Geology and evolution of the McDermitt caldera, northern Nevada and southeastern Oregon, western USA

Christopher D. Henry1, Stephen B. Castor1, William A. Starkel2, Ben S. Ellis3, John A. Wolff2, Joseph A. Laravie4, William C. McIntosh5, and Matthew T. Heizler6

1Nevada Bureau of Mines and Geology, University of Nevada, 1664 N. Virginia Street, Reno, Nevada 89557, USA
2School of the Environment, Washington State University, PO Box 642812, Pullman, Washington 99164, USA
3Institute of Geochemistry and Petrology, ETH Zurich, NW Clausiusstrasse 25, 8092 Zurich, Switzerland
4Great Basin GIS, 803 Clover Drive, Spring Creek, Nevada 89815, USA
5New Mexico Bureau of Geology and Mineral Resources, New Mexico Institute of Mining and Technology, 801 Leroy Place, Socorro, New Mexico 87801, USA
6Great Basin Geology, 4039 S. Silverado Drive, Reno, Nevada 89509, USA

ABSTRACT

The McDermitt caldera (western USA) is commonly considered the point of origin of the Yellowstone hotspot, yet until now no geologic map existed of the caldera and its geology and development were incompletely documented. We developed a comprehensive geologic framework through detailed reconnaissance geologic mapping, extensive petrographic and chemical analysis, and high-precision 40Ar/39Ar dating. The caldera formed during eruption of the 16.39 ± 0.02 Ma n = 3 McDermitt Tuff (named here), which is strongly zoned from peralkaline, aphyric, high-Si rhyolite (comendite) to metaluminous, abundantly anorthoclase-phryic, trachydacite, or Fe-rich anesite (icelandite).

The McDermitt caldera formed in an area that had undergone two episodes of Eocene intermediate volcanism at 47 and 39 Ma and major middle Miocene volcanism that led continuously to caldera formation. Eruption of the Steens Basalt, the oldest Miocene activity, began before 16.69 Ma. Exclusively mafic magmas erupted initially. Rhyolite lavas erupted as early as 16.69 Ma, contemporaneous with upper parts of the Steens Basalt, and the proportion of rhyolite increased steadily until eruption of the McDermitt Tuff. Precaldera silicic activity was diverse and almost entirely effusive, including metaluminous to mildly peralkaline, sparsely anorthoclase-phryic rhyolite to peralkaline, aphyric rhyolite (comendite) to metaluminous, abundantly anorthoclase-phryic, trachydacite, or Fe-rich anesite (icelandite).

The McDermitt caldera is similar to calderas of the middle Cenozoic ignimbrite flareup of the Great Basin, especially in strong compositional zoning and large volume of erupted tuff, collapse along a distinct ring-fracture system, abundance of megabreccia in intracaldera tuff, and resurgence. The greatest differences are that McDermitt is larger in area than all except a few flareup calderas and underwent far less collapse (~1 km versus 3–4 km to as much as 6 km). The McDermitt caldera may be an analog for the buried calderas of the...
central Snake River Plain, which also erupted voluminous rhyolitic tuffs. However, the Snake River Plain ignimbrites are metaluminous and homogeneous in bulk chemistry, with plagioclase, sanidine, and minor quartz as common phenocrysts, and rarely contain pumice or lithics.

**INTRODUCTION**

The McDermitt caldera is a well-known and well-exposed but poorly documented part of the Yellowstone hotspot track in northern Nevada and southeastern Oregon, USA (Figs. 1 and 2) that is important for many reasons.

1. The caldera has long been cited as the point of initiation of the hotspot track (Pierce and Morgan, 1992). The interrelated Columbia River Basalt Group and Yellowstone hotspot constitute the most prominent intraplate Cenozoic magmatic province in the United States (Pierce and Morgan, 1992, 2009; Wolff et al., 2008; Reidel et al., 2013). Although the origin of the Yellowstone hotspot is debated, in part because of the opposite Newberry trend (Fig. 1; Humphreys et al., 2000; Christiansen et al., 2002; Jordan et al., 2004; Foulger et al., 2015), it is commonly considered to have initiated at or near the McDermitt caldera, partly because McDermitt was thought to be the oldest silicic caldera of the track (Pierce and Morgan, 1992, 2009; Parsons et al., 1994; Zoback et al., 1994; Camp and Ross, 2004). Recent work shows that several calderas in northwestern Nevada are slightly older, and silicic volcanism ca. 16 Ma or older that may record initiation of the hotspot is widespread through Nevada, southeastern Oregon, and southwestern Idaho (Fig. 2; Noble et al., 1970; Rytuba and McKee, 1984; Castor and Henry, 2000; Cummings et al., 2000; Christiansen et al., 2002; Bonnichsen et al., 2008; Brueseke et al., 2008; Shervais and Hanan, 2008; Wypych et al., 2011; Coble and Mahood, 2012, 2016; Streck et al., 2015; Benson and Mahood, 2015, 2016). This wide footprint may indicate the geographic extent of the initial Yellowstone plume head (Pierce and Morgan, 2009; Coble and Mahood, 2012). The oldest rhyolites erupted contemporaneously with late parts of Steens Basalt lavas and dikes, the oldest part of Columbia River Basalt activity, and are in and south of the southern part of Steens Basalt distribution.

2. The McDermitt caldera is a large collapse feature that formed during eruption of magma that was strongly zoned from peralkaline aphyric rhyolite to metaluminous porphyritic trachydacite (Rytuba and Conrad, 1981; Conrad, 1984; Rytuba and McKee, 1984; Starkel et al., 2012, 2015).
Figure 2. Shaded relief digital elevation model of the McDermitt caldera region showing known and proposed early Yellowstone hotspot calderas and other silicic volcanic centers (Noble et al., 1970; Rytuba and McKee, 1984; Castor and Henry, 2000; Manley and McIntosh, 2002; Bonnichsen et al., 2008; Brueseke and Hart, 2008; Brueseke et al., 2008, 2014; Wypych et al., 2011; Hausback et al., 2012; Henry et al., 2012a; Coble and Mahood, 2012, 2016; Streck et al., 2015; Benson and Mahood, 2015, 2016; Benson et al., 2017; Mahood and Benson, 2017), the northern Nevada rift and related magnetic anomalies (John et al., 2000; Glen and Ponce, 2002), and areas of Eocene magmatism (Axelrod, 1966; Castor et al., 2003; Colgan et al., 2006a, 2011; Lerch et al., 2008). Our interpretation of calderas in the High Rock–Virgin Valley area shown here (Castor and Henry, 2000; Hausback et al., 2012) differs from that of Coble and Mahood (2016). CA—California; ID—Idaho; NV—Nevada; OR—Oregon.
Despite many rocks being peralkaline, the caldera area underwent at least three episodes of biotite-bearing, metaluminous rhyolite eruptions that raise interesting questions about the petrogenesis of all silicic magmatism at McDermitt.

3. The caldera is well exposed, has been negligibly affected by later extension, and nicely illustrates many aspects of caldera development, including abundant precaldera magmatism, voluminous caldera-forming eruption, a complex ring-fracture fault system, and post-collapse magmatism and resurgence. Both intracaldera and outflow McDermitt Tuff, the caldera-forming tuff, are intensely rheomorphic to the degree that their pyroclastic origin is substantially obscured in many outcrops. However, the occurrence of abundant, voluminous rhyolitic lavas aids in the discrimination of pyroclastic and effusive deposits. The McDermitt Tuff has both similarities and differences with voluminous rhyolitic ignimbrites of the central Snake River Plain, where source calderas are entirely buried, but may have an analog in the McDermitt caldera.

4. Although mercury is the best-known element of economic significance related to the caldera and the only one that has had significant mining, the caldera also contains known deposits or concentrations of uranium, lithium, gallium, gold, and zirconium. Most of these deposits are, or probably are, related to caldera igneous activity and structures. The caldera contains significantly more mineral potential than has been exploited.

Despite this significance, the basic geology of the caldera and its spatial, temporal, and geochemical evolution are not well known. Our efforts to understand McDermitt geology include detailed and reconnaissance geologic mapping of the caldera and construction of a 1:70,000-scale map (Plate 1), 40Ar/39Ar dating of the sequence of igneous rocks and several of the mineral deposits, and geochemical analysis of the igneous rocks (this study; Starkel, 2014; Starkel et al., 2012, 2014).

## STUDY METHODS

The major objectives of this study included (1) deciphering the igneous and structural evolution of the McDermitt caldera, (2) assessing published multicaldera interpretations and proposed ash-flow tuff sources (Rytuba and McKee, 1984), (3) estimating the area and amount of collapse relevant to general caldera evolution models, (4) providing a comprehensive geologic framework for thorough petrogenetic studies (Starkel, 2014), and (5) analyzing the association of numerous mineral deposits and occurrences in and around the caldera with caldera igneous activity and structure. To accomplish these objectives, we mapped the geology of the caldera at an overall scale of 1:70,000 (Plate 1) through a combination of detailed 1:24,000-scale mapping in selected areas and less detailed mapping throughout. We also conducted petrographic, geochemical, and geochronologic analyses of samples that span the Cenozoic.

Phenocryst contents and abundances were determined by point counts and visual estimates of more than 100 thin sections of representative samples of all rock types (Table 1). Chemical analyses were obtained of 102 rocks spanning all the intermediate to silicic igneous activity around the McDermitt caldera by X-ray fluorescence and inductively coupled plasma-mass spectrometry (ICP-MS) at Washington State University (http://cahnrs.wsu.edu/soe/facilities/geolab/) or by ICP-optical emission spectrometry (major oxides) and ICP-MS (trace elements) at ALS Chemex (http://www.alsglobal.com/services-and-products/geochemistry/geochemistry-testing-and-analysis; method ME-MS81d) (Supplemental Table S1). Seven samples were analyzed at both laboratories; analyses agree well for the same constituents. We also used six published analyses, which were of major oxides only, from Wallace et al. (1980) and Castor et al. (1982).

Determining the timing of igneous activity in and around the McDermitt caldera was a critical aspect of this study. Published K-Ar dates demonstrate that the McDermitt caldera is middle Miocene (Greene, 1976; Rytuba and McKee, 1984) but have large uncertainties, and some are demonstrably inaccurate, so are of limited use. Therefore, we obtained single crystal 40Ar/39Ar dates on sanidine and anorthoclase from 26 samples, which provide the most precise reproducible ages, and step-heating dates on 5 plagioclase, 1 matrix, and 3 potassium feldspars formed during mineralization (Table 2). Table 2 also lists the summed age data of three or four individual sanidine dates on each of the regional tuffs of the Trout Creek Mountains and Oregon Canyon, respectively. All ages in this report are calculated relative to the Fish Canyon Tuff sanidine standard at 28.201 Ma (Kuiper et al., 2008) and a total 8K decay constant of 5.483e-10/a (Min et al., 2000).

Initial dating was done with a single collector MAP215–50, whereas dating since 2012 was done with a multicolonlector Argus VI instrument, which is significantly more sensitive and precise (Heintz et al., 2014; Heizler et al., 2011; Heizler et al., 2014; Hereford et al., 2016). With these advances in mass spectrometry, the greatest limiting factors in precise dating are the quality of dated material and reduction of reactor fluence gradients. ARGUS VI data yield lower standard deviation, but higher MSWD (mean square of weighted deviates) values compared to MAP215–50 data (Supplementary Table S2). That is, the mean is more tightly constrained (accurate) with the ultrahigh precision analysis of the ARGUS VI, but the data are overdispersed relative to the cited precision. Much of the overdispersion can be accounted for when considering both vertical and circumferal fluence gradients in the irradiation trays. Despite tightly packaging crystals in sub–2 mm irradiation pits that are not deeper than 4 mm, the grain to grain difference in fluence is measurable. This variation causes the high MSWD values for the ARGUS VI data. There is probably geological variation at the sub-per mil level that is now detectable, but until fluence gradients can be further reduced, assessing geological scatter is somewhat ambiguous. If we assume that all scatter is reactor related, and if all grains from all holes (monitors and unknowns) are analyzed, the weighted mean values can be used to calculate age and J-factor and the uncertainty would be proportional to the sum of the inverse variance of each analysis. However, we take a more conservative approach to determining uncertainty by increasing the calculated weighted mean error by the square root of the MSWD value. This is conservative because a significant contribution to the high MSWD values is grain to grain variation of neutron flux.
Some thin surficial Quaternary units have been omitted.

GEOLOGIC MAP OF THE MCDERMITT CALDERA, HUMBOLDT COUNTY, NEVADA, AND HAREY AND MALHEU COUNTIES, OREGON

Christopher D. Henry*, Stephen B. Castor, William A. Starkel, Ben S. Ellis
John A. Wolf3, Joseph A. Lavelle4, William G. McIntosh5, and Matthew T. Hinzler 3

1 Nevada Bureau of Mines and Geology, University of Nevada, Reno
2 School of the Environment, Washington State University, Pullman
3 Institute of Geology and Petrology, B.C. Ministry of Energy and Mines, Victoria, Canada
4 New Mexico Bureau of Geology and Mineral Resources, New Mexico Tech
5 New Mexico Bureau of Geology and Mineral Resources, New Mexico Tech

2017

CONTOUR INTERVAL 50 METERS

30'x60' quadrangle (1985)

4 1070

Plate 1. Geologic map of the McDermitt caldera by Henry et al. (2016). To view Plate 1 at full size, please visit http://doi.org/10.1130/GES01454.55.
### TABLE 1. COMPOSITIONAL AND PETROGRAPHIC CHARACTERISTICS OF IGNEOUS ROCKS OF THE McDERMITT CALDERA

| Rock type | Map unit | Age (Ma) | Rock name* | SiO₂ † | FeO | Aℓ | Na₂O+K₂O | Total | Plagioclase | Sanidine | Anorthoclase | Quartz | Amphibole | Clinopyroxene | Biotite | Olivine | Other |
|-----------|---------|----------|------------|--------|-----|----|----------|-------|-------------|----------|-------------|--------|----------|--------------|--------|---------|-------|
| Postcaldera late basaltic lavas | Tbl | 14.9 | HAOT | 50 | 9.9 | 0.24 | 2.7 | 0–20 | 0–20, 1 x 4 | 0–2, 1/1 | 0–1, 1/1 |
| Fe-rich andesite (icelandite) domes and lavas southwestern caldera | Tbl | 16.1 | FA | 56 | 11 | 0.6 | 6.2 | 5 | 4, 15 | s1, to 8 |
| Aurora lavas | Tbl | 16.41 | FA | 60–64 | 9.4–8.5 | 0.82–0.78 | 7.8–8.8 | 2–3 | 2–3, 2 | 2–4, 2 | fa, s1, 1/1 |
| Black Mountain | Tiv, Tbl | 6.7 | FA | 60–62 | 9.4–9.3 | 0.70–0.74 | 7.2–7.3 | 5–7 | 5, 2 | fa, s1, 1/1 |
| Round Mountain | Tiv | 16.6 | FA | 61–62 | 9.4–8.8 | 0.64–0.73 | 7.4–6.7 | 5–15 | 4–15, 4–2, ≤1 | 0–1, ≤1 |
| Rhyolite of McDermitt Creek (McDermitt Tuff?) | Trm | 16.41 | R | 70–73 | 4.5–3.7–2.8 | 0.85–0.91 | 9.5–9.0 | 1–2, 1–2 | 15–25, 1–6 | <1 | fa, trace | ap |
| McDermitt Tuff | 16.39 | TR | 74–77 | 3.8–2.8 | 1.00–1.12 | 8.2–9.4 | 0–5 | trace | 0–5, to 2 | naa, trace, 0.2 |
| Pre-caldera peralkaline rhyolite lava | Tpr | 16.40 | PR | 76 | 2.5–3.3 | 1.0–1.16 | 8.6–9.7 | 5–30 | 5–30, to 2.5 | trace–1, to 1 | fa, s1, to 1 |
| Biotite rhyolites biotite rhyolite lava of Hoppin Peaks | Tbr | 16.36 | MR | 75 | 1.7–1.9 | 0.86 | 8.6–8.8 | 3–6 | trace–2, 1–2 | 1–2, 1–2 | trace–1, 1–1.5 | fa, trace–1, s1, 1/1 |
| Biotite rhyolite of Mendicuri Suri | Tbr | 16.49 | MR | 76 | 0.9 | 0.93 | 9 | 6 | 1.5, 2 | 2, 2–3 | 1, 1.5 | hbl, trace, 0.3 | <1, 0.5 |
| Quartz-sandine rhyolite | Trq | 16.61 | MR | 76–77 | 1.6–1.9 | 0.88–0.90 | 8.6–8.4 | 15–20 | 2, 0.2–2 | 5–7–15 | trace | 0.5 |
| Biotite rhyolite of Moonlight Mine | Tbr | 16.62 | MR | 71–75 | 2.6–1.9 | 0.84–0.98 | 10.4–9.4 | 10–16 | 6–10, 1.5 | 1–3, 2.5 | trace | <1 |
| Distal ash-flow tuffs tuff of Trout Creek Mountains | Ttt | 16.49 | PR | 74–76 | 3.6–4.5 | 1.07–1.24 | 9.0–9.4 | 14–20 | 6–7, 2 | 5–12, 2–3 | naa, <1, 0.8 | trace–0.3, s1 |
| Tuff of Oregon Canyon | Tto | 16.56 | PR–TD | 77–67 | 2.7–5.7 | 1.00–0.96 | 9.8–8.7 | 6–15 | 4–8, 2–4 | 1–4, 1–2.5 | 0–1, 1 |
| Anorthoclase-phyric ash-flow tuff | Tta | 16.65 | R | 74 | 2.1 | 0.93 | 9.5 | 10 | 8, 1–3 | trace, 0.5 | <1, 1 |
| Aphryic and sparsely porphyritic rhyolite lavas aphyric rhyolite | Tar, Taro, Tarv | 16.35 | AR | 75–76 | 3.1 | 1.12–1.17 | 9.0–9.5 | 0 |
| Anorthoclase rhyolite | Taw, Tra, Tram, Trau | 16.35 | &R | 75–76 | 3.2 | 1.07–1.16 | 9.0–9.3 | 1–3 | 1–2, ≤2 | <1, 0.5 |
| Anorthoclase rhyolite | 16.46 | &R | 72 | 1.8 | 0.93 | 10.3 | 4–5 | 4, ≤1.5 | trace, ≤0.4 |
| Precaldera intermediate lavas basaltic andesite to dacitic lava | Tpi | 16.56 | BA | 55–62 | 11.4–6.5 | 0.60–0.65 | 5.8–7.8 | 0–5 | 0–5, 1–15 | trace–1, s1 | fa |
| Steens Basalt | Tpb | 16.40 | FA | 62–67 | 9.4–4.8 | 0.57–0.87 | 6.2–8.5 | 5–15 | 4–10, 1–3 | 0–2, 1–2 | ≤1, 0.8 |
| Steens Basalt | 16.56 | B | 49–54 | 12.0–10.3 | 0.37–0.57 | 3.7–5.3 | 0–5 | ≤4, to 2 | ≤1 | ≤1 |
| Steens Basalt | 16.69 | PB | 46–51 | 13.7–11.6 | 0.29–0.42 | 2.9–4.5 | ~40 | ≤0.9, to 60 | olivine |
| Eocene younger Eocene andesite | Tv2 | 39.2 | A | 60–62 | 6.6–6.3 | 0.62–0.65 | 8.1–7.7 | 30 | 25, 5 | <1, 1 | 1, 3, 1 | ap |
| Older Eocene dacite and andesite | Tv1 | 46.7 | A–D | 62–64 | 5.3–4.0 | 0.52–0.57 | 6.5–6.9 | 35 | 23, 3–4 | hbl, 5, 1.3 | 4–5, 4, zr |

*R—rhyolite, PR—peralkaline rhyolite, AR—aphryic rhyolite, MR—metametavolcanic to peraluminous rhyolite, BA—basaltic andesite, FA—Fe-rich andesite (icelandite), B—basalt, PB—porphyritic basalt, TD—trachydacite, D—porphyritic dacite, A—andesite, HAOT—high–alumina olivine tholeiite.

† Weight percent, major elements normalized to 100% volatile free.

‡ AI—anapatite or peralkalinity index (molar Na+K/Al).

**Total—mean or range in volume percent followed by size in mm. Abbreviations: tr—trace, hbl—hornblende, fa—fayalite, all—allanite, ap—apatite, zr—zircon, aen—aenigmatite, mon—monazite, naa—sodic amphibole.
To improve sample quality, steps in addition to typical magnetic and density mineral separation are taken. All feldspars were leached with 5% HF for ~1 h to remove adhering matrix, alteration products, or other non-feldspar material. Any grains containing such material or alteration products along cleavage planes. Any grains containing such material could be identified with binocular and petrographic microscope examination were discarded, but not all non-feldspar material is.

| Area | Sample | Unit | Mineral | Age (Ma) | ±2σ | K/Ca | ±2σ | n1 | MSWD |
|------|--------|------|---------|----------|------|------|------|-----|------|
| NW   | MD-8   | Tuff of Whitehorse Creek | sanidine | 15.67 | 0.05 | 49.8 | 126 | 18/20 | 0.94 |
| N    | H11-199| Tia, Aurora lava, icelandite | anorthoclase | 16.412 | 0.016 | 4.1 | 3.1 | 13/14 | 3.61 |
| NE   | H04-55 | Trm, rhyolite of McDermitt Creek (Greene, 1976) | anorthoclase | 16.417 | 0.018 | 1.3 | 1.0 | 15/15 | 3.09 |
| NE   | 10-McD-225 | Trm, rhyolite of McDermitt Creek (Greene, 1976) | plag/ anorthoclase | 16.39 | 0.10 | 0.8 | 0.9 | 9/15 | 0.82 |
| NW   | H15-67 | Tmt, northwestern outflow | anorthoclase | 16.403 | 0.009 | 10.6 | 3.2 | 23/25 | 2.00 |
| NW   | H12-259| Tmv, northwestern tuff dike | anorthoclase | 16.374 | 0.012 | 6.8 | 5.3 | 12/18 | 2.60 |
| N    | H05-23P| Tmt, northern outflow | anorthoclase | 16.386 | 0.008 | 6.0 | 3.3 | 12/13 | 2.97 |
| N    | H04-40 | Tmt, northern intracaldera, highly rheomorphic | anorthoclase | 16.348 | 0.014 | 6.4 | 1.9 | 14/18 | 4.00 |
| W    | H05-26 | Tral, anorthoclase rhyolite lava | sanidine | 16.301 | 0.024 | 4.3 | 3.4 | 15/16 | 6.19 |
| S    | H07-191| Tmt, southern outflow | anorthoclase | 16.346 | 0.010 | 7.4 | 5.1 | 8/8 | 4.05 |
| N    | H04-39 | Tra, anorthoclase rhyolite lava | anorthoclase | 16.35 | 0.04 | 15/22 | 0.75 |
| E    | H04-51 | Tbr1, biotite rhyolite lava of Hoppin Peak, lowest | sanidine | 16.384 | 0.069 | 24.4 | 13.3 | 15/15 | 8.46 |
| E    | H04-49 | Trau, anorthoclase rhyolite lava | anorthoclase | 16.397 | 0.061 | 7.4 | 4.4 | 14/15 | 8.16 |
| W    | C95-107| Tpr, peralkaline rhyolite lava, Moonlight Mine | sanidine | 16.399 | 0.016 | 133 | 159 | 14/14 | 16.39 |
| SE   | H12-250| Tpr, peralkaline rhyolite lava, southeast caldera wall | sanidine | 16.427 | 0.015 | 137 | 26 | 15/15 | 13.64 |
| SW   | 11-McD-290| Tpr, peralkaline rhyolite lava, south of caldera | sanidine | 16.404 | 0.018 | 256 | 743 | 7/7 | 9.98 |
| NE   | H04-47 | Tra, anorthoclase rhyolite lava, Oregon Canyon | sanidine | 16.40 | 0.04 | 17.2 | 19/19 | 2.19 |
| N    | C04-07 | Tral, anorthoclase rhyolite lava | anorthoclase | 16.462 | 0.018 | 6.3 | 1.4 | 12/13 | 8.94 |
| NE   | H05-28 | Tbrm, biotite rhyolite dome of Mendi Suri | sanidine | 16.487 | 0.015 | 83.2 | 47.7 | 3/3 | 7.79 |
| E    | 09-McD-33| Trm, anorthoclase rhyolite lava (McConnell Canyon) | sanidine | 16.496 | 0.079 | 2.5 | 2.4 | 13/14 | 17.93 |
| N    | mean of 3 | Tt, tuff of Trout Creek Mountains | sanidine | 16.49 | 0.03 | ~500 | 35/35 |
| N    | mean of 4 | Tt, tuff of Oregon Canyon | sanidine | 16.57 | 0.02 | ~200 | 49/49 |
| N    | H04-44 | Tio, tuff of Oregon Canyon | sanidine | 16.517 | 0.013 | 203 | 247 | 18/19 | 7.49 |
| NE   | H16-12 | Trq, quartz-sandine rhyolite lava dome | sanidine | 16.607 | 0.015 | 27.8 | 14.6 | 24/26 | 5.50 |
| W    | C95-103A| Tbr, biotite rhyolite dome, Moonlight Mine | sanidine | 16.619 | 0.020 | 26.7 | 13.4 | 13/14 | 55.34 |
| W    | C05-280 | Tia, anorthoclase-phric tuff | anorthoclase | 16.651 | 0.008 | 8.8 | 4.3 | 12/16 | 15.88 |
| NE   | H05-26 | Tra, anorthoclase rhyolite lava | anorthoclase | 16.691 | 0.021 | 5.9 | 1.9 | 11/13 | 29.78 |

---

**PREVIOUS WORK**

Greene (1972, 1976) first recognized the caldera from geologic mapping of the northeastern part (Jordan Meadow 15’ quadrangle; Greene, 1972) and reconnaissance in surrounding areas; he recognized many of the essential rock units, much of the northern caldera fault and flexure boundary, and that intra-
TABLE 2. 40Ar/39Ar DATES OF IGNEOUS ROCKS AND MINERALIZATION, McDERMITT CALDERA (continued)

| Area* | Sample | Unit | Mineral | Age (Ma) | λ | K/Ca | σ | n | MSWD | Lab | Lat | Long | Mass spectrometer |
|-------|--------|------|---------|----------|---|------|---|---|------|-----|-----|------|-----------------|
| S     | WLC43-180.2 | K-feldspar in Li deposits | 14.87 | 0.05 | 54.4 | 7/12 | 15.24 | 0.26 | 173 | 78 | 14.92 | 0.03 | 61056 | 41.70949 | -118.06058 | Argus VI |
| W     | C95-105  | Adularia, Moonlight Mine U mineralization | 16.32 | 0.10 | 94.9 | 9/13 | 16.38 | 0.25 | 5990 | 41.7880 | -118.1597 | Argus VI |
| NE    | McD     | Adularia, McDermitt Mine Hg mineralization | no plateau | 16.67 | 0.14 | 295 | 1.2 | 16.56 | 0.12 | 64958 | Argus VI |
| Postcaldera rocks | SE | WLC67-13.5 | Tbl, late basalt (high-alumina olivine tholeite) | matrix | 14.90 | 0.74 | 92.6 | 10/12 | 15.89 | 1.10 | 292 | 6 | 15.18 | 1.16 | 61065 | 41.71017 | -118.07190 | Argus VI |
| SC    | H07-198  | Tv1, Fe-rich andesite (icelandite) vent | plagioclase | 16.22 | 0.13 | 89.8 | 11/13 | 16.32 | 0.23 | 283 | 26 | 16.11 | 0.19 | 64197 | 41.81757 | -118.05543 | Argus VI |
| SW    | 09-McD-53 | Ti, Fe-rich andesite (icelandite) lava | plagioclase | 16.08 | 0.10 | 93.8 | 9/10 | 16.12 | 0.03 | 273 | 13 | 15.76 | 0.10 | 62899 | 41.74001 | -118.10789 | Argus VI |
| Miocene precaldera rocks | N | C04-06 | Tpi, precaldera intermediate lava, icelandite | plagioclase | 16.53 | 0.16 | 80.6 | 8/10 | 16.54 | 0.19 | 295 | 12 | 16.57 | 0.15 | 55652 | 42.04800 | -117.87212 | MAP215-50 |
| Eocene volcanic rocks | W | 11-McD-311 | Tv2, andesite lava | plagioclase | 39.18 | 0.12 | 94.2 | 7/8 | 39.09 | 0.22 | 320 | 87 | 39.02 | 0.17 | 61052 | 41.87447 | -118.15990 | Argus VI |
| W     | MD-6     | Tv1, dacite lava | plagioclase | 46.74 | 0.41 | 80.4 | 8/10 | 46.91 | 0.27 | 277 | 15 | 46.30 | 0.28 | 55651 | 41.91674 | -118.18456 | MAP215-50 |

Note: Ages in bold are best estimates of eruption or deposition age. Weighted mean (wtd mn) 40Ar/39Ar ages of feldspars (plag—plagioclase) calculated with inverse variance as weighting factor. Uncertainty is after Taylor (1982) (times square root mean square of weighted deviates, MSWD, when MSWD > 1). Error also includes uncertainty of J-factor and reactor interference correction factors. All 40Ar/39Ar analyses were done at the New Mexico Geochronological Research Laboratory (NMGRl; methodology in McIntosh et al., 2003). Neutron flux monitor Fish Canyon Tuff sandine (FC-1) with assigned age of 28.201 Ma (Kuiper et al., 2006) Minerals were separated from crushed, sieved samples by standard magnetic and density techniques, feldspars were leached with dilute HF to remove matrix, and all were hand-picked. Decay constants are after Min et al. (2000); ttotal = 5.463 x 10^{-10} yr^-1. Isotopic abundances are after Steiger and Jäger (1977); 40K/39K = 1.167 x 10^-6.

*Area of caldera or caldera wall; N—north, S—south, E—east, W—west, NW—northwest, NE—northeast, SE—southeast, SW—southwest, SC—south central. Italics indicate outside map area. NAD27—North American Datum of 1927.

**Laboratory number (NMGRl analysis number, tied to Supplemental Table S2 [see text footnote 2]).

** GEOLOGY OF THE McDERMITT CALDERA

Pre-Cenozoic Rocks

The oldest exposed rocks in the map area are Cretaceous granodiorite (Figs. 4A, 4B) and abundant crosscutting felsic to mafic dikes. Large irregular bodies of diorite cut the granodiorite, and aplite dikes cut both. The granodiorite is undated but almost certainly Cretaceous and possibly the same age as plutons in the Santa Rosa Range 20 km to the east, where U-Pb zircon dates reveal 2 pulses ca. 101 Ma and 94 Ma (Brown and Hart, 2013). Triassic–Jurassic metasedimentary rocks that are the host rocks for the plutons in adjacent ranges (Compton, 1960; Wyld et al., 2003; Purcell and Hart, 2012) are not exposed in the McDermitt map area but probably occur in the subsurface.

Eocene Volcanic Rocks (Tv1, Tv2)

Two sequences of Eocene andesite to dacite lavas that crop out in the northwestern caldera wall are the oldest Cenozoic rocks in the McDermitt area. Both sequences directly overlie an irregular surface cut into Cretaceous granitic rock and are overlain by middle Miocene Steens Basalt (Fig. 4A). The older group (Tv1; see Plate 1 for map unit symbols), with a plagioclase 40Ar/39Ar date of 46.74 ± 0.41 Ma (MD-6, Table 2), crops out in a 3 x 2 km area –3 km...
Figure 3. Google Earth image of McDermitt caldera showing caldera and faults interpreted in this study, approximate limit of outflow McDermitt Tuff (Tmt; from Rytuba and Curtis, 1983; Rytuba et al., 1983a, 1983b; Minor, 1986; Minor et al., 1989a, 1989b; and this study), and proposed calderas of Rytuba and McKee (1984) with names in yellow type and calderas of Whitehorse area of Mahood and Benson (2017) with names in blue type.
south of Disaster Peak (Fig. 3; Plate 1). The unit consists of porphyritic andesite to dacite lava flows and minor dikes (~62%–64% SiO₂) with phenocrysts of plagioclase, biotite, and hornblende, and interbedded volcaniclastic rocks containing angular to rounded blocks of volcanic rocks and sparse granitic rocks to 40 cm in diameter. The total sequence is as much as 200 m thick.

The younger group (Tv2), with a plagioclase ⁴⁰Ar/³⁹Ar date of 39.18 ± 0.12 Ma (1 1-McD-31 1, Table 2), consists of several, poorly exposed flows that total ~125 m in thickness and crop out in a narrow 4 × 0.4 km belt just south of Tv1. A single analyzed sample has 60% SiO₂ and phenocrysts of plagioclase, biotite, and clinopyroxene.

The Eocene rocks are calc-alkaline, metaluminous, and I-type, which distinguishes them from all the Miocene rocks, which are A-type (Fig. 5). Lower peralkalinity mostly reflects much higher Al₂O₃, because the Eocene and Miocene rocks have similar total alkalies (Fig. 6). The Eocene rocks also have lower FeO, MnO, TiO₂, Y, and heavy rare earth elements (HREEs), and variably higher CaO, Sr, Cr, and Th than Miocene rocks but are similar in other major oxides and trace elements.

The Eocene rocks are part of the Eocene–Oligocene southwestward sweep of magmatism across the Great Basin, probably resulting from rollback of the shallow Laramide slab (Humphreys, 1995; Henry and John, 2013; Best et al., 2016). Late Eocene intermediate volcanic rocks crop out throughout northern Nevada (Fig. 2), as far west as the Warner Range in California (Colgan et al., 2011), and at least as far east as the Tuscarora volcanic field (Castor et al., 2003). The nearest similar rocks are 39 Ma, 60%–63% SiO₂, plagioclase-hornblende-biotite trachyandesite intrusions in the Pine Forest Range, 50 km to the west-southwest (Colgan et al., 2006a). Andesite to dacite lavas erupted in several episodes in the 40–39 Ma Tuscarora volcanic field, ~150 km to the east-southeast (Castor et al., 2003). Other late Eocene rocks are 36–35 Ma basaltic lava flows in the Black Rock and Santa Rosa Ranges (Lerch et al., 2008; Brueseke and Hart, 2008).
Figure 5. (A) Total alkali (Na₂O + K₂O)-SiO₂ plot of igneous rocks of the McDermitt caldera. (B) Peralkalinity versus alumina index of McDermitt caldera rocks. (C) Zr versus 10⁴ * Ga/Al plot.
Figure 6. Harker variation diagrams of McDermitt igneous rocks. (A) TiO$_2$. (B) Al$_2$O$_3$. (C) FeO*. (D) Mol (Na$_2$O + K$_2$O)/Al$_2$O$_3$. (E) MnO. (F) Zr (ppm). (G) La/Yb.
Middle Miocene Precaldera Volcanism

The following discussion is mostly arranged by age, oldest to youngest. However, distinct categories of rocks, (1) Steens Basalt and precaldera intermediate lavas, (2) sparsely porphyritic to aphyric rhyolite lavas, (3) distal ash-flow tuffs, and (4) biotite-bearing rhyolites, are grouped because of their petrographic, compositional, and possible genetic relationships.

Volcaniclastic Sedimentary Deposits (Tvs)

A probable middle Miocene, volcaniclastic sedimentary sequence (Tvs) that consists of conglomerate, sandstone, mudstone, and siltstone crops out in the caldera wall east of Horse Creek. This unit filled a topographic low at least 100 m deep cut into Cretaceous granitic rock and underlies the Steens Basalt (Tsb). Conglomerate layers contain small (≤10 cm, most ≤1 cm) clasts of granitic rock, probable Eocene volcanic rocks, and possible Miocene volcanic rocks. The presence of accretionary lapilli-bearing, shard-bearing, and biotite-bearing ash beds in the upper part of the unit and fining-upward bed sets indicate deposition during active volcanism, which was probably the earliest part of middle Miocene activity.

Steens Basalt (Tsb, Tb) and Precaldera Intermediate Lavas (Tpi)

Middle Miocene volcanism began in the McDermitt caldera area before 16.891 ± 0.021 Ma, the oldest dated rock of this study and ca. 300 ka before eruption of the McDermitt Tuff and caldera collapse (Table 2). Precaldera rocks are predominantly Steens Basalt (Tsb, Tb) and intermediate composition lavas (Tpi), which are extensively exposed around the northern half of the caldera and regionally (Fig. 4B; Greene, 1972, 1976; Brueseke et al., 2007; Bondre and Hart, 2008; Brueseke and Hart, 2008; Jarboe et al., 2010; Camp et al., 2013), along with numerous interbedded local to distal silicic lavas and tuffs. Erupted rocks generally became more silicic through time, although rhyolite lavas are locally interbedded with Steens Basalt low in the stratigraphic section. These rhyolitic units are particularly important because they are dated more precisely than the more mafic rocks, and so provide tight age constraints on Steens magmatism (Henry et al., 2006; Mahood and Benson, 2016, 2017).

Definition of Steens Basalt. Camp et al. (2013) reviewed the controversy over what constitutes the Steens Basalt, stratigraphic unit versus geochemical rock type (e.g., Brueseke et al., 2007), and formally proposed stratigraphic limits to the unit. Based on 2 measured sections 13 km northwest and 15 km north of the northern margin of the McDermitt caldera, Camp et al. (2013) restricted the Steens Basalt to rocks below the tuffs of Oregon Canyon or Trout Creek Mountains, distinct marker units that crop out along the northern margin. Average single-collector MAP215-50 sanidine 40Ar/39Ar dates for these tuffs are 16.57 ± 0.02 Ma and 16.49 ± 0.03 Ma, respectively (n = 4 and 3, respectively, Table 2; Fig. 7), and a new multicollector Argus VI date on the tuff of Oregon Canyon is 16.517 ± 0.013 Ma (H04-44; Table 2).

Steens Basalt (Tsb, Ts) around the McDermitt caldera. Mafic lavas around the McDermitt caldera underlie the tuffs of Oregon Canyon and Trout Creek Mountains, fit the formal definition of Camp et al. (2013), and are here termed the Steens Basalt. We divide the Steens Basalt into two units based on stratigraphy and the presence of coarsely plagioclase-phryic lavas common to the Steens Basalt (Fig. 4C). The lower unit Tsb consists of numerous thin (≤2 m) basalt to basaltic andesite lavas that include both aphyric and coarsely plagioclase-phryic lavas. The porphyritic rocks have plagioclase phenocrysts as much as 6 cm long as well as 2–3-mm-diameter olivine and clinopyroxene. Unit Tsb is ~300 m thick at Disaster Peak, the best-exposed and most complete section; Tsb is equivalent to unit Ts (Steens Basalt) of Minor et al. (1989b) at Catlow Peak along the northwest continuation of the Disaster Peak–Orevada View ridge (Fig. 3). An overlying unit Ts ~75 m thick is predominantly aphyric basalt flows and is stratigraphically equivalent to unit Taf (andesitic flows) of Minor et al. (1989b), which is basalt according to the analyses of Camp et al. (2013). Detailed petrographic and geochemical studies demonstrate that both Tsb and Ts of this study (Starkel, 2013; Starkel et al., 2014) and Ts and Taf of Minor et al. (1989b) at Catlow Peak (Camp et al., 2013) are Steens Basalt. Minor et al. (1989b) mapped tuff of Oregon Canyon capping their unit Taf at Catlow Peak.

Unit Tsb crops out only around the northernmost wall of the caldera as far south as Horse Creek. Unit Ts occurs in the same area as well as across most of the northern caldera wall. Rocks of both units may underlie younger rocks farther south.

Preceding Intermediate lavas (Tpi). Effusive volcanism continued after emplacement of the tuff of Oregon Canyon but became more silicic than the Steens Basalt. Unit Tpi consists of intermediate (~56%–66% SiO2; Figs. 5 and 6), aphyric to sparsely plagioclase-phryic lavas that overlie the tuff of Oregon Canyon or tuff of Trout Creek Mountains where the former is absent. The flows may be as much as 425 m thick in the northern caldera wall, although the sequence may be repeated by unrecognized faults resulting from caldera collapse, and the bounding tuffs crop out discontinuously. Unit Tpi is equivalent to unit Tif (intermediate flows) of Minor et al. (1989b) northwest of the caldera.

Unit Tpi and other Miocene intermediate rocks of the McDermitt caldera generally match the characteristics of andesite, i.e., a high FeO*, TiO2, MnO, low Al2O3, and simple phenocryst assemblage, compared to calc-alkaline andesites (Fig. 6; Carmichael, 1964), and the term commonly has been applied to the McDermitt rocks (Wallace et al., 1980; Noble et al., 1988). McDermitt rocks are significantly more potassic and span a wider SiO2 range than the original Icelandites (59%–65% SiO2; Carmichael, 1964). However, others have applied the term to a wider SiO2 range and moderately potassic rocks (e.g., 54%–63%; van der Zander et al., 2010). Many McDermitt rocks fit either SiO2 range, so we use the term here despite its imperfect match with the original Icelandites.

Unit Tpi consists mostly of Icelandites or trachyandesites with 8%–10% FeO* (Wallace et al., 1980; Figs. 5 and 6). As with all Miocene rocks, they plot in the A-type field on a Ga/Al vs Zr plot; they are along linear trends with other
Miocene rocks in most major oxide and trace element plots, e.g., FeO* and Zr. However, they have distinctly lower MnO and Ba and higher Sr contents at a given SiO2 than do the least silicic rocks of the McDermitt Tuff (Fig. 6).

Plagioclase from an icelandite lava that underlies the tuff of Trout Creek Mountains at Flattop in the northeastern caldera wall yielded a 40Ar/39Ar date of 16.53 ± 0.16 Ma (C04-06, Table 2; Fig. 4D). The precaldera intermediate lavas can be no younger than ca. 16.41 Ma, the average age of the peralkaline rhyolite lava (Tpr; Table 2), which overlies intermediate lavas in the eastern caldera wall and at the Moonlight Mine (Plate 1).

Two small outcrops of mafic icelandite (~56%–57% SiO2) that appear beneath rhyolite lavas in the southeastern caldera wall are included in unit Tpi. These rocks are demonstrably older than 16.41 Ma, and the tuffs of Oregon Canyon and Trout Creek Mountains are not exposed in this area. Their presence indicates that precaldera intermediate lavas (Tpi) occur at least beneath the southern part of the caldera. Younger rhyolites cover all rocks farther south. However, aeromagnetic highs that coincide with the western faulted edge of the Double H Mountains (U.S. Geological Survey, 1972) suggest that intermediate or mafic rocks probably underlie the rhyolites.

Figure 7. Timing of middle Miocene igneous activity in and around the McDermitt caldera. Activity began with eruptions of Steens Basalt between ca. 16.85 and 16.74 Ma (Camp et al., 2013; Mahood and Benson, 2016) and progressed continuously to eruption of the McDermitt Tuff ca. 16.39 ± 0.02 Ma, then was followed probably more sporadically by postcaldera icelandites to ca. 16.1 Ma. The tuff of Whitehorse Creek erupted ca. 15.7 Ma, possibly from a caldera northwest of the McDermitt caldera (Fig. 3). The youngest igneous activity in the caldera area was eruption of high-alumina olivine tholeiite lavas (Tbl) ca. 14.9 Ma. Most mineralization was demonstrably or probably contemporaneous with the major pulse of igneous activity, but formation of lithium deposits appears to have occurred much later, ca. 14.9 Ma.
Sources for Steens Basalt and precaldera intermediate lavas. No dikes or other unequivocal sources for the Steens Basalt or the precaldera intermediate lavas have been found in the McDermitt map area. However, Glen and Ponce (2002) identified three prominent, curvilinear, approximately north-trending magnetic bands, the central of which extends through the McDermitt caldera and suggests an underlying dike swarm (Fig. 2). The nearest known dike swarm, which extends south from Steens Mountain to the Pueblo Mountains, coincides with the western magnetic lineation. Minor (1986) and Minor and Wager (1989) found several 1–6-m-wide, porphyritic basalt dikes in the northeastern Bilk Creek Mountains that they interpreted to be feeders to the Steens Basalt, but no magnetic band is present there.

Sparsely Porphyritic to Aphyric Rhyolite Lavas (Units of Tra and Tar)

Sparsely anorthoclase-phyric (Tra, Tral, Tra2, Traw, Tram, Trau) to aphyric rhyolite (Tar, Taro, Tarv) lavas that are interbedded with middle and upper parts of the Steens Basalt or precaldera intermediate lavas (Tpi) crop out in the northern, eastern, and southern caldera walls. The different units designate lavas at different locations or stratigraphic positions. The anorthoclase rhyolites have 2%–8% phenocrysts, mostly of anorthoclase and lesser clinopyroxene. The aphyric rhyolites commonly have nonhydrated glass, especially in upper breccias. The lavas are 10 m to as much as 200 m thick; they erupted through the precaldera activity, and dates range from 16.691 ± 0.021 to 16.35 ± 0.04 Ma (Fig. 7; Table 2), all on anorthoclase rhyolites. In keeping with the general upward increase in SiO2, the typically lower SiO2 (72%–75%) anorthoclase rhyolites mostly underlie the higher SiO2 (75%–76%) aphyric rhyolites where they occur together. The anorthoclase rhyolites are metaluminous or peraluminous to slightly peralkaline, whereas the aphyric rhyolites, with one exception, are peralkaline (Figs. 5 and 6).

The oldest anorthoclase rhyolite is a 16.691 ± 0.021 Ma, 72% SiO2 lava (sample H05-26, Table 2) from the lower rhyolite unit of Ryutaba and Curtis (1983), the lowest exposed unit along the range front ~13 km north of the caldera (Fig. 3). The rhyolite is directly overlain by mafic lavas, including some with plagioclase laths as much as 1 cm long, and is ~120 m stratigraphically below the tuff of Oregon Canyon. No similar rhyolite lava occurs in the stratigraphically equivalent Catlow Peak section (Camp et al., 2013). The 16.651 ± 0.008 Ma anorthoclase-phyric ash-flow tuff (Tta; see following discussion of Distal Ash-Flow Tuffs) overlies basalts of unit Ts and may be a pyroclastic deposit related to the anorthoclase-phyric lavas. All other anorthoclase or aphyric rhyolites overlie tuff of Oregon Canyon and have ages ranging from 16.48 ± 0.08 or 16.462 ± 0.018 to 16.35 ± 0.04 Ma (Table 2). The only anorthoclase rhyolite lava in the western caldera wall is an undated flow (Traw) in unit Tpi along Horse Creek.

Pyroclastic-fall deposits consisting of coarse, poorly sorted pumice, locally with obsidian blocks, are nearly ubiquitous between flows (Fig. 8A). The pyroclastic eruptions were probably precursors to lava extrusion. Many fall deposits are strongly welded, either from agglutination or welding during emplacement, both indicating near-vent deposition (Sparks and Wright, 1979; Stevenson and Wilson, 1997), or through fusion of tephra due to heating and compaction by overlying lavas. The fall deposits are not indicators that the flows are lava-like ignimbrites, because the flows invariably have basal breccias that overlie the fall deposits with a sharp contact (Bonnichsen and Kauffman, 1987; Henry and Wolff, 1992).

The only identified source for any of these lavas is a circular vent of geochemically distinct aphyric rhyolite (Tarv vent for unit Taro) ~900 m in diameter in the northeastern caldera wall. The vent has radial, inward-dipping flow bands and abundant, locally glassy, carapace breccia. The other lavas probably also have nearby sources, as suggested by the coarseness of interbedded pyroclastic-fall deposits (Fig. 8A).

The anorthoclase-phyric and aphyric rhyolites mostly plot as two coherent and contiguous groups, suggesting that the aphyric rhyolites are more evolved versions of the porphyritic rocks (Figs. 5, 6, and 9). With the exception of obsidian nodules from unit Taro, the aphyric rhyolites are highly enriched in incompatible elements, e.g., Zr and REEs, have large Eu anomalies, and are depleted in compatible elements such as Ba. Obsidian from Taro (C04-10A) has low Zr and REEs, which suggest affinity to nearby biotite rhyolite despite being aphyric.

The aphyric to sparsely porphyritic rhyolite lavas of this study are shown as ash-flow tuff, units 1 and 2, in the southern and southeastern caldera wall, and as ash-flow tuff, units 4-5 undivided, in the northern caldera wall (Ryutaba, 1 976; Ryutaba and Glanzman, 1978, 1979; Glanzman and Ryutaba, 1979). However, the rocks are lavas.

Distal Ash-Flow Tuffs

Anorthoclase-phyric ash-flow tuff (Tta). Three rhyolitic ash-flow tuffs from uncertain sources are also interbedded with Steens-type rocks. The oldest tuff is an anorthoclase-phyric rhyolite (16.651 ± 0.008 Ma, Fig. 7; C05-280, Table 2) that crops out only in a small area ~7 km south of Disaster Peak in the western caldera wall, where it overlies a thin sequence of units T7 and T8 that overlies Cretaceous granodiorite. This is also the oldest middle Miocene ash-flow tuff in the region, not identified in any previous studies (Castor and Henry, 2000; Coble and Mahood, 2012, 2016). The source is unknown. Its anorthoclase phenocrysts and composition suggest that it may be related to the anorthoclase-phyric rhyolite lavas, but the tuff has higher Al2O3 and lower FeO and is more similar to sample HP5-26 than to other anorthoclase-phyric rhyolites (Fig. 6).

Tuff of Oregon Canyon (Tto). The 16.57 ± 0.02 Ma (n = 4 samples by MAP215–50) and 16.517 ± 0.013 Ma (Argus VI, H04-44, Table 2) tuff of Oregon Canyon is a mildly peralkaline, mostly high-SiO2 rhyolite distinguished from the anorthoclase-phyric tuff by age, composition, and the presence of sanidine, quartz, and aenigmatite (a Na-Fe-Ti silicate found in peralkaline rocks) phenocrysts (Figs. 5 and 6; Table 1). Ryutaba and Curtis (1983) named the tuff of Oregon Canyon for its occurrence in Oregon Canyon ~15 km north of the caldera (Fig. 3). The tuff crops out locally in the northern caldera wall, and Minor et al. (1989b) and Camp et al. (2013) showed it capping the Steens
Basalt at Catlow Peak northwest of the caldera. The tuff of Oregon Canyon is highly enriched in incompatible elements and depleted in compatible elements (Starkel, 2014). Zoning to less evolved compositions is indicated by an upper trachydacite that is only locally preserved and pumice in the rhyolite that contains ~67% SiO₂ (Fig. 6). Rytuba and McKee (1984) proposed a caldera source partly overprinted by the northeastern part of the McDermitt caldera. However, we do not find evidence for a caldera there, or an indication for a local source for the tuff (see discussion of Washburn caldera). Reported sanidine ⁴⁰Ar/³⁹Ar dates of the tuff of Oregon Canyon are 16.654 ± 0.025 Ma (weighted mean of eight aliquots of one sample from Catlow Peak; Jarboe et al., 2010) and 16.573 ± 0.014 Ma (weighted mean of eight samples; Mahood and Benson, 2016, 2017), all calculated with respect to 28.201 Ma for Fish Canyon sanidine. Our data, coupled with these published results, reveal an age range of 0.137 Ma (0.8%, 16.517 ± 0.013 Ma to 16.654 ± 0.025).
Figure 9. Chondrite-normalized rare-earth element plots of McDermitt igneous rocks; normalization is from Sun and McDonough (1989). (A) All samples. (B) Peralkaline rhyolites. (C) McDermitt Tuff; the two least silicic samples are labeled with %SiO₂. (D) Rhyolite of McDermitt Creek. (E) Aphyric rhyolites. (F) Biotite rhyolites. (G) Postcaldera icelandites. (H) Precaldera intermediate lavas.
The New Mexico Geochronology Research Laboratory (NMGRL) ARGIS VI and MAP215–50 data vary by ca. 50 ka and are nearly equal at 2σ, and this small difference is likely not significant considering there is only a single ARGIS VI analysis (i.e., error not adjusted for Student’s t-test). The overall data demonstrate the ongoing problem of high-precision comparison of data between laboratories. The relatively old age of 16.654 ± 0.0025 Ma from Jarboe et al. (2010) is consistent with offsets between the NMGRL and the Berkeley Geochronology Center (BGC), that may be related to similar inaccuracies of BGC data reported by Niespolo et al. (2016).

**Tuff of Trout Creek Mountains (Ttt).** The 16.49 ± 0.03 Ma (n = 3) tuff of Trout Creek Mountains (Ttt) is the youngest regional tuff in the McDermitt map area, underlies the McDermitt Tuff around much of the southwestern caldera wall, and thins to ~1 m in the northeastern wall. The tuff of Trout Creek Mountains is one of the most strongly peralkaline rocks in the region, with particularly low Al2O3 and high FeO*, Zr, Y, and REEs (Fig. 6), and contains sanidine, quartz, sodic amphibole, aegirine-augite, and aenigmatite phenocrysts (Table 1). Rytuba and McKee (1984) proposed a source almost entirely buried beneath sodic amphibole, aegirine-augite, and aenigmatite phenocrysts (Table 1). The presence of biotite, hornblende (at Mendi Suri), and plagioclase in high-SiO2 rhyolite, and trace zircon and metaluminous to aegirine-augite, and aenigmatite phenocrysts (Table 1).

**Biotite Rhyolites**

Sparsely porphyritic biotite rhyolites were emplaced at four separate areas and times (Fig. 7) along or near the western, northeastern, and eastern margins of the caldera. The presence of biotite, hornblende (at Mendi Suri), and plagioclase in high-SiO2 rhyolite, and trace zircon and metaluminous to peraluminous rocks in the region, with particularly low Al2O3, high FeO*, Zr, Y, and REEs (Fig. 6), and contains sanidine, quartz, sodic amphibole, aegirine-augite, and aenigmatite phenocrysts (Table 1). Rytuba and McKee (1984) proposed a source almost entirely buried beneath sodic amphibole, aegirine-augite, and aenigmatite phenocrysts (Table 1).

**Biotite rhyolite of the Moonlight Mine (Tbr).** Biotite rhyolite lava crops out discontinuously along the southwestern caldera margin from the Moonlight Mine south to Thacker Pass (Fig. 11A). It is at least 150 m thick at the mine, and the base is nowhere exposed. Similar rock occurs as lithics in McDermitt Tuff and as megabreccia (notably in the megarheomorphic folds of Hargrove and Sheridan, 1984) through the same area and as far north as Horse Creek. With a sanidine 40Ar/39Ar date of 16.619 ± 0.020 Ma (C95-103A, Table 2), this is the second oldest, demonstrably locally derived silicic rock in the caldera area. The Moonlight rhyolite has 10%–16% phenocrysts, the most among the biotite rhyolites, consisting of plagioclase, sanidine, biotite, and minor quartz (Fig. 10A; Table 1). It is rhyolite with 70%–75% SiO2 (Figs. 5 and 6). The rock is altered through much of its distribution, particularly at the Moonlight Mine, with coarse hydrothermal quartz, adularia as disseminations and replacing feldspar phenocrysts, pyrite, arsenopyrite, fluorite, and various uranium minerals, especially uranium-rich zircon (Castor and Henry, 2000).

**Quartz-sandine rhyolite lava dome (Tgr).** A porphyritic, high-SiO2 rhyolite lava dome crops out northeast of the caldera in one large contiguous area against Quaternary in the adjacent basin and in a smaller western window below underlying rocks. The dome is at least 5 km across and 250 m thick, and the base is not exposed. The dome underlies older aphyric rhyolite (Taro) and precaldera intermediate lavas (Tpi, Plate 1). Vitrophyre that crops out at low elevation in its southernmost extent, adjacent to Quaternary cover, may mark the base of the dome. Vitrophyre that is exposed beneath precaldera intermediate lavas in the small western window probably marks the former top of the dome. The rock is strongly flow banded; layers are planar and subhorizontal in much of the dome and turn steeply upward and are folded in its upper part. A few major ledges give the impression of separate flow lobes, but no breccia or vitrophyre that could mark separate flows have been found.

Vitrophyres consist of smoky quartz, sandine, plagioclase, and a trace of clinopyroxene (Table 1). Although it does not contain biotite, the quartz-sandine rhyolite is petrographically and compositionally similar (plagioclase, low FeO*, Zr, and Al, and high Al2O3). The 40Ar/39Ar age and K/Ca (16.607 ± 0.015 Ma, 27.8 ± 14.8, H16-12, Table 2) to the biotite rhyolites in most respects, so we include it with them. The quartz-sandine rhyolite is distinguishable from the biotite rhyolite of the Moonlight Mine (Tbr) in sandine 40Ar/39Ar age and K/Ca (16.607 ± 0.015 Ma, 27.8 ± 14.8, H16-12, Table 2).

**Biotite rhyolite of Mendi Suri (Tbrm).** Biotite rhyolite crops out as two light colored ~1.5-km-wide and ≥140 m exposed thickness bodies that are either shallow intrusions into, or lava domes underlying, unit Tpi (precaldera intermediate lavas) at Mendi Suri (Figs. 3 and 8B), ~5 km northeast of the northeastern margin of the caldera. The rock is massive to strongly and steeply flow banded. Several north-northeast–striking dikes of biotite rhyolite cut the northern body. The southern body has an upper vitrophyre breccia, and clasts of similar rhyolite occur in nearby conglomerate. Obsidian nodules (C04-10A, Supplemental Table S1 [see footnote 1]) compositionally similar to the distinctive main body (H05-28) erode out of the top of an aphric rhyolite (Taro) ~4 km south of the domes; this suggests that the rocks are related despite the difference in phenoocrysts. The Mendi Suri domes contain 6%–10% phenocrysts of sanidine, plagioclase, quartz, biotite, and a trace of hornblende (Fig. 10B). Both the main body and the obsidian nodules have ~76% SiO2, distinctive flat REE patterns, large Eu anomalies, high concentrations of most incompatible elements (Rb, Y, Nb, Ga, Pb, and Th), and the lowest Zr and TiO2 contents of any rhyolites (Figs. 5, 6, and 9). A sanidine 40Ar/39Ar date of 16.487 ± 0.015 Ma (H05-28, Table 2).

**Biotite rhyolite of Hoppin Peaks (Tbrh).** Three lava flows totaling ~230 m thick cap Hoppin Peaks along the eastern margin of the caldera. The flows, which were interpreted as strongly rheomorphic intracaldera tuff by Rytuba and McKee (1984), consist mostly of strongly flow banded and folded stony rhyolite with basal breccia characteristic of lavas (Fig. 8C). The stratigraphically lowest and highest flows have basal vitrophyres as much as 10 m thick in the lowest flow. All flows contain 3%–6% phenocrysts of quartz, sandine, plagioclase, and biotite, and the lowest flow contains a trace of pyroxene. All flows are compositionally similar with 75% SiO2 (Figs. 5 and 6). Sanidine 40Ar/39Ar dates on the lowest and highest flows are 16.384 ± 0.069 (H04-51) and 16.362 ± 0.024 Ma (H04-53), respectively (Table 2). These dates indicate that the rhyolite of Hoppin Peaks erupted closely preceding eruption of the McDermitt Tuff.
Geochemistry. The biotite rhyolites are distinct from other Miocene rocks in being metaluminous to peraluminous, having higher Al₂O₃, lower FeO*, MnO, and HREEs, and mostly lower TiO₂ and Zr (Figs. 5, 6, and 9). The four suites are distinct from each other. The Moonlight Mine suite is mostly low-SiO₂ rhyolite, whereas the other three have 75%–76% SiO₂. Moonlight rocks also have lower Rb and Y and higher TiO₂ and Zr, more like the alkaline Miocene rocks.

The single sample from Mendi Suri (H05-28) and the compositionally similar but aphyric obsidian nodules (C04-10A) have very low TiO₂, FeO*, P₂O₅, Sr, Zr, Ba, and REEs, high Rb and Nb, and the deepest Eu anomalies. The four suites also plot as distinct groups on a plot of Ga/Al versus Zr. The quartz-sanidine rhyolite (Trq) is distinct from all other rhyolites in having the highest light REE concentrations and steepest REE patterns (Figs. 6 and 9).

Peralkaline Rhyolite Lava (Tpr)

What is probably a single lava flow of peralkaline rhyolite crops out in three areas around the southern margin of the caldera. The most prominent outcrops are as cliffs along the escarpment into Kings River Valley centered on the Moonlight Mine, where a single flow is as much as 180 m thick (Fig. 11A), continuous along strike for ~8 km, and has both basal and upper hydrated vitrophyre and breccia. A similar thick flow with an upper vitrophyric breccia forms cliffs for 3.5 km along the escarpment south of Thacker Pass. The flow in the southeastern caldera wall crops out for 1.5 km, is covered by aphyric rhyolites to the north and south, is as much as 55 m thick, and has a well-exposed basal breccia of hydrated vitrophyre that overlies welded fall...
deposits (Fig. 8D). In all areas, the main part of the flow is light tan, massive to flow banded, and crystalline. The rock everywhere contains 12%–17% phenocrysts of euhedral and inclusion-free sanidine, commonly smoky quartz, clinopyroxene, aenigmatite, monazite, magnetite, ilmenite, and trace apatite (Figs. 10C, 10D). Probable vapor-phase sodic amphibole is common. All analyzed samples are indistinguishable with 76% SiO₂ and low Al₂O₃ and high FeO characteristic of peralkaline rhyolites (Figs. 5, 6, and 9). Sanidine yielded indistinguishable ⁴⁰Ar/³⁹Ar dates in the western (16.399 ± 0.016 Ma, C95–107), southwestern (16.404 ± 0.018 Ma, 1 McD-290), and southeastern areas (16.427 ± 0.015 Ma, Fig. 7; H12-250, Table 2); mean = 16.410 ± 0.015 Ma. The source of the unit is unknown. Similarity in rock type, phenocryst assemblage, composition, and age suggests that the three outcrops are part of a single flow across an ~450 km² area. The presence of quartz and sanidine phenocrysts, rather than anorthoclase, distinguishes this unit from the other precaldera peralkaline rhyolites.

These rhyolites are moderately peralkaline, although two samples probably have lost Na₂O and gained K₂O (Figs. 5, 6, and 9). Despite the great difference in phenocrysts, the peralkaline and aphyric rhyolites are compositionally similar, with overlapping concentrations of incompatible and compatible elements and large Eu anomalies.

McDermitt Tuff and Related Rocks (Tmt, Tmb, Tmtu, Tmv)

Collapse of the McDermitt caldera occurred during eruption of what we herein name the McDermitt Tuff, which is zoned from peralkaline, aphyric, high-SiO₂ rhyolite to metaluminous, abundantly anorthoclase-phyric, high-
SiO₂ trachydacite and icelandite (Figs. 5, 6, 11, and 12; Table 1; Conrad, 1984; Starkel et al., 2012; Starkel, 2014). The tuff ponded to a thickness of at least 610 m in the caldera, where it is commonly strongly rheomorphic (see also Hargrove and Sheridan, 1984). This rheomorphism generally folded together the different petrographic and compositional zones (Fig. 12A); a probable original vertical zonation is at best poorly preserved in intracaldera tuff (Tmt). Megablocks and mesobreccia lenses of precaldera rocks (Tmb) are common in intracaldera tuff and are especially well exposed along the western escarpment of the Montana Mountains. A sparsely porphyritic tuff (Tmtu) that overlies mesobreccia at the top of the Moonlight Mine section is probably a late variant of the McDermitt Tuff. Vitrophyric pyroclastic dikes (Tmv), probably feeders to late tuff eruptions, are present along the western side of the caldera. Outflow tuff that is variably rheomorphic crops out in the northern caldera wall in the Oregon Canyon and Trout Creek Mountains and in the southern wall in the Double H Mountains, where the vertical zonation is locally well preserved.

Introduction of the name McDermitt Tuff is necessitated by the confusing application of previous informal names. The McDermitt Tuff includes the alkali rhyolite of Jordan Meadow of Greene (1976) and parts of the tuff of Long Ridge of Rytuba and McKee (1984), who provided only a 1:1,000,000-scale map that shows all volcanic rocks in the region as a single unit. The general distribution of the tuff of Long Ridge, members 1, 2, 3, and 4-5 undivided, are depicted on page-sized maps in Rytuba (1976), Rytuba and Glanzman (1978, 1979), and Glanzman and Rytuba (1979) that encompass approximately the same area as our Plate 1. Members 1 and 2 appear to be entirely sparsely porphyritic or aphyric rhyolite lavas (our units Tra and Tar). Member 3, which is shown only in the Double H Mountains, partly coincides with our McDermitt Tuff. Member 4-5 undivided, which is shown only north of the caldera, partly coincides with
Figure 12. Photomicrographs, McDermitt Tuff (Tmt, A-D) and Vitrophyric tuff dikes (Tmv, E, F). A is anorthoclase. C is clinopyroxene. (A) McDermitt Tuff (H04-32B, outflow in southern caldera wall). Interbanded aphyric (left) and abundantly porphyritic (right) phases. Cross-polarized light. Horizontal field is 30 mm. (B) MD-3A, intracaldera vitrophyre, Moonlight Mine section. Anorthoclase and clinopyroxene phenocrysts in matrix containing variably colored, aphyric pumice that are probably compositionally varied. Horizontal field is 3.2 mm. (C) MD-21, outflow in southern caldera wall. Euhedral anorthoclase and clinopyroxene phenocrysts in matrix of devitrified shards and pumice. Cross-polarized light. Horizontal field is 3.2 mm. (D) H05-23A, outflow in northern caldera wall. Anorthoclase, clinopyroxene, and probable sodic amphibole (N) phenocrysts in devitrified matrix. Cross-polarized light. Horizontal field is 0.9 mm. (E) Vitrophyre dike (H12-259, northwestern caldera wall). Anorthoclase and clinopyroxene phenocrysts in densely welded, shard- and fine pumice-rich matrix. Horizontal field is 3.2 mm. (F) Vitrophyre dike (H95-103, Moonlight Mine section). Anhedral anorthoclase phenocryst with abundant clinopyroxene inclusions. Cross-polarized light. Horizontal field is 3.2 mm.
our McDermitt Tuff where it caps Flattop, and the phenocryst assemblage of member 5 (anorthoclase and minor ferrohedenbergite and fayalite; Rytuba and McKee, 1984) most closely matches the McDermitt Tuff. However, member 4-S also includes sparsely porphyritic or aphyric rhyolite lavas that underlie precaldera intermediate lavas (Tpi) that underlie McDermitt Tuff. Rytuba and McKee (1984) identified member 4 as the tuff of Trout Creek Mountains, but described it as having 26% anorthoclase and 1% quartz phenocrysts, unlike the sanidine-quartz-rich tuff of Trout Creek Mountains. The page-sized maps of Rytuba (1976), Rytuba and Glanzman (1978, 1979), and Glanzman and Rytuba (1979) also showed most of what we map as intracaldera McDermitt Tuff as various units of amalgamated “rhyolite intrusive rocks, flows, and tuffs” (figure 2 of Rytuba and Glanzman, 1978); no tuff of Long Ridge is shown inside the McDermitt caldera. Based on our comprehensive new work and the confusion about what constitutes all of the tuffs of Rytuba and McKee (1984), the older names should be abandoned.

**Intracaldera Tuff (Tmt)**

Because of the intense rheomorphic intrafolding of the different petrographic-compositional types, we map most intracaldera tuff as a single unit (Tmt). More detailed 1:24,000-scale mapping of the Calavera Canyon 7S' quadrangle, approximately centered on Calavera Canyon and Horse Creek, where intracaldera tuff is well exposed along the western escarpment, allows greater subdivision of the petrographic variants (Starkel, 2014). The western escarpment exposes a complete section of intracaldera tuff as well as precaldera units: Steens Basalt (Tsb, Ts) and precaldera intermediate rocks (Tpi), biotite rhyolite of the Moonlight Mine (Tbr), and peralkaline rhyolite lava (Tpr), to overlying intracaldera sedimentary deposits (Tis) (Fig. 11A). It is notable that no other tuff underlies the McDermitt Tuff where the base is exposed. The tuff is thickest, 610 m, at Garden Creek immediately south of Calavera Canyon, and generally >500 m thick along the escarpment (Fig. 3; Plate 1). Because the precaldera rocks dip variably into the caldera along the western margin (see discussion of Western Margin herein), the tuff may thicken eastward (Plate 1, section A-A). A drillhole (WNC-58C; D.P. Bryan, Western Lithium Corporation, written commun., 2013) in the southern part of the caldera encountered 544 m of intracaldera tuff and bottomed in it. Elsewhere, only the uppermost parts of intracaldera tuff are exposed. Nevertheless, the tuff is at least 290 m (Rock Creek) and 220 m thick (McDermitt Creek) in canyons in the southern and northern parts of the caldera, respectively.

The greatest volume of McDermitt Tuff is aphyric, high-SiO2 rhyolite (77%; Figs. 5 and 6). Phenocryst content, mostly anorthoclase with minor clinopyroxene and fayalite, increases continuously to ~30 vol% with decreasing SiO2 content (Fig. 12; Table 1). Anorthoclase in higher-SiO2 rhyolite is generally more euhedral and less inclusion rich than those in lower-SiO2 rocks (cf. Figs. 12B, 12C with 12E, 12F). The most silicic rocks also contain sparse sodic amphibole phenocrysts consistent with peralkalinity (Fig. 12D). High-SiO2 trachydacite with 68% SiO2 is the most common low-SiO2 end member, but icelandite with as little as 64% SiO2 is present at the top of the section in the southwestern rim (HG95-83; Supplemental Table S1 [see footnote 1]). Incompatible elements Rb, Y, Zr, Nb, Hf, Ta, Pb, Th, and U increase with increasing SiO2, whereas compatible elements Sr and Ba decrease. All REEs are incompatible, except for Eu, which is incorporated in anorthoclase (Fig. 9; Starkel et al., 2012). The least silicic rocks have positive Eu anomalies and are similar to precaldera intermediate rocks (Tpi) except for distinctly different MnO trends. The McDermitt Tuff makes a linear trend starting at the edge of the I- and S-type granite field on a Ga/Al versus Zr plot because of increasing Zr and Ga and decreasing Al2O3 with increasing SiO2.

Most tuff is devitrified, but vitrophyre zones are common around megabreccia and mesobreccia (Fig. 11B). Pumice is generally present in these zones and constitutes 10 vol% to as much as 50 vol%. Mixed pumice populations, with aphyric pumice in porphyritic rock, and vice versa, are common (Fig. 12B). Lithics vary from sparse to highly abundant and from millimeter size to blocks as large as 10 m in diameter. Clasts represent a nearly complete selection of precaldera rocks, including Cretaceous granitic rocks, probable Eocene biotite dacite, coarsely plagioclase-phycric to aphyric Steens Basalt, particularly abundant biotite rhyolite, and peralkaline rhyolite. The largest blocks are mapped as megabreccia. Lenses of mesobreccia with clasts as large as ~1 m in diameter in a poorly exposed matrix are especially prominent in and at the top of the Moonlight Mine section (Figs. 11A, 13A, and 13C).

The McDermitt Tuff is moderately to intensely rheomorphic throughout the caldera (Figs. 11C–11H). Sparse, weakly rheomorphic rocks are nearly undeformed or have pumice stretched only in one direction. Strain partitioning, where one layer is significantly stretched and an adjacent, apparently indistinguishable layer is not strained, is developed in what were probably originally stratigraphically high parts of the tuff (Fig. 11C). The megahemorrhagic folds of Hargrove and Sheridan (1984) represent moderately to highly strained rock where large-scale folds have developed, and aphyric and abundantly porphyritic rocks are interbedded (Fig. 11D). Most of the folded rock is easily recognizable as ash-flow tuff with stretched but recognizable pumice ± lithics (Fig. 11E). Lenses of more highly sheared rock in the folds probably took up most of the strain and indicate an extreme example of strain partitioning (Fig. 11F). Hargrove and Sheridan (1984, p. 8632) interpreted that biotite rhyolite was lava that was interbedded and cofolded with McDermitt Tuff in a “plastic manner”; however, we find that biotite rhyolite existed entirely as megablocks and mesobreccia and that compaction around the blocks probably made a minor contribution to apparent folding. In the most uniformly strained rocks, foliation appears more like that of lava (Fig. 11G). Pumice either was not present or is so elongate that its original discontinuity is difficult to recognize. Most intracaldera tuff appears to have deformed to this degree, because rock exposed in canyons of Pole, Washburn, and McDermitt Creeks (south to north through the caldera) all have this appearance. Flow folds do not appear to have regular orientations, a conclusion also reached by Hargrove and Sheridan (1984). Lineations along foliation surfaces in the Moonlight Mine section (Fig. 11H) trend regularly ~110°.
nearly perpendicular to the caldera wall there, which suggests that some rheo-

morphism may have been slumping off the growing wall. The vitrophyre pyro-
clastic dikes (Tmv) cut vertically through highly rheomorphic rock, indicating
that rheomorphism occurred during emplacement, therefore, during caldera
collapse. This interpretation slightly contrasts from Hargrove and Sheridan
(1994, p. 8629), who concluded that “…large-scale folds in the deformed se-
quence developed shortly after caldera collapse” because the folds underlie un-
deformed rocks. The folds demonstrably predate nearly flat-lying mesobreccia
lenses that overlie the megareomorphic folds, similar folds at the Moonlight
Mine section, and the upper tuff (Tmtu) at the Moonlight Mine section.

Strong rheomorphism of McDermitt Tuff throughout the caldera suggests
that slumping off a growing caldera wall was not the only operative process.
Deposition and slumping over irregular precaldera topography or irregular col-
lapse of the subsiding caldera block may also be significant.

Figure 13. (A) Vitrophyre pyroclastic dike (Tmv, Moonlight Mine section) cuts steeply
through rheomorphic intracaldera McDermitt Tuff and through left side of mesobreccia
knob, dark area just below box, which shows location of C. Mesobreccia knob is ~6 m high.
(B) Vitrophyre pyroclastic dike with subhorizontal columnar joints. Same as dike in A. Height
from columnar joints to top of outcrop is ~ 45 m high. (C) Mesobreccia (location shown
in A) containing angular to subrounded clasts of peralkaline rhyolite, Steens Basalt, and
other pre-McDermitt Tuff rock types.
Because of the intense rheomorphism, some intracaldera McDermitt Tuff is not immediately recognizable as ash-flow tuff. Much of the intracaldera tuff is massive to steeply foliated aphyric rhyolite. Bands of variably anorthoclase-phyric rhyolite in sharp contact with aphyric rhyolite constitute as much as 50% of the tuff in some locations (Fig. 12A) but are essentially absent through large areas elsewhere. Greene (1972, 1976, p. 36) recognized this characteristic, stating “The alkali rhyolite of Jordan Meadow consists of interlayered aphyric, sparsely porphyritic, and abundantly porphyritic alkali rhyolites.”

Megablocks and Mesobreccia Lenses (Tmb)

Large megablocks and mesobreccia lenses of finer blocks are common in McDermitt Tuff, and mesobreccia is between McDermitt Tuff (Tmt) and upper tuff (Tmtu) along the western escarpment. Mesobreccia (Fig. 13C) contains angular to subrounded clasts of peralkaline rhyolite (Tpr), Steens Basalt (Ts, Tsb), and other pre-McDermitt Tuff rock types.

Vitrophyre Pyroclastic Dikes (Tmv)

Vitrophyre pyroclastic dikes are present in two areas. Numerous dikes cut intracaldera McDermitt Tuff just south of Calavera Canyon in the Moonlight Mine section. The dikes mostly form northeast-striking, steeply northwest-dipping planar bodies, 1–10 m thick, commonly with columnar joints approximately perpendicular to walls (Figs. 11A, 13A, and 13B). The dikes are locally slightly nonplanar, particularly where they bend around or through mesobreccia. Common pumice, lithics, and densely welded, eutaxitic textures in the dikes demonstrate their pyroclastic origin (Fig. 12E). The dikes cut sharply through highly rheomorphic tuff, and therefore postdate rheomorphism. Some bodies identified as dikes may be vertically dipping tuff. A single northern dike ~3 km southeast of Disaster Peak cuts the Steens Basalt (Ts) in the caldera wall also strikes northeast, but dips more gently northwest, ~40°–45°, and is as much as 30 m thick. All dikes are glassy, moderately to abundantly porphyritic with 68%–72% SiO2 (H12-259, H95-103; Figs. 12E, 12F; Supplemental Table S1 [see footnote 1]). All these features suggest that the dikes fed late eruptions of McDermitt Tuff but prior to eruption of the upper tuff (Tmtu), which is much less porphyritic.

Upper Tuff (Tmtu)

A sparsely anorthoclase-phyric, strongly eutaxitic, lithic-bearing, densely welded, and moderately rheomorphic ash-flow tuff as much as 35 m thick overlies an equally thick mesobreccia layer (Tmb) at the top of the Moonlight Mine section (Fig. 11A). The rock is distinctive because of pervasive alteration of feldspars and variable silicification. This variant is a late eruption that may be significantly younger (years to thousands of years?) than the underlying tuff; however, field relations only provide a relative age, and dating is precluded by alteration. Petrographically similar, probably correlative, late outflow tuff also is present in the Double H Mountains south of the map area (H07-189, H07-190; Supplemental Table S1 [see footnote 1]).

Outflow Tuff

Northern caldera wall. McDermitt Tuff crops out discontinuously along the northern caldera wall but may have been more continuous before erosion. Excellent cliff exposures at Flattop have thin (1–3 m) layers of abundantly porphyritic trachydacite within thick aphyric to sparsely porphyritic rhyolite that are folded into mostly low-amplitude, long-wavelength folds (<5 x 100–300 m) with limb dips no more than 10° (Fig. 14A). The tuff locally steepens into ramps with near vertical dips. Lower contacts of the more porphyritic rock are gradational over 1–2 m, whereas upper contacts are sharp. Sparse, abundantly porphyritic pumice as much as 20 cm long and common aphyric, mafic to intermediate lithic fragments as much as 5 cm in diameter occur in the lower parts of the aphyric to sparsely porphyritic rock (Fig. 14B). The sparsely porphyritic phase contains the probable sodic amphibole phenocrysts (Fig. 12D). The tuff at Flattop overlies a sparsely anorthoclase-phyric rhyolite lava (Tra2) with a similar phenocryst assemblage as part of the tuff. However, this underlying rock is not part of the McDermitt Tuff because it is distinctly older (16.462 ± 0.018 Ma; C04-07; Table 2; Fig. 7) and stratigraphically distinct. The rhyolite lava has an upper vitrophyric breccia overlain by thin tuffaceous sedimentary deposits that record a significant time and depositional break from the overlying tuff.

South of caldera in Double H Mountains. Outflow McDermitt Tuff crops out as much as 13 km south of the caldera margin in the Double H Mountains (Fig. 3; Plate 1). The tuff overlies aphyric rhyolite lava (Tar) throughout this area. Outflow McDermitt Tuff shows an abrupt transition from thin (<10 m) welded, normal ash-flow tuff to thick (>70 m) strongly rheomorphic tuff where it was deposited over the steep margin of an aphyric rhyolite lava (Figs. 14C, 14D). The lava has a thick, upper vitrophyric breccia, so is distinctly an older unit, not a lower phase of the tuff. The thin outflow tuff has an aphyric base overlain by a sparsely porphyritic layer 1–2 m thick, in turn overlain by more abundantly porphyritic tuff. Where the tuff thickens, flow bands and folds develop and progress to isoclinal, recumbent, and sheath folds, all indicating transport toward 140°. Aphyric and more porphyritic bands are intermittently folded, similar to rheomorphic tuff in the caldera. Sparse lithics to 5 cm are the only preserved indicator of a pyroclastic origin. Still farther south, beyond the map area, bands of abundantly porphyritic rock as much as 3 m thick are complexly interfolded with aphyric to sparsely porphyritic rock. This rheomorphic unit is overlain by two thin cooling units of nonrheomorphic, sparsely anorthoclase-phyric tuff that may correlate with the late, post-mesobreccia tuff (Tmtu) at the Moonlight section.
Overall distribution of outflow McDermitt Tuff. Delineating the original distribution of outflow tuff is hampered by uncertainty in correlation with mapped units in areas of published geologic maps, by lack of mapping in other areas, and by uncertainty in the amount of erosion of distal tuff (Fig. 3). The McDermitt Tuff is probably present in much of the southern Oregon Canyon and Trout Creek Mountains and northern Bilk Creek Mountains north and west of the caldera, based on descriptions of the probably correlative tuff of Long Ridge given by Rytuba and Curtis (1983), Rytuba et al. (1983a, 1983b), Minor (1986), Minor and Wager (1989), and Minor et al. (1989a, 1989b). No detailed mapping is available for the southern Bilk Creek Mountains southwest of the caldera. McDermitt Tuff is not present in the eastern caldera wall or in the Santa Rosa Range farther east (Brueseke and Hart, 2008; our observations).
Age

The $^{40}$Ar/$^{39}$Ar dates were determined on anorthoclase from six samples, one intracaldera tuff, three outflow tuffs, and two pyroclastic dikes (Fig. 7; Table 2). The 6 dates average 16.36 ± 0.04 Ma, but individual dates do not overlap within their 2σ sigma uncertainties. All dates were done on the Argus VI multicollector mass spectrometer. However, dates generally correlate positively with K/Ca of analyzed anorthoclase and with SiO$_2$ of host samples. Some anorthoclase in H95-103, which has the youngest apparent age and lowest K/Ca, is particularly inclusion rich (Table 2; Fig. 12F). Whether the differences in K/Ca result from differences in anorthoclase composition or the presence of inclusions is unknown. Magnetic separation and hand-picking should remove grains with abundant clinopyroxene inclusions, but might not eliminate grains with few, small inclusions. In addition, analyses of H15-67, H12-259, and H05-23P were done most recently using concurrent HF leaching and sonification to remove impurities; this probably generated cleaner anorthoclase samples. These three samples yielded the oldest dates and average 16.388 ± 0.015 Ma (Fig. 7), which is probably the best estimate of eruption age, although analytically indistinguishable from the mean of all six. The slightly older age of these three samples is more consistent with dates on post–McDermitt Tuff units. The scatter in anorthoclase dates contrasts with the tighter clustering of sanidine dates from the probably single lava flow of peralkaline rhyolite (Tpr) and illustrates the greater precision and accuracy of sanidine dates. Mahood and Benson (2017) and Benson et al. (2017) reported upper and lower K/A and 16.39 ± 0.10 Ma (10-McD-225) on a mix of plagioclase and anorthoclase samples. These dates indicate that the rhyolite of McDermitt Creek, from north and northwest of the McDermitt caldera map area.

Volume

We estimate a total volume of McDermitt Tuff of between 600 and 1100 km$^3$ dense rock equivalent, ~50% to 85% of which is intracaldera (Table 3). These estimates use two approaches: (1) using known and estimated distributions and thicknesses of intracaldera and outflow tuff, and (2) using the area and estimated total collapse of the caldera (John et al., 2008; Henry and John, 2013). Method 1 uses a caldera area of ~930 km$^2$, average intracaldera tuff thickness between 500 and 800 m, the distribution of outflow tuff as discussed here, and average outflow thickness between 50 and 100 m. The lower estimate of 600 km$^3$, based on only 500 m of intracaldera tuff, is probably too low. A significant uncertainty is the thickness of intracaldera tuff, because a complete section is only exposed along the western margin. This approach also does not consider dispersed ash.

Method 2, using the same caldera area and estimated collapse between 1000 and 1200 m (the greatest estimated collapse from along the southern margin), yields 930–1116 km$^3$. Uncertainty in thickness of intracaldera tuff, and so the amount of collapse, is significant, but this approach accounts for all erupted magma including dispersed ash. Considering all the data and their significant uncertainties, a total volume of ~1000 km$^3$ seems reasonable.

Postcaldera Magmatism

Rhyolite of McDermitt Creek (Trm)

The rhyolite of McDermitt Creek (Greene, 1972, 1976) is a group of low-SiO$_2$ (71%–73%) porphyritic rhyolite lavas that crop out in the northeastern part of the caldera. Much of the rock resembles highly rheomorphic McDermitt Tuff, and both units are indistinguishable in age. However, the rhyolite of McDermitt Creek overlaps McDermitt Tuff, locally with a basalt flow breccia and vesicular rock (Greene, 1976), and contains no interlayered aphric rock. The lavas are more coarsely porphyritic than McDermitt Tuff, with anorthoclase phenocrysts as much as 8 mm long, and contain plagioclase, which is absent in the tuff. Two single crystal $^{40}$Ar/$^{39}$Ar dates are 16.417 ± 0.018 Ma (H04-55) on anorthoclase and 16.39 ± 0.10 Ma (10-McD-225) on a mix of plagioclase and anorthoclase (Fig. 7; Table 2). These dates indicate that the rhyolite of McDermitt Creek was emplaced immediately after eruption of McDermitt Tuff and caldera collapse.

Chemical analyses also support the rhyolite of McDermitt Creek not being part of the McDermitt Tuff (Figs. 5, 6, and 9). The rhyolite has lower HREEs at the same SiO$_2$ content and lower MnO, similar to MnO in the biotite rhyolites and along the lower precaldera iceindstedt trend.

| TABLE 3. AREA, THICKNESS, AND VOLUME OF McDERMITT TUFF |
| Area (km$^2$) | Tuff thickness (m) | Volume (km$^3$) | Intracaldera |
|----------------|------------------|----------------|-------------|
|                | Minimum | Maximum | Minimum | Maximum | Minimum (%) | Maximum (%) |
| Caldera        | 930     | 500     | 800     | 465     | 744         | 63          | 85          |
| Outflow        | 2720    | 50      | 100     | 136     | 272         |             |             |
| Total          | 3650    | 100     | 100     | 136     | 272         |             |             |
| Amount of collapse | 1000 | 1200 | 930 | 1116 | 50 | 67 |
Icelandite Lavas and Vents (Tiv, Til, Tia)

Post-collapse volcanic vents (Tiv) or lavas (Til, Tia) of icelandite that range from ~56% to 64% SiO$_2$ crop out in 6 areas in the caldera. Vents at Round Mountain and Black Mountain retain subcircular, resistant rims, ~700 m in diameter, of dense, locally glassy, icelandite and an interior basin, both surrounded by flows and agglutinated scoria (Fig. 15A). Round Mountain and Black Mountain also correlate well with regional (1:250,000 scale) aeromagnetic highs (Fig. 16), as do the lavas mapped near Jordan Meadow Mountain, which suggests a possible vent there. Lavas as much as 120 m thick extend ~1.5 km outward from Black Mountain. Greene (1976) suggested that the area mapped as lavas 2 km northeast of Black Mountain is another vent, but this area does not correlate with an aeromagnetic high. The rocks at Black Mountain and Round Mountain are similar, with 60%–62% SiO$_2$, and sparse phenocrysts of plagioclase and clinopyroxene. Jordan Meadow Mountain is a slightly northeast-elongated area of icelandite lavas. They are petrographically similar to the Round Mountain rocks, but we have no chemical analyses. Plagioclase from the Round Mountain vent yielded an $^{40}$Ar/$^{39}$Ar date of 16.22 ± 0.13 Ma (H07-198 plateau; 16.32 ± 0.23 Ma isochron; Table 2).

The Aurora lavas (Tia), the host for U mineralization at the Aurora deposit (Wallace et al., 1980; Wallace and Roper, 1981; Roper and Wallace, 1981), are a sequence of icelandite lavas, flow breccias, and volcaniclastic deposits just inside the northeastern caldera wall. The lavas overlie McDermitt Tuff and underlie tuffaceous lacustrine sediments (unit Tis). Aurora lavas include the most silicic postcaldera rocks (60%–64% SiO$_2$) and contain plagioclase (anorthoclase in the most silicic rocks; Fig. 15B), clinopyroxene, and fayalite.

Figure 15. (A) Icelandite vent complex at Round Mountain; top, flat part is ~1.3 km across. Looking west at resistant rim of dense icelandite surrounding core of less resistant scoria breccia. (B) Photomicrograph, Aurora lava (Tia, H11–199). Anorthoclase (A) phenocrysts with resorbed rims in matrix of anorthoclase laths, clinopyroxene, Fe-Ti oxides, and finely devitrified glass. Leaching with dilute HF and hand-picking removed rims for single crystal $^{40}$Ar/$^{39}$Ar dating. Crossed-polarized light. Horizontal field is 2.3 mm. (C) Late basaltic lava (Tbl) banked against aphyric or very sparsely porphyritic rhyolite lava (Tar) in southeastern caldera wall at Pole Creek north of Sentinel Rock (upper left), which is a sparsely porphyritic rhyolite lava. Variably welded pyroclastic-fall deposits composed of pumice and obsidian are well exposed between the two rhyolite flows. Small cliff at left edge of photo is ~20 m high. (D) Intracaldera sedimentary deposits (Tis) composed of generally well-bedded, variably tuffaceous sandstone, siltstone, and mudstone exposed in a hectorite pit near Thacker Pass, southern part of the caldera. Face is ~3 m high.
Figure 16. Aeromagnetic contours (U.S. Geological Survey, 1972) overlain on geology of the McDermitt caldera (simplified from Plate 1 with transparent colors). The resurgent uplift, which exposed intracaldera McDermitt Tuff in the middle of the caldera, coincides with a magnetic high that, in turn, coincides with vents at Black and Round Mountains and probable subsurface distribution of late Icelandite (Tiv).
A single crystal anorthoclase date is 16.4 ± 0.20 Ma (H11-199, Fig. 7; Table 2). This is indistinguishable from the age of the McDermitt Tuff and indicates the Aurora lavas erupted immediately after ash-flow eruption. Distinctly less silicic lavas (56% SiO₂) with vitreous plagioclase laths as much as 15 mm long overlie poorly exposed intracaldera sediments (Tis) in the southwestern part of the caldera. The plagioclase yielded a ⁴⁰Ar/³⁹Ar date of 16.08 ± 0.10 Ma (16.12 ± 0.03 Ma isochron; 09-McD-53, Table 2). The source of these lavas is unknown.

The icelandite lavas overlap in SiO₂ with the precaldera icelandites, but most have slightly higher alkalis and peralkalinity index than the precaldera icelandites (Figs. 5, 6, and 9). Aurora lavas plot along the McDermitt Tuff trend in MnO and Ba, and have positive Eu anomalies like the least silicic tuff samples. In contrast, the young southern sample (09-McD-53) plots along the precaldera icelandite trend and does not have a positive Eu anomaly.

**Late Basalt (TbL; High-Alumina Olivine Tholeiite)**

Flows and a small plug of aphyric to moderately porphyritic basalt are the youngest igneous rocks in the caldera area. The flows are poorly exposed in the southern part of the caldera, where they overlie intracaldera sediments (Tis) and were deposited against aphyric rhyolite in the southeastern caldera wall (Fig. 15C). The ~300-m-diameter plug intrudes intracaldera sediments in the northern part of the caldera. The only chemical analysis, of a southern aphyric flow from drill core, shows high-alumina olivine tholeiite (Hart et al., 1984; Brueseke and Hart, 2008) with distinctive low Ti, K, P, Rb, Nb, and REEs (WLC07-13.5, Supplemental Table S1 [see footnote 1]) markedly unlike Steens Basalt. An imprecise matrix ⁴⁰Ar/³⁹Ar date on the same sample is 14.90 ± 0.74 Ma (Table 2).

**Intracaldera Sedimentary Deposits (Tis)**

A heterogeneous sequence of mostly poorly exposed, poorly lithified, predominantly tuffaceous sedimentary deposits (Fig. 15D) accumulated in a lacustrine basin in the caldera following collapse. They are preserved, and possibly were deposited, in an irregular doughnut-shaped rim or moat around the margin of the caldera. Stratigraphic relations with dated rocks indicate the sediments were deposited over a relatively long time. Through most of the caldera, the sediments overlie intracaldera McDermitt Tuff. In a small area in the northeast, they overlie rhyolite of McDermitt Creek (Trm) or Aurora lavas (Tia), both dated as 16.4 Ma. They underlie 16.1 Ma icelandite in the southwestern part of the caldera and are interbedded with ca. 14.9 Ma late basalt (TbL) near Thacker Pass. Along most of the northern and northwestern structural margins of the caldera, the sedimentary rocks overlie precaldera rocks in the caldera wall. The sediments close to the structural margin around most of the caldera dip −10° inward and locally to −25°. Our few measurements also indicate that they dip gently off the central resurgent dome-topographic high in the middle of the caldera.

The best-exposed deposits are coarse, commonly silicified conglomerate and breccia at the base of the unit near the western margin of the caldera. These basal deposits are poorly sorted with angular to subangular clasts as much as 1 m diameter of McDermitt Tuff and precaldera volcanic rocks, particularly biotite rhyolite, and lesser granite. These coarse deposits are as much as 15 m thick near the western margin and probably resulted from erosion of the nearby caldera wall soon after collapse, similar to mesobreccia but postdating all tuff eruption.

Well-bedded, planar to cross-laminated tuffaceous sandstone, siltstone, mudstone, tephra, and minor conglomerate and fossiliferous limestone overlie the coarse deposits. These finer deposits are sparsely exposed in a few canyons and in mine and exploration workings in the northern part of the caldera, where they are locally silicified by hydrothermal alteration, but poorly exposed to commonly covered by Quaternary fan deposits elsewhere, especially around Thacker Pass in the south. The tuffaceous sediments have been altered to smectite, zeolites, and potassium feldspar (Glanzman and Ryutaba, 1979) and are locally silicified to opal or chalcedony (Yates, 1942; Wallace and Roper, 1981; Hetherington and Cheney, 1985; McCormack, 1986).

Where exposed, the rocks occur mostly as float and commonly form light and dark bands on aerial photographs that correspond with varying densities of sagebrush and other vegetation. The more easily observed northern deposits consist of glass shards, fine pumice, and mineral and rock fragments. Petrified wood, diatoms, fish, and leaves occur in tuffaceous sediments and unidentified shells in limestone (Yates, 1942; Greene, 1976; Glanzman and Ryutaba, 1979; Castor et al., 1982).

The sediments are thickest in the northern caldera, where drilling encountered as much as 210 m at the Aurora U deposit (Wallace and Roper, 1981), and in the southern caldera at Thacker Pass, where drilling by Western Lithium Corporation encountered as much as 190 m. No more than ~120 m are preserved in the western rim, where the deposits are substantially eroded.

A common interpretation is that the sediments were deposited post-resurgence in a moat between the caldera wall and the resurgent uplift in the middle of the caldera (Rytuba and Glanzman, 1979; Glanzman and Ryutaba, 1979; Castor et al., 1982; Garside, 1982). If true, the sediments probably never covered much of the resurgent uplift and certainly have been removed now. Interbedding with late basalt (TbL) is consistent with deposition over a long period, some of which would have been post-resurgence. However, the inward dip of sediments near the caldera wall and the gentle dip off the uplift suggest that some sediments may have been deposited before resurgence. The intracaldera sedimentary deposits are similar to the Creede Formation of intracaldera sediments in the Creede caldera of Colorado. The Creede Formation has a thin sequence of coarse breccia overlain by much thicker lacustrine sediments, dominantly post-resurgence deposition, but possibly some preresurgence or synresurgence deposition (Larsen and Crosse, 1986; Larsen and Lipman, 2018). More structural, stratigraphic, and facies data on the sediments are needed to evaluate the full timing of deposition relative to resurgence at McDermitt.
CALDERA STRUCTURE

Caldera Collapse–Ring-Fracture System

We recognize a single, keyhole-shaped McDermitt caldera that is ~45 km north-south by 30–22 km east-west (Fig. 3; Plate 1). Although some previous studies also recognized a single McDermitt caldera (e.g., Wallace and Roper, 1981), the best-known study interpreted five overlapping nearly circular calderas in the same area, although it showed them only in a generalized, 1:1,000,000-scale, page-sized figure (Rytuba and McKee, 1984). Because of these different interpretations, we describe the geometry of the caldera, its structural margin, and field relationships around the caldera in detail. Our interpretation of a single caldera is based on continuity of a single intracaldera tuff (Tmt), of a single outflow tuff correlative with the intracaldera tuff, of the ring fault system, and of intracaldera sedimentary deposits (Tis), as well as evidence discounting the other proposed calderas (see Discussion). Older middle Miocene tuffs are absent where the base of the McDermitt Tuff is exposed in the caldera in the western escarpment, although the base is not exposed throughout the rest of the caldera. Only the tuffs of Oregon Canyon and Trout Creek Mountains underlie McDermitt Tuff in adjacent caldera walls. The continuity of intracaldera McDermitt Tuff and sedimentary deposits were discussed herein; here we describe the ring fault system.

Minor extension of the McDermitt caldera preserved its original shape. A single major caldera fault is well exposed or reasonably interpreted beneath post-collapse intracaldera sedimentary deposits (Tis) around the entire caldera (Fig. 3; Plate 1). Smaller displacement concentric faults, which commonly encompass zones where precaldera rocks are warped down into the caldera, are also present outside the major fault, especially along the northern and western margin. Moderate to high-amplitude aeromagnetic highs and lows that mark various precaldera rocks outline the caldera around the northern, western, and southern margins (Fig. 16).

Northern Margin

The major ring-fracture fault is marked by the juxtaposition of intracaldera sedimentary deposits (Tis) against precaldera rocks in the caldera wall. This relationship is especially clear in the Brez Mine, where coarse sediments and minor megabreccia are against rhyolite in what is partly a buttress unconformity (Wallace and Roper, 1981; Fig. 17A). Blocks in megabreccia and smaller clasts in sedimentary deposits are flow banded and folded spherulitic sparsely porphyritic rhyolite, which crops out in the adjacent wall. This major fault is exposed for ~4.5 km length along the northeastern margin and is covered beneath intracaldera sedimentary deposits farther west and along much of the western caldera margin.

Minimum displacement across this major fault zone of 420–580 m is estimated from the ~1980 m elevation of the base of outflow McDermitt Tuff at Flattop (Fig. 3) and the elevation of the top of intracaldera tuff in drillholes at the Aurora U deposit, which varies from ~1400 to 1560 m (Table 4; Wallace and Roper, 1981; J. Schloderer, Oregon Energy, written commun., 1985). If intracaldera McDermitt Tuff is ~600 m thick, the fault zone has >1 km displacement, possibly significantly more.

An anastomosing zone of faults, with tens of meters to ~100 m displacement mostly down toward the caldera, parallels this main fault away from the caldera in the northern wall. The zone is ~1 km wide near the Brez Mine and as much as 6 km wide in the northwestern caldera wall. Individual faults form zones as wide as 70 m, e.g., at sample location H07-186 (tuff of Oregon Canyon). Rocks in the overall fault zone dip gently, ~10° or less, into the caldera (Fig. 17A). Dip decreases across faults away from the caldera, and rocks in the wall more than a few kilometers outboard dip gently northward. Local tilting in some fault zones generated significantly steeper dips; e.g., the tuff of Oregon Canyon dips 60° southward in the fault zone at H07-186 (Plate 1).

Western Margin

A similar major fault zone with subsidiary parallel faults continues around the northwestern caldera margin to Horse Creek, where erosion along the major west-dipping Montana Mountains fault zone exposes much deeper levels (Fig. 17B). The caldera structural margin in and south of the canyon combines wide areas of steep downwarp into the caldera and discrete but variably exposed faults. In the northern part of Horse Creek, a major fault along Cherry Creek juxtaposes various Miocene volcanic rocks down against Cretaceous granodiorite (Kg). An inner fault separates intracaldera McDermitt Tuff (Tmt) and sedimentary deposits (Tis) from older Miocene and Eocene rocks and is partly overlapped by sedimentary deposits. Rocks between the two faults dip into the caldera, locally as steeply as 50°, probably as a result of slumping during collapse (e.g., Steens Basalt, Tsb; Ts; Fig. 17B). The outer fault appears to die out southward, but the steep inward dip continues (e.g., the Tvs dip as steep as 44°). The inner fault continues but is poorly exposed; it separates the Steens Basalt on the west from precaldera intermediate lavas (Tpi), peralkaline rhyolite (Tpr), and McDermitt Tuff (Tmt) on the east from north to south.

A major fault, probably the continuation of the inner fault from the north, is well exposed in mine workings at the Moonlight Mine, dips ~50° E, and separates precaldera biotite rhyolite of the Moonlight Mine (Tbrm) in the hanging wall from Cretaceous granodiorite (Kg) (Fig. 17C). We interpret this as a caldera collapse fault rather than an antithetic Basin and Range fault, because adularia in U-mineralized breccia in the fault zone yields a 40Ar/39Ar date of 16.32 ± 0.10 Ma (C95-105, Table 2), which provides a minimum age for the fault.

Total collapse along the western margin is difficult to estimate because corerelative rocks are not exposed on opposite sides of faults and much of the displacement was accommodated by steep downwarping. However, Cretaceous granodiorite (Kg) crops out at ~1945 m west of Horse Creek and is not exposed below precaldera rocks at the Moonlight Mine at ~1400 m. This indicates at least 545 m downdrop into the caldera, consistent with the ~450–610 m total thicknesses of intracaldera McDermitt Tuff along this margin (Table 4).
Figure 17. (A) North wall of caldera looking west from the Bretz Mine (Fig. 3). Rocks in the Oregon Canyon Mountains on skyline are flat-lying to gently northward dipping. Dip changes to gently southward dipping, into the caldera, across a series of small-displacement faults. (B) West wall of caldera looking north across Horse Creek. Main caldera fault here has Cretaceous granodiorite in the wall and McDermitt Tuff over slumped rocks including steeply inward dipping Steens Basalt in the caldera. The skyline is ~500 m above Horse Creek. (C) Caldera fault at the Moonlight Mine dips ~50° into the caldera, has dip-slip striae, and separates Cretaceous granodiorite from pre-collapse biotite rhyolite of the Moonlight Mine, both highly altered with U mineralization. Exposed fault face is approximately 2 m high. (D) South wall of caldera (looking southwest) with outflow McDermitt Tuff overlying aphyric rhyolite in the wall and intracaldera sedimentary deposits in foreground mostly covered by Quaternary fan deposits. (E) East wall of caldera looking north from near Pole Creek; scarp is ~200 m high. Aphyric rhyolite in the wall dips gently away from the caldera. Intracaldera McDermitt Tuff in the background is overlain by intracaldera sedimentary deposits, mostly covered by Quaternary fan deposits, and by late basalt.
The southeastern caldera margin mostly consists of a single, north-northeast–striking fault that extends ~23 km from the southern margin. The fault is expressed as a low topographic scarp (100 to as much as 270 m) with mostly aphyric to sparsely porphyritic rhyolite (Tar) and locally underlying peralkaline rhyolite (Tpr) in the caldera wall at Fig. 17D). Intracaldera McDermitt Tuff is locally exposed at the base of the topographic scarp and is separated from aphyric rhyolite (Tar) in the caldera wall by a nonexposed fault. Intracaldera sedimentary rocks (Tis) that overlie caldera faults are exposed nearby but are mostly covered by Quaternary deposits.

Total collapse along the southern margin is at least 1100–1200 m, based on outcrop of outflow tuff in the caldera wall at 1800–1940 m (Table 4). Drillhole WLC-58C (Plate 1) bottomed in intracaldera tuff at 738 m elevation.

### Eastern Margin

The southeastern caldera margin mostly consists of a single, north-northeast–striking fault that extends ~23 km from the southern margin. The fault is expressed as a low topographic scarp (100 to as much as 270 m) with mostly aphyric to sparsely porphyritic rhyolite (Tar) and locally underlying peralkaline rhyolite (Tpr) and icelandite lava (Tpi) in the caldera wall (Fig. 17E). Poorly exposed intracaldera sedimentary deposits (Tis), mostly covered by Quaternary fans, occur at the south and north ends of this segment; rhyolite of McDermitt Creek (Trm) is in the middle. Late basaltic lava (Tbl) banked against aphyric rhyolite (Tar) in the caldera wall at the south end. Two faults parallel this north-northeast segment at its northern end ~2 and 4 km to the east.

Total collapse along this margin must be >>250 m (Table 4), the difference between the highest elevation of precaldera rhyolite of Hoppin Peaks (Trbh, ~1875 m) to the east and the surface elevation in the caldera (~1624 m). Adding the thickness of rhyolite of McDermitt Creek (Trm) and intracaldera tuff (Ttmt) would greatly increase this estimate, probably to at least 1000 m.

The caldera margin turns abruptly east-northeast for ~11 km to the Cordero and McDermitt Mines at the northern end of the north-northeast segment (Fig. 3; Plate 1). The main north-northeast segment and the two parallel outboard segments truncate at the east-northeast segment. This east-northeast segment may have reactivated older (pre-Cenozoic?) structures during collapse. It parallels the vitrophyre pyroclastic dikes (Tmv), and both are strongly discordant to other collapse structures. The eastern end of the east-northeast segment disappears beneath Quaternary deposits. However, the caldera wall presumably continues beneath Quaternary deposits to connect to the eastern end of the northern caldera margin (Fig. 3; Plate 1). The difference in elevation between rhyolite of Hoppin Peaks (Trbh) and intracaldera sedimentary rocks (Tis) in the McDermitt Mine (<1400 m) requires at least 475 m of collapse. Adding rhyolite of McDermitt Creek and intracaldera tuff would greatly increase this estimate. Fisk (1968) reported a rhyolite dike along the northeast segment at the Cordero Mine; however, he also said that the dike bottomed at ~210 m depth, which complicates interpretation of the body as intrusive or extrusive.

### Resurgence

Resurgence of the McDermitt caldera created a structural and topographic high where intracaldera McDermitt Tuff crops out at its highest elevations in the center of the caldera (Fig. 18; Plate 1, section B–B’). Intracaldera tuff crops out at slightly more than 2000 m elevation immediately north and south of Round Mountain, ~700 m higher than the 1300 m top of tuff in drillhole WLC-58C in Thacker Pass in the southern part of the caldera (Fig. 3). This structural-topographic expression is most apparent in a 3× exaggerated Google Earth (https://www.google.com/earth/) image (Fig. 18). Without this clear geologic and topographic expression, doming would be difficult to confirm because of the lack of good structural markers. Foliation in intracaldera tuff, which is the primary structural marker in most calderas (e.g., John et al., 2008), is unsuitable at McDermitt because the tuff is so strongly rheomorphic that paleoshortening is rarely preserved. Obtaining attitudes in the poorly exposed and commonly slumped intracaldera sedimentary deposits (Tis) is difficult, so they are of limited use. Whether doming preceded, followed, or overlapped with sediment deposition remains uncertain.

Doming was accomplished partly by upwarping of intracaldera tuff and probably partly by faults approximately tangential to the uplift. Upwarped is apparent around most of the dome where the surface slopes steeply upward (Fig. 18). In addition, east-striking, down-to-the-south faults separate tuff on the north from sediments on the south in Thacker Pass. Because of their orientation, perpendicular to north-striking extensional faults, these faults were most likely active during resurgence. If so, resurgence postdated some sediment deposition.

The topographic high drops north of Long Ridge into the McDermitt Creek drainage, then rises slightly north of the creek in the northern part of the caldera. It is unknown whether this topographic pattern results from variation in original doming or from erosion along McDermitt Creek, which is the largest drainage in the caldera and the only one that crosses the entire resurgence uplift.

The resurgent uplift coincides well with an aeromagnetic high and with the highest anomalies at the late icelandite vents at Black and Round Mountains.

### Table 4. Amount of Collapse, McDermitt Caldera

| Part of caldera             | Minimum (m) | Maximum (m) |
|-----------------------------|-------------|-------------|
| North                       | 580         | >1000       |
| West                        | 545         | >610        |
| South                       | 1100–1200   | ≥1100–1200  |
| Southeast                   | 250         | >>250       |
| Northeast segment           | 475         | >475        |

*Henry et al. | Geology and evolution of the McDermitt caldera*
This pattern suggests that post-collapse intrusion or rise of icelandite magma caused resurgence, and some magma actually erupted. Based on dates on icelandites, resurgence could have occurred almost immediately after ash-flow eruption and caldera collapse or as much as ca. 0.1 Ma or more later (Table 2; Fig. 7).

MINERALIZATION

The McDermitt caldera has a wide range of associated mineral deposits and elemental enrichment, far more mineralization than in other known Yellowstone hotspot calderas (Table 5; Figs. 3 and 7). This discussion focuses on how mineralization is related to caldera activity and structure; references cited in Table 5 provide more complete descriptions of many deposits, and Vikre et al. (2016) summarized publicly available data on most. The Cordero, McDermitt, Bretz, and Opalite Mines were the largest mercury producers in the U.S. between 1933 and 1989. Uranium deposits were evaluated extensively but yielded only minor production in the 1950s from the Moonlight Mine. Gallium and gold anomalies have also been investigated. Deposits of hectorite, a high-value Li-smectite used for specialty drilling fluids, yielded small production and are currently being developed on a larger scale (Odom, 1992; Western Lithium Corporation, 2014; now LithiumAmericas; Fig. 3). The lithium is being evaluated for use in Li-ion electric batteries (http://lithiumamericas.com/companies/lithium-nevada-corp/).

All non-lithium deposits and prospects are along or just inside the ring-fracture system of the caldera (Fig. 3), and most reports emphasize the importance of caldera faults as conduits for hydrothermal fluids. The Moonlight Mine exemplifies this relationship. Uranium occurs as breccia fill along the western ring-fracture fault at the mine, and the age of related adularia (16.32 ± 0.10 Ma; C95-105; Table 2) indicates that mineralization occurred soon after caldera collapse. Intrusions are interpreted to be the heat source for alteration at the Moonlight Mine. For example, Rytuba (1976, 1994) and Rytuba et al. (1979) variably cited rhyolite domes or a peralkaline rhyolite dike as intrusions along the western ring fracture and as host for the Moonlight and Horse Creek uranium deposits. However, we find these rocks to be precaldera biotite (Tbr) and peralkaline rhyolite (Tpr) lavas. Although ring-fracture intrusions are common in many calderas and are logical sources of heat and possibly metals, ring-fracture intrusions are notably scarce at McDermitt.
The timing of other uranium mineralization is defined only as being younger than host rocks (Table 5). However, similar element assemblages, especially enrichment in zirconium at Moonlight, Horse Creek, Big Bend Spring, and Old Man Spring, and their spatial association in the western part of the caldera, suggest genetic and possibly temporal ties. Zirconium enrichment appears to decrease upward, with the greatest enrichment at the Moonlight Mine and the least at Old Man Spring, where mineralization probably extends into basal intracaldera sedimentary deposits (Castor et al., 1982). All of these occurrences have U-rich, hydrothermal zircon, which hosts most of the uranium at the Moonlight Mine (Castor and Henry, 2000). All U-Zr mineralization may have formed at the same time, shortly after tuff eruption and caldera collapse and early during deposition of the intracaldera sedimentary deposits (Tis).

The timing of other uranium mineralization is defined only as being younger than host rocks (Table 5). However, similar element assemblages, especially enrichment in zirconium at Moonlight, Horse Creek, Big Bend Spring, and Old Man Spring, and their spatial association in the western part of the caldera, suggest genetic and possibly temporal ties. Zirconium enrichment appears to decrease upward, with the greatest enrichment at the Moonlight Mine and the least at Old Man Spring, where mineralization probably extends into basal intracaldera sedimentary deposits (Castor et al., 1982). All of these occurrences have U-rich, hydrothermal zircon, which hosts most of the uranium at the Moonlight Mine (Castor and Henry, 2000). All U-Zr mineralization may have formed at the same time, shortly after tuff eruption and caldera collapse and early during deposition of the intracaldera sedimentary deposits (Tis).

The timing of mercury mineralization at the McDermitt Mine, which is hosted by intracaldera sedimentary deposits (Tis), is poorly constrained. Step-heating of fine-grained (≤50 µm) adularia intergrown with opal did not yield a plateau (McD; Table 2, Supplemental Table S2 [see footnote 2]), but...
indicates that mineralization occurred approximately coincident with caldera formation. Individual steps yielded ages ranging from 16.15 ± 0.10 to 16.97 ± 0.10 Ma; the largest single step, with 25% of released 39Ar, is 16.25 ± 0.09 Ma. All steps but one were 32%–38% radiogenic, probably reflecting the presence of some residual opal with abundant atmospheric Ar in the mineral concentrate. An isochron of all steps is 16.67 ± 0.14 Ma (McD; Table 2), but this age is older than those of ash-flow eruption, caldera collapse, and deposition of intracaldera sedimentary deposits. A K-Ar date on alunite of 12.6 ± 0.7 Ma was interpreted to indicate a second, younger hydrothermal episode (Rytuba and McKee, 1984; Noble et al., 1988). However, the date more likely indicates supergene formation of alunite in a low-sulfidation epithermal system in which adularia is the primary K alteration phase (J.K. McCormack, 2016, personal communication; McCormack, 1986).

Stratiform lenses containing Li-enriched illite or smectite are found in intracaldera sedimentary deposits (Tis) throughout the caldera (Glanzman and Rytuba, 1979; Odom, 1992; Western Lithium Corporation, 2014; Fig. 3). The relative contributions of low-temperature diagenetic versus higher temperature hydrothermal fluids in concentrating lithium are uncertain. Glanzman and Rytuba (1979) and Rytuba and Glanzman (1979) proposed that the lithium mineralization resulted from leaching of lithium from nonwelded glass in the sedimentary deposits and concentration in potassium feldspar-rich alteration zones; they reported that alteration mineralogy progresses from almost fresh glass, to a mix of zeolites, to analcime, and to potassium feldspar, and interbedded mudstones are altered to dioctahedral smectite and trioctahedral smectite including hectorite. Our recent work additionally found interlayered illite/smectite and illite. Except for illite, this mineral zonation is typical of that of forms during closed hydrologic system diagenesis, where groundwater evolves through rock-water interaction and evaporation in a closed basin such as in the McDermitt caldera to become highly alkaline and saline (Surdam, 1977; Langella et al., 2001). Most closed-system diagenesis occurs at near surface temperature, ≤50 °C (Langella et al., 2001), although diagenesis in a volcanically active area could be at higher temperatures. Glanzman and Rytuba (1979) did not report a temperature of diagenesis, but interpreted that hydrothermal alteration overprinted the diagenetic sequence at five locations in the caldera: at the McDermitt, Breit, and Opalite Mines, a location 2 km south of the Breitze Mine, and a general area along the western part of the caldera. Based on studies of the diagenesis of Paleogene–Neogene deposits of the Gulf Coast, the presence of illite could indicate temperatures of 100 °C or more (Boles and Franks, 1979; Bourdelle et al., 2013). However, Turner and Fishman (1991) found that pore-water chemistry can be more important than temperature in illite formation and that illite can form at near-surface temperatures in saline alkaline lakes similar to the McDermitt caldera setting.

Potassium feldspar from the Western Lithium Corporation Kings Valley Stage I lens at Thacker Pass yielded a 40Ar/39Ar date of 14.87 ± 0.05 Ma (WLC43-180.2; Table 2). This date indicates that potassium feldspar formation occurred long after most caldera-related magmatism, although possibly contemporaneous with eruption of late basalt lavas that are interbedded with the intracaldera sedimentary deposits in the southern part of the caldera. The implications of this date for the timing of lithium deposition are less certain. If hectorite and Li-rich smectite formed during diagenesis, then at least some lithium concentration resulted from diagenesis. The date seems consistent with the sedimentary deposits taking a considerable time to accumulate and become altered by low-temperature groundwater diagenesis. A low-temperature origin is consistent with proposed mechanisms of formation of Li-rich brines in closed basins (Davis et al., 1986) and extreme enrichment and posterosive depletion of lithium in some rhyolites of the western United States (Hofstra et al., 2013). The presence of illite is the primary evidence for higher temperatures, but its significance is debated. Problems with an igneous-driven hydrothermal system include distribution of lithium enrichment throughout the caldera, lack of association with igneous rocks except locally with the small-volume and areally restricted basalts, and association with probable low-temperature diagenesis.

Regardless of the exact origin and conditions of lithium mineralization, the different types of deposits and elemental concentrations and multiple episodes of igneous activity at the McDermitt caldera suggest considerable potential for overprinting by multiple hydrothermal systems and diagenesis. Further determination of deposit types, mineralogic sites of concentrated elements, age, formation temperature, and potential source rocks (e.g., modern determination of lithium concentrations in unaltered rocks) is needed to evaluate the origin of deposits.

**EXTENSION**

The McDermitt caldera is within the Basin and Range Province, but the minor tilting of the caldera demonstrates that it has undergone only minor extension, which probably began ca. 12 Ma based on thermochronology of the Santa Rosa Range to the east and the Pine Forest Range to the west (Figs. 2 and 3; Colgan et al., 2004, 2006a, 2006b). Quaternary faults bound both the western and eastern sides of the Montana Mountains (Plate 1). The west-dipping Montana Mountains fault zone forms the western edge of the Montana Mountains and the Double H Mountains to the south, and dies out northward into the Trout Creek Mountains (Plate 1; U.S. Geological Survey, 2016). The east-dipping Hoppin Peaks fault zone forms the eastern edge of the Montana Mountains and is an antithetic set to the larger displacement, west-dipping Santa Rosa Range fault system that forms the western front of the Santa Rosa Range.

The caldera is gently east tilted, based on the overall higher elevations of the west side of the caldera block, relative elevations of poorly stratified or exposed rocks across the caldera, and especially on obvious gentle, –10°–12° east dips of better-layered volcanic rocks to the south and north, e.g., in the Double H Mountains (our observations) and in the Trout Creek Mountains of Orevada View (Plate 1) and farther north (Minor et al., 1989b).
DISCUSSION

How Many Calderas?

Our interpretation of a single caldera contrasts with the conclusions of Rytuba and McKee (1984), who interpreted three partly overlapping calderas composing a McDermitt caldera complex (Fig. 3; Table 6) plus two adjacent and partly overlapping calderas, one younger and one older than the caldera complex. Our interpretation of a single caldera is based on continuity of a single intracaldera tuff (Tmt), the ring fault system, and intracaldera sedimentary deposits (Tis), and absence of evidence for the other proposed calderas. Here we discuss significant aspects of the geology that we interpret either similarly or differently in comparison to Rytuba and McKee (1984). A limitation is that published depictions of the previous interpretation consist only of a 1:1,000,000-scale page-sized map (Rytuba and McKee, 1984), and generalized, page-sized geologic maps do not show the individual calderas (Rytuba, 1976; Rytuba and Glanzman, 1978, 1979; Glanzman and Rytuba, 1979).

Proposed McDermitt Caldera Complex

Rytuba and McKee (1984) divided the McDermitt caldera as identified here into three separate calderas: from north to south, the Long Ridge, Jordan Meadow, and Calavera calderas (Fig. 3; Table 6). An immediate problem is that their only caldera depiction, Figure 1 of Rytuba and McKee (1984), shows the Long Ridge caldera blending continuously into the Jordan Meadow caldera with no boundary, and their text identifies no map boundary between intracaldera tuffs, so it is impossible to know where one interpreted caldera ends and the other starts. In contrast, we find intracaldera McDermitt Tuff, indistinguishable in outcrop characteristics, phenocryst assemblages, composition (Starkel et al., 2012), and precise 40Ar/39Ar age, to be continuously exposed from north of McDermitt Creek to near Thacker Pass, through all three proposed calderas.

In the most clearly described relationships, Rytuba and McKee (1984) proposed that the Calavera caldera collapsed during eruption of the tuff of Double H (formerly member 1 of the tuff of Long Ridge; Rytuba and Glanzman, 1978, 1979), an aphyric comendite that they showed in the southern and southeastern caldera wall where we mapped aphyric rhyolite lava (Tar). Rytuba and McKee (1984) placed the northern boundary of the caldera through Calavera Canyon; that boundary is shown overlapped by the southwestern margin of the younger Jordan Meadow caldera slightly to the northeast, and the Jordan Meadow boundary is shown as buried or projected through continuous intracaldera McDermitt Tuff (Fig. 3). As initially discussed by Castor and Henry (2000), precaldera rocks [biotite rhyolite of the Moonlight Mine (Tbr) and peralkaline rhyolite (Tpr)], intracaldera McDermitt Tuff (Tmt), and post-collapse intracaldera sedimentary deposits (Tis) are not displaced and do not change thickness across Calavera Canyon or anywhere within several kilometers to the north or south; this precludes a caldera boundary fault there (Fig. 11A). Hargrove and Sheridan (1984) came to a similar conclusion.

Proposed Adjacent Overlapping Calderas

Washburn caldera: Proposed source of the tuff of Oregon Canyon. Rytuba and McKee (1984) interpreted a Washburn caldera, the source of the tuff of Oregon Canyon, partly overprinted by and partly exposed northeast of the northeastern part of the McDermitt caldera (Fig. 3). However, we find no evidence for any caldera in this area (see following). Rytuba and McKee (1984, p. 8618) interpreted the northwestern Washburn caldera structural margin to be an arcuate, 6-km-wide zone of down-to-the-southeast faults that “dis-

| Proposed caldera | Proposed caldera-forming tuff | This study | Key points |
|------------------|-------------------------------|------------|-----------|
| Whitehorse       | Tuff of Whitehorse Creek      | Yes, but possibly for tuff of Trout Creek Mountains, not for tuff of Whitehorse Creek | Tuff of Whitehorse Creek 47 m thick in caldera, suggests deposited in existing caldera. |
| Hoppin Peaks     | Biotite rhyolite of Hoppin Peaks | Not a caldera | Proposed intracaldera tuff consists of 3 rhyolite lavas totaling 210 m thick, no structural margin, and proposed margin appears to go through middle of lava flows. |
| McDermitt caldera complex | Tuff of Long Ridge member 5 | Part of single McDermitt caldera | Intracaldera McDermitt Tuff, indistinguishable in outcrop characteristics, phenocryst assemblage, composition, and 40Ar/39Ar age, crops out continuously through entire McDermitt caldera. Structural margin is continuous around entire caldera. Proposed Calavera margin goes uninterrupted through precaldera, intracaldera, and postcaldera rocks. Tuff of Long Ridge members 1, 2, and 3 are rhyolite lavas. |
| Jordan Meadow    | Tuff of Long Ridge member 2, 3 | Part of single McDermitt caldera | No structural margin and no intracaldera tuff or postcaldera fill despite exposure of rocks older than tuff of Oregon Canyon in proposed caldera. Tuff of Oregon Canyon farther from proposed caldera appears to be closer to source, which may be in the Whitehorse caldera area (Benson and Mahood, 2015). |
| Calavera         | Tuff of Double H (tuff of Long Ridge member 1) | Part of single McDermitt caldera |
| Washburn         | Tuff of Oregon Canyon         | Not a caldera | |
place the tuff of Oregon Canyon downward 250 m.” Figure 5a of Rytuba and McKee (1984), a generalized geologic cross section, shows the following. (1) The northwestern 4 faults have a few tens to ~100 m displacement of the tuff of Oregon Canyon, which maintains constant thickness across the faults. (2) The tuff was displaced ~600 m down across the southeasternmost fault, which coincides with the caldera boundary shown in Figure 3 but is in an area for which no published geologic map is available. Rytuba and McKee (1984) seemed to indicate that the tuff is nowhere exposed in the proposed source caldera. (3) Intracaldera tuff is interpreted to be covered beneath ~600 m of caldera-filling rhyolite lavas (the upper rhyolite, Turf, of Rytuba and Curtis, 1983), which are ~150 m thick northwest of the fault. The abrupt change in thickness of these post-tuff lavas is cited as evidence that “define the northwest margin of the caldera.” However, these lavas are not exposed south of the southeasternmost fault in the proposed caldera. (4) A “400-m-thick sequence of iron-rich andesite flows, ponds along the northwestern ring zone, indicates continued subsidence of the caldera”; these are displaced ~200 m across the southeastern fault but do not change thickness across it: in addition, “These lavas are restricted to areas adjacent to the Washburn caldera.” (Rytuba and McKee, 1984, p. 8618). Rytuba and McKee (1984) also interpreted the northeast-striking part of the McDermitt caldera margin north of Hoppin Peaks to be a reactivated part of the Washburn caldera.

In contrast to point 2, Peterson et al. (1988) did not show the southeasternmost fault, but showed an ~1.5-km-long, thin (≤100 m?) lens of tuff of Oregon Canyon ~2 km south of Mendi Suri, which would be close to the middle of the proposed Washburn caldera. If present, the tuff would indicate little displacement across the southeasternmost fault and remarkably thin intracaldera tuff. However, we found that area to consist entirely of precaldera intermediate lavas (unit Tpi). In contrast to point 3, we found no great thickness of rhyolite lavas that could correlate with the upper rhyolite. In contrast to point 4, Rytuba et al. (1983a, 1983b) showed the same iron-rich andesite map units cropping out at least 15 km to the northwest. We find similar icelandite lavas around most of the McDermitt caldera.

The most telling evidence against a Washburn caldera is that the quartz-sanidine rhyolite lava dome (Trq), which predates tuff of Oregon Canyon (Fig. 7; Table 2), occupies the proposed center of the caldera. The presence of this older rock and absence of tuff of Oregon Canyon in the section preclude this area being a caldera for the tuff.

Sparse evidence suggests that the source for the tuff of Oregon Canyon could be significantly north or west of the McDermitt caldera. The tuff at the proposed northernmost caldera wall just northwest of the southeasternmost fault of point 1 is ~35 m thick with an ~8-m-thick poorly welded zone at the top, has few lithics ≤2 cm in diameter, and shows no evidence of rheomorphism. The tuff of Oregon Canyon in Oregon Canyon 10 km to the northwest appears to be much more proximal, ~60 m thick, more coarsely (to 15 cm diameter) and abundantly lithic, and strongly rheomorphic with flow folds and ramps to vertical. These characteristics suggest a source farther north, and Benson et al. (2017) interpreted a source caldera that is overprinted by the Whitehorse Creek caldera (see following; Figs. 2 and 3). A less likely alternative is that the tuff of Oregon Canyon correlates with the petrographically and compositionally similar Idaho Canyon Tuff, which erupted from the Virgin Valley caldera 65 km to the west (Noble et al., 1970; Castor and Henry, 2000). However, Coble and Mahood (2016) argued that the two tuffs are not the same based on small compositional differences.

**Whitehorse caldera: Proposed source of the tuff of Whitehorse Creek.** The ~15–20-km-diameter Whitehorse caldera is ~20 km northwest of the McDermitt caldera and is interpreted to have formed during eruption of the tuff of Whitehorse Creek (Figs. 2 and 3; Rytuba et al., 1981; Rytuba and McKee, 1984). The caldera interpretation is reasonable; a shallow topographic basin contains tuffaceous sediments including diatomite with five rhyolite domes along the probable ring fracture (Rytuba et al., 1981; Benson et al., 2017). However, the caldera being the source of the tuff of Whitehorse Creek is less certain. This tuff, for which we obtained a sanidine 40Ar/39Ar date of 15.67 ± 0.04 Ma (MD-8, Table 2), crops out in the eastern part of the caldera where it laps onto the caldera wall (Rytuba et al., 1981, 1983b). The tuff of Whitehorse Creek is nearly flat lying in the caldera but dips moderately, to 28°, into the caldera where it is compacted against the southeastern caldera wall (Rytuba et al., 1983b; our observations); this indicates that the caldera wall existed when the tuff was deposited. Moreover, Rytuba et al. (1981) reported that a drillhole to a depth of 226 m in the southern part of the caldera encountered 136.5 m of caldera-fill sedimentary deposits, 47 m of intracaldera tuff of Whitehorse Creek, and 43 m of rhyolite lava to the bottom of the hole. Although a caldera-forming tuff could compact against its own caldera wall, the thinness of the tuff of Whitehorse Creek inside the caldera suggests that the tuff was deposited in an existing caldera. The 47 m thickness of intracaldera tuff is very small compared to the thickness of intracaldera McDermitt Tuff and seems insufficient to generate caldera collapse. The Whitehorse caldera may have formed during eruption of an older tuff, the most likely being the 16.49 Ma tuff of Trout Creek Mountains, based on that unit’s distribution. From new geologic mapping and geochronology, Benson et al. (2017) came to a similar conclusion and even interpreted calderas for both the tuffs of Trout Creek Mountains and Oregon Canyon to be nested beneath the Whitehorse caldera (Fig. 2).

**Hoppin Peaks caldera: Proposed source of the tuff of Hoppin Peaks.** Rytuba and McKee (1984) interpreted a Hoppin Peaks caldera, the source of their tuff of Hoppin Peaks, east of the McDermitt caldera and largely buried beneath Quinn River Valley (Fig. 3). Much of the cited evidence consists of a closed 20 mg gravity low in the valley. However, the rhyolite of Hoppin Peaks consists of three separate rhyolite lavas that lack any evidence of being pyroclastic, e.g., there are no nonwelded facies, no pumice, fiamme, or lithics in any rock, and they are cumulatively only 210 m thick. The proposed structural margin of Rytuba and McKee (1984, Fig. 1 therein) appears to go through the middle of the lava outcrop. The gravity anomaly is better interpreted to indicate basin fill in an extensional basin (Figs. 2 and 3).
Research Paper

Timing of Steens Basalt and Precaldera Intermediate Magmatism and Initiation of the Yellowstone Hotspot

The Steens Basalt has been difficult to date precisely, but new \(^{40}\)Ar/\(^{39}\)Ar dating indicates initiation between ca. 16.85 and 16.74 Ma. Camp et al. (2013; see also Brueseke et al., 2007; Jarboe et al., 2010; Barry et al., 2013) reported 7 new \(^{40}\)Ar/\(^{39}\)Ar dates on plagioclase separates and whole rocks and 41 published Steens Basalt dates from many sources, and 14 dates on silicic rocks that overlie the Steens Basalt (including some of our data), all recalculated to the 28.201 Ma age of the Fish Canyon Tuff sanidine monitor (Kuiper et al., 2008) used in this study. Based on these data, Camp et al. (2013) concluded that eruption of their formally proposed Steens Basalt may have begun ca. 16.85 Ma and lasted ca. 0.3 Ma (Fig. 7). Based on \(^{40}\)Ar/\(^{39}\)Ar dates on silicic tuffs interbedded with Steens Basalt and assumed basalt accumulation rates, Mahood and Benson (2016, 2017) estimated a slightly younger time of initiation, ca. 16.74 Ma.

Our data, with four dates between 16.691 ± 0.021 and 16.607 ± 0.015 Ma on three lavas and one tuff probably all in the middle or upper part of the Steens Basalt, and the 16.56 ± 0.02 Ma (n = 4) or 16.517 ± 0.013 Ma dates on the capping tuff of Oregon Canyon, are consistent with this Steens Basalt eruption timing (Fig. 7). Except for the tuff of Oregon Canyon, none of these units are in the two sections examined by Camp et al. (2013) near the McDermitt caldera, so precise correlation into those sections is not possible. Precaldera intermediate lavas (Tpi) overlie the tuff of Oregon Canyon and continued at least until the 16.53 ± 0.16 Ma date on icelandite in the north wall. Icelandite continued to erupt inside the caldera until 16.08 ± 0.10 Ma, although whether this rock represents continued magmatism related to the precaldera lavas or late erosion from the McDermitt magma chamber is unknown.

Although not the oldest caldera (Fig. 2), the McDermitt volcanic field has some of the oldest silicic volcanism in the region, ca. 16.7 Ma. The Virgin Valley caldera, source of the 16.56 Ma Idaho Canyon Tuff, is the oldest currently identified caldera (Castor and Heny, 2000; Hausback et al., 2012; Coble and Mahood, 2016). A possible caldera for the tuff of Oregon Canyon in the Whitehorse area northwest of McDermitt could be slightly older (Benson and Mahood, 2015; Benson et al., 2017). Rhyolite lavas as old as ca. 16.7 Ma are also abundant in the Santa Rosa–Calico volcanic field immediately to the east (Fig. 2; Brueseke and Hart, 2008), to the west in the Hawks Valley–Lone Mountain volcanic field (Wypych et al., 2011), and far to the northeast in the Silver City, Idaho, area (Bonnichsen et al., 2008).

The timing and, aside from a major ash-flow eruption, style of magmatism at McDermitt are similar to that in the Santa Rosa–Calico volcanic field east of McDermitt (Fig. 2; Brueseke and Hart, 2008; Brueseke et al., 2008). Santa Rosa–Calico magmatism started with eruption of the Steens Basalt, was followed by a wide range of mafic through silicic lavas, and mostly ended ca. 16.2 Ma, with some minor, mostly mafic magmatism much later, ca. 14 Ma (Brueseke and Hart, 2008). Significant differences are that the voluminous Santa Rosa–Calico magmatism did not culminate with major ash-flow eruption and caldera collapse, although a small (~2.5 x 3.5 km) caldera erupted a local tuff ca. 15.5 Ma, and high-alumina olivine tholeiite magmatism occurred only early in the field.

Silicic volcanism as old as ca. 16 Ma is even more widespread, from the High Rock caldera complex to the west, to the Lake Owyhee volcanic field, and approximately located Dinner Creek center (Oregon) and Silver City Range (Idaho), and east to the Jarbidge Rhyolite (Fig. 2; Noble et al., 1970; Cummings et al., 2000; Bonnichsen et al., 2008; Wypych et al., 2011; Hausback et al., 2012; Brueseke et al., 2014; Streck et al., 2015; Benson and Mahood, 2016; Coble and Mahood, 2016). The Jarbidge Rhyolite is distinctly old relative to its longitude along the Yellowstone track, petrologically distinct, and may be more related to regional extension (Brueseke et al., 2014). Nevertheless, early Yellowstone silicic activity is widespread, extending over an area of ~40,000 km², but is concentrated in discrete centers with abundant silicic activity. Silicic rocks are absent in major sections and in known or probable source areas of the Steens Basalt away from the silicic centers, e.g., at Steens Mountain and in the Catlow Peak section in the Trout Creek Mountains just northwest of the McDermitt caldera (Fig. 3; Camp et al., 2013). Although basaltic activity began in the northern Nevada rift at about the same time as regional Steens magmatism, silicic volcanism did not begin until much later, ca. 15.7 Ma (John et al., 2000).

Hart et al. (1984) identified a distinctive high-alumina olivine tholeiite magma type that is widely distributed through northeastern California, southeastern Oregon, southwestern Idaho, and northern Nevada; it began erupting ca. 10. 5 Ma, and continued to later than 0.1 Ma throughout most of this area. Recognition of high-alumina olivine tholeiite lavas and dikes as old as ca. 16.8 Ma in the Santa Rosa field (Brueseke and Hart, 2008) and 14.90 ± 0.74 Ma at McDermitt (Table 2) confirms the geographic distribution, but significantly increases its time of initiation. These results further support high-alumina olivine tholeiite being a small volume but integral component of regional magmatism (Hart et al., 1984).

Continuity of Magmatism and Peralkaline versus Metaluminous-Peraluminous Activity

Field relations and high-precision \(^{40}\)Ar/\(^{39}\)Ar dating indicate that magmatism around the McDermitt caldera was essentially continuous from the beginning of Steens Basalt eruptions ca. 16.85 and 16.74 Ma (Brueseke et al., 2007; Jarboe et al., 2010; Camp et al., 2013; Barry et al., 2013; Mahood and Benson, 2016, 2017; this study) through eruption of the McDermitt Tuff and caldera collapse ca. 16.39 Ma, to post-collapse volcanism. Erupted rocks generally became more silicic through time. The oldest eruptions were ~47–49 wt% SiO$_2$ lavas of the Steens Basalt (Camp et al., 2013, and references therein; Starkel, 2014; Starkel et al., 2014). Minor rhyolite erupted as early as 16.691 ± 0.021 Ma, and rhyolite predominated by ca. 16.5 Ma. The precaldera intermediate lavas (Tpi) are between the Steens Basalt and the rhyolite lavas in composition (56%–66% SiO$_2$; Supplemental Table S1 [see footnote 1]) and strati-

---

Henry et al. | Geology and evolution of the McDermitt caldera
graphic position; they overlie the tuff of Oregon Canyon (16.56 ± 0.04 Ma, n = 4, MAP215–50; 16.517 ± 0.013 Ma, Argus VI) and partly underlie the tuff of Trout Creek Mountains (16.49 ± 0.03 Ma, n = 3, MAP215–50).

Peralkaline and biotite-bearing (metaluminous-peraluminous) rhyolites overlapped closely in space and time around the McDermitt caldera, but are somewhat antithetic in detail. The variably peralkaline, sparsely anor- thoclase-phyrty to aphyric rhyolite lavas (various Tra, Tar) that erupted semicontinuously from 16.69 to eruption of the McDermitt Tuff at 16.39 Ma are nearly absent in the southwestern wall, which consists mostly of the 16.619 ± 0.020 Ma biotite rhyolite of the Moonlight Mine (Tbr). The closest overlap between peralkaline and metaluminous activity is in the eastern caldera wall, where 16.38 ± 0.07 Ma biotite rhyolite of Hoppin Peaks was deposited closely after the underlying 16.397 ± 0.061 Ma anorthoclase and aphyric rhyolite. The 16.410 ± 0.015 Ma (n = 3) sanidine-quartz-phyrty peralkaline rhyolite (Tpr) appears to have been a distinct event unlike any of the other rhyolites. We have not found any evidence that the different magma systems interacted during eruption. However, their close spatial and temporal proximity would seem to allow for the possibility (e.g., Allan et al., 2012).

Emplacement of icelandite lavas after caldera collapse suggests (1) that the caldera magma chamber retained some magma after tuff eruption, or (2) that regional, less silicic magmatism was reestablished after interruption by the caldera system density filter, which prevented the rise of higher density, more mafic magmas through lower density, more silicic magmas of the caldera. The similarity between postcaldera and precaldera icelandites hampers choosing between these alternatives. That the immediately postcaldera Aurora lava icelandites have MnO (Fig. 6E) and Ba contents that plot on the McDermitt Tuff trend supports alternative 1. That the much younger southern icelandite lavas plot along the precaldera icelandite trend may support alternative 2 once sufficient time elapsed. Determining the regional distribution and timing of similar magmatism could help distinguish between the alternatives.

Comparison of the McDermitt Caldera with Other Silicic Calderas

Here we compare the McDermitt caldera to other calderas associated with voluminous, silicic magmatism but spanning the range from strongly peralkaline to calc-alkaline (metaluminous), and (2) evaluate to what extent the McDermitt caldera is an analog for the volcanic centers associated with voluminous rhyolites but buried beneath younger rocks of the central Snake River Plain (Table 7). Comparisons are made with strongly peralkaline calderas, for which erupted tuffs have peralkalinity indices >1.1 to 2.0 (Mahood, 1984), and with calderas of the ignimbrite flareup of western North America. The ignimbrite flareup formed a semicontinuous belt of ignimbrites and source calderas from the Great Basin of the United States to the southern end of the Sierra Madre Occidental in Mexico (Swanson and McDowell, 1984; Ferrari et al., 2007; McDowell and McIntosh, 2012; Henry et al., 2012b; Best et al., 2013a, 2016). Overall understanding of the geometry, evolution, and geotectonic setting of calderas is based heavily on study of the ignimbrite flareup, particularly in the United States (Table 7; Lipman, 1984, 2007; Best et al., 2013a, 2013b, 2016; Henry and John, 2013; Lipman and Bachmann, 2015).

Important characteristics of the McDermitt caldera are (1) generation by eruption of a strongly zoned, probably high-temperature peralkaline rhyolite to metaluminous icelandite, (2) large area (~930 km²) and volume (~1000 km³) but relatively shallow collapse (~1 km), (3) major resurgence, (4) extensional setting that was not undergoing active extension (Colgan et al., 2004, 2006a, 2006b; Colgan and Henry, 2009), and possibly (5) location in accreted island arc or transitional terrane crust (Fig. 1).

The McDermitt caldera is most like calderas of the ignimbrite flareup, especially in producing pyroclastic deposits with strong compositional zoning, large volume, and resurgence, as well as most subsidiary characteristics (Table 7). McDermitt and many ignimbrite flareup calderas also have similar crustal settings and tectonic environments. The McDermitt caldera is in thin, accreted island arc crust, as are the westernmost of the ignimbrite flareup calderas (Henry and John, 2013; John et al., 2015). The tectonic setting of McDermitt may have been more extensional than that of the ignimbrite flareup, but both occurred during a neutral environment that lacked any active deformation (Best and Christiansen, 1991; John et al., 2008; Colgan and Henry, 2009; Henry and John, 2013). The most notable differences are that the McDermitt caldera is larger in area than all except a few ignimbrite flareup calderas, but has undergone far less collapse, ~1 km versus typically 3–4 km. The proportion of intracaldera versus outflow tuff is similar for McDermitt and ignimbrite flareup calderas and appears to be independent of other variables, including amount of collapse. The paucity of ring-fracture volcanism at McDermitt is also anomalous, especially given that ring-fracture rhyolites are abundant around the other early hotspot calderas, Virgin Valley and High Rock (Castor and Henry, 2000; Hausback et al., 2012; Coble and Mahood, 2016). The similarities and differences with ignimbrite flareup calderas apply even to the rare peralkaline ignimbrite flareup caldera. The 25.4 Ma Questa caldera of northern New Mexico is ~15 km in diameter, underwent more than 2 km of collapse during eruption of 500 to possibly 1000 km³ of slightly zoned peralkaline rhyolite, and has resurgent and ring-fracture intrusions of peralkaline rhyolite indistinguishable in age and composition to the erupted tuff (Lipman et al., 1986; Lipman, 1988; Zimmerer and McIntosh, 2012).

The McDermitt caldera is less similar to those associated with strongly peralkaline magmas (Mahood, 1984; Table 7). The McDermitt Tuff shares strong compositional zoning and is at the low end of the alkalinity range, but the caldera is much larger in area, erupted volume, and amount of collapse, and underwent resurgence, which is rare among strongly peralkaline calderas.

These comparisons suggest that magma system volume and derivative eruptible volume may be the most important factors in determining most caldera characteristics. Intrinsic variables such as peralkalinity and inferred high temperature strongly encouraged the intense rheomorphism shown by the McDermitt Tuff and most strongly peralkaline tuffs, but seem to have had little influence on caldera character.
TABLE 7. CHARACTERISTICS OF THE McDERMITT CALDERA AND OTHER SILICIC CALDERAS

|                         | McDermitt Caldera | Buried Yellowstone hotspot calderas | Strongly peralkaline calderas | Ignimbrite flareup, western North America |
|-------------------------|-------------------|------------------------------------|-------------------------------|------------------------------------------|
| Precaldera volcanism    | Major, diverse; Steens Basalt to intermediate to rhyolite lavas | ? | Shield built from abundant low viscosity lavas | Mostlty major (to minor and possibly none) |
| Ash-flow tuff character | Strongly zoned, peralkaline high-Si rhyolite (comendite) to metaluminous trachydacite (celandite) | High-temperature, generally homogeneous, low- or high-Si rhyolites | Commonly strongly zoned from pantellerite to crystal-rich trachyte; high temperature | Generally zoned, calc-alkaline, metaluminous to mildly peraluminous high-Si rhyolite to dacite andesite; low-temperature alkalic systems are rare |
| Phenocrysts             | Anorthoclase, pyroxene, fayalite | Plagioclase, sanidine, augite, pigeonite, rare quartz and hypersthene, ilmenite, magnetite, zircon, apatite | Aikali feldspar, scarce quartz | Quartz, sanidine, plagioclase, biotite, hornblende, clinopyroxene, apatite, zircon |
| Peralkalinity index    | 1.12–0.80 | -0.7–0.9 | >1.1 to 2.00 | -0.7–0.9 |
| Structural collapse area (km²) | 930; larger than all except a few Ignehntrate flareup | Calderas are inferred but not exposed | Mostly ~10–100, rarely to 300 | ~100–600, rarely 1600 (Window Butte, Great Basin), 2500? (La Garita, southern Rocky Mountains) |
| Topographic wall       | No more than 1 km outside structural margin | ? | Mostly very little outside structural margin | Varies from many kilometers outside structural margin to almost none |
| Vertical collapse and/or subsidence | ~1 km | ? | Hundreds of meters, rarely to 1 km | Hundreds of meters to 6 km |
| Tuff volume (km³)      | ~1000 | To at least 1000 | Mostly 2–10, rarely to 100s | ~100 to ~600; maximum >600 = Fish Canyon Tuff (southern Rocky Mountains), 5900 Wah Wah Springs (Great Basin) |
| Megabreccia and mesobreccia | Common | ? | Intracaldera tuff rarely exposed | Abundant, probably universally present |
| Outflow distance       | <20 km | ? | Few kilometers to tens of kilometers, commonly limited by island or rift setting | Tens of kilometers to as much as 250 km where flowed in paleovalleys |
| Infracaldera tuff (%)  | 50–85 | ? | Poorly known because rarely exposed | Variable, 80 to ~50 |
| Rheomorphism           | Common, intense | Common, intense | Common, intense | Rare, and only with most alkalic |
| Postcaldera igneous activity | Volumetrically minor icelandite as central vents and lavas, few if any ring-fracture bodies | Common, voluminous trachyte lavas from central vent | Major to none, ring-fracture intrusions and lava domes very common | Major to none, ring-fracture intrusions and lava domes very common |
| Resurgent uplift       | Yes, through center of caldera | ? | Rare and with little structural uplift: Pantelleria and Mount Suswa only | Mostly yes, in some cases strong doming |
| Resurgent Intrusion    | Yes, probably icelandite that also erupted as late vents and lavas | ? | Probably trachyte | Common; some compositionally similar to less evolved part of erupted tuff |
| Ring-fracture Intrusions | Absent or sparse | ? | Variable, common along outer concentric fractures | Abundant to common to minor |
| Part of Caldera Complex | No | ? | Commonly part of nested complex | Both complexes and individual calderas are common |
| Related hydrothermal activity and mineralization | Abundant, diverse (U, Hg, Au, Li) | Minor to none? | Probably none | Highly variable; none, to major hydrothermal systems without deposits, to a few major deposits (Au, Cu) |
| Tectonic setting       | In extensional setting, but negligible total extension until ~4 Ma after caldera activity | Extensional, at least partly during active extension | Commonly associated with active but low-magnitude extension; e.g., East Africa Rift | Neutral tectonic environment, with no measurable extension or any contemporaneous deformation |
| Crustal setting        | Accreted island arc terrane | North American craton | Variable | North American craton, to transitional-riifted craton, to accreted island arc terrane |
| Implications for Magma Chamber | Large area but small subsidence may imply thin (few kilometers) tabular magma chamber | ? | Shallow | Variable area with small to very large subsidence implies generally thick, tabular magma chambers |

Note: From this study, Mahood (1984), Swanson and McDowell (1984), McIntosh et al. (1992), Christiansen (2001), Christiansen (2005), Lipman (2007), Christiansen and McCurry (2008), Lipman and McIntosh (2008), Bonnichsen et al. (2008), John et al. (2008), McDowell and McIntosh (2012), Henry and John (2013), Ellis et al. (2013), Best et al. (2013a, 2013b, 2016), and Lipman and Bachmann (2015).  
?—no data; buried.
The characteristics of the McDermitt caldera have several possible implications for the underlying magma chamber. Given the general interpretation that caldera area approximates magma chamber area (Lipman, 1997, 2007; Christiansen, 2005), the McDermitt magma chamber probably was large and underlay the entire caldera. The small amount of interpreted collapse suggests that the erupted part was only a thin cap to the chamber, overlying a substantial volume of more intermediate composition or noneruptible, crystal-rich mush. Subsequent resurgence implies that (1) sufficient magma remained in the chamber to generate resurgence, or (2) a deeper magma source remained sufficiently active to quickly provide significant magma volumes that were indistinguishable from the least evolved tuff.

McDermitt: An Analog for the Central Snake River Plain?

The new mapping and geochemical and petrological constraints described here make the McDermitt volcanic field one of the best studied of the entire Yellowstone system. With this information we can evaluate to what extent McDermitt represents a potential analog for the central Snake River Plain (CSRP) supereruption-producing centers, which are concealed beneath later basalt (Fig. 1; Table 7; Bonnichsen et al., 2008).

Similarity in erupted volumes suggests that McDermitt is a good analog for the CSRP calderas in terms of caldera style and geometry. The Kimberly drillcore in the CSRP east of McDermitt encountered more than 1.3 km of Castleford Crossing ignimbrite (Fig. 1; Knott et al., 2016), by far the thickest ignimbrite recorded from the province, and consistent with an intracaldera deposit. Other drillholes have mostly encountered outflow deposits. For example, among numerous units encountered in core in the eastern SRP the thickest was a 200-m-thick lava and the thickest ignimbrite was ~30 m (Anders et al., 2014). This core bottomed in rhyolite, allowing the potential for greater thickness beneath, but the recovered drillcore produced 9 different units over a thickness of ~360 m, consistent with an extra-caldera sequence.

The relatively thin occurrence of intracaldera McDermitt Tuff (with the associated caveat of only having reliable observations from the western caldera margin) at first appears similar to the intracaldera thickness of the Yellowstone ignimbrites. While the 400 m maximum thickness contour of Member A, Lava Creek Tuff, could perhaps be considered similar to the 600 m of measured McDermitt Tuff, we note that the isopachs of deposit thickness illustrated in Christiansen (2001) for the major Yellowstone ignimbrites are still interpolations, because much of the current exposure is composed of postcaldera lavas (e.g., Stelten et al., 2015), and locations where both the base and top of an ignimbrite sheet can be seen are scarce, even with the density of drillcores at Yellowstone.

Petrologically, the CSRP rhyolites are significantly unlike the McDermitt volcanic system, which shows more similarities with the other early hotspot centers (High Rock and Virgin Valley; Castor and Henry, 2000; Haubensack et al., 2012; Coble and Mahood, 2016). The McDermitt system produced voluminous intermediate compositions (e.g., icelandites, andesites), which are also abundant in the Santa Rosa–Calico field (Brueseke and Hart, 2008, 2009) but are not reported from the younger volcanism of the SRP. Furthermore, the variety in compositions of rhyolites in terms of peraluminosity and mineral assemblage (the common occurrence of biotite-bearing rhyolite) is at odds with the rhyolites of the Snake River Plain–Yellowstone system, where hydrous phases are rarely found (Christiansen, 2001; Drew et al., 2013; Troch et al., 2017). At McDermitt the intermediate compositions are effusive, and so such compositions could have erupted in the younger systems but remain obscured within the poorly exposed calderas. However, two main lines of evidence argue against this: (1) no such intermediates are observed at Yellowstone where intracaldera lavas are abundant (Christiansen, 2001; Stelten et al., 2015); and (2) samples from drillcores in the CSRP exhibit the same basalt-rhyolite compositional bi-modality as the surficial record (Anders et al., 2014; Knott et al., 2016). These petrologic differences may reflect different crustal environments, i.e., accreted terrane for the McDermitt caldera, and craton for the CSRP (Fig. 1).

The McDermitt Tuff shows some similarities but also some important differences in comparison with most CSRP ignimbrites. The anhydrous mineralogy and relatively high inferred magmatic temperature of the McDermitt Tuff are somewhat similar to those of CSRP ignimbrites, which show a general decrease in magmatic temperatures toward Yellowstone (Nash et al., 2006; Brannen et al., 2008). The moderate to dense welding of the McDermitt Tuff is more similar to the ignimbrites from the Yellowstone volcanic field than to the lava-like ignimbrites of the CSRP. The McDermitt Tuff also contains abundant pumice (typically expressed as fluff) and abundant lithic clasts of a wide variety of rock types (Figs. 11B and 12B), both features that are atypical of the CSRP ignimbrites. While the CSRP ignimbrites do contain lithic clasts, typically these occur as small, centimeter-scale clasts at low proportions (<5%–10%) and are almost uniquely composed of antecedent CSRP rhyolites with glassy lithologies seemingly over-represented with respect to surface lithologies (Ellis and Brannen, 2010). The relative paucity of clasts in the CSRP ignimbrites has been attributed to the inferred distance to a proposed caldera; however, the general scarcity of pumice (in both ignimbrite and fallout deposits) cannot readily be explained in such a way. In terms of deposit architecture, the large compositional zonation exhibited by the McDermitt Tuff is unknown in the CSRP, where deposits show extremely limited heterogeneity in bulk compositions (Ellis et al., 2013; Wolff et al., 2015).

Brannen et al. (2008) suggested that the extent to which the unusual characteristics of Snake River Plain–type volcanism are displayed is at a maximum around the Bruneau-Jarbidge region and decreases to the east and west. Our detailed study of the McDermitt volcanic system confirms this idea, illustrating both similarities and differences with the rhyolites of the Snake River Plain across a range of petrology, geochemistry, and physical appearance.

CONCLUSIONS

The McDermitt caldera formed ca. 16.39 Ma in an area that had undergone abundant prior igneous activity, the most important being major middle Miocene volcanism that led continuously to caldera formation, as well as two pre-
ceding episodes of Eocene intermediate volcanism at 47 and 39 Ma. Diverse middle Miocene activity began with eruption of voluminous basalt lavas of the Steens Basalt, probably between ca. 16.85 and 16.74 Ma (Camp et al., 2013; Mahood and Benson, 2016, 2017). The oldest rhyolites erupted by 16.69 Ma, and subsequent activity became progressively more dominated by rhyolite. Pre-caldera rhyolites were particularly diverse, ranging from voluminous peralkaline (aphric to sparsely anorthoclase phryic to moderately quartz-sani- dine ± sodic amphibole phryic) to common but less voluminous metalumi- nous (moderately quartz-sanidine-plagioclase-biotite phryic). Although not the oldest caldera of the Yellowstone hotspot, the McDermitt area has the oldest documented rhyolitic activity.

The McDermitt caldera formed during eruption of magma zoned from aphric, high-Si, peralkaline rhyolite to abundantly anorthoclase phryic, high-Si trachydacite and icelandite with as little as 64% SiO₂. The resultant caldera is ~40 × 32–22 km in area, and total collapse was no more than ~1 km. Total erupted volume was ~1000 km³, 50%–85% of which accumulated within the caldera. Continuity of intracaldera tuff, the caldera ring-fracture system, and intracaldera sedimentary deposits demonstrate a single caldera. The cal- dera underwent post-collapse resurgence and several episodes of dominantly icelandic magmatism until ca. 16.1 Ma, when almost all magmatism ceased. Late high-alumina olivine tholeiite lavas erupted ca. 14.9 Ma, but are proba- bly unrelated to the caldera system. Understanding the petrogenesis of the caldera magmatic system as well as the diverse precaldera and postcaldera rocks will require extensive geochemistry, mineralogic characterization, and isotopic analyses.

The McDermitt caldera and other ca. 16 Ma and older rhyolite centers ex- tend over a large area in northern Nevada and southeastern Oregon (Figs. 1 and 2). Their wide distribution is consistent with initiation of the Yellowstone hotspot over an equally large area. Curiously, however, these rhyolite-domi- nated centers only partly overlap with, and partly are south of, the imme- diately preceding to contemporaneous Steens Basalt lavas. Full delineation of the Steens Basalt in the subsurface in northern Nevada and of ca. 16 Ma and older rhyolite centers is needed.

The McDermitt caldera is highly mineralized, with significant deposits of Hg, U, Zr, and Li as well as lesser accumulations of Au, Ga, and Mo. Most, possibly all, mineralization is related to the caldera magmatic and structural system, although considerably more work is needed to fully understand how the deposits relate to that system.

ACKNOWLEDGMENTS

We thank Western Lithium Corporation (now LithiumAmerica; especially geologists Dennis Bryan and Bo Elgyb) and Oregon Energy (John Hasleby, Gavin Doyle, and John Scholdere) for sharing core and drillhole data, John McCormack for discussion about the geology of mercury deposits and for providing sample McD for dating, and Jerry Walker and Dan Larsen for discussions about intracaldera sedimentary deposits. We also thank reviewers Vic Camp, Peter Lipman, and Matt Coble, and themed issue editor Matt Bruske. Our geologic mapping was done sporadically between the late 1970s and 2016 with a wide range of funding, some long forgotten. The U.S. Department of Energy Yucca Mountain Project Office (Las Vegas, Nevada; contract 1950125–03) provided funding in 1995. Starkel received funding from the Emap component of the U.S. Geologi- cal Survey National Cooperative Geologic Mapping Program, the Hugh E. Kinstry Fund of the Society of Economic Geologists Foundation, Inc., the American Philosophical Society, and Washington State University (Pullman).
rite in Gulf Coast sandstones (Texas, U.S.A.): American Mineralogist, v. 98, p. 914–926, doi: 10.2138/am.2013.4238.

Branney, M.J., Bonnichsen, B., Andrews, G.D.M., Ellis, B., Barry, T.L., and McCurry, M., 2008, ‘Snake River (SR)-type’ volcanism at the Yellowstone hotspot track: Distinctive products from unusual, high-temperature silicic super-eruptions: Bulletin of Volcanology, v. 70, p. 293–314, doi:10.1007/s00445-007-0140-7.

Brown, K., and Hart, W.K., 2013, Late Cretaceous arc flare-up in northwestern Nevada: Elemental, zircon Hf Isotope, and U-Pb zircon geochronology of the Santa Rosa Range and Blood Run Hills granitoids: Geological Society of America Abstracts with Programs, v. 45, no. 6, p. 12.

Bruesee, M.E., and Hart, W.K., 2008, Geology and Petrology of the Mid-Miocene Santa Rosa–Calico Volcanic Field, Northern Nevada: Nevada Bureau of Mines and Geology Bulletin 113 46 p.

Bruesee, M.E., and Hart, W.K., 2009, Intermediate composition magma production in an intracontinental setting: Unusual andesites and dacites of the mid-Miocene Santa Rosa–Calico volcanic field, northern Nevada: Journal of Volcanology and Geothermal Research, v. 188, p. 197–213, doi:10.1016/j.jvolgeores.2008.12.015.

Bruesee, M.E., Heizer, M.T., Hart, W.K., and Mertzman, S.A., 2007 Distribution and geochronology of Oregon Plateau (U.S.A.) flood basalt volcanism: The Steens Basalt revisited: Journal of Volcanology and Geothermal Research, v. 161, p. 167–214, doi:10.1016/j.jvolgeores.2006.12.004.

Bruesee, M.E., Hart, W.K., and Heizer, M.T., 2008, Diverse mid-Miocene silicic volcanic association with the Yellowstone-Newberry thermal anomaly: Bulletin of Volcanology, v. 70, p. 343–360, doi:10.1007/s00445-007-0142-5.

Bruesee, M.E., Calcoo, J.S., Hames, W., and Larson, P.B., 2014, Mid-Miocene rhyolite volcanism in northeastern Nevada: The Jarbidge Rhyolite and its relationship to the Cenozoic evolution of the northern Great Basin (USA): Geological Society of America Bulletin, v. 126, p. 1047–1067, doi:10.1130/B30738.1.

Camp, V.E., and Ross, M.E., 2004, Mantle dynamics and genesis of mafic magmatism in the intermontane Pacific Northwest: Journal of Geophysical Research, v. 109, B08204, doi:10.1029/2003JB002838.

Camp, V.E., Ross, M.E., Duncan, R.A., Jarboe, N.A., Coe, R.S., Hanan, B.B., and Johnson, J.A., 2013, The Steens Basalt: Earliest lavas of the Columbia River Basalt Group, in Reidel, S.P., et al., eds., The Columbia River Flood Basalt Province: Geological Society of America Special Paper 497, p. 87–116, doi:10.1130/2013.2497(04).

Carmichael, I.S.E., 1964, The petrology of Thingmuli, a Tertiary volcano in eastern Iceland: Journal of Petrology, v. 5, p. 435–460, doi:10.1016/0022-0396(64)90021-9.

Castor, S.B., and Henry, C.D., 2000, Geology, geochemistry, and origin of volcanic rock-hosted uranium deposits in northwestern Nevada and southeastern Oregon: Ore Geology Reviews, v. 16, p. 1–40, doi:10.1006/isop.1998.0001-29.

Castor, S.B., Mitchell, T.P., and Guade, J.G., 1982, National Uranium Resource Evaluation, Vya Quadrangle, Nevada, Oregon, and California: U.S. Department of Energy Open-File Report PGJ/F136(82), 26 p.

Castor, S.B., Boden, D.R., Henry, C.D., Cline, J.S., Hofstra, A.H., McIntosh, W.C., Tosdal, R.M., and Woden, J.P., 2003, Geology of the Eocene Tescarca volcanic-hosted, epithermal precious metal district, Elko County, Nevada: Economic Geology and the Bulletin of the Society of Economic Geologists, v. 98, p. 339–356.

Childs, J.F., 2007, Cordero gold-silver project technical report for Tech Industries LLC: http://www.silverpredator.com/documents/Cordero-43-101-09-04-07-Final.pdf.

Christiansen, E.H., 2005, Contrasting processes in silicic magma chambers: Evidence from very large and small volume ignimbrites: Geological Magazine, v. 142, p. 669–681, doi:10.1017/S0016686005001445.

Christiansen, E.H., and McCurry, M., 2008, Contrasting origins of Cenozoic silicic volcanic rocks from the western Cordilleran of the United States: Bulletin of Volcanology, v. 70, p. 251–267, doi:10.1007/s00445-008-0485-6.

Christiansen, R.L., 2001, The Quaternary and Pliocene Yellowstone Plateau Volcanic Field of Wyoming, Idaho, and Montana: U.S. Geological Survey Professional Paper 729-G, 145 p.

Christiansen, R.L., Foulger, G.R., and Evans, J.R., 2002, Upper mantle origin of the Yellowstone hotspot: Geophysical Society of America Bulletin, v. 114, p. 1245–1256, doi:10.1130/0016-7606(2002)114<1245:UMOHOT>2.0.CO;2.

Coble, M.A., and Mahood, G.A., 2012, Initial impingement of the Yellowstone plume located by widespread silicic volcanism contemporaneous with Columbia River flood basalts: Geology, v. 40, p. 655–658, doi:10.1130/G32862.1.

Coble, M.A., and Mahood, G.A., 2016, Geology of the High Rock caldera complex, northwest Nevada, and implications for intense riftogenic volcanism associated with flood basalt magmatism and the initiation of the Snake River Plain–Yellowstone trend: Geosphere, v. 12, p. 58–113, doi:10.1130/GES01162.1.

Colgan, J.P., and Henry, C.D., 2009, Rapid middle Miocene collapse of the Sevier orogenic plateau in north-central Nevada: International Geology Review, v. 51, p. 920–961, doi:10.1177/0020662909333672.

Colgan, J.P., Dumitrutu, T., and Miller, E.L., 2004, Diachroneity of Basin and Range extension and Yellowstone hotspot volcanism in northwestern Nevada: Geology, v. 32, p. 121–124, doi:.10.1130/G20037.1.

Colgan, J.P., Dumitrutu, T., McWilliams, M., and Miller, E.L., 2006a, Timing of Cenozoic volcanism and Basin and Range extension in northwestern Nevada: New constraints from the northern Pine Forest Range: Geological Society of America Bulletin, v. 118, p. 126–139, doi:10.1130/1051-0784(2006)118[126:TOCVAI]2.0.CO;2.

Colgan, J.P., Dumitrutu, T.A., Reiners, P.W., Wood, J.L., and Miller, E.L., 2006b, Cenozoic tectonic evolution of the Basin and Range Province in northwestern Nevada: American Journal of Science, v. 306, p. 616–654, doi:10.2475/08.2006.02.

Colgan, J.P., Egger, A.E., John, D.A., Cousins, B., Fleck, R.J., and Henry, C.D., 2011, Olivine and Cenozoic arc volcanism in northern California: Evidence for segmentation of the subducting Farallon plate: Geosphere, v. 7, p. 733–755, doi:10.1130/GES00690.1.

Compton, R.R., 1960, Contact metamorphism in Santa Rosa Range, Nevada: Geological Society of America Bulletin, v. 71, p. 1383–1416, doi:10.1130/0016-7606(1960)71<1383:CMSSR>2.0.CO;2.

Conrad, W.K. 1984, The mineralogy and petrology of compositionally zoned ash flow tuffs, and related silicic volcanic rocks, from the McDermitt Caldera Complex, Nevada-Oregon: Journal of Geophysical Research, v. 89, p. 8639–8664, doi:10.1029/JB089iB10p08639.

Cummings, M.L., Evans, J.G., Ferns, M.L., and Lees, K.R., 2000, Stratigraphic and structural evolution of the middle Miocene synvolcanic Oregon-Idaho graben: Geological Society of America Bulletin, v. 112, p. 668–682, doi:10.1130/0016-7606(2000)112<668:SASSTC>2.0.CO;2.

Cobb, M.A., and Mahood, G.A., 2016, Geology of the High Rock caldera complex, northwest Nevada, and implications for intense riftogenic volcanism associated with flood basalt magmatism and the initiation of the Snake River Plain–Yellowstone trend: Geosphere, v. 12, p. 58–113, doi:10.1130/GES01162.1.
Glanzman, R.K., and Ryutva, J.J., 1979, Zeolite-Clay Mineral Zonation of Volcaniclastic Sedi-
ments within the McDermitt Caldera Complex of Nevada and Oregon: U.S. Geological Survey
Open-File Report 79-1888, 52 p.

Glanzman, R.K., Ryutva, J.J., and McCarthy, J.H., 1978, Lithium in the McDermitt caldera, Nevada
and Oregon: Energy, v. 3, p. 347–353, doi:10.1036/0380-5442(78)90031-2.

Glen, J.M.G., and Ponce, D.A., 2002, Large-scale fractures related to inception of the Yellowstone
hotspot: Geology, v. 30, p. 647–650, doi:10.1029/091761-20020306670789102.0.C02.

Greene, R.C., 1972, Preliminary geologic map of Jordan Mountain quadrangle, Nevada-Oregon:
U.S. Geological Survey Miscellaneous Field Studies Map MF-341, scale 1:48,000.

Greene, R.C., 1976, Volcanic Rocks of the McDermitt Caldera, Nevada-Oregon: U.S. Geological
Survey Open-File Report 76-753, 80 p.

Hargrove, H.R., and Sheridan, M.F., 1984, Welded tuffs deformed into megahemorphic folds
during the collapse of the McDermitt caldera, Nevada-Oregon: Journal of Geophysical Re-
search, v. 89, p. 8629–8638, doi:10.10J08JB0108629.

Hart, W.K., Aronson, J.L., and Mertzman, S.A., 1984, Areal distribution and age of low-K, high-
alumina olivine tholeiite magmas in the northwestern Great Basin: Geological Society of
America Bulletin, v. 95, p. 186–195, doi:10.1007/s004450050186-ADAAO2.0.C02.

Haasback, B., Smith, J., Henry, C.D., Hilton, R.P., McIntosh, W.C., Heizler, M.T., and Noble, D.C.,
2012, The High Rock caldera complex, NW Nevada: Geologic mapping, volcanology, geo-
chemistry, and ultra-high precision 4Ar/3Ar dating of early Yellowstone hotspot magma-
tism: American Geophysical Union fall meeting, abs. V3B–2856.

Heintz, M.L., Yancey, T., Miller, B.V., and Heizer, M.T., 2014, Tephrochronology and geochemistry
of Eocene and Oligocene volcanic ashes of east and central Texas: Geological Society of
America Bulletin, v. 127, p. 779–780, doi:10.1130/B31145.1.

Heizer, M.T., 2011, Introducing the ARGUS VI mass spectrometer to geo and thermochronology:American Geophysical Union fall meeting, abs. V51A–2508.

Heizer, M.T., McIntosh, W.C., Ross, J., and Hamilton, D., 2014, 10±3 Ohm Faraday multi-col-
lection: Striving for accuracy to 25 match ultrachap persistent 4Ar/3Ar measurements: Gold-
schmidt Abstracts 2014, p. 952.

Henry, C.D., and John, D.A., 2013, Magmatism, ash-flow tuffs, and calderas of the ignimbrite
Henry, C.D., and Wolff, J.A., 1992, Distinguishing strongly rheomorphic tuffs from extensive
Henry, C.D., Castor, S.B., Starkel, W.A., Ellis, B.S., Wolff, J.A., Heizler, M.T., and McIntosh, W.C.,
2014, 10e13 Ohm Faraday multi-col-
Heizler, M.T., McIntosh, W.C., Ross, J., and Hamilton, D. 2014, 10e13 Ohm Faraday multi-col-
anthology of the Oregon High Lava Plains: Implications for the plume interpretation of
Yellowstone: Journal of Geophysical Research, v. 109, B10202, doi:10.1029/2009JB007276.

Knott, T.R., Branney, M.J., Reichow, M.K., Finn, D.R., Coe, S.S., Storey, M., Barford, D., and Mc-
Curry, M., 2016, Mid-Miocene record of large-scale Snake River-type explosive volcanism
and associated subsidence on the Yellowstone hotspot track: The Cassia Formation of Idaho,
USA: Geological Society of America Bulletin, v. 128, p. 1121–1146, doi:10.1130/11301244.1.

Kuiper, K.F., Deino, A., Hilgen, F.J., Krijgsman, W., Renne, P.R., and Wijbrans, J.R., 2008, Synchro-
nization of Eocene lavas using oxygen isotopes of foraminiferal 
shells: Journal of Geophysical Research, v. 113, B10201, doi:10.1029/2007JB004539.

Langella, A., Cappelletti, P., and de Gennaro, R., 2001, Zeolites in closed hydrologic systems:
Reviews in Mineralogy and Geochemistry, v. 45, p. 235–260, doi:10.2138/rmg.2001.45.7.

Larsen, D., and Crossley, J.L., 1996, Depositional environments and paleoceanography of an ancient
volcanic ash layer: Oligocene Cretaceous, Colorado: Geological Society of America Bul-
letin, v. 108, p. 526–544, doi:10.1130/0016-7606(1996)108<0526:DEPAOA>2.3.CO;2.

Larsen, D., and Lipman, P., 2016, Exploring the ancient volcanic and lacustrine environments of the
Oligocene Creede caldera and environs, San Juan Mountains, Colorado, in Keller, S.M., and
Morgan, M.L., eds., Unfolding the Geology of the West: Geological Society of America
Field Guide 44, p. 1–40, doi:10.1130/2016.0044.01.

Lerch, D., Colgan, J.P., Miller, E.L., and McWilliams, M., 2008, Tectonic and magmatic evolution of
the northwestern Basin and Range transition zone; mapping and geochronology from the Black
Rock Range, NV: Geological Society of America Abstracts with Programs, v. 31, no. 7, p. 70.

Lipman, P.W., 1984, The roots of ash flow calderas in western North America: Windows into
the tops of granitic batholiths: Journal of Geophysical Research, v. 89, p. 8801–8841, doi:10.1029/
JB089iB10p08801.

Lipman, P.W., 1986, Evolution of silicic magma in the upper crust: the mid-Tertiary Latr volcanic
field, central Oregon: Geophysical Society of America Abstracts with Programs, v. 44, no. 3, p. 21.

Lipman, P.W., 2007, Incremental assembly and prolonged consolidation of Cordilleran magma
chambers: Evidence from the Southern Rocky mountain volcanic field: Geosphere, v. 3, p.
42–70, doi:10.1130/GS00061.1.

Lipman, P.W., and Bachmann, O., 2015, Ignimbrites to batholiths: Integrating perspectives from
geochemical, geophysical, and geochronological data: Geosphere, v. 11, p. 705–743, doi:10.1130/
GS01091.1.

Lipman, P.W., and McIntosh, W.C., 2008, Eruptive and noneruptive calderas, northeastern San
Juan Mountains, Colorado: Where did the ignimbrites come from?: Geological Society of
America Abstracts with Programs, v. 40, no. 7, p. 42–43.

Lipman, P.W., Mehnert, H.H., and Naeser, C.W., 1986, Evolution of the Latr volcanic field, north-
ern New Mexico, and its relation to the Rio Grande Rift, as indicated by potassium-argon
and fission track dating: Journal of Geophysical Research, v. 91, p. 6329–6345, doi:10.1029/
JB091iB06p06329.
Noble, D.C., McCormack, J.K., McKee, E.H., Silberman, M.L., and Wallace, A.B., 1988, Time of
Noble, D.C., McKee, E.H., Smith, J.G., and Korringa, M.K., 1970, Stratigraphy and geochronology
Mahood, G.A., and Benson, T.R., 2002, The Juniper Mountain volcanic center, Owyhee County, south-western Idaho: Age relations and physical volcanology, in Bonnichsen B., et al., Tectonic and Magmatic Evolution of the Snake River Plain Volcanic Province: Idaho Geological
Mcintosh, W.C., Heizer, M., Peters, L., and Esser, R., 2003, 40Ar/39Ar geochronology at the New
McDowell, F.W., and McIntosh, W.C., 2012, Timing of intense magmatic episodes in the northern
Min, K., Mundil, R., Renne, P.R., and Ludwig, K.R., 2000, A test for systematic errors in 40Ar/39Ar
Minor, S.A., Vander Meulen, D.B., Rytuba, J.J., and Vercoutere, T.L., 1989b, Geologic map of The
Minor, S.A., Wager, M., and Harwood, C.S., 1989a, Geologic Map of the Trident Peak SW Quad-
Manley, C.R., and McIntosh, W.C., 2002, The Juniper Mountain volcanic center, Owyhee County, south-western Idaho: Age relations and physical volcanology, in Bonnichsen B., et al., Tectonic and Magmatic Evolution of the Snake River Plain Volcanic Province: Idaho Geological
McCormack, J.K., 1986, Paragenesis and origin of sediment-hosted Mercury ore at the McDer-
mament, B.P., Perkins, M.E., Christensen, J.N., Lee, D.-C., and Halliday, A.N., 2006, The Yellowstone
McIntosh, W.C., Chapin, C.E., Ratté, J.C., and Sutter, J.F., 1992, Time-stratigraphic framework for the Eocene-Oligocene Mogollon-Datil volcanic field, southwest New Mexico: Geologi
cal Society of America Bulletin, v. 104, p. 851–871, doi: 10.1130/0016-7606(1992)104<0851:TSFTE>2.3.CO;2.
McIntosh, W.C., Heizer, M., Peters, L., and Esser, R., 2003, 4Ar/39Ar geochronology at the New
Mexico Bureau of Geology and Mineral Resources: New Mexico Bureau of Geology and
Minor, S.A., Wager, M., and Harwood, C.S., 1989a, Geologic Map of the Trident Peak SW Quadrangle, Humboldt County, Nevada: U.S. Geological Survey Open-File Report 89-561, scale 1:24,000.
Minor, S.A., Wager, M., and Harwood, C.S., 1989a, Geologic Map of the Trident Peak SW Quadrangle, Humboldt County, Nevada: U.S. Geological Survey Open-File Report 89-561, scale 1:24,000.
Minor, S.A., Turner, R.L., Plouff, D., and Leszczynski, A.M., 1988, Mineral resources of the Disaster Peak wilderness study area, Harney and Malheur Counties, Oregon, and Humboldt County, Nevada: U.S. Geological Survey Bulletin 1742, Chapter A., 13 p.
Minor, S.A., Wager, M., and Harwood, C.S., 1989a, Geologic Map of the Trident Peak SW Quadrangle, Humboldt County, Nevada: U.S. Geological Survey Open-File Report 89-447, scale 1:24,000.
Minor, S.A., Vander Meulen, D.B., Rytuba, J.J., and Vercoutere, T.L., 1989b, Geologic map of The V quadrangle, Harney County, Oregon: U.S. Geological Survey Open-File Report 89-351, scale 1:24,000.
Myers, G., 2005, Technical report of the Aurora uranium project, Malheur County, Oregon, for Quincy Energy Corp., 81 p., www.sedar.com/displaycompanydocuments.do?lang=EN&issuerno=00021197.
Nash, B.P., Perkins, M.E., Christensen, J.N., Lee, D.-C., and Halliday, A.N., 2006, The Yellowstone hotspot in space and time: Nd and Hf isotopes in silicic magmas: Earth and Planetary Science Letters, v. 247, p. 143–156, doi:10.1016/j.epsl.2006.04.030.
Niespolo, E.M., Rutte, D., Deino, A.L., and Renne, P.L., 2016, Intercalibration and age of the Alder Creek sanidine 4Ar/39Ar standard: Quaternary Geochronology, doi:10.1016/j.quageo.2016.09.004.
Noble, D.C., McKee, E.H., Smith, J.G., and Korringa, M.K., 1970, Stratigraphy and geochronology of Miocene volcanic rocks in northwestern Nevada: U.S. Geological Survey Professional Paper 700D, p. D23–D32.
Noble, D.C., McCormack, J.K., McKee, E.H., Silverman, M.L., and Wallace, A.B., 1988, Time of mineralization in the evolution of the McDermitt caldera complex, Nevada-Oregon, and the relation of middle Miocene mineralization in the northern Great Basin to coeval regional basaltic magmatic activity: Economic Geology and the Bulletin of the Society of Economic Geologists, v. 83, p. 859–863, doi:10.2113/gsecongeo.83.4.859.
Ryuba, J.J., John, D.A., Foster, A., Ludington, S.D., and Kotlyar, B., 2003, Hydrothermal enrichment of gallium in zones of advanced argillic alteration—Examples from the Paradise Peak and McDermitt ore deposits: U.S. Geological Survey Bulletin 2209C, 16 p.

Sherwin, J.W., and Hanan, B.B., 2008, Lithospheric topography, tilted plumes, and the track of the Snake River–Yellowstone hot spot: Tectonics, v. 27, TC5004, doi:10.1029/2007TC002181.

Sparks, R.S.J., and Wright, J.V., 1979, Weided air-fall tuffs, in Chapin, C.E., and Elston, W.E., eds., Ash-Flow Tuffs: Geological Society of America Special Paper 180, p. 155–166, doi:10.1130/SP180-p155.

Starkel, W.S., 1993, Mapping, geologic evolution and petrogenesis of the McDermitt caldera center, northern Nevada and southern Oregon, USA (Ph.D. thesis): Pullman, Washington State University, 393 p.

Starkel, W.A., Wolff, J.A., Ellis, B.S., Henry, C.D., and Rowe, M.C., 2012, Petrogenesis of the eruptive products at the mid-Miocene McDermitt caldera center, northern Nevada and southern Oregon: American Geophysical Union fall meeting, abs. V31C-2807.

Streck, M.J., Ferns, M.L., and McIntosh, W.C., 2015, Large, persistent rhyolitic magma reservoirs in the Cascades: Eruption dynamics and magmatic evolution: Journal of Geology, v. 123, p. 279–301, doi:10.1086/682449.

Stillings, L.L., and Morissette, C., 2012, Lithium clays in sediments from closed-basin, evaporative lakes in the southwestern United States: Geological Society of America Abstracts with Programs, v. 44, no. 7, p. 210.

Streeter, M.J., Ferns, M.L., and McIntosh, W.C., 2015, Large, persistent rhyolitic magma reservoirs above Columbia River Basalt flow storage sites: The Dinner Creek Tuff eruptive center, eastern Oregon: Geosphere, v. 11, p. 226–235, doi:10.1130/GEOS1086.1.

Sun, S., and McDonough, W.F., 1989, Chemical and isotopic systematics of oceanic basalts: Implications for mantle composition and processes, in Saunders, A.D., and Norry, M.J., eds., Magmatism in the Ocean Basins: Geological Society of London Special Publication 42, p. 313–345, doi:10.1144/GSL.SP.1989.042.01.19.

Surdam, R.C., 1977, Zeolites in closed hydrologic systems, in Mumpton, E.A., ed., Mineralogy and Geology of Natural Zeolites: Mineralogical Society of America Short Course Notes, v. 4, p. 95–91.

Swanson, E.R., and McDowell, F.W., 1984, Calderas of the Sierra Madre Occidental volcanic field, western Mexico: Journal of Geophysical Research, v. 89, p. 8787–8798, doi:10.1029/JB089iB10p08787.

Taylor, J.R., 1982, An Introduction to Error Analysis: The Study of Uncertainties in Physical Measurements: Mill Valley, California, University Science Books, 279 p.

Trotch, J., Ellis, B.S., Mark, D.F., Bindeman, I.N., Kent, A.J.R., Guillong, M., and Bachmann, O., 2012, Rhyolite generation prior to a Yellowstone supereruption: Insights from the Island Park–Mount Jefferson rhyolite series: Journal of Petrology, doi:10.1093/petrology/egz071.

Turner, C.E., and Fishman, N.S., 1991, Jurassic Lake T’oo’dichi’i: A large alkaline, saline lake, Morison Formation, eastern Colorado Plateau: Geological Society of America Bulletin, v. 103, p. 538–568, doi:10.1130/0016-7606(1991)103<0538:JTODDA-2>3.0.CO;2.

U.S. Geological Survey, 1972, Aeromagnetic map of the Vya and part of the McDermitt 1° by 2° quadrangles: U.S. Geological Survey Open-File Report 72-393, scale 1:250,000.

U.S. Geological Survey, 2016, Quaternary fault and fold database for the United States: http://earthquakes.usgs.gov/regional/quets/faults/.

van der Zinder, I., Sinton, J.M., and Mahoney, J.J., 2010, Late shield-stage silicate magmatism at Wai’aane volcano: Evidence for hydrous crustal melting in Hawaiian volcanoes: Journal of Petrology, v. 51, p. 671–701, doi:10.1093/petrology/egq094.

Vikre, R.G., et al., 2016, Geology and mineral resources of the Sheldon–Hart Mountain National Wildlife Refuge Complex (Oregon and Nevada), the southeastern Oregon and north-central Nevada, and the southern Idaho and northern Nevada (and Utah) sagebrush focal areas: U.S. Geological Survey Scientific Investigations Report 2016-5089-B, 224 p., doi:10.3133/sir20165089B.

Wallace, A.B., and Roper, M.W., 1981, Geology and uranium deposits along the northeastern margin, McDermitt caldera complex, Oregon, in Goodell, P.C., and Waters, A.C., eds., Uranium in Volcanic and Volcanioclastic Rocks: American Association of Petroleum Geologists Studies in Geology 13, p. 73–79, doi:10.1306/33421C7.

Wallace, A.B., Drexler, J.W., Grant, N.K., and Noble, D.C., 1980, Icelandite and aenigmatite-bearing pantellerite from the McDermitt caldera complex, Nevada–Oregon: Geology, v. 8, p. 380–384, doi:10.1130/0091-7613(1980)8<380:IAAPPF>2.CO;2.

Western Lithium Corporation, 2014, Updated NI43-101 technical report, Kings Valley project, Humboldt County, Nevada: http://www.westernlithium.com/lithium-project/reports/.

Wolff, J.A., Ramos, F.C., Hart, G.L., Patterson, J.D., and Brandon, A.D., 2008, Columbia River flood basalts from a centralized crustal magmatic system: Nature Geoscience, v. 1, p. 177–180, doi:10.1038/ngeo124.

Wolff, J.A., Ellis, B.S., Ramos, F.C., Starkel, W.A., Boroughs, S., Olin, P.H., and Bachmann, O., 2015, Remelting of cumulates as a process for producing chemical zoning in silicic tuffs: A comparison of cool, wet and hot, dry rhyolitic magma systems: Lithos, v. 236–237, p. 275–286, doi:10.1016/j.lithos.2015.09.002.

Wyld, S.J., Rogers, J.W., and Copeland, P., 2003, Metamorphic evolution of the Luning–Fence–maker fold-thrust belt, Nevada: Illite crystallinity, metamorphic petrology, and 4Ar/39Ar geochronology: Journal of Geology, v. 111, p. 17–38, doi:10.1086/344663.

Wyppych, A., Hart, W., Scarberry, K., McHugh, K., Pasquale, S.A., and Legge, R.W., 2011, Geologic map of the Hawks Valley–Lone Mountain region, Harney County, Oregon: U.S. Geological Survey Department of Geology and Mineral Industries Open-File Report 0–11–12, 28 p., scale: 1:36,000.

Yates, R.G., 1942, Quicksilver deposits of the Opalite district, Malheur County, Oregon, and Humboldt County, Nevada: U.S. Geological Survey Bulletin 2209C, 16 p.

Zimmerer, M.J., and McIntosh, W.C., 2012, The geochronology of volcanic and plutonic rocks at the Questa caldera: Constraints on the origin of caldera-related silicic magmas: Geological Society of America Bulletin, v. 124, p. 1394–1408, doi:10.1130/B30544.1.

Zoback, M.L., McKee, E.H., Blakely, R.J., and Thompson, G.A., 1994, The northern Nevada rift: Regional tectono-magmatic relations and middle Miocene stress direction: Geological Society of America Bulletin, v. 106, p. 371–382, doi:10.1130/0016-7606(1994)106<0371:TNRRRT-2>3.0.CO;2.