Wt. average	0.03	0.08	0.00	0.07	0.00	−0.02	0.01	−0.01	0.01

Results are expressed in the linearized delta notation $\delta^x\text{Pb}/^x\text{Pb}' = 10^3 \ln(^x\text{Pb}/^x\text{Pb}_{\text{NBS981}}/^x\text{Pb}_{\text{NBS981}})$.

NBS981 values from Eisele et al. (2003) ($^{207}\text{Pb}/^{206}\text{Pb} = 0.817945$; $^{208}\text{Pb}/^{206}\text{Pb} = 2.02869$).

APPENDIX D

Individual laser ablation feldspar and glass Pb isotope measurements

Sample ^a	²⁰⁸ Pb beam	$\delta^{207}\text{Pb}/^{206}\text{Pb}'$	$2\sigma_m$	$\delta^{208}\text{Pb}/^{206}\text{Pb}'$	$2\sigma_m$
<i>Early Bishop Tuff</i>					
R99LV51-g16t1sn	0.86	0.10	0.09	0.52	0.06
R99LV51-g17t1sn	1.23	0.17	0.10	0.94	0.05
R99LV51-g19t1sn	0.84	0.38	0.11	0.89	0.14
R99LV51-g20t1sn	1.07	0.53	0.20	1.20	0.15
R99LV51-g21t1sn	1.20	0.31	0.13	0.92	0.07
R99LV51-g22t1sn	1.01	−0.09	0.09	0.40	0.07
R99LV51-g23t1sn	0.88	0.28	0.07	0.82	0.08
R99LV51-g24t2sn	0.75	0.25	0.13	0.81	0.06
R99LV51-g25t2sn	0.75	0.13	0.12	0.73	0.11
R99LV51-g26t1sn	1.18	0.12	0.04	0.77	0.05
R99LV51-g27t1sn	0.83	0.25	0.11	0.80	0.11
<i>JS03LV18p1-g3t1sn</i>					
JS03LV18p1-g3t1sn	0.59	0.10	0.08	0.83	0.10
JS03LV18p1-g7t1sn	0.79	0.28	0.07	1.04	0.06
JS03LV18p1-g8t1sn	0.71	0.09	0.10	0.60	0.11
JS03LV18p1-g10t1sn	0.79	0.12	0.05	0.82	0.11
JS03LV18p1-g10t2sn	0.76	−0.08	0.12	0.21	0.06
JS03LV18p1-g12t1sn	0.64	0.15	0.10	0.70	0.07
JS03LV18p1-g13t1sn	0.84	0.09	0.15	0.76	0.11
JS03LV18p1-g13t2sn	0.71	−0.04	0.08	0.51	0.08
JS03LV18p1-g14t1sn	0.81	0.18	0.07	0.91	0.08
JS03LV18p1-g15t1sn	0.65	0.04	0.12	0.74	0.13
<i>Late Bishop Tuff</i>					
JS03LV20-16/17p1-g4t1sn	1.12	0.08	0.12	0.84	0.06

Appendix D (*continued*)

Sample ^a	²⁰⁸ Pb beam	$\delta^{207}\text{Pb}/^{206}\text{Pb}'$	$2\sigma_m$	$\delta^{208}\text{Pb}/^{206}\text{Pb}'$	$2\sigma_m$
JS03LV20-16/17p1-g5t1sn	1.05	-0.09	0.07	0.50	0.10
JS03LV20-16/17p1-g11t1sn	0.81	0.08	0.07	0.86	0.07
JS03LV20-16/17p1-g14t1sn	0.44	0.23	0.14	1.04	0.12
JS03LV20-16/17p1-g15t1sn	0.81	0.07	0.07	0.76	0.10
JS03LV20-16/17p1-g16t1sn	0.41	0.07	0.16	0.81	0.07
JS03LV20-16/17p1-g13t1pl	0.54	-0.06	0.18	0.73	0.19
R99LV58-g5t1sn	0.59	0.21	0.14	0.97	0.11
R99LV58-g10t1sn	0.40	0.08	0.16	0.83	0.08
R99LV58-g11t1sn	0.47	-0.04	0.10	0.74	0.13
R99LV58-g13t1sn	0.52	0.77	0.29	1.25	0.24
R99LV58-g14t1sn	0.53	0.27	0.09	0.96	0.15
R99LV58-g29t1sn	0.98	0.02	0.12	0.83	0.12
R99LV58-g32t1sn	0.64	0.07	0.14	0.77	0.14
R99LV58-g33t1sn	0.93	0.15	0.09	0.86	0.09
R99LV58-g34t1sn	0.90	0.10	0.09	0.79	0.06
R99LV58-g35t1sn	0.85	0.17	0.09	0.94	0.06
R99LV58-g37t1sn	0.99	0.07	0.09	0.82	0.06
R99LV58-g9t1pl	0.39	0.09	0.14	0.90	0.17
R99LV58-g30t1pl	0.40	0.13	0.19	0.94	0.11
R99LV58-g31t1pl	0.53	0.12	0.10	1.02	0.08
R99LV58-g36t1pl	0.47	0.13	0.12	0.91	0.12
<i>Glass Mountain YA</i>					
JS01LV04-g14t1sn	0.46	0.34	0.19	1.07	0.15
JS01LV04-g1t1pl	0.56	0.85	0.14	1.66	0.08
JS01LV04-g3t1pl	0.56	0.53	0.22	0.95	0.27
JS01LV04-g4t1pl	0.57	1.11	0.21	1.59	0.15
JS01LV04-g5t2pl	0.65	0.15	0.09	0.84	0.12
JS01LV04-g6t1pl	0.57	0.19	0.12	0.91	0.11
JS01LV04-g7t2pl	0.48	0.45	0.10	1.24	0.05
JS01LV04-g13t1pl	0.39	0.19	0.15	1.06	0.12
JS01LV04-g16t1pl	0.44	0.17	0.20	1.00	0.10
JS01LV04-g17t1pl	0.60	0.22	0.09	1.03	0.10
<i>Glass Mountain YG</i>					
R00LV62-g1t1sn	0.90	0.78	0.24	1.32	0.28
R00LV62-g2t1sn	0.95	0.31	0.12	1.03	0.06
R00LV62-g3t1sn	0.82	0.39	0.11	1.08	0.09
R00LV62-g5t1sn	0.90	0.10	0.15	0.90	0.17
R00LV62-g6t1sn	0.97	0.30	0.35	0.76	0.38
R00LV62-g7t1sn	1.23	0.17	0.09	1.01	0.07
R00LV62-g10t1sn	1.00	0.17	0.17	0.93	0.12
R00LV62-g11t1sn	0.82	0.15	0.13	0.85	0.09
R00LV62-g12t1sn	0.86	0.15	0.08	0.82	0.10
R00LV62-g4t1sn/anorthoclase	0.78	0.23	0.08	0.97	0.10
R00LV62-g8t1sn/anorthoclase	1.09	0.21	0.08	0.96	0.06
R00LV62-g9t1sn/anorthoclase	0.86	0.15	0.08	0.90	0.09
R00LV62-glass I	0.82	0.25	0.12	1.15	0.10
R00LV62-glass II	0.75	0.25	0.04	1.04	0.09
R00LV62-glass III	0.73	0.20	0.12	1.20	0.10
<i>Glass Mountain OD</i>					
R00LV63-g2t1sn	1.52	1.04	0.08	1.78	0.04
R00LV63-g3t1sn	1.36	1.00	0.05	1.96	0.07
R00LV63-g4t1sn	1.37	0.98	0.14	1.91	0.10
R00LV63-g5t1sn	1.56	1.24	0.10	2.17	0.09
R00LV63-g8t1sn	1.26	0.96	0.11	1.72	0.08
R00LV63-g9t1sn	1.59	0.94	0.06	1.87	0.04
R00LV63-g11t1sn	1.77	1.05	0.09	1.97	0.09
R00LV63-g13t1sn	1.65	0.94	0.10	1.80	0.06
R00LV63-g7t1sn/anorthoclase	1.56	1.10	0.08	2.05	0.06
R00LV63-g12t1sn/anorthoclase	1.27	0.89	0.19	1.56	0.30
R00LV63-g1t1pl	0.76	2.10	0.37	2.90	0.24

(continued on next page)

Appendix D (continued)

Sample ^a	²⁰⁸ Pb beam	$\delta^{207}\text{Pb}/^{206}\text{Pb}'$	$2\sigma_m$	$\delta^{208}\text{Pb}/^{206}\text{Pb}'$	$2\sigma_m$
R00LV63-g16t1pl	0.97	1.26	0.08	2.31	0.09
R00LV63-g18t1pl	0.72	1.65	0.15	2.55	0.12
<i>Glass Mountain OL</i>					
JS01LV05-g1t1sn	1.41	1.34	0.05	2.44	0.05
JS01LV05-g2at1sn	1.17	1.44	0.08	2.43	0.07
JS01LV05-g2bt1sn	0.99	1.40	0.09	2.38	0.05
JS01LV05-g3t1sn	1.47	1.33	0.09	2.18	0.09
JS01LV05-g4t1sn	1.47	1.47	0.05	2.53	0.09
JS01LV05-g7t1sn	1.30	1.24	0.12	2.29	0.07
JS01LV05-g6t1sn/anorthoclase	1.63	1.43	0.05	2.59	0.06
JS01LV05-g12t1sn/anorthoclase	1.42	1.36	0.03	2.47	0.03
JS01LV05-g5t1pl	0.80	1.58	0.08	2.74	0.07
JS01LV05-g9t1pl	0.75	1.26	0.08	2.27	0.08
<i>Glass Mountain OC</i>					
R00LV60-g1t1sn	1.25	1.34	0.06	2.47	0.10
R00LV60-g2t1sn	1.20	1.15	0.11	2.19	0.08
R00LV60-g3t1sn	1.09	0.90	0.35	1.81	0.39
R00LV60-g4t1sn	1.08	1.21	0.10	2.19	0.04
R00LV60-g5t1sn	1.40	1.28	0.10	2.47	0.07
R00LV60-g6t1sn	1.19	1.14	0.11	2.22	0.11
R00LV60-g7t1sn	1.31	1.25	0.10	2.27	0.06
R00LV60-g8t1sn	1.20	1.29	0.07	2.36	0.09
R00LV60-g9t1sn	1.08	1.22	0.11	2.12	0.12
R00LV60-g10t1sn	1.14	1.25	0.07	2.28	0.08
R00LV61-g1t1sn/anorthoclase	1.24	1.33	0.08	2.36	0.06
R00LV61-g2t1sn/anorthoclase	1.34	1.34	0.08	2.44	0.08
R00LV61-g3t1sn	1.34	1.19	0.09	2.22	0.04
R00LV61-g4t1sn	1.34	1.14	0.06	2.18	0.06
R00LV61-g5t1sn	1.26	1.23	0.10	2.20	0.10
R00LV61-g6t1sn	1.19	1.16	0.08	2.21	0.08
R00LV61-g7t1sn	1.36	1.22	0.10	2.22	0.08
R00LV61-g9t1sn	1.34	1.19	0.09	2.27	0.07
R00LV61-g10t1sn	1.46	1.33	0.05	2.36	0.03
R00LV61-g11t1sn	1.31	1.16	0.08	2.07	0.06
R00LV61-g12t1sn	1.07	1.11	0.12	2.05	0.10
R00LV61-g13t1sn	1.46	1.26	0.10	2.49	0.07

Results are expressed in the linearized delta notation $\delta^x\text{Pb}/^x\text{Pb}' = 10^3 \ln(^x\text{Pb}/^x\text{Pb}_{\text{spl}}/^x\text{Pb}/^x\text{Pb}_{\text{LVref}})$. LVref is $^{207}\text{Pb}/^{206}\text{Pb} = 0.817945$; $^{208}\text{Pb}/^{206}\text{Pb} = 2.02869$, a representative lead isotope composition for a primitive Long Valley reservoir based on typical mafic lavas (Cousens, 1996). Feldspar identification performed by oil immersion. Analyses are identified by rhyolite, grain (g#), trough (s#), and feldspar type (“pl”, plagioclase; “sn”, sanidine; and “anorthoclase”).

^a Nomenclature of “early” and “late” Bishop Tuff (EBT and LBT, respectively) is primarily used for reference, but also signifies pre-eruptive temperature differences (Hildreth, 1977); Glass Mountain rhyolites after Metz (1987).

APPENDIX E

Results of U/Th/Pb ion microprobe analyses of zircon from Glass Mountain rhyolites

Spot number	$^{206}\text{Pb}/^{238}\text{U} (\times 10^4)^a$	χ^b	Age (ka) ^c	σ_m	Th/U	U (ppm) ^d
<i>Glass Mountain rhyolite OC-I (R00LV60)^c</i>						
GMII-r5g1s1	3.97	0.90	2409	71	0.25	7537
GMII-r5g1s1pol	3.49	0.99	2327	33	0.32	9229
GMII-r5g2s1	3.82	0.90	2302	63	0.26	6123
GMII-r5g2s1pol	3.38	0.96	2179	36	0.32	6782
GMII-r5g3s1	3.79	0.89	2275	88	0.23	4347
GMII-r5g3s1pol	3.20	0.92	2005	38	0.26	4873
GMII-r5g4s1	4.34	0.80	2331	92	0.29	6530
GMII-r5g4s1pol	3.31	0.99	2213	40	0.32	8632
GMII-r5g5s1	3.84	0.96	2463	61	0.34	8496

Appendix E (continued)

Spot number	$^{206}\text{Pb}/^{238}\text{U} (\times 10^4)^a$	χ^b	Age (ka) ^c	σ_m	Th/U	U (ppm) ^d
GMII-r5g5s1pol	3.41	0.99	2274	40	0.41	10176
GMII-r6g3s1	4.21	0.78	2208	75	0.29	5184
GMII-r6g4s1	3.39	0.95	2162	46	0.29	4883
GMII-r6g6s1	3.40	0.99	2266	36	0.30	7931
GMII-r6g7s1	3.92	0.92	2424	83	0.21	4379
GMII-r6g7s1pol	3.23	0.98	2146	42	0.25	5442
GMII-r7g1s1	3.93	0.91	2410	79	0.30	4543
GMII-r7g1s1pol	3.29	0.99	2193	37	0.36	5161

Analyses are identified by row (r#), grain (g#), spot (s#), near rim location (“rim”), and sessions in which an additional ~5–10 μm was removed by re-polishing (“pol”).

^a Raw data.

^b Fraction of radiogenic ^{206}Pb based on common $^{207}\text{Pb}/^{206}\text{Pb}$ correction.

^c Ages corrected for initial ^{206}Pb and ^{238}U -series disequilibrium (see text for details).

^d U concentrations are ~5% (1s) based on comparison to measurement of zircon standard 91500.

^e Sample nomenclature after Metz (1987).

APPENDIX F

Spot number	$^{206}\text{Pb}/^{238}\text{U} (\times 10^4)^a$	χ^b	Age (ka) ^c	σ_m	Th/U	U (ppm) ^d
<i>Glass Mountain rhyolite OL south (JS01LV05)^e</i>						
<i>Interior spot analyses</i>						
GMV-r1g1s1	3.27	0.89	1977	38	0.31	4998
GMV-r1g2s1	3.18	0.97	2070	44	0.33	7103
GMV-r1g3s1	3.16	0.99	2099	47	0.50	8889
GMV-r1g4s1	5.52	0.58	2169	71	0.34	6582
GMV-r1g5s1	3.46	0.98	2284	55	0.31	7798
GMV-r1g6s1	3.35	0.98	2205	37	0.40	9597
GMV-r1g7s1	3.04	0.96	1982	77	0.30	5443
GMV-r1g13s1-pol	7.53	0.43	2173	61	0.35	7467
GMV-r1g21s1	3.16	0.99	2101	42	0.37	9493
GMV-r1g23s1	3.05	0.98	2023	50	0.27	5723
GMV-r1g24s1	3.02	0.97	1981	73	0.38	4706
<i>Depth profiles (from <2 μm surface analyses inward)</i>						
GMVI-r1g16s1 (<2 μm surface)	3.03	0.94	1932	54	0.23	3244
s1 (near surface interior)	3.14	0.97	2061	53	0.34	7330
GMVI-r1g17s1 (<2 μm surface)	3.03	0.96	1971	64	0.22	2622
s1 (near surface interior)	3.38	0.99	2227	67	0.59	12746
s2 (near surface interior)	3.08	0.96	1988	64	0.40	10433
GMVI-r1g20s1 (<2 μm surface)	4.17	0.77	2158	50	0.32	6850
s1 (near surface interior)	3.38	0.99	2232	58	0.55	11604
<i>Xenocryst</i>						
GMVI-r1g15s1 (<2 μm surface)	3.62	0.97	2348	90	0.33	8815
s1 (near surface interior)	3.87	0.99	2567	97	0.49	10808
s1 (near surface interior cont.)	4.28	0.99	2825	159	0.49	10304
s2 (near surface interior)	4.49	0.85	2567	87	0.31	9453

Analyses are identified by row (r#), grain (g#), spot (s#), near rim location (“rim”), and sessions in which an additional ~5–10 μm was removed by re-polishing (“pol”).

^a Raw data.

^b Fraction of radiogenic ^{206}Pb based on common $^{207}\text{Pb}/^{206}\text{Pb}$ correction.

^c Ages corrected for initial ^{206}Pb and ^{238}U -series disequilibrium (see text for details).

^d U concentrations are ~5% (1s) based on comparison to measurement of zircon standard 91500.

^e Sample nomenclature after Metz (1987).

REFERENCES

- Bacon C. R., and Lowenstern J. B. (2005) Late Pleistocene granodiorite source for recycled zircon and phenocrysts in rhyodacite lava at Crater Lake, Oregon. *Earth Planet. Sci.* **233**, 277–293.
- Baker J., Peate D., Waight T., and Meyzen C. (2004) Pb isotopic analysis of standards and samples using a ^{207}Pb – ^{204}Pb double spike and thallium to correct for mass bias with a double-focusing MC-ICP-MS. *Chem. Geol.* **211**, 275–303.
- Bindeman I. N., and Valley J. W. (2002) Oxygen isotope study of the Long Valley magma system, California: isotope thermometry

- and convection in large silicic magma bodies. *Contrib. Mineral. Petrol.* **144**, 185–205.
- Charlier B. L. A., Wilson C. J. N., Lowerstern J. B., Blake S., van Calsteren P. W., and Davidson J. P. (2004) Magma Generation at a Large Hyperactive Silicic Volcano (Taupo, New Zealand) Revealed by U–Th and U–Pb Systematics in Zircons. *J. Petrol.* **46**, 3–32.
- Chen J. H., and Tilton G. R. (1991) Applications of lead and strontium isotopic relationships to the petrogenesis of granitoid rocks, central Sierra Nevada batholith, California. *Geol. Soc. Am. Bull.* **103**, 439–447.
- Cherniak D. J., and Watson E. B. (1994) A study of strontium diffusion in plagioclase using Rutherford backscattering spectroscopy. *Geochim. Cosmochim. Acta* **58**, 5179–5190.
- Christensen J. N., and DePaolo D. J. (1993) Time scales of large volume silicic magma systems: Sr isotopic systematics of phenocrysts and glass from the Bishop Tuff, Long Valley, California. *Contrib. Mineral. Petrol.* **113**, 100–114.
- Christensen J. N., and Halliday A. (1996) Rb–Sr ages and Nd isotopic compositions of melt inclusions from the Bishop Tuff and the generation of silicic magma. *Earth Planet. Sci. Lett.* **144**, 547–561.
- Clemens J. D. (1998) Observations on the origins and ascent mechanisms of granitic magmas. *J. Geol. Soc. London* **155**, 843–851.
- Cousens B. L. (1996) Magmatic evolution of Quaternary mafic magmas at Long Valley Caldera and the Devils Postpile, California: effects of crustal contamination on lithospheric mantle-derived magmas. *J. Geophys. Res.* **101**, 27673–27689.
- Dalrymple G. B., Grove M., Lovera O. M., and Harrison T. M. (1999) Age and thermal history of the Geysers plutonic complex (felsite unit), Geysers geothermal field, California: a $^{40}\text{Ar}/^{39}\text{Ar}$ and U–Pb study. *Earth Planet. Sci.* **173**, 285–298.
- Davidson J. P., and Tepley F. J. (1997) Recharge in volcanic systems: evidence from isotope profiles of phenocrysts. *Science* **275**, 826–829.
- Davidson J., Charlier B., Hora J. M., and Perloth R. (2005) Mineral and isotopic fingerprinting: pitfalls and promises. *Geology* **1**, 29–32.
- Davies G. R., and Halliday A. H. (1998) Development of the Long Valley rhyolitic magma system: Strontium and neodymium isotope evidence from glasses and individual phenocrysts. *Geochim. Cosmochim. Acta* **62**, 3561–3574.
- Davies G. R., Halliday A. H., Mahood G. A., and Hall C. M. (1994) Isotopic constraints on the production rates, crystallisation histories and residence times of pre-caldera silicic magmas, Long Valley, California. *Earth Planet. Sci. Lett.* **125**, 17–37.
- Duffield W. A., and Ruiz J. (1998) A model that helps explain Sr-isotope disequilibrium between feldspar phenocrysts and melt in large-volume silicic magma systems. *J. Volcanol. Geotherm. Res.* **87**, 7–13.
- Dunbar N. W., and Hervig R. L. (1992) Petrogenesis and volatile stratigraphy of the Bishop Tuff: evidence from melt inclusion analysis. *J. Geophys. Res.* **97**, 15129–15150.
- Eisele J., Abouchami W., Galer S. J. G., and Hofmann A. W. (2003) The 320 kyr Pb isotope evolution of Mauna Kea lavas recorded in the HSDP-2 drill core. *Geochem. Geophys. Geosystems* **4**, doi:10.1029/2002GC00033.
- Gagnevin D., Daly J. S., Waight T. E., Morgan D., and Poli G. (2005) Pb isotopic zoning of K-feldspar megacrysts determined by Laser Ablation Multi-Collector ICP-MS: insights into granite petrogenesis. *Geochim. Cosmochim. Acta* **7**, 1899–1915.
- Giletti B. J., and Casserly J. E. D. (1994) Strontium diffusion kinetics in plagioclase feldspars. *Geochim. Cosmochim. Acta* **58**, 3785–3793.
- Halliday A. H., Mahood G. A., Holden P., Metz J. M., Dempster T. J., and Davidson J. P. (1989) Evidence for long residence times of rhyolitic magma in the Long Valley magmatic system: the isotopic record in precaldra lavas of Glass Mountain. *Earth Planet. Sci. Lett.* **94**, 274–290.
- Heumann A., and Davies G. R. (1997) Isotopic and chemical evolution of the post-caldera rhyolitic system at Long Valley, California. *J. Petrol.* **38**, 1661–1678.
- Heumann A., Davies G. R., and Elliott T. (2002) Crystallization history of rhyolites at Long Valley, California, inferred from combined U-series and Rb–Sr isotope systematics. *Geochim. Cosmochim. Acta* **66**, 1821–1837.
- Hildreth, W. (1977) The Magma Chamber of the Bishop Tuff: Gradients in Temperature, Pressure, and Composition, Ph.D. Thesis, University of California, Berkeley.
- Hildreth W. (1981) Gradients in silicic magma chambers: implications for lithospheric magmatism. *J. Geophys. Res.* **86**(10), 10153–10192.
- Hildreth W. (2004) Volcanological perspectives on Long Valley, Mammoth Mountain, and Mono Craters: several contiguous but discrete systems. *J. Volcanol. Geotherm. Res.* **136**, 169–198.
- Hildreth W., and Mahood G. A. (1986) Ring-fracture eruption of the Bishop Tuff. *Geol. Soc. Am. Bull.* **97**, 396–403.
- Hirata T., and Nesbitt R. W. (1995) U–Pb geochronology of zircon: evaluation of the laser probe-inductively coupled plasma mass spectrometry technique. *Geochim. Cosmochim. Acta* **59**, 2491–2500.
- Johnson C. M. (1991) Large-scale crust formation and lithosphere modification beneath Middle to Late Cenozoic calderas and volcanic fields, Western North America. *J. Geophys. Res.* **96**, 13485–13507.
- Kalsbeek F. (1992) The statistical distribution of the mean squared weighted deviation—comment: Isochrons, errorchrons, and the use of MSWD-values. *Chem. Geol.* **94**, 241–242.
- Lipman P. W. (1988) Evolution of silicic magma in the upper crust: the mid-Tertiary Latir volcanic field and its cogenetic granitic batholith, northern New Mexico, USA. *Roy. Soc. Edinburgh Trans.* **79**, 265–288.
- Machado N., and Gauthier G. (1996) Determination of $^{207}\text{Pb}/^{206}\text{Pb}$ ages on zircon and monazite by laser-ablation ICPMS and application to a study of sedimentary provenance and metamorphism in southeastern Brazil. *Geochim. Cosmochim. Acta* **60**, 5063–5073.
- Mahon K. I. (1996) The new “York” regression: application of an improved statistical method to geochemistry. *Int. Geol. Rev.* **38**, 293–303.
- Mathez E. A., and Waight T. E. (2003) Lead isotopic disequilibrium between sulfide and plagioclase in the Bushveld Complex and the chemical evolution of large layered intrusions. *Geochim. Cosmochim. Acta* **67**, 1875–1888.
- Metz, J.M. (1987) Physical and chemical evolution of Glass Mountain: precaldra high silica rhyolites from the Long Valley magma system, Ph.D. Thesis, Stanford University.
- Metz J. M., and Mahood G. A. (1985) Precursors to the Bishop Tuff Eruption: Glass Mountain, Long Valley, California. *J. Geophys. Res.* **90**, 11121–11126.
- Metz J. M., and Bailey R. A. (1993) Geologic map of Glass Mountain. In *Miscellaneous Investigations Series I-1995*. Mono country, California.
- Morgan D. J., and Blake S. (2006) Magmatic residence times of zoned phenocrysts: introduction and application of the binary element diffusion modeling (BEDM) technique. *Contrib. Mineral. Petrol.* **151**, 58–70.
- Nielsen D., Mahood G., Halliday A., and Essene E. (1991) Petrogenesis of precaldra mafic and intermediate lavas, Long Valley, CA. *Geol. Soc. Am.*, 83.
- Ormerod, D.S. (1988) Late- to post-subduction magmatic transition in the Western Great Basin, USA, Ph.D. Thesis, The Open University.

- Quidelleur X., Grove M., Lovera O. M., Harrison T. M., Yin A., and Ryerson F. J. (1997) The thermal evolution and slip history of the Renbu Zedong thrust, southeastern Tibet. *J. Geophys. Res.* **102**, 2659–2679.
- Ramos F. C., and Reid M. R. (2005) Distinguishing Melting of Heterogeneous Mantle Sources from Crustal Contamination: insights from Sr Isotopes at the Phenocryst Scale, Pissgah Crater, California. *J. Petrol.* **46**, 999–1012.
- Reid M. R., and Coath C. D. (2000) In situ U–Pb ages of zircons from the Bishop Tuff: no evidence for long crystal residence times. *Geology* **28**, 443–446.
- Reid M. R., and Ramos F. C. (1996) Chemical dynamics of enriched mantle in the southwestern United States: Thorium isotope evidence. *Earth Planet. Sci.* **138**, 67–81.
- Renne, P.R., Simon, J.I., Mundil, R., and Ludwig, K.R. (2006) Implications of pre-eruptive zircon saturation in silicic magmas for high-precision U/Pb geochronology. EOS Trans. AGU, Fall Meet. Suppl. 87, Abstract V21A-0564.
- Schärer U. (1984) The effect of initial ^{230}Th disequilibrium on young U–Pb ages: the Makalu case, Himalaya. *Earth Planet. Sci. Lett.* **67**, 191–204.
- Schmitt A. K., and Simon J. I. (2004) Boron isotopic variations in hydrous rhyolitic melts: a case study from Long Valley, California. *Contrib. Mineral. Petrol.* **146**, 590–605.
- Shaw H. R. (1985) Links between magma-tectonic rate balances, plutonism, and volcanism. *J. Geophys. Res.* **90**, 11275–11288.
- Simkin T. (1993) Terrestrial volcanism in space and time. *Ann. Rev. Earth Planet. Sci.* **21**, 427–452.
- Simon J. I., and Reid M. R. (2005) The pace of rhyolite differentiation and storage in an ‘archetypical’ silicic magma system, Long Valley, California. *Earth Planet. Sci. Lett.* **235**, 123–140.
- Smith, R.L. (1979) Ash-flow magmatism. In Geological Society of America Special Paper 180 (ed. C.E. Chapin and W.E. Elston), pp. 5–42.
- Spera F. J., and Crisp J. A. (1981) Eruption volume, periodicity, and caldera area: relationships and inferences on development of compositional zonation in silicic magma chambers. *J. Volcanol. Geotherm. Res.* **11**, 169–187.
- van den Bogaard P., and Schirnick C. (1995) $^{40}\text{Ar}/^{39}\text{Ar}$ laser probe ages of Bishop Tuff quartz phenocrysts substantiate long-lived silicic magma chamber at Long Valley, United States. *Geology* **23**, 759–762.
- Walder A. J., Abell I. D., Platzner I., and Freedman P. A. (1993) Lead isotope ratio measurements of NIST 610 glass by laser ablation inductively coupled mass spectrometry. *Spectrochimica Acta* **48B**, 397–402.
- Wendt I. (1993) Isochron or mixing line? *Chem. Geol.* **104**, 301–305.
- Willigers B. J., Baker J. A., Krogstad E. J., and Peate D. W. (2002) Precise and accurate in situ Pb–Pb dating of apatite, monazite, and sphene by laser ablation multiple-collector ICP-MS. *Geochim. Cosmochim. Acta* **66**, 1051–1066.
- Wilson C. J. N., and Hildreth W. (1997) The Bishop Tuff: new insights and implications from eruptive stratigraphy. *J. Geol.* **105**, 407–439.
- Wolff J. A., Ramos F. C., and Davidson J. P. (1999) Sr isotope disequilibrium during differentiation of the Bandelier Tuff, New Mexico: constraints on the crystallization of a large rhyolitic magma chamber. *Geology* **27**, 495–498.
- Wolff J. A., and Ramos F. (2003) Pb isotope variations among Bandelier Tuff feldspar: no evidence for a long-lived silicic magma chamber. *Geology* **31**, 533–536.
- Woodhead J. D., and Hergt J. M. (2001) Strontium, Neodymium and Lead isotope analyses of NIST glass certified reference materials: SRM 610, 612, 614. *Geostandard. Newslett.* **25**, 261–266.
- York D. (1969) Least-squares fitting of a straight line with correlated errors. *Earth Planet. Sci. Lett.* **5**, 320–324.
- Young E. D., Ash R. D., Galy A., and Belshaw N. S. (2002) Mg isotope heterogeneity in the Allende meteorite measured by UV laser ablation-MC-ICPMS and comparisons with O isotopes. *Geochim. Cosmochim. Acta* **66**, 683–698.

Associate editor: Yuri Amelin