Oxygen isotopic investigation of silicic magmatism in the Stillwater caldera complex, Nevada: Generation of large-volume, low-δ^{18}O rhyolitic tuffs and assessment of their regional context in the Great Basin of the western United States

Kathryn E. Watts^1,4, David A. John^1, Joseph P. Colgan^2, Christopher D. Henry^3, Ilya N. Bindeman^4, and John W. Valley^5

1U.S. Geological Survey, Menlo Park, California 94025, USA
2U.S. Geological Survey, Denver, Colorado 80225, USA
3Nevada Bureau of Mines and Geology, Reno, Nevada 89557, USA
4Department of Geoscience, University of Wisconsin–Madison, Madison, Wisconsin 53706, USA
5Department of Earth Sciences, University of Oregon, Eugene, Oregon 97403, USA

© 2019 The Authors. Gold Open Access: This paper is published under the terms of the CC-BY license.

GSA Bulletin; July/August 2019; v. 131; no. 7/8; p. 1133–1156; https://doi.org/10.1130/B35021.1; 12 figures; 4 tables; Data Repository item 2019060; published online 14 February 2019.

successive caldera-forming eruptions from ca. 30 to 25 Ma generated a large nested caldera complex in western Nevada that was subsequently dissected by Basin and Range extension, providing extraordinary cross-sectional views through diverse volcanic and plutonic rocks. A high-resolution oxygen isotopic study was conducted on units that represent all major parts of the Job Canyon, Louderback Mountains, Poco Canyon, and Elevenmile Canyon caldera cycles (29.2–25.1 Ma), and several Cretaceous plutons that flank the Stillwater caldera complex. We provide new oxygen and strontium isotopic data for 12 additional caldera centers spanning 1133–1156 km² and >150 published oxygen and strontium isotope analyses for regional Mesozoic basaltic and andesitic rocks. Stillwater zircons span a large isotopic range (δ^{18}O_\text{zircon} = 3.6‰–8.2‰), and all caldera cycles possess low-δ^{18}O zircons. In some cases, they are a small proportion of the total populations, and in others, they dominate, such as in the low-δ^{18}O rhyolitic tuffs of Job Canyon and Poco Canyon (δ^{18}O_\text{zircon} = 4.0‰–4.3‰; δ^{18}O_\text{magma} = 5.5‰–6‰). These are the first low-δ^{18}O rhyolites documented in middle Cenozoic calderas of the Great Basin, adding to the global occurrence of these important magma types that fingerprint recycling of shallow crust altered by low-δ^{18}O meteoric waters. The appearance of low-δ^{18}O rhyolites in the Stillwater caldera complex is overprinted on a Great Basin–wide trend of miogeoclinal sediment contribution to silicic magmas that elevates δ^{18}O compositions, making identification of δ^{18}O depletions difficult. Though not a nominally low-δ^{18}O rhyolite, the tuff of Elevenmile Canyon possesses both low-δ^{18}O and high-δ^{18}O zircon cores that are overgrown by homogenized zircon rims that approximate the bulk zircon average, pointing to batch assembly of isotopically diverse upper crustal melts to generate one of the most voluminous (2500–5000 km³) tuff eruptions in the Great Basin. Despite overlapping in space and time, each caldera-forming cycle of the Stillwater complex has a unique oxygen isotope record as retained in single zircons. Most plutons that were spatially and temporally coincident with calderas have isotopic compositions that diverge from the caldera-forming tuffs and cannot be their cogeneric remnants.

INTRODUCTION

During the middle Cenozoic, tens of thousands of cubic kilometers of silicic magma erupted from dozens of calderas that span across Nevada in a broad northwest to southeast belt (Fig. 1; Best et al., 2013a). This ignimbrite flare-up may have been related to steepening, or roll-back, of the subducting Farallon plate beneath the western edge of North America (Coney and Reynolds, 1977; Coney, 1978; Humphreys, 1995). Although most calderas occur within intact crustal blocks, some were dissected and significantly tilted by Basin and Range extension during the late Cenozoic, providing valuable windows into the volcanic and plutonic processes that underpinned them, and perhaps by analogue, the life spans and eruptive potential of modern-day caldera volcanoes.

Caldera exposures are exemplified by the Stillwater–Clan Alpine complex in western Nevada (Figs. 1–2; abbreviated here and elsewhere as the Stillwater caldera complex). The western margin of the complex in the Stillwater Range is steeply tilted, exposing an ~10 km deep section through the upper crust, including several overlapping calderas and plutons that provide a unique opportunity to elucidate the petrogenesis of large-volume silicic magmas in the Great Basin ignimbrite flare-up. Here, we present the results of a comprehensive oxygen isotopic investigation, with particular emphasis on refractory zircon crystals, which demonstrate significant diversity in Stillwater magmas, including two low-δ^{18}O rhyolitic tuffs (Job Canyon and Poco Canyon; δ^{18}O_\text{magma} = 5.5‰–6‰). To our knowledge, these are the first low-δ^{18}O rhyolites described in the Great Basin ignimbrite flare-up. The low-δ^{18}O Ammonia Tanks Tuff (δ^{18}O_\text{magma} = 5.4‰–6‰) in
Regional granites (King et al., 2004)
- Mountain range sampled
- Sample locality number

- Cenozoic
- Cretaceous
- Jurassic
- Precambrian

Regional granites (Wooden et al., 1999)
- Cenozoic
- Cretaceous
- Jurassic

Regional granites (this study, Cretaceous)
- Sample locality

AC: Alameda Canyon (84 Ma)
LC: La Plata Canyon (87 Ma)
SS: Sand Springs pluton (89 Ma)

Humboldt mafic complex (Klister and Speed, 2000)
- Gabbro, basalt sample locality (172 Ma)

Regional calderas and their associated tuffs (this study, Eocene-Miocene)

AD: tuff of Arc Dome, 25.2 Ma
CC: tuff of Campbell Creek, 25.2 Ma
CM: tuff of Cove Mine, 34.4 Ma
CT: Cañada Tuff, 34.0 Ma
DC: tuff of Deep Canyon, 30.4 Ma
EC: tuff of Elevenmile Canyon, 25.1 Ma

FCM: Fish Creek Mountains Tuff, 24.9 Ma
FP: tuff of Fairview Peak, 19.5 Ma
HC: tuff of Hall Creek, 34.0 Ma
JC: tuff of Job Canyon, 29.2 Ma
LB: tuff of Louderback Mountains, 25.2 Ma
MH: Manhattan, Round Rock Formation, 24.8 Ma

MUL: lower tuff of Mount Jefferson, 27.3 Ma
MLU: upper tuff of Mount Jefferson, 27.8 Ma
NH: Nine Hill Tuff, 25.4 Ma
PC: tuff of Poco Canyon, 25.2 Ma
UD: Undertown Tuff, 25.0 Ma

Figure 1. Map showing the location of the Stillwater caldera complex and other middle Cenozoic calderas in the Great Basin, Nevada. Calderas are subdivided into the western Nevada, central Nevada, and Indian Peak–Caliente volcanic fields (Best et al., 2013a). Dashed pale-gray lines show the boundaries between volcanic fields, black solid lines show the locations of the $^{87}\text{Sr}/^{86}\text{Sr}_1 = 0.706$ and $^{87}\text{Sr}/^{86}\text{Sr}_2 = 0.708$ isopleths (Farmer and DePaolo, 1983), and dotted black lines show the paleodivide boundaries of Henry and John (2013) (“PD1”) and Best et al. (2013a) (“PD2”). The mountain ranges sampled by King et al. (2004) are shaded in gray and numbered according to their numbering scheme. Labels and ages for the calderas and granites included in this study are shown in the figure legend. Labels are included for two Cretaceous granite localities to the north of the Stillwater caldera complex, the New York Canyon (NYC) stock and the Rocky Canyon (RC) pluton, which may be pertinent to the basement crustal architecture (see text for detail). Stars show town localities in Nevada.
Figure 2. Simplified geologic map and cross section of the Stillwater caldera complex, modified after Colgan et al. (2018). (A) Approximate caldera margins for seven overlapping calderas are shown by the bold dashed lines; thin bold lines show faults, some of which bound the caldera margins. Distributions of Mesozoic basement rocks and rocks of the Stillwater caldera complex are shown by the filled polygons and legend. (B) Pretilt north-south cross section through three overlapping calderas in the Stillwater Range: the Job Canyon, Poco Canyon, and Elevenmile Canyon calderas.
southern Nevada (Fig. 1) was produced during the much younger regime of late Cenozoic Basin and Range extensional tectonics and magmatism (Bindeman and Valley, 2003). Low-δ¹⁸O rhyolites fingerprint hydrothermally altered crustal materials that have exchanged oxygen with low-δ¹⁸O meteoric waters (Taylor, 1986), thus revealing a shallow process of crustal recycling heretofore not recognized in the Great Basin ignimbrite flare-up.

In addition to providing important new constraints on Cenozoic silicic magmatism in the Great Basin, this study adds to the occurrence of low-δ¹⁸O rhyolites worldwide that signify an important part of the geochemical evolution of Earth’s continental crust as it interacts with the hydrosphere. Low-δ¹⁸O rhyolites have been documented in huge volumes (>10,000 km³) to the north of the Great Basin in caldera centers of the Snake River Plain—Yellowstone hotspot track (Hildreth et al., 1984; Bindeman and Valley, 2001; Boroughs et al., 2005; Bindeman et al., 2008; Ellis et al., 2010; Watts et al., 2011, 2012; Drew et al., 2013; Colón et al., 2015; Loewen and Bindeman, 2015; Blum et al., 2016; Troch et al., 2017). Other settings in which they have been documented include present or former rift environments, such as Iceland (Bindeman et al., 2012; Pope et al., 2013) and South Africa (Harris and Erkland, 1992), and modern subduction zone environments such as the Kamchatka Peninsula in Russia (Bindeman et al., 2004; Seligman et al., 2014), the southern Andes in South America (Grunder, 1987), and Crater Lake in the Oregon Cascades (Bacon et al., 1989; Ankney et al., 2017). Despite this ubiquity, the importance of low-δ¹⁸O rhyolites to our understanding of crustal magmatism likely remains underestimated due to the dearth of oxygen isotope data needed to identify them, and the propensity for such shallow rocks to be stripped from the geologic record. Geographic location also has a profound effect on the presence and magnitude of δ¹⁸O depletion of meteoric waters, and therefore, regardless of tectonic setting and mechanism, the processes associated with low-δ¹⁸O rhyolite genesis may be obscure in many regions, such as in near-equatorial latitudes or arid environments (e.g., Folkes et al., 2013).

The Great Basin, which possesses a world-class endowment of hydrothermally altered and mineralized crustal basement rocks as well as dozens of overlapping calderas centers with hydrothermal alteration (e.g., Best et al., 2013a; Henry and John, 2013), is a prime setting in which low-δ¹⁸O rhyolites would be expected to occur. The Stillwater caldera complex in particular, with caldera-forming tuffs, plutons, and an ancient hydrothermal system exposed to ~10 km depth (John, 1995; John and Pickthorn, 1996), may be a critical locality for evaluating processes of low-δ¹⁸O rhyolite genesis in the Great Basin and worldwide.

GEOLOGIC BACKGROUND

The western margin of the North American continent consists of three major crustal zones in the Great Basin, from Precambrian cratonic basement in eastern Nevada (⁸⁷Sr/⁸⁶Sr > 0.708), to transitional rifted crust and sedimentary rocks of the miogeocline in central Nevada (⁸⁷Sr/⁸⁶Sr ~0.706–0.708), to accreted oceanic terranes in western Nevada (⁸⁷Sr/⁸⁶Sr < 0.706) (Fig. 1; Farmer and DePaolo, 1983). Tectonic interaction between the continental margin and accreted terranes has resulted in a complex juxtaposition of Paleozoic–Mesozoic lithologic assemblages (cf. Crafford, 2008). Overprinted on this complex basement geology are three major pulses of silicic igneous activity, manifested by Jurassic and Cretaceous granitic plutons that span across the state of Nevada and a broad (~400 km) belt of middle Cenozoic calderas (Fig. 1).

The western Nevada, central Nevada, and the eastern Nevada (Grunder, 1987), and Crater Lake in the Oregon Cascades (Bacon et al., 1989; Ankney et al., 2017). Despite this ubiquity, the importance of low-δ¹⁸O rhyolites to our understanding of crustal magmatism likely remains underestimated due to the dearth of oxygen isotope data needed to identify them, and the propensity for such shallow rocks to be stripped from the geologic record. Geographic location also has a profound effect on the presence and magnitude of δ¹⁸O depletion of meteoric waters, and therefore, regardless of tectonic setting and mechanism, the processes associated with low-δ¹⁸O rhyolite genesis may be obscure in many regions, such as in near-equatorial latitudes or arid environments (e.g., Folkes et al., 2013).

The Great Basin, which possesses a world-class endowment of hydrothermally altered and mineralized crustal basement rocks as well as dozens of overlapping calderas centers with hydrothermal alteration (e.g., Best et al., 2013a; Henry and John, 2013), is a prime setting in which low-δ¹⁸O rhyolites would be expected to occur. The Stillwater caldera complex in particular, with caldera-forming tuffs, plutons, and an ancient hydrothermal system exposed to ~10 km depth (John, 1995; John and Pickthorn, 1996), may be a critical locality for evaluating processes of low-δ¹⁸O rhyolite genesis in the Great Basin and worldwide.

GEOLOGIC BACKGROUND

The western margin of the North American continent consists of three major crustal zones in the Great Basin, from Precambrian cratonic basement in eastern Nevada (⁸⁷Sr/⁸⁶Sr > 0.708), to transitional rifted crust and sedimentary rocks of the miogeocline in central Nevada (⁸⁷Sr/⁸⁶Sr ~0.706–0.708), to accreted oceanic terranes in western Nevada (⁸⁷Sr/⁸⁶Sr < 0.706) (Fig. 1; Farmer and DePaolo, 1983). Tectonic interaction between the continental margin and accreted terranes has resulted in a complex juxtaposition of Paleozoic–Mesozoic lithologic assemblages (cf. Crafford, 2008). Overprinted on this complex basement geology are three major pulses of silicic igneous activity, manifested by Jurassic and Cretaceous granitic plutons that span across the state of Nevada and a broad (~400 km) belt of middle Cenozoic calderas (Fig. 1).

The western Nevada, central Nevada, and the eastern Nevada (Grunder, 1987), and Crater Lake in the Oregon Cascades (Bacon et al., 1989; Ankney et al., 2017). Despite this ubiquity, the importance of low-δ¹⁸O rhyolites to our understanding of crustal magmatism likely remains underestimated due to the dearth of oxygen isotope data needed to identify them, and the propensity for such shallow rocks to be stripped from the geologic record. Geographic location also has a profound effect on the presence and magnitude of δ¹⁸O depletion of meteoric waters, and therefore, regardless of tectonic setting and mechanism, the processes associated with low-δ¹⁸O rhyolite genesis may be obscure in many regions, such as in near-equatorial latitudes or arid environments (e.g., Folkes et al., 2013).

The Great Basin, which possesses a world-class endowment of hydrothermally altered and mineralized crustal basement rocks as well as dozens of overlapping calderas centers with hydrothermal alteration (e.g., Best et al., 2013a; Henry and John, 2013), is a prime setting in which low-δ¹⁸O rhyolites would be expected to occur. The Stillwater caldera complex in particular, with caldera-forming tuffs, plutons, and an ancient hydrothermal system exposed to ~10 km depth (John, 1995; John and Pickthorn, 1996), may be a critical locality for evaluating processes of low-δ¹⁸O rhyolite genesis in the Great Basin and worldwide.
Oxygen isotopic investigation of silicic magmatism in the Stillwater caldera complex, Nevada

>150 km west and ~60 km east of the caldera (John, 1995). The caldera is underlain by altered, sparsely porphyritic, flow-banded rhyolite lava flows, domes, and intrusive rocks as much as ~1.5 km thick that are locally intercalated with and overlain by silicic welded tuffs, andesites, and minor sedimentary rocks.

Elevenmile Canyon Cycle

Rocks of the Elevenmile Canyon caldera overlie the Poco Canyon and Louderback Mountains calderas, outflow tuff of Poco Canyon, and sparsely porphyritic flow-banded rhyolite lava flows, domes, and intrusive rocks. The trachydacitic to rhyolitic (~65–77 wt% SiO₂) tuff of Elevenmile Canyon is up to ~5 km thick in the Stillwater Range and ~3–4 km thick in the southern Clan Alps Mountains. Outflow tuff extends ~60 km to the east and ~125 km to the west (Henry and John, 2013).

Circa 25 Ma Intrusive Rocks

The Poco Canyon and Elevenmile Canyon calderas in the Stillwater Range are underlain and possibly intruded by the ca. 25 Ma Freeman Creek pluton, which consists of an older granodiorite (~65–67 wt% SiO₂) phase and a younger granite (~72–77 wt% SiO₂) phase. Where exposed, the Freeman Creek pluton intrudes rhyolite lavas and unaltered volcanic rocks that underlie the Poco Canyon and Elevenmile Canyon calderas; it does not directly intrude the tuffs, although it may do so elsewhere in the subsurface. An ~7-km-long composite granite and rhyolite porphyry dike intrudes the IXL pluton, separating the Job Canyon and Poco Canyon calderas (John, 1995). At Chalk Mountain, rhyolite porphyry and granodiorite plutons (ca. 25 Ma) underlie the Louderback Mountains and Elevenmile Canyon calderas (Colgan et al., 2018). Where exposed, the granodiorite intrudes Mesozoic dolomite, but the area between Chalk Mountain and exposures of the Elevenmile Canyon caldera is covered by alluvium and it is not known if it directly intrudes the intracaldera tuff.

SAMPLES AND METHODS

Oxygen isotopic data were collected for 16 caldera centers in the western Nevada volcanic field (labeled in Fig. 1; tabulated in Table 1). Data collected for the Stillwater complex represent four caldera cycles (Job Canyon, Louderback Mountains, Poco Canyon, and Elevenmile Canyon) and three Cretaceous granitic plutons (Alameda Canyon, La Plata Canyon, and Sand Springs). The University of Oregon Stable Isotope Laboratory was used to analyze a variety of mineral phases using the established protocols of Bindeman (2008) for CO₂-laser fluorination. Refractory quartz (single to several crystals per analysis) and bulk zircon were the primary targets, with additional analyses of sanidine, plagioclase, and biotite. Minerals were analyzed in 1.5–2.5 mg aliquots under vacuum with a CO₂ laser and BrF₃ reagent to liberate O₂ gas, which was then purified with a mercury diffusion pump to remove traces of fluorine gas, and converted to CO₂ in a platinum-graphite converter. For reactive sanidine, an airlock chamber was used between analyses to prevent reaction of grains with BrF₃ prior to lasing. A Thermo Finnigan MAT 253 mass spectrometer was used to analyze the oxygen isotope composition of the CO₂ gas, and δ¹⁸O values are reported relative to Vienna standard mean ocean water (VSMOW). A University of Oregon garnet standard (UOG; δ¹⁸O = 6.52‰, VSMOW) was used to correct δ¹⁸O values in unknowns; UOG analyses ranged from −0.1‰ to −0.6‰ relative to the empirical standard value and had an external reproducibility (2 standard deviations) of 0.1‰–0.2‰. The UOG standard is calibrated to the University of Wisconsin garnet standard (UWG-2; δ¹⁸O = 5.80‰, VSMOW).

Strontium isotope data for three of the four calderas investigated in the Stillwater caldera complex (Job Canyon, Poco Canyon, and Elevenmile Canyon) and the three Cretaceous granitic plutons (Alameda Canyon, La Plata Canyon, and Sand Springs). Whole-rock powders were analyzed for their strontium isotopic compositions using a ThermoFinnigan TRITON thermal ionization multicollector mass spectrometer at Carleton University in Ottawa, Canada. Strontium isotopic ratios were corrected for radiogenic ingrowth using published ⁴⁰Ar/³⁹Ar ages, or when not available, ²⁶⁹Pb/²³⁶U ages (Table 1). Previously published strontium isotope data were corrected for the most up-to-date age data tabulated in Table 1.

Secondary ion mass spectrometry (SIMS) oxygen isotopic data were collected for single zircon crystals from 22 samples that represent all major units of the Job Canyon, Poco Canyon, Elevenmile Canyon, and Louderback Mountains caldera cycles, and one sample of the Cretaceous Sand Springs pluton. Prior to SIMS analysis, zircons were separated from crushed whole rocks using a water table, Frantz magnetic separator, and heavy liquids; handpicked under a binocular microscope; mounted in epoxy rounds within a 10 mm diameter in the center of each mount; and ground down to expose the approximate midsections of grain interiors. Zircon standards (KIM-5, R33, MADDER) were included in the center of each mount. A Tescan VEGA3 scanning electron microscope (SEM) at the U.S. Geological Survey (USGS) Menlo Park microanalytical facility was used to map zircon crystals using backscattered electrons (BSE) and cathodoluminescence (CL) to document internal textures and enable precise targeting of core, interior, and rim domains.

Oxygen isotope ratios were analyzed in situ at the WiscSIMS Laboratory at the University of Wisconsin, Madison, using a CAMECA ims 1280 ion microprobe. The WiscSIMS ion microprobe was operated with a ¹³³Cs+ beam current of ~2 nA, accelerating potential of 10 kV (20 kV impact energy), and tuned to an ~15 μm beam spot diameter (~1 μm deep) on the sample (Kita et al., 2009; Valley and Kita, 2009). Secondary ions of ⁶⁰⁸O, ⁶⁰⁹O, and ⁶¹⁰O were analyzed simultaneously using three Faraday cup detectors. The ion yield for ⁶¹⁰O was ~1.4 Gcps/nA. Ratios of ⁶¹⁰O/⁶⁰⁸O were background corrected by comparison to the nominally anhydrous KIM-5 zircon (Wang et al., 2014). Analyses of KIM-5 zircon (δ¹⁸O = 5.09‰, VSMOW; Valley, 2003) bracketed each set of unknowns, with ~4–5 KIM-5 analyses before and after each set of ~15 unknown spots. Values of instrumental mass fractionation based on KIM-5 varied from ~1.2‰ to ~0.3‰, and the external reproducibility (2 standard deviations) during the analytical session ranged from 0.1‰ to 0.4‰, averaging 0.2‰.

A subset of zircons analyzed for oxygen isotopes using the WiscSIMS ion microprobe was analyzed for ²⁶⁹Pb/²³⁶U ages and trace elements using a sensitive high-resolution ion microprobe with reverse geometry (SHRIMP-RG) operated by the USGS and Stanford University. Zircon mounts were re-imaged by SEM after SIMS analysis to evaluate each analytical spot and to map their precise locations. After polishing away the WiscSIMS pits, the grain domains just beneath each SIMS δ¹⁸O spot were analyzed with the SHRIMP-RG, using a slightly larger diameter ion beam (~20 μm). Zircon U-Pb ages were standardized against R33 (419.3 Ma; Black et al., 2004) and trace elements against MAD-green zircon (Barth and Wooden, 2010). For a full description of the SHRIMP-RG analytical methods used, see Colgan et al. (2018).

Hornblende geochemistry was determined by electron microprobe for the tuff of Elevenmile Canyon in order to constrain the crystalization pressures of Stillwater magmas; hornblende is either absent or highly altered in tuffs of the other major caldera-forming cycles in the Stillwater caldera complex. Polished thin
<table>
<thead>
<tr>
<th>Caldera</th>
<th>Type</th>
<th>Unit</th>
<th>Map ID</th>
<th>Sample(s)</th>
<th>Latitude</th>
<th>Longitude</th>
<th>$^{40}$Ar/$^{39}$Ar age (Ma)</th>
<th>$^{206}$Pb/$^{238}$U age (Ma)</th>
<th>$\delta^{18}$O average (‰, VSMOW)</th>
<th>Quartz</th>
<th>Zircon</th>
<th>Magma $^{1}$</th>
<th>$^{87}$Sr/$^{86}$Sr $^{2}$</th>
<th>Note</th>
</tr>
</thead>
<tbody>
<tr>
<td>Caetano</td>
<td>Single</td>
<td>Caetano Tuff</td>
<td>CT</td>
<td>Many, see Watts et al. (2016)</td>
<td>40.16</td>
<td>-116.82</td>
<td>34.00 ± 0.03 (14)</td>
<td>34.23–34.46 ± 0.2–0.3 (169)</td>
<td>10.7 ± 0.3 (25)</td>
<td>8.7 ± 0.3 (74)</td>
<td>10.2–10.3</td>
<td>0.7068–0.7072</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hall Creek</td>
<td>Single</td>
<td>Tuff of Hall Creek</td>
<td>HC</td>
<td>H11-167, H11-22</td>
<td>39.77</td>
<td>-116.82</td>
<td>33.92 ± 0.02 (4)</td>
<td>34.06 ± 0.40 (23)</td>
<td>10.4 ± 0.3 (4)</td>
<td>-</td>
<td>10.0</td>
<td>--</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td>Cove Mine</td>
<td>Single</td>
<td>Tuff of Cove Mine</td>
<td>CM</td>
<td>H03-87, 05-DJ-5, 06-DJ-13</td>
<td>40.45</td>
<td>-117.02</td>
<td>34.44 ± 0.01 (4)</td>
<td>10.1 ± 0.4 (7)</td>
<td>9.7 ± 0.6 (5)</td>
<td>-</td>
<td>8.7</td>
<td>0.7067–0.7061</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fish Creek Mountains</td>
<td>Single</td>
<td>Fish Creek Mountains Tuff</td>
<td>FCM</td>
<td>FCM-2, FCM-5</td>
<td>40.19</td>
<td>-117.28</td>
<td>24.91 ± 0.03 (3)</td>
<td>25.02 ± 0.38 (6)</td>
<td>9.1 ± 0.2 (5)</td>
<td>-</td>
<td>8.7</td>
<td>0.7058–0.7061</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mount Jefferson</td>
<td>Nested</td>
<td>Lower tuff of Mount Jefferson</td>
<td>MJL</td>
<td>H94-28</td>
<td>38.77</td>
<td>-116.90</td>
<td>27.25 ± 0.04 (17)</td>
<td>-</td>
<td>9.7 ± 0.1 (2)</td>
<td>-</td>
<td>9.3</td>
<td>--</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td>Mount Jefferson</td>
<td>Nested</td>
<td>Upper tuff of Mount Jefferson</td>
<td>H94-4</td>
<td>-</td>
<td>38.77</td>
<td>-116.90</td>
<td>27.01 ± 0.05 (4)</td>
<td>-</td>
<td>8.9 ± 0.2 (2)</td>
<td>-</td>
<td>8.5</td>
<td>--</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td>Manhattan</td>
<td>Nested?</td>
<td>Diamond King Formation</td>
<td>MH</td>
<td>H94-43</td>
<td>38.58</td>
<td>-117.03</td>
<td>24.66 ± 0.07 (1)</td>
<td>-</td>
<td>8.4 ± 0.1 (2)</td>
<td>-</td>
<td>8.0</td>
<td>--</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td>Arc Dome</td>
<td>Single</td>
<td>Tuff of Arc Dome</td>
<td>AD</td>
<td>17-AD-1</td>
<td>39.00</td>
<td>-117.42</td>
<td>25.15 ± 0.02 (5)</td>
<td>-</td>
<td>8.8 ± 0.3 (2)</td>
<td>-</td>
<td>8.4</td>
<td>--</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td>Underdown</td>
<td>Single</td>
<td>Underdown Tuff</td>
<td>UD</td>
<td>JC13-5</td>
<td>39.14</td>
<td>-117.44</td>
<td>24.91 ± 0.05 (4)</td>
<td>25.32 ± 0.21 (21)</td>
<td>9.0 ± 0.1 (2)</td>
<td>-</td>
<td>8.6</td>
<td>--</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td>Campbell Creek</td>
<td>Nested?</td>
<td>Tuff of Campbell Creek</td>
<td>CC</td>
<td>H02-24</td>
<td>39.36</td>
<td>-117.68</td>
<td>28.94 ± 0.06 (24)</td>
<td>8.8 ± 0.5 (2)</td>
<td>-</td>
<td>-</td>
<td>8.4</td>
<td>0.7060</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td>Nine Hill</td>
<td>Nested?</td>
<td>Nine Hill Tuff</td>
<td>NH</td>
<td>H03-138</td>
<td>39.50</td>
<td>-117.94</td>
<td>25.38 ± 0.08 (12)</td>
<td>-</td>
<td>8.6 ± 0.1 (2)</td>
<td>-</td>
<td>8.2</td>
<td>0.7054</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td>Fairview Peak</td>
<td>Single</td>
<td>Tuff of Fairview Peak</td>
<td>FP</td>
<td>H03-58</td>
<td>39.20</td>
<td>-118.09</td>
<td>19.48 ± 0.02 (3)</td>
<td>19.65 ± 0.22 (21)</td>
<td>7.9 ± 0.2 (2)</td>
<td>-</td>
<td>7.5</td>
<td>0.7050</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td>Elevenmile Canyon</td>
<td>Nested</td>
<td>Tuff of Elevenmile Canyon</td>
<td>EC</td>
<td>Many, see Table 2</td>
<td>39.44</td>
<td>-117.98</td>
<td>25.11 ± 0.03 (28)</td>
<td>25.00–25.82 ± 0.2–0.3 (113)</td>
<td>7.9 ± 0.3 (7)</td>
<td>5.9–6.1 ± 0.1–0.5 (73/112)*</td>
<td>7.4–7.6</td>
<td>0.7051–0.7054</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>Poco Canyon</td>
<td>Nested</td>
<td>Tuff of Poco Canyon</td>
<td>PC</td>
<td>Many, see Table 2</td>
<td>39.56</td>
<td>-118.25</td>
<td>25.27 ± 0.03 (8)</td>
<td>25.60–25.99 ± 0.2–0.3 (48)</td>
<td>6.5 ± 0.3 (4)</td>
<td>4.0–4.3 ± 0.2 (59/71)*</td>
<td>5.5–5.8</td>
<td>0.7049–0.7055</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>Louderback Mountains</td>
<td>Nested</td>
<td>Tuff of Louderback Mountains</td>
<td>LB</td>
<td>13-DJ-7, see Table 2</td>
<td>39.39</td>
<td>-118.00</td>
<td>25.20 ± 0.02 (2)</td>
<td>25.29–25.78 ± 0.3–0.5 (38)</td>
<td>-</td>
<td>7.1–7.2 ± 0.1–0.2 (1728)*</td>
<td>8.6–8.7</td>
<td>--</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td>Job Canyon</td>
<td>Nested</td>
<td>Tuff of Job Canyon</td>
<td>JC</td>
<td>10-DJ-2, see Table 2</td>
<td>39.63</td>
<td>-118.26</td>
<td>29.25 ± 0.5 (17)</td>
<td>-</td>
<td>4.5 ± 0.3 (19/20)*</td>
<td>-</td>
<td>6.0</td>
<td>0.7052–0.7057</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>Basement locality</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Alameda Canyon</td>
<td>--</td>
<td>Alameda Canyon pluton</td>
<td>AC</td>
<td>JC11-25, 16-DJ-31</td>
<td>39.72</td>
<td>-118.20</td>
<td>--</td>
<td>83.92 ± 0.83 (11)</td>
<td>11.5 ± 0.1 (3)</td>
<td>-</td>
<td>11.1</td>
<td>0.7050</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td>La Plata Canyon</td>
<td>--</td>
<td>La Plata Canyon pluton</td>
<td>LC</td>
<td>JC09-LC4, 16-DJ-34</td>
<td>39.43</td>
<td>-118.30</td>
<td>--</td>
<td>87.25 ± 0.50 (12)</td>
<td>10.4 ± 0.1 (3)</td>
<td>-</td>
<td>10.0</td>
<td>0.7051</td>
<td>--</td>
<td></td>
</tr>
<tr>
<td>Sand Springs</td>
<td>--</td>
<td>Sand Springs pluton</td>
<td>SS</td>
<td>JC11-27, 16-DJ-38</td>
<td>39.33</td>
<td>-118.33</td>
<td>--</td>
<td>88.6 ± 3.1 (13)</td>
<td>9.8 ± 0.2 (3)</td>
<td>6.6 ± 0.2 (20)</td>
<td>8.1–9.4</td>
<td>0.7043–0.7047</td>
<td>--</td>
<td>--</td>
</tr>
</tbody>
</table>

Note: Latitude and longitude are the approximate centers of individual calderas. Italics indicate uncertain caldera location. $^{40}$Ar/$^{39}$Ar sanidine ages were compiled from John et al. (2008), Henry and John (2013), Watts et al. (2016), and Colgan et al. (2018); uncertainties quoted are 2σ; $n$—number of samples. $^{206}$Pb/$^{238}$U zircon ages were compiled from Watts et al. (2016), Colgan et al. (2018), and this study; uncertainties quoted are 95% confidence intervals; $n$—number of analyses. O isotope data were determined by laser fluorination for quartz and secondary ion mass spectrometry (SIMS) for zircon; uncertainties quoted are 1 standard deviation of combined analyses; $n$—number of analyses. VSMOW—Vienna standard mean ocean water.

$^{*}$Multimodal $\delta^{18}$O spectra; see Figure 5 and Supplementary Data Table DR2. Sr isotope data were compiled from John (1995), Varve (2013), Watts et al. (2016), and this study. John (1995) data were recalculated for the most up-to-date age data shown in the table. $^{*}$—indicates no data.

$^{1}$Magmas $\delta^{18}$O values were calculated based on average $\Delta^{18}O_{quartz-melt} = 0.4$‰ and $\Delta^{18}O_{melt-zircon} = 1.5$‰ (fractionation factors from Trail et al. [2009] for 750–850°C: $\Delta^{18}O_{quartz-melt} = 0.4$‰–0.5‰ and $\Delta^{18}O_{melt-zircon} = 1.4$‰–1.7‰).
sections of five samples of the tuff of Elevenmile Canyon were analyzed with a JEOL 8900 electron microprobe at the USGS Menlo Park microanalytical facility. Mapping of individual hornblende crystals with a petrographic microscope and SEM enabled targeting of core and rim domains. The electron microprobe was operated at a 15 kV accelerating voltage, with a 10 nA defocused (5-µm-diameter) beam. A series of USGS mineral and major-oxide standards was used to monitor instrument drift and correct compositional offsets. Data for two additional samples of the tuff of Elevenmile Canyon, published in the M.S. thesis of Stepner (2017), are synthesized with our new data.

RESULTS

Oxygen and Strontium Isotope Data for Caldera-Forming Tuffs in the Western Nevada Volcanic Field in the Context of Regional Basement

Magma \(\delta^{18}O\) values were calculated from quartz or zircon analyses for 16 caldera-forming tuffs in the western Nevada volcanic field. Units for which we have both quartz and zircon data indicate consistency in calculated magmatic (melt) values using fractionation factors from Trail et al. (2009) for \(\delta^{18}O_{\text{quartz}} = 0.4\%e\) and \(\delta^{18}O_{\text{zircon}} = 1.5\%e\) (Table 1). Calculated magmatic \(\delta^{18}O\) values for the tuffs span from ~10%e to 5.5‰ across a longitudinal profile from ~118.2°W to ~118.3°W (Figs. 1 and 3; Table 1). Whole-rock \(\delta^{87}Sr/\delta^{86}Sr\) ratios for the same units were included for nine of these tuffs and span from ~0.707 to 0.705. These new data were plotted with available oxygen isotope data for regional granites (and strontium, where it exists for equivalent units) across the state of Nevada (Figs. 1 and 3; Supplementary Data Table DR1). Most of these regional granites are Cretaceous or Jurassic in age, with magmatic \(\delta^{18}O\) values calculated from quartz and zircon analyses (King et al., 2004), and fewer approximated from whole-rock \(\delta^{18}O\) analyses (Wooden et al., 1999). Magmatic \(\delta^{18}O\) values approximated from whole-rock analyses may be affected by alteration, whereas those calculated from zircon are free of these effects (e.g., Lackey et al., 2008).

A clearly defined gradient in both oxygen and strontium isotopes is apparent as the longitudinal profile steps west from the ~0.708 Sr isopleth, which marks the western boundary of the Precambrian craton in eastern Nevada, across a crustal transition zone, and into accreted arc terranes at the ~0.706 Sr isopleth (Fig. 3). The oxygen isotope gradient is broadly apparent in the regional Mesozoic granites, as first recognized by Solomon and Taylor (1989) and later refined by King et al. (2004), who concluded that granites east of the 0.706 Sr isopleth have a crustal signature indicative of high-\(\delta^{18}O\) sedimentary components derived from the continental margin, with the edge at approximately the 0.708 Sr isopleth. For the middle Cenozoic calderas that are the focus of this investigation, the gradient closely mirrors that of the Mesozoic granitic basement, and it is bracketed by the high-\(\delta^{18}O\) Caetano Tuff in central Nevada (\(\delta^{18}O_{\text{magmas}} = 10.2\%e\); \(\delta^{87}Sr/\delta^{86}Sr = 0.7068–0.7072\) and two low-\(\delta^{18}O\) rhyolites of the Stillwater caldera complex in western Nevada, the tuffs of Poco Canyon and Job Canyon (\(\delta^{18}O_{\text{magmas}} = 5.5\%e–6.6\%e\); \(\delta^{87}Sr/\delta^{86}Sr = 0.7049–0.7057\)) (Figs. 1 and 3).

Nestcd calderas of the Stillwater complex have highly diverse magmatic \(\delta^{18}O\) values; tuffs span from 5.5%e to 8.7%e (Fig. 3A; Table 1). In comparison, \(\delta^{87}Sr/\delta^{86}Sr\) ratios for the same units appear to be less heterogeneous, spanning from 0.7049 to 0.7057 (Fig. 3B; Table 1). The Stillwater complex is the only documented locality with low-\(\delta^{18}O\) magmas in the region. The Mount Jefferson caldera system, southeast of the Stillwater complex, also displays heterogeneity in successively erupted tuffs; the upper tuff of Mount Jefferson is ~1%e lower that the lower tuff of Mount Jefferson (Fig. 3A; Table 1). Just south of the Mount Jefferson calderas, the Manhattan caldera may partly overlap the mineralized Round Mountain caldera, which hosts one of the largest epithermal gold deposits in the Great Basin (cf. Henry and John, 2013). The Manhattan caldera is distinctly lower in \(\delta^{18}O\) compared to other calderas at this longitude in the O-Sr profile (Fig. 3A). In contrast, nested calderas of the central Nevada volcanic field studied by Larson and Taylor (1986) have consistently high and homogeneous \(\delta^{18}O\) values (Fig. 3A), based on their analysis of bulk mineral separates. The high level of \(\delta^{18}O\) diversity observed in nested calderas at Stillwater, Mount Jefferson, and potentially Manhattan, is not observed in different units within single (isolated) calderas in the western Nevada volcanic field, such as the Caetano caldera (Watts et al., 2016).

Three Cretaceous granitic plutons with outcrops around the Stillwater caldera complex, the Alameda Canyon to the north and the La Plata Canyon and Sand Springs to the south (Fig. 1), have high and variable magmatic \(\delta^{18}O\) values: ~11%e for Alameda Canyon, ~10%e for La Plata Canyon, and ~8%e–9%e for Sand Springs (Fig. 3A; Table 1). Only Sand Springs overlaps the upper range of Stillwater tuffs. Whole-rock \(\delta^{87}Sr/\delta^{86}Sr\) ratios for Alameda Canyon (0.7050) and La Plata Canyon (0.7051) overlap Stillwater tuffs, whereas Sand Springs has distinctly lower \(\delta^{87}Sr/\delta^{86}Sr\) (0.7043–0.7047) (Fig. 3B; Table 1). Therefore, none of these plutons is an isotopic match in oxygen and strontium for the Stillwater tuffs. However, other Cretaceous plutons in the vicinity do overlap the isotopic compositions of the Stillwater tuffs, such as the New York granite stock, ~40 km north in the Stillwater Range (Fig. 1), which has a whole-rock \(\delta^{18}O\) value of 8.9%e and \(\delta^{87}Sr/\delta^{86}Sr\) ratio of 0.7056 (Figs. 3A–3B; Supplementary Data Table DR1), and the Rocky Canyon granitic pluton, ~80 km north intruding the Triassic Koipato Group in the West Humboldt Range (Fig. 1), which has a \(\delta^{87}Sr/\delta^{86}Sr\) ratio of 0.7051 (Fig. 3B; Supplementary Data Table DR1). The Jurassic Humboldt magmatic complex, which has extensive gabbroic and basaltic outcrops within ~20–40 km of the northern margin of the Stillwater caldera complex (Fig. 1), has unequivocally mantle-derived \(\delta^{87}Sr/\delta^{86}Sr\) ratios (0.7041–0.7043) that are far lower than those found for any Stillwater units (Fig. 3B; Supplementary Data Table DR1).

Zircon U-Pb Ages and Inheritance in the Stillwater Caldera Complex in the Context of Regional Basement

Zircon \(^{206}Pb/^{238}U\) ages have been determined by SHRIMP-RG for the 22 samples of the Stillwater caldera complex investigated for zircon O isotopes by SIMS (Table 2). As described by Colgan et al. (2018), U-Pb ages are bimodal, with the Job Canyon pulse at ca. 29 Ma and the Louderback Mountains, Pocoi Canyon, and Elevenmile Canyon pulses at ca. 25 Ma (Table 2). One sample from the Poco Canyon cycle and two from the Job Canyon cycle have inherited zircon grains, with U-Pb ages that are Eocene (35 Ma), Late Cretaceous (76–71 Ma, 98–95 Ma), Late Jurassic (150 Ma), and Middle Jurassic (166 Ma), and Neoproterozoic (545 Ma) (Fig. 4A). Of the three Cretaceous granitic plutons analyzed, Alameda Canyon (84 Ma), La Plata Canyon (87 Ma), and Sand Springs (89 Ma), Alameda Canyon and La Plata Canyon have inherited grains that are Late Cretaceous (96 Ma), Early Cretaceous (104–102 Ma), and Triassic (203 Ma) (Fig. 4A). Two inherited Job Canyon zircons overlap Sand Springs zircons in both U-Pb age and \(\delta^{18}O\) at ~6.6%e (Fig. 4A, inset). In contrast, one Job Canyon grain is much older at 150 Ma, though overlapping in

---

\(^{2}\)GSA Data Repository item 20190606, O and Sr isotope and age data for regional basement rocks in the Great Basin, zircon O isotope, age and trace element data for the Stillwater caldera complex, and hornblende chemistry data for the tuff of Elevenmile Canyon, is available at http://www.geosociety.org/datarpository/2019 or by request to editing@geosociety.org.
Figure 3. Longitudinal profiles for (A) magmatic δ¹⁸O and (B) ⁸⁷Sr/⁸⁶Sr values for calderas of the western Nevada volcanic field, with labels that correspond to those in Figure 1 and data in Table 1. Unfilled boxes indicate uncertain caldera locations. Included in these plots are published δ¹⁸O data for regional granites, and, when available for the same units, ⁸⁷Sr/⁸⁶Sr (see Supplementary Data Table DR1 for compilation of all plotted data). The low-δ¹⁸O magma threshold (δ¹⁸O = 6‰) is based on closed-system differentiation of mantle basalt (Bindeman, 2008). (C) Generalized pre-Jurassic crustal profile at ~40°N latitude in central Nevada, shown for the same longitudinal scale, adapted after Farmer and DePaolo (1983) and King et al. (2004). Depths are pre-Jurassic; they do not account for Jurassic–Cretaceous crustal shortening or Cenozoic extension. GT—Golconda thrust, RMT—Roberts Mountains thrust, M—mantle.
TABLE 2. SUMMARY OF OXYGEN ISOTOPE DATA FOR THE STILLWATER CALDERA COMPLEX

<table>
<thead>
<tr>
<th>Caldera cycle</th>
<th>Unit</th>
<th>Sample(s)</th>
<th>Age (Ma)</th>
<th>Zircon range*</th>
<th>Zircon average†</th>
<th>Laser fluorination data, δ18O (‰, VSMOW)</th>
<th>Zircon</th>
<th>Quartz</th>
<th>Sanidine</th>
<th>Plagioclase</th>
<th>Biotite</th>
<th>Magma§</th>
<th>∆18Oquartz-zircon T# (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Job Canyon</td>
<td>Pre-Job andesite/dacite</td>
<td>13-DJ-11</td>
<td>29.38 ± 0.38</td>
<td>7.0–8.6 (17/17)</td>
<td>7.21 ± 0.11 (12/17)</td>
<td>––</td>
<td>––</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>8.71</td>
<td>–</td>
</tr>
<tr>
<td>Job Canyon</td>
<td>Pre-Job dacite tuff</td>
<td>10-DJ-5</td>
<td>29.32 ± 0.97</td>
<td>6.5–7.7 (25/26)</td>
<td>6.98 ± 0.24 (25/25)</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>8.48</td>
<td>–</td>
</tr>
<tr>
<td>Job Canyon</td>
<td>Tuff of Job Canyon</td>
<td>10-DJ-2</td>
<td>29.25 ± 0.47</td>
<td>4.2–6.8 (20/20)</td>
<td>4.49 ± 0.26 (19/20)</td>
<td>4.10, 4.69</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>5.99</td>
<td>–</td>
<td></td>
</tr>
<tr>
<td>Job Canyon</td>
<td>Post-Job andesite/dacite</td>
<td>10-DJ-3; 13-DJ-46 (LF)</td>
<td>28.54 ± 0.51</td>
<td>4.0–8.1 (20/20)</td>
<td>5.41 ± 0.25 (10/20)</td>
<td>–</td>
<td>–</td>
<td>6.55, 6.56</td>
<td>5.33, 5.42</td>
<td>6.91</td>
<td>–</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Job Canyon</td>
<td>IXL pluton JC08-IXL4</td>
<td>28.45 ± 0.35</td>
<td>5.8–6.8 (14/14)</td>
<td>6.24 ± 0.27 (14/14)</td>
<td>6.00, 8.68, 8.51</td>
<td>–</td>
<td>–</td>
<td>7.74, 721, 787</td>
<td>–</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Job Canyon</td>
<td>IXL pluton JC08-IXL2</td>
<td>28.07 ± 0.33</td>
<td>6.2–6.6 (15/15)</td>
<td>6.34 ± 0.11 (15/15)</td>
<td>6.27</td>
<td>–</td>
<td>–</td>
<td>7.84</td>
<td>–</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Louderback Mountains</td>
<td>Louderback Mountains rhyolite</td>
<td>12-DJ-34</td>
<td>25.52 ± 0.25</td>
<td>4.5–7.4 (14/14)</td>
<td>7.24 ± 0.11 (12/14)</td>
<td>–</td>
<td>–</td>
<td>8.74</td>
<td>–</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Louderback Mountains</td>
<td>Tuff of Louderback Mountains</td>
<td>13-DJ-7</td>
<td>25.29 ± 0.26</td>
<td>3.9–7.3 (13/14)</td>
<td>7.09 ± 0.23 (5/14)</td>
<td>–</td>
<td>–</td>
<td>8.59</td>
<td>–</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Louderback Mountains</td>
<td>Tuff of Louderback Mountains JC11-26</td>
<td>25.05 ± 0.26</td>
<td>4.9–6.1 (15/15)</td>
<td>5.77 ± 0.17 (14/15)</td>
<td>–</td>
<td>–</td>
<td>7.27</td>
<td>–</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Louderback Mountains</td>
<td>Tuff of Louderback Mountains JC11-32</td>
<td>25.04 ± 0.32</td>
<td>5.4–6.2 (14/14)</td>
<td>5.73 ± 0.25 (14/14)</td>
<td>–</td>
<td>–</td>
<td>7.23</td>
<td>–</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Poco Canyon</td>
<td>Breccia-rich tuff of Poco Canyon</td>
<td>10-DJ-6</td>
<td>25.99 ± 0.20</td>
<td>3.9–4.7 (20/20)</td>
<td>4.25 ± 0.21 (20/20)</td>
<td>–</td>
<td>–</td>
<td>5.75</td>
<td>–</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Poco Canyon</td>
<td>Lower tuff of Poco Canyon</td>
<td>12-DJ-38</td>
<td>25.74 ± 0.19</td>
<td>3.6–6.9 (29/30)</td>
<td>4.03 ± 0.21 (25/29)</td>
<td>4.07</td>
<td>6.16, 6.28</td>
<td>–</td>
<td>5.53</td>
<td>757, 825</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Poco Canyon</td>
<td>Tuff of Poco Canyon</td>
<td>10-DJ-4</td>
<td>25.60 ± 0.25</td>
<td>4.0–6.3 (22/22)</td>
<td>4.32 ± 0.15 (14/22)</td>
<td>4.92, 5.25</td>
<td>6.66, 6.88</td>
<td>6.35</td>
<td>–</td>
<td>5.82</td>
<td>702, 765</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Poco Canyon</td>
<td>Rhyolite of Coyote Canyon</td>
<td>JC13-9</td>
<td>25.24 ± 0.25</td>
<td>4.0–7.1 (15/15)</td>
<td>4.27 ± 0.17 (15/15)</td>
<td>–</td>
<td>–</td>
<td>5.77</td>
<td>–</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Poco Canyon</td>
<td>Freeman Creek pluton, granite</td>
<td>JC08-IXL8</td>
<td>24.93 ± 0.37</td>
<td>5.5–6.4 (17/17)</td>
<td>5.66 ± 0.14 (15/17)</td>
<td>4.74, 5.47</td>
<td>–</td>
<td>–</td>
<td>7.16</td>
<td>–</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Elevenmile Canyon</td>
<td>Tuff of Lee Canyon</td>
<td>12-DJ-36</td>
<td>25.57 ± 0.26</td>
<td>4.2–7.1 (20/20)</td>
<td>6.03 ± 0.29 (16/20)</td>
<td>6.13</td>
<td>7.88, 8.21</td>
<td>6.81</td>
<td>–</td>
<td>7.53</td>
<td>804, 872</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Elevenmile Canyon</td>
<td>Tuff of Hercules Canyon</td>
<td>12-DJ-33</td>
<td>25.37 ± 0.19</td>
<td>5.4–7.2 (20/20)</td>
<td>6.02 ± 0.28 (16/20)</td>
<td>5.78, 5.87</td>
<td>7.68, 7.91</td>
<td>–</td>
<td>7.40, 7.50</td>
<td>7.52</td>
<td>873, 945</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Elevenmile Canyon</td>
<td>Tuff of Elevenmile Canyon H00-104A</td>
<td>25.23 ± 0.26</td>
<td>4.9–7.9 (19/19)</td>
<td>6.03 ± 0.51 (13/19)</td>
<td>5.70, 5.71</td>
<td>7.38, 8.06</td>
<td>7.05</td>
<td>–</td>
<td>7.53</td>
<td>834, 903</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Elevenmile Canyon</td>
<td>Tuff of Elevenmile Canyon 14-DJ-79</td>
<td>25.20 ± 0.30</td>
<td>4.2–6.5 (14/14)</td>
<td>5.95 ± 0.31 (12/14)</td>
<td>–</td>
<td>–</td>
<td>7.45, 704, 767</td>
<td>–</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Elevenmile Canyon</td>
<td>Tuff of Elevenmile Canyon 12-DJ-35</td>
<td>25.16 ± 0.23</td>
<td>5.2–7.7 (19/19)</td>
<td>5.80 ± 0.33 (11/19)</td>
<td>5.37, 5.66</td>
<td>–</td>
<td>–</td>
<td>7.30</td>
<td>–</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Elevenmile Canyon</td>
<td>Tuff of Elevenmile Canyon 12-DJ-37</td>
<td>25.12 ± 0.31</td>
<td>5.9–8.2 (16/16)</td>
<td>6.11 ± 0.13 (8/16)</td>
<td>–</td>
<td>–</td>
<td>7.61</td>
<td>–</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Elevenmile Canyon</td>
<td>Tuff of Elevenmile Canyon 14-DJ-75A</td>
<td>25.00 ± 0.26</td>
<td>3.9–7.1 (23/23)</td>
<td>5.66 ± 0.15 (8/23)</td>
<td>6.02, 6.20</td>
<td>8.28</td>
<td>–</td>
<td>7.36</td>
<td>708, 772</td>
<td>–</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Note: Ages are error-weighted mean zircon 206Pb/238U ages; data are from Colgan et al. (2018); “–” indicates no data. SIMS—secondary ion mass spectrometry; VSMOW—Vienna standard mean ocean water.

*Numbers in parentheses are number of analyses out of total number of analyses used to define range (Mesozoic and Cenozoic xenocrysts excluded); see Supplementary Data Table DR2 for full data set.

†Dominant (and/or rim) peak in spectra was used to calculate average; uncertainties quoted are 1 standard deviations of combined analyses; n—number of analyses out of defined range used to calculate average; see Supplementary Data Table DR2 for full data set.

§Magma δ18O values were calculated based on average ∆18Oquartz-melt = 0.4‰ and ∆18Omelt-zircon = 1.5‰ (fractionation factors from Trail et al. [2009] for 750–850 °C; ∆18Oquartz-melt = 0.4‰–0.5‰ and ∆18Omelt-zircon = 1.4‰–1.7‰). The calibration of Lackey et al. (2008) for whole-rock wt% SiO2 (~65–78 wt%) results in calculated magma δ18O values that are ~0‰–0.5‰ higher.

#Equilibrium ∆18Oquartz-zircon temperatures were calculated with fractionation factors from Trail et al. (2009) and Valley et al. (2003); the latter reference yields ~60–70 °C higher temperatures.
δ18O, and another is much younger at 35 Ma and distinctly higher in δ18O at ~8.3‰ (Fig. 4A, inset). One inherited Poco Canyon zircon overlaps Sand Springs in U-Pb age but is slightly higher in δ18O at ~7.4‰ (Fig. 4A, inset).

Age data for regional basement rocks and Cenozoic calderas define three major pulses, or flare-ups, of silicic igneous activity in the middle Cenozoic, Cretaceous, and Middle–Late Jurassic (Fig. 4B). Inherited zircons in the Stillwater caldera complex and Stillwater basement plutons fall broadly into these flare-up time periods, and most inherited zircons are Cretaceous in age (Figs. 4A–4B). Ages of specific basement units discussed in the previous section on oxygen and strontium isotopes are as follows: New York Canyon granite stock (ca. 71–69 Ma), Rocky Canyon granitic pluton (ca. 92–90 Ma), Humboldt mafic complex (ca. 172–169 Ma), and Koipato Group volcanics (ca. 250–248 Ma) (Fig. 4B; age data from Johnson, 1977; Wooden et al., 1999; Kistler and Speed, 2000; Vikre, 2014). Of these, all but the Koipato Group have ages that overlap within error with the inherited zircon ages documented in the Stillwater caldera complex (Figs. 4A–4B).

Oxygen Isotope Evolution in the Stillwater Caldera Complex

Oxygen isotope ratios collected by SIMS for 22 samples and ~400 analytical spots for single zircons are synthesized in Figure 5 and Table 2 (full data in Supplementary Data Table DR2). Laser fluorination data for bulk zircon and major phenocryst phases of quartz, sanidine, plagioclase, and biotite are included for 13 samples (Fig. 5; Table 2).

Job Canyon cycle magmatism began with andesite and dacite lava flows and small explosive eruptions (Fig. 5A). These units have high δ18O values and 206Pb/238U ages. The caldera-forming eruption of the tuff of Job Canyon was the first low-δ18O rhyolite produced, with δ18O of 4.5‰ and δ18O of 6.0‰. Bulk zircon analyzed by laser fluorination is in agreement with the SIMS zircon data (δ18O = 4.1‰–4.7‰). Next erupted was one of the most heterogeneous units documented in the Stillwater caldera complex, the post–Job Canyon andesite/dacite lava, with δ18O values that span from 4.0‰ to 8.1‰. Bulk zircon analyzed by laser fluorination is in high-temperature equilibrium (i.e., >700 °C) with the magmatic δ18O composition calculated from zircon rims. In addition to having the largest overall δ18O range in zircons, the post–Job Canyon andesite/dacite also has the largest δ18O variation within single zircons (>2‰ zoning between core, interior,
Oxygen isotopic investigation of silicic magmatism in the Stillwater caldera complex, Nevada

Figure 5. Oxygen isotope evolution of the Stillwater caldera complex, with units of each caldera cycle plotted in order of their error-weighted mean $^{206}\text{Pb}/^{238}\text{U}$ zircon ages; tie lines between panels show age overlaps. Secondary ion mass spectrometry (SIMS) zircon oxygen isotope data are shown by error-weighted probability density functions (PDFs), constructed using Isoplot (Ludwig, 2012). On the right side of each panel are rank-order plots that show the individual zircon data used to construct PDFs for an example sample. Data are summarized in Table 2 and included in full in Supplementary Data Table DR2. Laser fluorination oxygen isotope data for bulk zircon and other mineral phases are shown by the symbols plotted over the PDFs. The calculated magmatic $\delta^{18}\text{O}$ values for each sample are connected by thin black lines. The pre-caldera (i.e., pre-Job andesite/dacite) $\delta^{18}\text{O}$ zircon boundary (~7.2‰) is shown by the horizontal gray dashed line, and the threshold for low-$\delta^{18}\text{O}$ zircons (<4.5‰) is shown by the horizontal gray bar.
and rim domains). Intrusion of the IXL pluton marked a return to normal-δ¹⁸O values in the Job Canyon cycle, with δ¹⁸O_magma of 6.2‰–6.3‰ and δ¹⁸O_magma of 7.7‰–7.8‰. The IXL pluton samples are the most homogeneous of all Job Canyon cycle units, with unimodal probability density functions (PDFs) and zircons that are unzoned or very subtly zoned in δ¹⁸O. Laser fluorination data for bulk zircon from the IXL pluton are identical within error to the SIMS data, and laser fluorination data for quartz are in high-temperature equilibrium with zircon.

The next major caldera cycle in the Stillwater complex resumed after a long hiatus (>2 m.y.) with the eruption of the tuff of Louderback Mountains (Fig. 5B). The tuff of Louderback Mountains has a multimodal PDF, with both low-δ¹⁸O and high-δ¹⁸O zircons, and it is distinctly lacking in zircons with δ¹⁸O of ~6‰, which defined zircons at the end of the Job Canyon cycle. The Louderback Mountains rhyolite, a lava or intrusion underlying the tuff of Louderback Mountains, has zircons with a very similar distribution to the tuff of Louderback Mountains; zircon rims in this sample are 7.2‰–7.4‰. The dominant PDF peak and rim analyses in the Louderback Mountains rhyolite correspond to δ¹⁸O_magma of 7.1‰–7.2‰ and δ¹⁸O_magma of 8.6‰–8.7‰. Two intrusive units that underlie the Louderback Mountains caldera, the Chalk Mountain pluton and porphyry, have zircon populations that are very similar to one another, but very different from the tuff of Louderback Mountains. The Chalk Mountain units are unimodal, with the exception of one zircon with slightly lower δ¹⁸O in the pluton. The dominant peak and rim analyses correspond to δ¹⁸O_magma of 5.7‰–5.8‰ and δ¹⁸O_magma of 7.2‰–7.3‰.

The Poco Canyon caldera cycle succeeded the Louderback Mountains caldera cycle (Fig. 5C). All samples of the tuff of Poco Canyon are low-δ¹⁸O rhyolites, with δ¹⁸O_magma of 4.0‰–4.3‰ and δ¹⁸O_magma of 5.5‰–5.8‰. Two samples of the tuff (lower and upper cooling units) have multimodal PDFs with a small proportion (~15%–35%) of normal-δ¹⁸O zircons. Laser fluorination data for quartz and sanidine are in high-temperature equilibrium with the dominant PDF peaks for Poco Canyon zircons. The rhyolite of Coyote Canyon, which is a lava that preceded the tuff, has a unimodal zircon PDF that is identical to the dominant PDF peaks in the tuff of Poco Canyon. The granitic phase of the Freeman Creek pluton, which underlies the Poco Canyon caldera, has zircons that are significantly higher in δ¹⁸O, with δ¹⁸O_magma of 5.7‰ and δ¹⁸O_magma of 7.2‰. Laser fluorination data for bulk zircon overlap the SIMS zircon data for one analysis, but a duplicate analysis was ~0.9‰ lower. This fact, combined with the bimodal PDF for the granitic phase of the Freeman Creek pluton, indicates that it may be more heterogeneous than the current data show.

The Elevenmile Canyon caldera cycle is the youngest and most voluminous of the Stillwater caldera complex and overlapped the Poco Canyon cycle in space and time; however, the Elevenmile Canyon cycle has an oxygen isotope record that is highly disparate from the Poco Canyon cycle (Fig. 5D). All samples of the tuff of Elevenmile Canyon (and equivalent tuffs of Lee and Hercules Canyon from previous nomenclature of John [1995] and Henry and John [2013]) have multimodal PDFs with δ¹⁸O values that spans from 4.2‰ to 8.2‰. Low-δ¹⁸O zircons were found in three of the six tuff samples studied. The dominant PDF peaks have δ¹⁸O values of 5.9‰–6.1‰ (δ¹⁸O_magma of 7.4‰–7.6‰), which is consistent with the laser fluorination data for bulk zircon, and in high-temperature equilibrium with the laser fluorination data for

Figure 6. Oxygen isotope data for zircons of the tuff of Elevenmile Canyon determined by ion microprobe: (A) error-weighted probability density function for all zircon analyses, and (B) rank-order plot for zircons with two analytical spots per grain. Horizontal dashed lines connect core, interior, and/or rim analyses for individual grains. Vertical black line shows the average and standard deviation of only rim analyses. Zircon cathodoluminescence (CL) images annotated with δ¹⁸O values for core and rim analyses are shown for a few representative grains. White bar at the bottom of each CL image is 50 μm.
quartz, sanidine, and plagioclase. Single zircons in the tuff of Elevenmile Canyon are commonly zoned, with lower- and higher-δ¹⁸O cores and interiors overgrown by homogeneous rims with an average δ¹⁸O of 5.9‰ ± 0.2‰ (n = 12 rims; 1 standard deviation), which approximates the dominant PDF peak for all tuff of Elevenmile Canyon zircon analyses combined (n = 112; Fig. 6). The granodiorite phase of the Freeman Creek pluton has a multimodal zircon PDF that is very similar to the tuff of Elevenmile Canyon, with δ¹⁸O core of 5.8‰ for the dominant peak. Laser fluorination data for bulk zircon in the Freeman Creek pluton granodiorite are consistent with the SIMS data.

Trace elements were collected for a subset of the zircons analyzed for oxygen isotopes from the Job Canyon, Poco Canyon, and Elevenmile Canyon caldera cycles (Fig. 7; Supplementary Data Table DR2). Hafnium contents of zircons vary from ~5000 to 12,000 ppm for high-, normal-, and low-δ¹⁸O zircons (Figs. 7A, 7C, and 7E). Low-δ¹⁸O zircons in the Job Canyon cycle have relatively restricted Hf (~6000–7000 ppm), while those in the Poco Canyon cycle span almost the full range of Hf represented in the Stillwater caldera complex (~6000–12,000 ppm). Low-δ¹⁸O zircons in the Elevenmile Canyon cycle are few in comparison; the two for which we have Hf data are ~9000 ppm. Each cycle

Figure 7. Hafnium, δ¹⁸O, and rare earth element (REE) data for single zircons in the Stillwater caldera complex determined by ion microprobe (Supplementary Data Table DR2). Labels correspond to those in Figure 1. Vertical dashed gray line in the left panels shows the threshold for low-δ¹⁸O zircons (<4.5‰). Tie lines connect analytical spots within the same grains, with grain numbers labeled for those with a corresponding cathodoluminescence (CL) image in the left panels. White bar at the bottom of each CL image is 50 µm: g—granite, gd—granodiorite. In panels E and F, grains interpreted to be inherited from the Poco Canyon (PC) cycle are indicated.
possesses zircons with significant zoning in both $\delta^{18}$O and Hf. The greatest zoning is observed in the post-Job Canyon andesite/dacite, which has highly heterogeneous core and interior zircon domains that are overgrown by homogenized rims with $\delta^{18}$O of ~5.4‰ and Hf of ~9000 ppm (Fig. 7A). Chondrite-normalized rare earth element (REE) patterns in zircons from tuffs and plutons in the three cycles (Figs. 7B, 7D, and 7F), combined with the Hf and $\delta^{18}$O data, indicate that zircons in the Job Canyon and Poco Canyon cycles have tuffs and plutons with distinctly different geochemical signatures (i.e., higher $\delta^{18}$O, higher Hf, and lower REEs in the IXL pluton compared to the tuff of Job Canyon, and higher $\delta^{18}$O, lower Hf, and higher REEs in the Freeman Creek pluton granite compared to the tuff of Poco Canyon). In contrast, zircons in the tuff of Elevenmile Canyon and Freeman Creek pluton granodiorite overlap in $\delta^{18}$O, Hf, and REEs. The greatest zoning is observed in $\delta^{18}$O and Hf values that are within the range of Poco Canyon and Louderback Mountains zircons, suggesting that they are inherited from one of these cycles.

**Hornblende Barometry Constraints for the Tuff of Elevenmile Canyon Magma Chamber**

Hornblende chemistry was determined by electron microprobe for the tuff of Elevenmile Canyon to constrain crystallization pressures and estimate the crustal depths at which silicic magma chambers resided in the Stillwater caldera complex (Fig. 8; Supplementary Data Table DR3). The samples span ~20 km across the Louderback, Clan Alpine, and Desatoya Mountains and represent the trachydacitic-rhyolitic compositional range of the tuff (~62–73 wt% SiO$_2$). There are two distinct compositional groups of hornblende, a low-Al group with ~6–7.5 wt% Al$_2$O$_3$ and a high-Al group with ~10.5–12 wt% Al$_2$O$_3$ (Fig. 8A). The high-Al group has higher Mg, Ca, Ti, Na, and K and lower Si, Fe, and Cl relative to the low-Al group. Two samples have hornblende crystals that fall within each group, but the two groups do not overlap, and the high-Al hornblendes were only documented in samples with whole-rock SiO$_2$ less than 68 wt%. Single hornblende crystals are unzoned or very subtly zoned from core to rim (e.g., Figs. 8B–8C).

Two Al-in-amphibole geobarometers (Ridolfi et al., 2010; Mutch et al., 2016) were used to estimate crystallization pressures for the two hornblende groups. The results between the two geobarometers are disparate for the same hornblende Al contents (vary by about a factor of two), and because it is not clear which geobarometer may be more accurate (and because Al-in-amphibole geobarometry is prone to large uncertainties in general; Putirka, 2016), we make very general estimates for crystallization pressures and depths for average compositions for the two hornblende groups. We note that the Mutch et al. (2016) barometer was calibrated for low-pressure granitic rocks with the same mineral assemblage that is found in Stillwater rhyolites. For the low-Al group, the intermediate total Al content (1.2 atoms per formula unit, apfu) corresponds to 1.0–2.2 kbar and ~4–8 km depth (based on a 3.7 km/kbar pressure-depth conversion) (Fig. 8A). For the high-Al group,

---

Figure 8. Hornblende geochemical data for the tuff of Elevenmile Canyon determined by electron microprobe (Supplementary Data Table DR3). (A) Aluminum content vs. Mg#. Whole-rock SiO$_2$ contents of host tuff samples are shown in parentheses. Pressures (P) calculated with the barometers of Ridolfi et al. (2010) ("Ref 1") and Mutch et al. (2016) ("Ref 2") are connected by horizontal dashed lines to single hornblende analyses that span the compositional ranges of the two hornblende groups. Depths were estimated from calculated pressures with a 3.7 km/kbar conversion. Crystallization temperatures (T) of hornblende from the Ridolfi et al. (2010) calibration are included to the right. (B–C) Representative photomicrographs of high- and low-Al hornblende crystals in one of the tuff samples, showing a lack of zoning between core and rim domains.
the intermediate total Al content (2.0 apfu) corresponds to 3.6–5.3 kbar and ~13–19 km depth (Fig. 8A). Because the majority (~80%) of hornblende crystals fall within the low-Al group and do not define a continuum with the high-Al group, we interpret the ~4–8 km depth estimate as representative of the tuff of Elevenmile Canyon magma chamber, and the high-Al hornblendes as xenocrysts.

**DISCUSSION**

We begin the discussion by describing regional isotopic trends in Cenozoic and Mesozoic silicic magmas in the Great Basin, which we use to frame our study of the Stillwater caldera complex, one of the largest nested caldera centers of the middle Cenozoic ignimbrite flare-up. We consider potential crustal sources in the petrogenesis of isotopically diverse Stillwater magmas, beginning with the first Job Canyon caldera cycle and then subsequent caldera cycles, followed by an evaluation of possible low-$\delta^{18}$O rhyolite genesis models. We conclude the discussion by assessing relationships between caldera-forming tuffs and plutons in the Stillwater caldera complex.

**Significance of Regional Oxygen and Strontium Isotopic Trends in the Great Basin**

The isotopic trends defined by our new data for voluminous caldera-forming tuffs of the western Nevada volcanic field reflect the basement crustal architecture, with the highest $\delta^{18}$O values and $^{87}$Sr/$^{86}$Sr ratios just west of the North American craton boundary at the 0.708 Sr isopleth, where the thickest packages of continental clastic rocks accumulated in the miogeoclinal (Figs. 1 and 3). The contribution of the clastic (meta)sediment package to silicic magmas appears to wane (lower $\delta^{18}$O, $^{87}$Sr/$^{86}$Sr) westward across the transition zone and into accreted oceanic terranes at the 0.706 Sr isopleth. The same trends in oxygen and strontium isotopes are broadly apparent in Jurassic and Cretaceous granites (Fig. 3), indicating that these trends are pre-Jurassic features; they reflect the Precambrian craton margin, passive-margin sedimentation, and Paleozoic tectonic events (Antler and Sonoma orogenies) that thickened the miogeoclinal sediment prism (Dickinson, 2006). The tighter trends resolved in the Cenozoic calderas may be a function of averaging of isotopic heterogeneities in the crustal block over tens of millions of years of silicic magmatism during the Jurassic and Cretaceous, with the Cenozoic calderas representing discrete point sources through the homogenized crustal block.

In addition to a trend of increasing sedimentary components in magmas from west to east, King et al. (2004) proposed an increase in the availability of sedimentary components through time as successive orogenic events thickened the crust, with Late Cretaceous granites having higher $\delta^{18}$O values and $^{87}$Sr/$^{86}$Sr ratios (and thus a greater crustal affinity) than Early Cretaceous and Jurassic granites, similar to the conclusions of Barton (1990). Our data demonstrate that middle Cenozoic calderas just west of the 0.708 Sr isopleth have highly elevated $\delta^{18}$O values (e.g., $\delta^{18}$O$_{quart}$ of 8.7‰ and $\delta^{18}$O$_{quart}$ of 10.2‰ for the Caetano Tuff) that are similar to values of Late Cretaceous granites emplaced during the Sevier orogeny at this longitude (cf. King et al., 2004). This indicates that orogenic events that thickened the crust in the Cretaceous continued to exert a control on the isotopic signatures of Cenozoic silicic magmas. Metapelitic xenoliths, zircon inheritance, and Sr-Nd-O isotopic modeling support derivation of the Caetano Tuff from anatexis of Proterozoic metasedimentary basement crust (Watts et al., 2016). In contrast to the interpretation by King et al. (2004), for a greater mantle component in Cenozoic magmas due to crustal thinning during Basin and Range extension, it is now recognized that episodes of extension significantly postdated middle Cenozoic calderas (e.g., Colgan et al., 2008), and our oxygen and strontium isotope data do not support a greater mantle component in their source magmas as compared to Mesozoic granites at the same longitude (Fig. 3).

Several Cretaceous granitic plutons west of the 0.706 Sr isopleth that were analyzed in this study (Alameda Canyon, La Plata Canyon, Sand Springs) have much higher $\delta^{18}$O values ($\delta^{18}$O$_{quart}$ of 9.8‰–11.5‰ and $\delta^{18}$O$_{quart}$ of 8.1‰–11.1‰) than the O-Sr trends would predict for this location on the longitudinal profile (Fig. 3A). King et al. (2004) also documented several Cretaceous granites with high-$\delta^{18}$O quartz values at approximately this longitude, which they attributed to subsolidus alteration of quartz rather than reflecting the isotopic composition of the host magmas. Our data indicate that these elevated $\delta^{18}$O values may be primary (or close to primary), based on (1) the consistency of triplicate analyses for quartz in each of the three plutons, and (2) the high-temperature equilibrium of quartz with zircon for the Sand Springs pluton ($\delta^{18}$O$_{quartz-zircon}$ = 3.2‰, which corresponds to 580 °C or 635 °C equilibrium based on the fractionation factors of Trail et al. [2009] or Valley et al. [2003], respectively). The location of the plutons is coincident with backarc basinal sedimentary terranes that were thrust onto the continental shelf during the Jurassic Luning-Fencemaker fold-and-thrust events (Wyld et al., 2003), indicating that their genesis may have involved contamination by these sedimentary sources. Though the $\delta^{18}$O compositions of the plutons are elevated, their $^{87}$Sr/$^{86}$Sr ratios (0.7043–0.7051) are consistent with the radiogenic signatures of other granites and caldera-forming tuffs at this longitude (Fig. 3B).

In broader applicability, the O-Sr longitudinal trends documented here, which represent isotopic sampling of the crust on a Great Basin-wide scale, may be informative for a wide range of topics and processes beyond the scope of this study. For example, correlation of tuff deposits and linking individual tuffs to source calderas has been challenging in some cases due to large tuff volumes, widespread distributions due to paleotopography, and varying levels of exposure and concealment by sedimentary basins. Our new data are consistent with the hypothesized location of the tuff of Cave Mine caldera ~20 km to the north and between the Fish Creek Mountains and the Caetano Tuff calderas (Fig. 1; Henry and John, 2013), as it is on-trend and intermediate to the Fish Creek Mountains and Caetano Tuff in the O-Sr longitudinal profile (Fig. 3). Recent mapping, geochemistry, petrography, and geochronology indicate that the Nine Hill caldera is within the northern part of the Elevenmile Canyon caldera in the Stillwater caldera complex (Fig. 2), in contrast to previous interpretations for its source caldera beneath the Carson Sink to the west of the Stillwater Range (Best et al., 1989; Henry and John, 2013). Our new oxygen and strontium isotope data for Nine Hill place it isotopically between the Campbell Creek and Fairview Peak calderas (Fig. 3), and thus are consistent with new mapping and interpretations.

O-Sr longitudinal trends also bear on hypothesized links between crustal structure and metallogeny in the Great Basin. For example, middle Cenozoic Carlin-type gold deposits, which cumulatively form the second largest concentration of gold on Earth, may have had gold and other metals derived from Proterozoic sediments of the miogeoclone in central and eastern Nevada (Vikre et al., 2011). Because the middle Cenozoic caldera-forming tuffs track miogeoclinal sedimentary components in their source magmas, they offer important constraints for testing metallogenesis models. Two very large (~465 metric tons) Carlin-type gold deposits (Cortez and Cortez Hills) are located along the northeastern margin of the Caetano caldera in central Nevada (cf. Watts et al., 2016). The Caetano Tuff has the largest miogeoclinal sedimentary contribution of the 16 calderas studied (Fig. 3), and therefore these isotopic parameters may be useful for delineating favorable areas for Carlin-type mineralization.
Sources of High-, Normal-, and Low-δ18O Crustal Components in the First Caldera Cycle of the Stillwater Caldera Complex

The first caldera cycle in the Stillwater caldera complex, ca. 29 Ma Job Canyon, produced high-, normal-, and low-δ18O magmas, and, therefore, there must have been normal- to high-δ18O and low-δ18O crustal components over relatively restricted spatial scales at the onset of Stillwater magmatism. Inherited zircons in Job Canyon cycle rocks are broadly reflective of major flare-up events in the Mesozoic (Fig. 4). Late Cretaceous (76–71 Ma and 97–95 Ma) grains dominate the inherited zircon populations and overlap the ages of the three Stillwater basement plutons and the oxygen isotopic composition of zircons from the Sand Springs pluton (Fig. 4A). Inherited ages also overlap with two of the Late Cretaceous granitic plutons in the vicinity (New York Canyon and Rocky Canyon) that were found to have similar whole-rock δ18O and 87Sr/86Sr compositions to normal- to high-δ18O Stillwater magmas (Figs. 3–4). In contrast, the La Plata Canyon and Alamenda Canyon plutons have δ18O values that are higher, and the Sand Springs pluton has δ18O values that overlap normal- to high-δ18O Job Canyon magmas, but 87Sr/86Sr ratios that are lower (Fig. 3). A few older zircons in the Job Canyon dacite tuff, including Middle–Late Jurassic (166–150 Ma) and Neoproterozoic (545 Ma), increase the diversity of the inherited zircons. The 150 Ma granite has an oxygen isotopic composition that is identical within error to the Late Cretaceous inherited grains and the Sand Springs pluton zircons (Fig. 4A, inset). Therefore, it appears that a variety of normal- to high-δ18O crustal components could have contributed to the initial oxygen isotope signatures of Stillwater magmas (probably most similar to the New York Canyon stock, based on combined δ18O and 87Sr/86Sr), but the source of the low-δ18O component is unclear. The absence of low-δ18O zircons in any of the investigated Mesozoic granitic plutons (this study; King et al., 2004; Blum et al., 2016) indicates that whatever the low-δ18O source, it was not tapped during previous episodes of crustal melting.

Here, we evaluate potential rock reservoirs that could have been altered by meteoric waters to yield a low-δ18O crustal component in the first low-δ18O caldera-forming tuff of the Stillwater complex, the tuff of Job Canyon (see next discussion section for evaluation of the Poco Canyon and Elevenmile Canyon cycles). Stillwater calderas overlie three distinct Mesozoic basement terranes (Sand Springs, Quartz Mountain, and Jungo) that consist of clastic, volcanic, and carbonate rocks that were accreted to the continental margin, metamorphosed, and intruded by Jurassic and Cretaceous plutons (Fig. 9; cf. Crafford, 2007, 2008; Vetz, 2011, 2014). Reasonable initial sources for a low-δ18O crustal component could be the Middle Jurassic Humboldt maﬁc complex or the Early Triassic Koipato Group rocks, as they both have extensive exposures of hydrothermally altered rocks to the north of the Stillwater caldera complex (Kistler and Speed, 2000; Johnson, 2000; Vetz, 2014), and they are projected beneath or within the basement terranes that host the Stillwater complex (Fig. 9). Each of these could have been thrust-thickened and transported downward, potentially ~7–14 km, during crustal shortening associated with the Jurassic Luning-Fencemaker fold-and-thrust events (Wydell et al., 2003). Despite these facts, the isotopic data and age inheritance observed in the Stillwater caldera complex are not consistent with either of these potential sources. The Humboldt maﬁc complex has δ18O/δ18O of 0.7041–0.7043, and the Koipato Group has δ18O/δ18O of 0.709–0.712 (Kistler and Speed, 2000; Vetz, 2011), which are vastly different from magmas of the Stillwater caldera complex, which have δ18O/δ18O of 0.7049–0.7057. The age of the Humboldt maﬁc complex is close to overlapping with one inherited grain found in the pre–Job Canyon dacite tuff, whereas no inherited grains correspond to the age of the Koipato Group rocks. Local Cretaceous granitic plutons are also commonly hydrothermally altered, some with mineralization (Wilden and Speed, 1974; Johnson et al., 1986; Quade and Tingley, 1987). Though none of the Cretaceous plutons in the immediate vicinity is an isotopic match in δ18O and δ18O/δ18O for Stillwater magmas, the New York Canyon stock and Rocky Canyon pluton to the north are very similar. Perhaps most importantly, zircon inheritance in the Job Canyon cycle is dominated by Cretaceous granites. Therefore, Job Canyon cycle magmas unequivocally traveled through and incorporated Cretaceous crust.

Potential depths of partial melting and assimilation of hydrothermally altered Cretaceous plutons into the Job Canyon magma chamber must be considered in this hypothesis. Unlike the hydothermally altered Triassic-Jurassic basement rocks, which may have been transported downcrust during the Jurassic Luning-Fencemaker fold-and-thrust events, no sy- or post-Late Cretaceous episodes of crustal shortening occurred in this part of Nevada to transport altered Cretaceous plutons to deeper crustal levels. The Sevier orogeny peaked during the Late Cretaceous, but the front of this fold-and-thrust belt was much further east in the Great Basin and would not have resulted in crustal shortening in basement rocks beneath the Stillwater caldera complex in western Nevada (cf. Dickinson, 2006). Notwithstanding this issue, granitic plutons can be altered by meteoric waters to significant depths. For example, the lowest exposed parts of the IXL pluton (~10 km paleodepth) in the Job Canyon caldera have highly altered whole-rock δ18O values of ~0‰ (Fig. 10).

Because assimilation would have happened at shallow depths in the crust, thermal constraints must also be considered. Pre–Job Canyon andesite eruptions indicate that maﬁc magmas were heating the crustal block prior to the formation of the tuff of Job Canyon magma chamber. These andesites are some of the earliest preserved evidence of slab roll-back magmatism in the Stillwater region. Decompression of the asthenosphere during slab roll-back is presumed to have led to the formation of large volumes of basaltic magmas that perturbed the local geothermal gradient and promoted large-scale crustal melting (cf. Best et al., 2016). No basaltic erupted in or around the Stillwater caldera complex at its inception ca. 30–29 Ma, but basaltic andesites (54–58 wt% SiO2) erupted peripherally to the Stillwater complex from ca. 34 to 32 Ma near the Fairview Peak caldera and West Gate (~0–25 km south) and near the Deep Canyon caldera and Edwards Creek Valley (~5–20 km northeast). The presence of maﬁc enclaves (56–58 wt% SiO2) in the IXL pluton (John, 1995) provides direct evidence for basaltic andesite in the shallow crust beneath Stillwater calderas.

The spatial extent and depth of midercrustal sills, crustal melt zones, or other salient features are unresolved by geophysics due to late Cenozoic extension and magmatism that have profoundly modified the structure of the crust in the Great Basin. Geophysical imaging of the Altiplano-Puna volcanic complex, which is perhaps the closest modern-day analogue to the Great Basin ignimbrite ﬂare-up (Best et al., 2016), reveals a composite partial melt zone that is continuous from depths of ~4 to 25 km (Ward et al., 2014), and modeling of the thermal structure of the crust indicates a temperature of ~500 °C at ~10 km depth (Gottsmann et al., 2017). Numerical thermodynamic modeling for low-δ18O rhyolites in the Snake River Plain, which are comparable in volume to the tuff of Job Canyon (>55 km3), have shown that basaltic sills can generate superheated silicic magmas that melt and assimilate ~20%–50% of even relatively cold (400 °C) country rocks to form low-δ18O rhyolites (Simakin and Bindeman, 2012). Additional thermomechanical models by Colón et al. (2018) highlighted the importance of rheological discontinuities in the upper crust to focus melting. Heterogeneous distributions of Cretaceous plutonic rocks beneath the Stillwater caldera.
complex may have been the source of such discontinuities that aided the melting process.

Model of Low-δ¹⁸O Rhyolite Genesis in the Stillwater Caldera Complex

The appearance of a large-volume, low-δ¹⁸O rhyolite during the first caldera cycle diverges from the caldera “cannibalization” paradigm, in which younger calderas are expected to exhibit δ¹⁸O depletions through time as a result of remelting or “cannibalizing” hydrothermally altered parts of older calderas with which they overlap (e.g., Bindeman and Valley, 2001; Bindeman et al., 2008; Watts et al., 2011, 2012). However, other aspects of the oxygen isotope evolution of the Stillwater caldera complex are strikingly similar to the caldera cannibalization model, most notably, the extreme isotopic diversity of zircons within and between caldera cycles (Fig. 5). All caldera cycles have units with low-δ¹⁸O zircons (<4.5‰), but only the Poco Canyon caldera cycle has units that are dominated by low-δ¹⁸O zircons. The appearance of the low-δ¹⁸O tuff of Poco Canyon after the first Job Canyon caldera cycle, and the potential spatial overlap of its caldera with the

Figure 9. Relative stratigraphic positions and lithologic characteristics of Mesozoic basement terranes and intrusions projected beneath the Stillwater caldera complex in western Nevada. The Stillwater caldera complex is located at the intersection of the Sand Springs, Quartz Mountain, and Jungo terranes. The Jungo terrane is thrust over the Humboldt assemblage by the Luning-Fencemaker thrust (LFT) and is unconformably underlain by the Koipato Group volcanic rocks, which in turn are unconformably underlain by the Golconda terrane and the Golconda thrust (GT) (cf. Crafford, 2007, 2008; Vetz, 2011; Wyld et al., 2003; Dickinson, 2006). Note that the Koipato Group rocks are part of the Humboldt assemblage (Crafford, 2007). Stars in the inset map show the locations of granitic intrusions (SS—Sand Springs; LC—La Plata Canyon; AC—Alameda Canyon; NYC—New York Canyon; RC—Rocky Canyon); crosses show outcrops of the Humboldt mafic complex (HM).
Figure 10. Contoured oxygen isotope data for (A) hydrothermally altered rocks in the Job Canyon caldera, with (B) individual units and features labeled. Whole-rock oxygen isotope data from John and Pickthorn (1996) were contoured using Aabel geochemical plotting software. In panel (A), sample locations are shown by the black dots, and the area of the IXL pluton is outlined in a heavy black line. The caldera is tilted to nearly vertical, with the paleosurface to the west and the bottom of the exposed pluton to the east.
Job Canyon Caldera (Fig. 2; John, 1995; Colgan et al., 2018) present the possibility that its genesis was related to cannibalization of hydrothermally altered rocks from the Job Canyon cycle. Note that although the low-$\delta^{18}O$ tuff of Job Canyon was produced during the first caldera cycle, most magmas of that cycle were normal- to high-$\delta^{18}O$, including the last large-volume manifestation of silicic magmatism, the IXL pluton with unimodal $\delta^{18}O_{min}$ of $6.2‰$–$6.3‰$. This evidence supports derivation of Job Canyon cycle magmas from isotopically diverse crust, with rare pockets of low-$\delta^{18}O$ material, perhaps altered Cretaceous plutons, as described in the previous section. In contrast, the low-$\delta^{18}O$ source of the tuff of Poco Canyon was continuous enough that all parts of the tuff and the precaldera rhyolite of Coyote Canyon were contoured to approximately proportional areas of $\delta^{18}O$ values (Fig. 10). We note that the meteoric-hydrothermal system is characterized by near-neutral pH assemblages (K-feldspar stable propylitic, illitic, and intermediate argillic assemblages) that decrease in intensity up section, with an earlier phase of acid alteration (sericitic and advanced argillic) that is present locally at the top of the tuff and associated with postcaldera andesite dikes. Integrated $\delta^{18}O_{whole-rock}$ averages over depth profiles from $6$ to $10$ km, $5$ to $10$ km, and $0$ to $10$ km are of $0.9‰$, $-1.5‰$, and $-2.0‰$, respectively (Table 3). Assuming a normal-$\delta^{18}O$ starting magma with $7.5‰$ estimated from normal-$\delta^{18}O$ zircons in the tuff of Poco Canyon and normal-$\delta^{18}O$ Stillwater magmas, and modeled for the full range of initial $\delta^{18}O$ values ($0.7049$–$0.7057$) in the Stillwater caldera complex (Fig. 11; Table 3), the results indicate $\approx20%$ mass contribution from the Job Canyon intracaldera block using the $\approx4$–$10$ km depth (6 km thickness) and $\approx5$–$10$ km depth (5 km thickness) sections or $\approx30%$ using the $\approx6$–$10$ km depth (4 km thickness; IXL pluton only).

Volume estimates of the tuff of Poco Canyon ($\approx250$–$500$ km$^3$) and the Job Canyon intracaldera block over the defined $\approx4$–$10$ km depth range ($\approx80$–$480$ km$^3$) were then used to calculate the percentage of the Job Canyon intracaldera block volume in the tuff of Poco Canyon (Table 4). For intermediate volumes for the tuff of Poco Canyon ($375$ km$^3$) and Job Canyon intracaldera block ($250$ km$^3$), the required percentage of the Job Canyon intracaldera block volume is $\approx30%$ (Table 4). The lowest required percentage is $\approx10%$, assuming a maximum Job Canyon intracaldera block volume and a minimum tuff of Poco Canyon volume (Table 4). Conversely, the highest required percentage is $>100%$, if using a minimum Job Canyon intracaldera block volume and a maximum tuff of Poco Canyon volume (Table 4). The mostly feasible results indicate that the hypothesis for cannibalized Job Canyon components in the

---

**Figure 11. Modeling of the Job Canyon intracaldera block as a component in the low-$\delta^{18}O$ tuff of Poco Canyon magma.** (A) Schematic representation of the Job Canyon intracaldera block as a cylinder with a diameter of 5–10 km and a thickness of 4–6 km (4–10 km paleodepth range). (B) Sr-O isotopic mixing models constructed between a hypotethetical normal-$\delta^{18}O$ starting magma and the low-$\delta^{18}O$ Job intracaldera block; nine curves are modeled for high (0.7057), intermediate (0.7053), and low (0.7049) $\delta^{18}O$ values. Isotopic mixing curves were calculated with the following equation for a two component system: $R_{mix} = (R_A C_{F_A} + R_B C_{F_B} (1 - F_B) [C_{F_A} + C_B (1 - F_B)])$, where $R_{mix}$ is the isotopic ratio of the mixed magma composition, $R$ is the isotopic ratio for each element, $C$ is the element concentration, $F$ is the mass fraction, and components A and B represent the starting magma and the Job Canyon intracaldera block assimilant; tick marks demarcate $10%$ increments. Starting magma parameters were estimated from the least-evolved tuff of Poco Canyon and high-$\delta^{18}O$ zircons in the tuff of Poco Canyon (150 ppm Sr, 0.7049–0.7057 $\delta^{18}O$, 48.57 wt% O, $\delta^{18}O = 7.5%e$; Job Canyon intracaldera block assimilant parameters were estimated from the average of the IXL pluton, pre–Job Canyon andesite-dacite, and tuff of Job Canyon samples (500 ppm Sr, 0.7049–0.7057 $\delta^{18}O$, 47.87 wt% O, $\delta^{18}O = 0.9%e$ to $-2.0%e$). See Tables 3–4 for additional details.
low-δ18O Poco Canyon magma is viable. The required volume could be substantially lower if (1) the starting magma had a δ18O composition lower than 7.5‰, (2) the low-δ18O assimilated part of the Job Canyon block was more depleted than the ~2.0‰ average, or (3) the volume of the Job Canyon block was close to or exceeding the upper bounds of the estimated volume range. For example, if the starting magma was 7.0‰, the assimilant was ~6‰, and the volume of the Job Canyon block was at the upper modeled bound, ~4% of its total volume would be required.

We highlight a few critical observations to consider in alternative models of low-δ18O rhyolite genesis. First, no inherited Job Canyon zircons were found in the tuff of Poco Canyon. Because the Job Canyon zircons are ca. 29 Ma in age, their presence in the ca. 25 Ma Poco Canyon magma should have been apparent, unless they were completely dissolved (we note that errors on individual 206Pb/238U ages of Stillwater zircons range from ~0.2 to 2.5 m.y., average of 0.3–0.5 m.y.). In this study, we did document one inherited grain in the tuff of Poco Canyon (98 Ma), which overlaps inherited grains in the Job Canyon cycle (Fig. 4A). We cannot rule out the possibility that the tuff of Poco Canyon also tapped a hydrothermally altered Cretaceous crustal source, as advocated for the tuff of Job Canyon. However, as described above, the low-δ18O zircon record in the Poco Canyon cycle is clearly disparate from the Job Canyon cycle, and the paucity of inherited Mesozoic or Job Canyon grains in the Poco Canyon cycle is a problem for either hypothesis. Second, field evidence for partial melting of intracaldera Job Canyon rocks is lacking: for example, obvious melt lenses or dikes through the IXL pluton have not been found despite detailed mapping. Although we have a nearly complete ~10 km vertical section through the Job Canyon caldera, it is a single slice of a three-dimensional caldera. The volume of the intracaldera block that is modeled to be involved in the cannibalization process is <30% and could be <5%–10% if using less conservative model parameters; statistically, it is likely to be represented in other parts of the three-dimensional caldera that are not exposed in the single vertical slice. Third, another possible low-δ18O crustal source in the tuff of Poco Canyon, if not Mesozoic basement or Job Canyon cycle rocks, may be older Oligocene calderas that underlie the Stillwater caldera complex. The ca. 30.4 Ma Deep Canyon caldera is exposed ~25 km to the east of the Job Canyon and Poco Canyon calderas (Fig. 2). Perhaps other calderas like it beneath the Poco Canyon caldera could have contributed low-δ18O source rocks. The lack of xenocrysts makes all of these possibilities difficult to assess.

Though the tuff of Elevenmile Canyon was not a nominally low-δ18O rhyolite, it has a remarkably diverse zircon record that is instructive for low-δ18O rhyolite genesis models. Multimodal δ18O spectra characterize zircons from all samples of the tuff of Elevenmile Canyon across the ~60 km caldera (Figs. 2 and 5D). Zircon cores and interiors are highly homogeneous (δ18Ocore of ~4‰–8‰), whereas zircon rims have a homogenized value (δ18Ores of ~6‰) (Fig. 6B). Two low-δ18O zircons (<4.5‰) for which we also have trace element data indicate that they are an identical match to zircons from the tuff of Poco Canyon, with the distinctive negative Eu anomalies in REE patterns (Figs. 7E–7F). The tuff of Louderback Mountains is another potential source of the low-δ18O zircons, based on our published trace element data (Colgan et al., 2018). This evidence indicates that the tuff of Elevenmile Canyon magma chamber may have cannibalized Poco Canyon and/or Louderback Mountains components in the parts of its magma chamber that overlapped these calderas. The pattern apparent in the tuff of Elevenmile Canyon, with isotopically diverse zircon cores overgrown by homogenized rims that approximate the bulk zircon average (Fig. 6), is remarkably similar to observations for the large-volume, low-δ18O Kilgore Tuff of the Heise volcanic field in the eastern Snake River Plain (Watts et al., 2011), and supports our view that large silicic magma chambers are assembled from isotopically diverse batches of melt in the shallow crust.

Finally, aspects of the regional caldera centers to which we extended our oxygen isotopic
investigation are pertinent to low-$\delta^{18}$O rhyolite genesis models. It is instructive that in addition to the Stillwater caldera complex, the Mount Jefferson caldera complex exhibits a clear depletion in $\delta^{18}$O through time. Quartz in the upper tuff of Mount Jefferson is $\sim1\%$ lower than the lower tuff of Mount Jefferson (Fig. 3A), supporting a model in which low-$\delta^{18}$O crust was assimilated into the upper tuff of Mount Jefferson magma chamber. In addition, the Manhattan caldera, which may overlap the mineralized Round Mountain caldera, is distinctly lower in $\delta^{18}$O compared to other tuffs in the area, and may have assimilated altered parts of the Round Mountain caldera or perhaps altered Cretaceous granites in the vicinity (Henry and John, 2013). In contrast, the nested calderas studied by Larson and Taylor (1986) in the central Nevada volcanic field produced high-$\delta^{18}$O magmas with a subtle $-0.4\%$ depletion from the early to middle eruptive sequence and a return to the initial higher values in the late eruptive sequence. These authors did not have the benefit of high-spatial-resolution zircon data; their interpretations may have underestimated contributions from isotopically diverse crustal components. Nonetheless, the nested calderas studied by Larson and Taylor (1986) are $\sim75$–100 km east of the paleodevide boundaries defined by Henry and John (2013) and Best et al. (2013a) (Fig. 1). Therefore, these calderas formed on the Nevadaplateau topographic crest in the middle Cenozoic (e.g., DeCelles, 2004). By analogue with the Altiplano region of the Central Andes, it may be that the lack of low-$\delta^{18}$O rhyolites in this part of Nevada was due to a lack of meteoric water in a high-elevation, arid environment (Folkes et al., 2013). Interestingly, the caldera-forming tuffs studied by Larson and Taylor (1986) in the central Nevada volcanic field have ages of ca. 32–24 Ma (Best et al., 2013b), which overlap those of the Stillwater caldera complex. Located well to the west of the paleodevide boundary, the Stillwater complex may have been better positioned for the establishment of caldera lakes and other surface-water features to promote hydrothermal circulation and alteration of shallow crustal rocks.

**Relationship between Caldera-Forming Tuffs and Granitic Plutons in the Stillwater Caldera Complex**

All investigated cycles of the Stillwater caldera complex have pluton or porphyry systems that are spatially coincident with calderas (Fig. 2; Colgan et al., 2018). Despite this spatial association, our new oxygen isotope data are clearly disparate for tuff and pluton samples for all caldera cycles except the tuff of Elevenmile Canyon, in which six analyzed samples of the tuff are very similar to the granodiorite phase of the Freeman Creek pluton (Fig. 5D). Trace element data collected for the same zircons analyzed for oxygen isotopes provide further support of this observation; only the tuff of Elevenmile Canyon and Freeman Creek pluton granodiorite have identical $\delta^{18}$O, Hf, and REE patterns (Fig. 7). The IXL pluton has discernibly younger zircon U-Pb ages than the tuff of Job Canyon (28.5–28.1 Ma for IXL pluton and 29.2 Ma for the tuff of Job Canyon), but U-Pb ages for all other spatially associated tuffs and plutons are indistinguishable within error (Table 2). It is clear that relying on field evidence and geochronology alone to assess volcanic-plutonic relationships is insufficient in the Stillwater caldera complex, and highlights the necessity of crystal-scale chemical and isotopic data to make robust conclusions about these processes in the Great Basin and worldwide. Our findings for the Stillwater caldera complex indicate that, in general, caldera-forming tuffs and plutons are not derived from the same magmatic sources. This is in direct contrast to our previous finding for the well-exposed 34.0 Ma Caetano caldera in north-central Nevada, where a coegenetic relationship between the caldera-forming tuff and large caldera intrusions is unequivocal based on many lines of whole-rock and crystal-scale evidence (Watts et al., 2016). We highlight that none of the plutonic or porphyry units appears to have assimilated low-$\delta^{18}$O crust, and we speculate that this may be due to a lower-temperature, waning phase of silicic magmatism that did not favor digestion of shallow crustal materials.

**CONCLUSIONS**

Voluminous silicic tuffs (>3000–5000 km$^3$) of the Stillwater caldera complex possess a remarkably diverse oxygen isotopic record. All caldera cycles have low-$\delta^{18}$O zircons, providing unequivocal evidence for recycling of hydrothermally altered crust in the formation and growth of silicic magma chambers that yielded climactic caldera-forming eruptions. Both existing hydrothermally altered crust from Mesozoic plutons, and cannibalization of hydrothermally altered intracaldera rocks in overlapping calderas, were likely important in generating low-$\delta^{18}$O rhyolites in the Stillwater caldera complex (Fig. 12). The low-$\delta^{18}$O tuffs of Job Canyon and Poco Canyon have a combined volume of $\sim400$ km$^3$ and are the first documented low-$\delta^{18}$O rhyolites in middle Cenozoic calderas of the Great Basin, extending the geographic distribution of these isotopically distinctive and important magma types. The 2500–5000 km$^3$ tuff of Elevenmile Canyon is one of the most voluminous tuffs produced during the Great Basin ignimbrite flare-up, and has a zircon record that points to rapid batch assembly of isotopically diverse melts in the upper crust, possibly including components from previous caldera cycles, and thus it is instructive for low-$\delta^{18}$O rhyolite genesis models, even though it is not a nominally low-$\delta^{18}$O rhyolite. We emphasize the importance of analyzing nominally normal-$\delta^{18}$O rhyolites with high-spatial-resolution methods, as processes associated with crustal recycling are likely to be obscure in the Great Basin and other regions with high-$\delta^{18}$O (meta) sedimentary contributions to magmas. We note that none of the single (isolated) calderas in our regional study produced low-$\delta^{18}$O rhyolites, but that the nested Mount Jefferson calderas exhibit $\delta^{18}$O depletions through time, and that the Manhattan caldera, which may also be nested, has a distinctly lower $\delta^{18}$O signature than other calderas in the region. This evidence supports a model in which overlapping calderas facilitate the generation of low-$\delta^{18}$O rhyolites. Nested calderas of the Stillwater complex exhibit a high level of diversity in space and time. Despite spatial and temporal overlap of calderas, each caldera-forming eruption has a unique isotopic fingerprint that is apparent from single zircon analyses. Furthermore, spatially associated plutons that intrude calderas are in most cases not genetically equivalent to the caldera-forming tuffs. Our work in the Great Basin has implications for caldera-related processes globally, particularly for mechanisms of shallow crustal recycling, which can be an important component in magmatism that produces hazardous eruptions from caldera volcanoes, but one that is not commonly identified. More isotopic studies are required to assess their global occurrence and significance.

**ACKNOWLEDGMENTS**

Aki Ishida and Noriko Kita are thanked for providing guidance during preparation and use of the WescSIMS ion microprobe, and Kim Klaußnen is thanked for her assistance in data collection. Peter Vikre and Steve Ludington provided helpful conversations about the regional basement geology in western Nevada. We thank Brian Cousins for performing strontium isotope analyses at Carleton University and for sharing his graduate student Kim’s time to travel and participate in analytical work at the WescSIMS laboratory. The following individuals are thanked for their guidance during analytical sessions: Jim Palandri at the University of Oregon Stable Isotope Laboratory, Jorge Vazquez and Matt Coble at the Stanford–USGS SHRIMP-RG laboratory, and Leslie O’Brien at the USGS Menlo Park microanalytical facility. We acknowledge Mark Stelten, Henrietta Cathey, and an anonymous reviewer for their careful reviews and

Geological Society of America Bulletin, v. 131, no. 7/8

1153

Downloaded from https://pubs.geoscienceworld.org/gsa/gsabulletin/article-pdf/131/7-8/1133/4729506/1133.pdf by guest
Figure 12. Schematic synthesis of petrogenetic processes in the Stillwater caldera complex. (A) Pre–Job Canyon andesite-dacite lavas and small-volume tuffs with zircon inheritance from Mesozoic basement crust, probably dominated by Cretaceous plutons. The tuff of Job Canyon magma chamber may have assimilated hydrothermally altered Cretaceous plutons to attain its low-δ¹⁸O signature. (B) Caldera-forming tuff of Job Canyon fills the Job Canyon caldera. It is represented over a thickness of ~2 km, but it is faulted and not a uniform thickness as shown in the schematic figure. Post–Job Canyon andesite-dacite lavas tap isotopically heterogeneous crust. Meteoric water penetrates the caldera. (C) IXL pluton intrudes the Job Canyon caldera. A large convective hydrothermal system driven by heat from the IXL pluton and dominated by meteoric water alters the intracaldera block, including parts of the pluton. (D) The tuff of Poco Canyon magma chamber partially overlapped the Job Canyon caldera and may have assimilated some of the hydrothermally altered Job Canyon intracaldera block. It may also have assimilated altered Cretaceous plutons. (E) Caldera-forming eruption of the tuff of Poco Canyon fills the Poco Canyon caldera. It is represented over a thickness of ~4.5 km, but it is faulted and not a uniform thickness as shown in the schematic figure. The tuff is underlain and overlain by rhyolite lava flows and intrusions. (F) Tuff of Elevenmile Canyon magmas are assembled as isotopically diverse batches that attain a homogenized value over time. The tuff of Elevenmile Canyon magma chamber overlaps the Poco Canyon caldera and may have assimilated some of the intracaldera Poco Canyon block. (G) Caldera-forming eruption of the tuff of Elevenmile Canyon is one of the largest in the Great Basin. Intracaldera tuff fills the caldera basin to a thickness of ~5 km. The composite Freeman Creek pluton intrudes rocks beneath the level of exposure of the tuff. The tuff is possibly cogenetic with the granodiorite phase of the Freeman Creek pluton. The granite phase of the Freeman Creek pluton intrudes the granodiorite phase and the composite pluton underlies both the Elevenmile Canyon and Poco Canyon calderas.
Oxygen isotopic investigation of silicic magmatism in the Stillwater caldera complex, Nevada


Coyner, A.R., and


