

Magmatism, ash-flow tuffs, and calderas of the ignimbrite flareup in the western Nevada volcanic field, Great Basin, USA

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ABSTRACT

The western Nevada volcanic field is the western third of a belt of calderas through Nevada and western Utah. Twenty-three calderas and their caldera-forming tuffs are reasonably well identified in the western Nevada volcanic field, and the presence of at least another 14 areally extensive, apparently voluminous ash-flow tuffs whose sources are unknown suggests a similar number of undiscovered calderas. Eruption and caldera collapse occurred between at least 34.4 and 23.3 Ma and clustered into five ~0.5–2.7-Ma-long episodes separated by quiescent periods of ~1.4 Ma. One eruption and caldera collapse occurred at 19.5 Ma. Intermediate to silicic lavas or shallow intrusions commonly preceded caldera-forming eruptions by 1–6 Ma in any specific area. Caldera-related as well as other magmatism migrated from northeast Nevada to the southwest through time, probably resulting from rollback of the formerly shallow-dipping Farallon slab. Calderas are restricted to the area northeast of what was to become the Walker Lane, although intermediate and effusive magmatism continued to migrate to the southwest across the future Walker Lane.

Most ash-flow tuffs in the western Nevada volcanic field are rhyolites, with approximately equal numbers of sparsely porphyritic ($\leq 15\%$ phenocrysts) and abundantly porphyritic (~20–50% phenocrysts) tuffs. Both sparsely and abundantly porphyritic rhyolites commonly show compositional or petrographic evidence of zoning to trachydacites or dacites. At least four tuffs have volumes greater than 1000 km³, with one possibly as much as ~3000 km³. However, the volumes of most tuffs are difficult to estimate, because many tuffs primarily filled their source calderas and/or flowed and were deposited in paleovalleys, and thus are irregularly distributed.

Channelization and westward flow of most tuffs in paleovalleys allowed them to travel great distances, many as much as ~250 km (original distance) to what is now the western foothills of the Sierra Nevada, which was not a barrier to westward flow of ash flows at that time. At least three tuffs flowed eastward across a north-south paleodivide through central Nevada. That tuffs could flow significant distances apparently uphill raises questions about the absolute elevation of the region and the elevation, relief, and location of the paleodivide.

Calderas are equant to slightly elongate, at least 12 km in diameter, and as much as 35 km in longest dimension. Exceptional exposure of two caldera complexes that resulted from extensional faulting and tilting show that calderas subsided as much as 5 km as large piston-like blocks; caldera walls were vertical to steeply inward dipping to depths ≥ 4 –5 km, and topographic walls formed by slumping of wall rock into the caldera were only slightly outboard (≤ 1 km) of structural margins.

Most calderas show abundant post-collapse magmatism expressed as resurgent intrusions, ring-fracture intrusions, or intracaldera lavas that are closely related temporally (~0–0.5 Ma younger) to caldera formation. Granitoid intrusions, which were emplaced at paleodepths ranging from < 1 to ~7 km, are compositionally similar to both intracaldera ash-flow tuffs and post-caldera lavas. Therefore in the western Nevada volcanic field, erupted caldera-forming tuffs commonly were the upper parts of large magma chambers that retained considerable volumes of magma after tuff eruption.

Several calderas in the western Nevada volcanic field hosted large hydrothermal systems and underwent extensive hydrothermal alteration. Different types of hydrothermal systems (neutral-pH alkali-chloride and acid or low-pH magmatic-hydrothermal) may

reflect proximity to (depth of) large resurgent intrusions. With the exception of the giant Round Mountain epithermal gold deposit, few known caldera-related hydrothermal systems are strongly mineralized. Major middle Cenozoic precious and base metal mineral deposits in and along the margins of the western Nevada volcanic field are mostly related to intrusive rocks that preceded caldera-forming eruptions.

INTRODUCTION

Huge volumes of dominantly rhyolitic ash-flow tuff that erupted from numerous calderas in the Great Basin, mostly during the middle Cenozoic, constitute the ignimbrite flareup (Fig. 1; defined by Coney, 1978, and recognized and documented by McKee, 1971; Lipman et al., 1972; Noble, 1972; Best et al., 1989; Best and Christiansen, 1991). The ignimbrite flareup in the Great Basin was part of a larger rhyolite ignimbrite province that includes calderas from Idaho through the Sierra Madre Occidental of western Mexico (Cather et al., 2009; Henry et al., 2010, 2012b [Challis volcanic field: Fisher et al., 1992; Janecke et al., 1997] [Southern Rocky Mountain volcanic field: Lipman et al., 1970; McIntosh and Chapin, 2004; Lipman and McIntosh, 2008] [Mogollon-Datil volcanic field: McIntosh et al., 1992; Chapin et al., 2004] [Trans-Pecos Texas: Henry and McDowell, 1986; Henry et al., 1994] [Sierra Madre Occidental: McDowell and Clabaugh, 1979; Ferrari et al., 2007; McDowell, 2007; McDowell and McIntosh, 2012]). This overall ignimbrite province was the largest Cenozoic magmatic event in western North America (Henry et al., 2010, 2012b). The most intense period of ash-flow eruptions in the Great Basin was between ca. 37 and 22 Ma, but voluminous ash-flow eruptions both preceded and followed, from ca. 45 to 8 Ma (Sawyer et al., 1994; Henry et al., 2010).

Papers in this themed issue separate the middle Cenozoic calderas of the Great Basin into the

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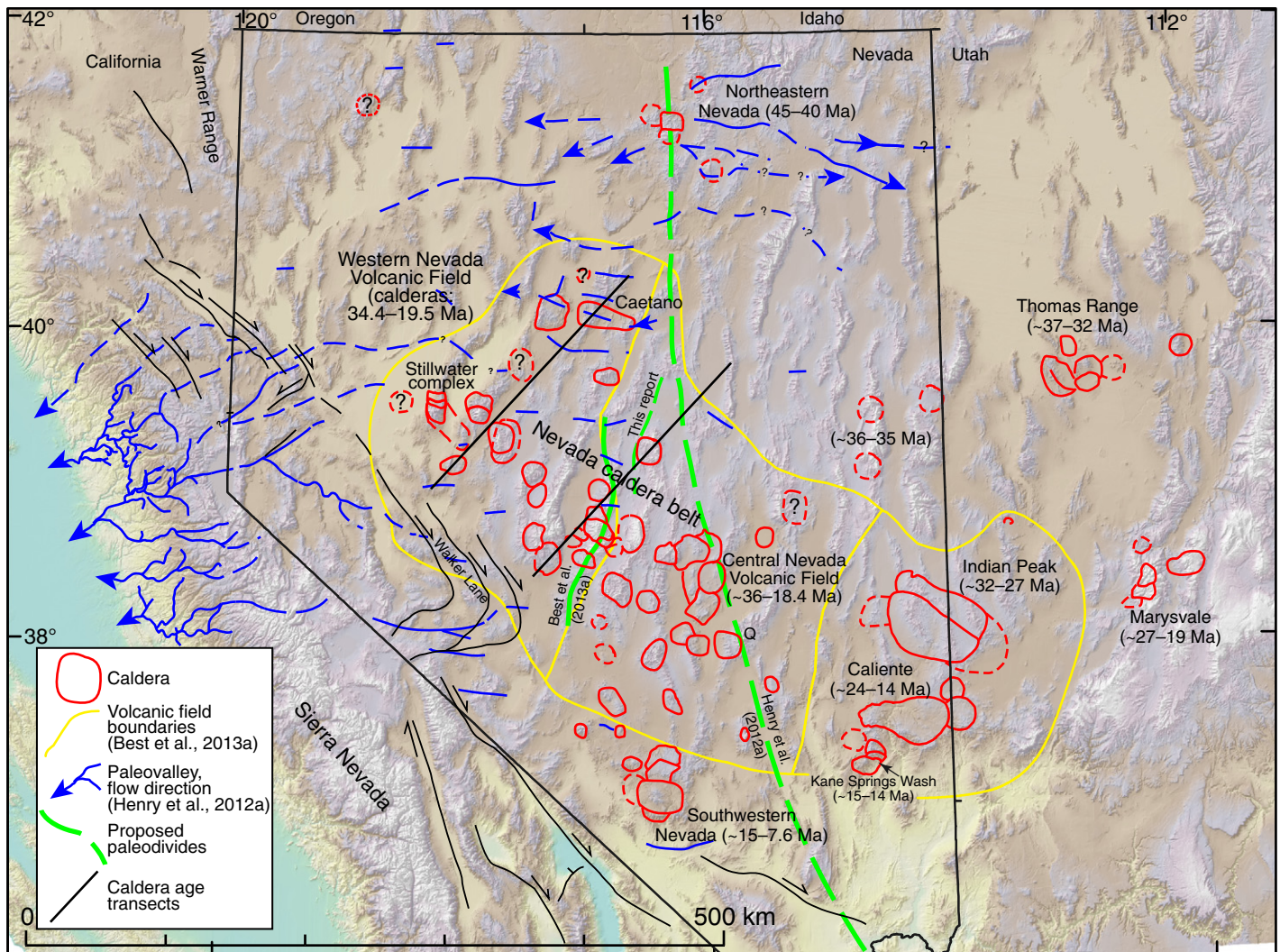


Figure 1. Index map of the Great Basin showing middle Cenozoic calderas, caldera groups, and their age ranges. This report focuses on the western Nevada volcanic field. Also shown are interpreted middle Cenozoic paleovalleys and faults of the Walker Lane. Not shown are middle Miocene calderas of the initial Yellowstone hotspot in northwestern Nevada and the Long Valley and Little Walker calderas of the Sierra Nevada. Northeast-striking lines are caldera age transects (Fig. 4). The Kane Springs Wash calderas are not part of the Caliente caldera complex. Q—Quinn Canyon Range.

Indian Peak–Caliente, central Nevada, and western Nevada volcanic fields (Fig. 1; Best et al., 2013a). Division between the Indian Peak–Caliente and central Nevada volcanic fields is based on the absence of calderas between the two fields. Division between the central Nevada and western Nevada volcanic fields is based on the apparent restriction of most ash-flow tuffs erupted from the western field to areas west of a proposed paleodivide (Henry, 2008; Best et al., 2009; Henry and Faulds, 2010; Henry et al., 2012a). The outlines of the three fields shown in Figure 1 are drawn around known and inferred calderas and surrounding areas within 20–50 km underlain by temporally equivalent effusive and intrusive rocks.

Recent geologic mapping, geochronology, study of mineral deposits, and other research have greatly expanded the understanding of calderas, magmatism, and tectonic history of northern and western Nevada (Proffett and Proffett, 1976; Ekren et al., 1980; Ekren and Byers, 1985, 1986a, 1986b, 1986c; Boden, 1986, 1992; Shawe, 1995, 1998, 1999a, 1999b, 2002; Shawe and Byers, 1999; Shawe et al., 1986; Shawe et al., 2000; John, 1992a, 1995, 2001; Faulds et al., 2005a, 2005b; Colgan et al., 2008; John et al., 2008; Henry and Faulds, 2010; Muntean et al., 2011; Henry et al., 2012a). The greatest strides have come in (1) identifying calderas,

(2) establishing a precise $^{40}\text{Ar}/^{39}\text{Ar}$ chronology of ignimbrite activity (including 250 total and

135 previously unpublished dates; Supplemental Tables 1¹ and 2²), (3) regionally correlating ash-flow tuffs, including correlation of many to source calderas and recognizing the wide

¹Supplemental Table 1. $^{40}\text{Ar}/^{39}\text{Ar}$ ages of ash-flow tuffs in the western Nevada volcanic field. If you are viewing the PDF of this paper or reading it offline, please visit <http://dx.doi.org/10.1130/GES00867.S1> or the full-text article on www.gsapubs.org to view Supplemental Table 1.

²Supplemental Table 2. Analytical data and probability plots of $^{40}\text{Ar}/^{39}\text{Ar}$ ages of ash-flow tuffs in the western Nevada volcanic field. If you are viewing the PDF of this paper or reading it offline, please visit <http://dx.doi.org/10.1130/GES00867.S2> or the full-text article on www.gsapubs.org to view Supplemental Table 2.

distribution of many across the region, (4) recognizing that ash-flow tuffs commonly flowed through and were deposited in paleovalleys, (5) comprehensively delineating the geometry of several calderas, and (6) understanding the relationship of major ore deposits to magmatism.

The understanding of geochemistry and petrology of the western Nevada volcanic field is less advanced than for the Indian Peak–Caliente and central Nevada volcanic fields, which have been intensively studied by M.G. Best, E.H. Christiansen, and colleagues (Best et al., 2013b, 2013c). Uncertainty in correlation of many tuffs, the plethora of locally assigned stratigraphic formation names, and a tendency to combine multiple regional and/or local ash-flow tuffs and lavas into composite formations are significant problems.

This paper is the first comprehensive treatment of middle Cenozoic volcanism in western Nevada during the peak of the ignimbrite flareup. It (1) provides a state-of-the-art summary of ignimbrite chronology, composition, caldera sources, and associated lavas and intrusions, (2) documents the character of caldera-forming ignimbrite volcanism developed on the western slope of the Sevier orogenic plateau and commonly on oceanic crust west of Precambrian basement, (3) includes numerous new certain and probable correlations of widely scattered outcrops of tuffs based on stratigraphy, $^{40}\text{Ar}/^{39}\text{Ar}$ ages, phenocryst assemblages, and compositions, (4) reinterprets two proposed calderas to show that they probably do not exist, and (5) uses extraordinary three-dimensional exposures of the Caetano caldera and the Stillwater caldera complex (Fig. 1) to document fundamental characteristics of calderas that are rarely exposed elsewhere. The understanding provided by geologic mapping and other studies of these exceptionally exposed calderas can be applied worldwide and can further constrain models of ash-flow caldera genesis. Finally, this paper complements similar reports that comprehensively document the central Nevada and Indian Peak–Caliente volcanic fields, which developed on a high Sevier plateau overlying thicker, Precambrian crust (Best et al., 2013b, 2013c). The combined three papers are the first comprehensive description of the Great Basin ignimbrite province and allow for detailed comparisons in the character of volcanism across the province (Best et al., 2013a).

CRUSTAL-TECTONIC SETTING AND EVOLUTION OF WESTERN NEVADA

Calderas of the Great Basin developed in a wide range of crustal settings. Isotopic data indicate that the caldera province extends westward from undisturbed Precambrian basement, across

a transitional zone rifted in the Neoproterozoic, into a region composed of accreted Phanerozoic oceanic and island arc terranes (Fig. 2). Most calderas of the western Nevada volcanic field developed in the accreted terranes, but several calderas in the northern and eastern part of the western Nevada volcanic field are in the transitional zone. The $\text{Sr}_i = 0.706$ isopleth, commonly used to estimate the edge of the Precambrian basement (Kistler, 1990; Tosdal et al., 2000), approximately coincides with the paleodivide proposed by Best and colleagues (Best et al., 2009, 2013a) in central Nevada and slightly west of our published interpretation (Henry, 2008; Henry et al. 2012a).

The few isotopic data for the western Nevada volcanic field are consistent with the regional trends. Twenty-eight initial Sr ratios from the Stillwater caldera complex developed in oceanic crust range from 0.7047 to 0.7057, with nearly all values between 0.7050 and 0.7054 (John, 1995; this study). Eleven analyses from ash-flow tuffs and interbedded lava flows in the Paradise Range also overlying oceanic crust range from

0.7049 to 0.7055 (John, 1992a). In contrast, four analyses from the Caetano caldera and underlying andesite lava flows above transitional crust range from 0.7070 to 0.7076 (B. Cousens, 2012, personal commun.).

Calderas of the Great Basin are restricted to areas of present-day thick crust (Fig. 2). Notably, the gap between pre–40 Ma and post–37 Ma caldera regions correlates with an east-west zone of thin crust, as well as with the Uinta-Cortez axis and western end of the Archean-Proterozoic suture (Roberts et al., 1965; Presnell, 1998; Tosdal et al., 2000; Rodriguez and Williams, 2008). However, correlation between present thick crust and middle Cenozoic calderas probably only indicates that the caldera belt resisted late Cenozoic extension (Henry et al., 2012b), so crust there has not been appreciably thinned.

The amount and distribution of pre-, syn-, and post-ignimbrite flareup extension have been intensely studied and debated. A major episode of extension began in the middle Miocene across most of northern Nevada, although the amount of extension varied greatly across the

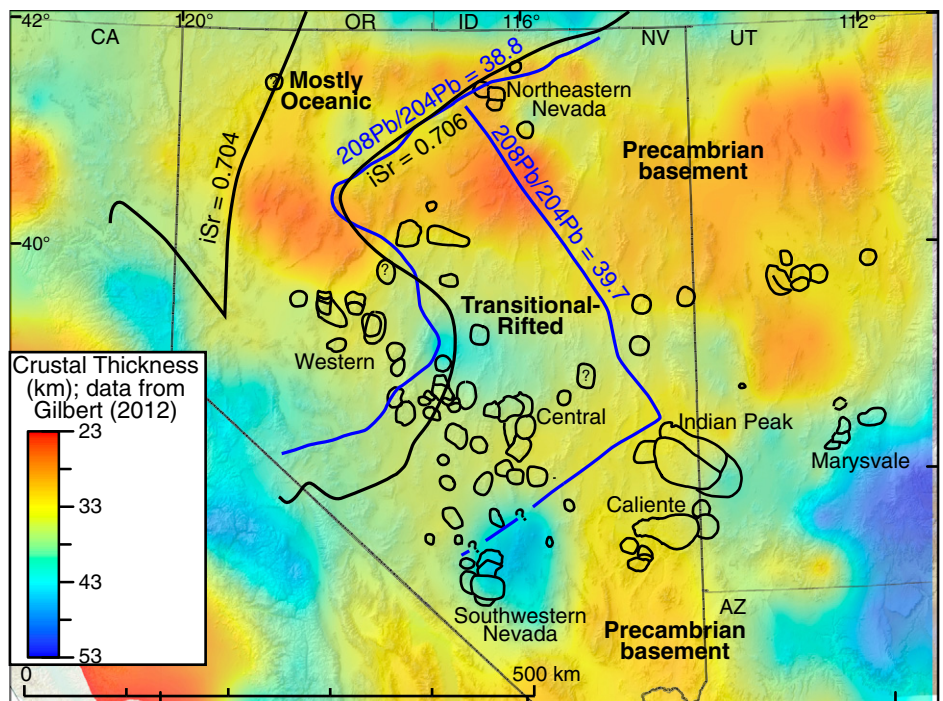


Figure 2. Crustal thickness and crustal isotope provinces of Nevada. Crustal thickness data from Gilbert (2012; contoured in ArcMap [www.esri.com]). Pb lines in blue and Sr lines in black from Tosdal et al. (2000). Calderas of the ignimbrite flareup appear to be restricted to areas of thicker crust (generally ≥ 34 km), possibly indicating that caldera belts resisted mid- to late Cenozoic extension and their crust has therefore not been thinned as much as in adjacent regions. Calderas also extend from intact Precambrian basement on the east, across a transition zone rifted in the Neoproterozoic, into a region composed of accreted Phanerozoic oceanic and island arc terranes on the west. Calderas of the western Nevada volcanic field extend from rifted basement into the accreted terranes.

region (Proffett, 1977; Best and Christiansen, 1991; Surpless et al., 2002; Colgan et al., 2008, 2010; Henry, 2008; Colgan and Henry, 2009; Henry and Faulds, 2010; Henry et al., 2011). Many geologists interpret major pre- and/or syn-ignimbrite flareup extension (Gans et al., 1989; Seedorff, 1991; Smith, 1992; McGrew and Snee, 1994; Dilles and Gans, 1995; Mueller et al., 1999; McGrew et al., 2000; Druschke et al., 2009). However, the lack of angular unconformities between Cenozoic tuffs, continuity of ash-flow tuffs in paleovalleys across areas of major extension, and the paucity of early or middle Cenozoic sedimentary deposits in the caldera belt indicate that pre- or syn-ignimbrite flareup extension was minor and/or exceedingly local (Best and Christiansen, 1991; Henry, 2008; Colgan and Henry, 2009; Long, 2012).

Most calderas in the western Nevada volcanic field are little tilted and therefore probably little extended. Correlation between calderas and thick crust is a manifestation of their lack of extension. However, two calderas or caldera clusters at transitions to thinner crust underwent major extension. The ~100% extension of the 34 Ma Caetano caldera took place mostly between 16 and 10 Ma (John et al., 2008; Colgan et al., 2008). A similar magnitude of extension in the 29–25 Ma Stillwater caldera complex was interpreted to have been mostly at 25–24 Ma (John, 1992b; Hudson et al., 2000). However, new, unpublished apatite fission-track dating by Joseph Colgan and David John indicates that major uplift of the east side of the southern Stillwater Range was ca. 16–12 Ma, so timing of major extension is equivocal.

CALDERA DISTRIBUTION AND TIMING

Major ash-flow tuffs and their known or suspected calderas in the western Nevada volcanic field range from 34 to 19.5 Ma, although only one is younger than 23.3 Ma (Table 1; Fig. 3). Intrusive and effusive magmatism began as early as 40 Ma, and small- to probably moderate-volume tuffs whose eruptions were unlikely to have induced caldera collapse are as old as 35.8 Ma. Major episodes of ash-flow eruptions occurred at 34.4–34.0 Ma (3 calderas), 32.9–32.6 Ma (2), 31.5–28.8 Ma (13), 27.4–26.8 Ma (8), 25.4–23.3 Ma (20), and one event at 19.5 Ma. The western and central Nevada volcanic fields form a continuous belt of calderas from the Stillwater Range on the northwest to the Quinn Canyon Range on the southeast (Fig. 1). Calderas are recognized in every range in this belt. The distribution of several widespread ash-flow tuffs suggests additional calderas especially in the northwestern part of the belt in, and possibly buried beneath

basins adjacent to, the Clan Alpine and Desatoya Mountains. At least four, mostly older calderas lie northeast of the main belt. The Nine Hill caldera is postulated to underlie the Carson Sink (Deino, 1985, 1989; Best et al., 1989; A.L. Deino, 2012, written commun.). Calderas are absent southwest of a distinct line that marks the northeastern margin of the Walker Lane (Fig. 1).

Based on identified calderas and the number of voluminous ash-flow tuffs whose eruption likely induced caldera collapse, at least 37 calderas are probably present in the western Nevada volcanic field (Fig. 1; Tables 1 and 2). Twenty-three calderas are reasonably well identified on the basis of thick (≥ 500 m) intracaldera ash-flow tuff, commonly containing megabreccia, or the presence of caldera walls. Detailed to reasonably thorough reconnaissance mapping exists for many calderas as documented below. Most calderas are ~20 km in diameter. The largest known calderas are those for the tuff of Elevenmile Canyon (~25 × 35–40 km), the tuff of Campbell Creek (~14–20 × 35 km), and the lower tuff of Mount Jefferson (~20 × 28 km).

Another ~14 calderas, dependent on possible correlations, are inferred in the northwestern part of the western Nevada volcanic field based on the widespread distribution of ash-flow tuffs in that part of the field and westward to western Nevada and the eastern Sierra Nevada (Faulds et al., 2005a, 2005b). Prominent examples are the very widespread Nine Hill Tuff (Deino, 1985; Best et al., 1989) and tuffs of Axehandle Canyon, Rattlesnake Canyon, and Sutcliffe (Faulds et al., 2005a, 2005b; Brooks et al., 2008; Henry and Faulds, 2010). Many calderas may be hidden beneath basins. At least one tuff, the tuff of Miller Mountain of the Candelaria Hills, which correlates with the Monotony Tuff, erupted from a caldera in the central Nevada volcanic field (Best et al., 2013c).

Southwesterly Migration of Caldera-Forming Magmatism

Northeast-to-southwest transects across the western and eastern parts of the western Nevada volcanic field illustrate the southwesterly migration of caldera-forming magmatism through time as well as one major anomaly (Fig. 4). Calderas are spread over a greater distance in the west than in the east, but the oldest calderas are at the northeast end of both transects. Along the western transect, these are the well-established ca. 34 Ma Caetano and Hall Creek calderas and the less certainly located, 34.4 Ma caldera for the tuff of Cove Mine. The younger, southwestern end of the transect has

several calderas at ca. 25 Ma and the 19.5 Ma Fairview Peak caldera.

Along the eastern transect, the ca. 35.4 Ma caldera for the Pancake Summit Tuff of the central Nevada volcanic field is at the northeastern end. The 32.9 Ma Northumberland caldera is the next to the southwest. The numerous calderas of the Toquima caldera complex are in the middle. The youngest calderas, the 24.8 Ma Manhattan, 23.3 Ma Toiyabe, and 24.5 Ma Peavine, are at the southwest end.

The area of the transects has undergone variable amounts of middle Miocene and younger extension, as much as 100% across the Caetano caldera and the southern Stillwater Range (John, 1995; Hudson et al., 2000; Colgan et al., 2008) but much less across all other calderas. Also, extension was west-northwest, nearly perpendicular to the northeast-oriented transects, so correction of north-south distances or southward rates for extension are not critical. Caldera magmatism migrated southwestward at ~20 km/Ma between ca. 34 and 29 Ma along the western transect, then slowed to no more than ~4 km/Ma between ca. 29 and 19 Ma, and ended at the edge of the future Walker Lane (Fig. 1). Migration was ~4 km/Ma along the eastern transect the entire time.

Fish Creek Mountains Anomaly

Formation of the 24.9 Ma Fish Creek Mountains caldera is an anomalously young event in an area that otherwise underwent extensive magmatism between ca. 40 and 33 Ma, including the 34 Ma Caetano caldera (John et al., 2008; Cousens et al., 2011). The Fish Creek Mountains caldera fits temporally with the 26–24 Ma calderas of the Stillwater Range and Clan Alpine Mountains, ~100 km to the southwest at the end of the transect (Fig. 4).

Rollback Origin of Cenozoic Magmatism

Southwesterly migration of caldera-forming eruptions is part of a larger pattern of southwesterly migration of all magmatism (Fig. 5). That pattern is the primary evidence for relating magmatism to the removal (rollback?) of the formerly shallowly dipping Farallon slab, a mechanism proposed by many (Coney and Reynolds, 1977; Best and Christiansen, 1991; Christiansen and Yeats, 1992; Humphreys, 1995; Humphreys et al., 2003; Dickinson, 2006; Henry et al., 2009a, 2010; Colgan et al., 2011a). The dip of the subducting Farallon slab shallowed between ca. 90 Ma, at the end of the Cretaceous assembly of the Sierra Nevada batholith, and 50 Ma. Related magmatism migrated as far east as Colorado (Dickinson and Snyder, 1978; Lipman, 1992; Dickinson, 2006).

TABLE 1. CORRELATION, AGE, SOURCE, AND VOLUME OF MAJOR ASH-FLOW TUFF UNITS IN THE WESTERN NEVADA VOLCANIC FIELD

| Tuff (names, west to east) | Age (Ma) | Source caldera (Figs. 10, 11) | Caldera size (km × km ²) | Tuff volume (km ³) | | Pre-caldera volcanism; age (Ma) | Post-caldera volcanism; age (Ma) | Comments | References |
|---|-----------|--|---|--------------------------------|---------|---|---|--|---|
| | | | | Minimum | Maximum | | | | |
| Tuff of Fairview Peak | 19.5 | Fairview Peak | 18 × 11; 160 | 115 | 640 | Sparse dacite to rhyolite domes; 20.3–20.1 | Abundant rhyolite to dacite domes and andesite lavas; 19.3–18.6 | One definite and one probable ash-flow tuff vent | Henry, 1996a, 1996b |
| Santiago Canyon Tuff – tuff of Copper Mountain – tuff of Toiyabe | 23.3 | Toiyabe | 14 × 20; 220 | Large | | Andesite lava flows overlie 24.5 Ma tuff of Peavine Canyon and underlie tuff of Toiyabe near caldera margin | Rhyolite | Sphene phenocrysts; caldera has several exposed ash-flow feeder vents | Brem et al., 1991; Whitebread and John, 1992; John, 1992a |
| Tuffs of Sheep Canyon and lone Canyon | 24 | Unknown | Unknown | Moderate? | | Unknown | Unknown | | John, 1992a; Whitebread et al., 1988 |
| Blue Sphinx Tuff – tuff of Goldyke | 24.3 | Unknown | Unknown | Small? | | Unknown | Unknown | Strongly resorbed quartz phenocrysts | Ekren et al., 1980; John, 1992a |
| Tuff of Peavine Canyon | 24.5 | Peavine | 15 × 18; 215 | ≤Large | | Unknown | Andesite lava flows overlie and partly fill caldera | Anorthoclase phenocrysts | Brem et al., 1991; John, 1992a |
| Tuff of Desatoya Peak | 24.7 | Central Desatoya Mountains | 12 × 17; 160 | 190 | | Nested inside ca. 29 Ma caldera formed by eruption of tuff of Campbell Creek | | | McKee and Conrad, 1987 |
| Round Rock Formation | 24.8 | Manhattan | 12 × 16; 135 | | | | Several rhyolite to andesite lavas and intrusions; ca. 24 (K-Ar) | | Shawe, 1999a |
| Fish Creek Mountains Tuff | 24.9 | Fish Creek Mountains | 12 × 13; 130 | 130 | 260? | | | McKee (1970) estimated volume of 310 km ³ | McKee, 1970; Cousens et al., 2011 |
| Underdown Tuff – tuff of Clipper Gap | 25.0 | Shoshone Mountains | ≥13 × 22; ≥230 | ≤Large | | None | None | | Bonham, 1970 |
| Tuff of Arc Dome | 25.2 | Probably buried in southern Reese River Valley | 12 × 18; 170 | Large? | | Unknown | Unknown | Location inferred from nearby thickness of tuff and geophysical data | Brem et al., 1991; John, 1992a |
| Eureka Canyon Tuff ? Bluestone Mine Tuff ? tuff of Gabbs Valley ? tuffs in the Paradise Range ? tuff of Railroad Ridge | 25.4–25.1 | Cian Alpine Mountains for tuff of Railroad Ridge | 15 × 10; 140 | Large | | | Voluminous rhyolite flow domes intrude margin of Railroad Ridge caldera | | Bingler, 1978a; Proffett and Dilles, 1984; Ekren and Byers, 1985, 1986a, 1986b, 1986c; Ekren et al., 1980; Hardyman et al., 1988; Dilles and Gans, 1995 |
| Tuff of Chimney Spring – tuff of Poco Canyon – New Pass Tuff | 25.2 | Poco Canyon | 15 (north-south); 175 (assuming circular caldera) | 350 | Large | Nested ca. 25 Ma calderas; ca. 29 Ma tuff of Job Canyon, intermediate lava flows, and IXL pluton | Abundant rhyolite and dacite lava domes and dikes; ca. 25 to 20 | Highly extended; original east-west extent unknown | John, 1995; Deino, 1985; Henry et al., 2004, 2009b; Faulds et al., 2005a, 2005b |
| Tuff of Elevenmile Canyon | 25.1 | Elevenmile Canyon | 20 (north-south) × 35–40 (east-west); 350 (assuming 100% extension) | 1050 | ≥1400 | Nested ca. 25 Ma calderas; lower cooling unit of tuff of Poco Canyon erupted shortly prior to Elevenmile Canyon caldera formation | Tuffs of Lee Canyon and Hercules Canyon in north part of caldera (25.1); small-volume ash flows and rhyolite lavas in south part of caldera (undated) | Highly extended; original east-west extent unknown | John, 1995, 1997; this study |
| Nine Hill Tuff – unit D, Bates Mountain Tuff | 25.4 | Buried beneath Carson Sink? | | Large | | Unknown | Unknown | Inferred source based on distribution of tuff; most widespread tuff in Great Basin | Deino, 1989; Best et al., 1989 |

(continued)

TABLE 1. CORRELATION, AGE, SOURCE, AND VOLUME OF MAJOR ASH-FLOW TUFF UNITS IN THE WESTERN NEVADA VOLCANIC FIELD (continued)

| Tuff (names, west to east) | Age (Ma) | Source caldera (Figs. 10, 11) | Caldera size (km ²) | Tuff volume (km ³) | | Pre-caldera volcanism; age (Ma) | Post-caldera volcanism; age (Ma) | Comments | References |
|---|----------|--|---|--------------------------------|---------|--|---|--|--|
| | | | | Minimum | Maximum | | | | |
| Singate Tuff – tuff of Round Mountain | 26.9 | Round Mountain | 3 x 4 remnant | Large | Unknown | Ongoing magmatism in Toiyama complex from 32.6, especially 27.4 to 26.9 | Unknown | | Profiett and Profiett, 1976; Boden, 1986, 1987, 1992, 1994; Shawe, 1995, 2002; Henry et al., 1996; Henry, 1997; Henry and Faulds, 2010 |
| Lenihan Canyon Tuff – upper tuff of Mount Jefferson | 27.0 | Upper Mount Jefferson | 13 x 22; 200 | 300 | Unknown | Ongoing magmatism in Toiyama complex from 32.6, especially 27.4 to 26.9 | Numerous ring-fracture rhyolites ≤1 km diameter; 26.95 ± 0.03, n = 5 | | Boden, 1986, 1987, 1992, 1994; Shawe, 1999b, 2002; Henry et al., 1996; Henry, 1997; Henry and Faulds, 2010 |
| Mickey Pass Tuff (Weed Heights Member) | 27.1 | Unknown, possibly Toiyama Range | Unknown | Large | Unknown | Unknown | Unknown | | Profiett and Profiett, 1976; Ekren et al., 1980 |
| Mickey Pass Tuff (main part of Guild Mine Member) – lower tuff of Mount Jefferson | 27.3 | lower Mount Jefferson | 20 x 28; 400 | 900 | 1600 | Ongoing magmatism in Toiyama complex from 32.6, especially 27.4 to 26.9 | Numerous rhyolite domes ≤1 km diameter along ring fracture; 27.30 ± 0.01, n = 4 | | Boden, 1986, 1987, 1992, 1994; Shawe, 2002; Shawe et al., 2000; Henry et al., 1996; Henry, 1997; Henry and Faulds, 2010 |
| Mickey Pass Tuff (basal part of Guild Mine Member) – tuff of Ryeocroft Canyon | 27.4 | Caldera remnant, Toiyama complex | 3 x 3 remnant | Large | | Ongoing magmatism in Toiyama complex from 32.6, especially 27.4 to 26.9 | Rhyolite domes of uncertain relationship to caldera remnant; 27.51 ± 0.25, 27.49 ± 0.10 | | Boden, 1986, 1987, 1992, 1994; Shawe, 1999b, 2002; Shawe et al., 2000; Henry et al., 1996; Henry, 1997; Henry and Faulds, 2010 |
| Tuff of Logan Spring – upper tuff of Corcoran Canyon | 27.4 | Caldera remnant, part of Ryeocroft caldera? | Overprinted | Large? | | Ongoing magmatism in Toiyama complex from 32.6, especially 27.4 to 26.9 | | | Boden, 1986, 1987, 1992, 1994; Shawe, 2002; Shawe et al., 2000; Henry et al., 1996; Henry, 1997; Henry and Faulds, 2010 |
| Tuff of Miller Mountain – Monotony Tuff; outflow from central Nevada field | 27.6 | Pancake Range caldera in central Nevada volcanic field | | | 4700 | | | | Speed and Cogbill, 1979; Robinson and Stewart, 1984; Petronis et al., 2009; Best et al., 2013c |
| Tuff of Campbell Creek – unit C, Bates Mountain Tuff | 28.9 | Desatoya Mountains | 14–20 x ≥35; 600 | 360 | 3000 | Andesite lavas, age unknown | Dacite to rhyolite intrusions compositionally related to tuff; 28.84 ± 0.05, 28.83 ± 0.04 | Strongly resorbed quartz phenocrysts; second-most widespread tuff in Great Basin | McKee and Conrad, 1987; Brooks et al., 2008; Henry et al., 2012 |
| Tuff of Deep Canyon | 29.0 | Clan Alpine Range | 15 x 20; 240 | 240 | 960 | Hornblende andesite lavas ~6 m.y. older than caldera | Numerous rhyolite flow domes along margin of Deep Canyon caldera | | Riehle et al., 1972; Hardyman et al., 1988 |
| Tuff E | 29.1 | Desatoya Mountains | Probably early eruption from Campbell Creek caldera | Large | | Tuff E probably predates Campbell Creek eruption | Campbell Creek caldera | | McKee and Conrad, 1987; Brooks et al., 2008; Henry et al., 2012 |
| Tuff of Job Canyon | ca. 29.4 | Job Canyon | 10 (north-south); 80 (assuming circular caldera) | 240 | Large | Large rhyolite flow dome complex (ca. 30? Ma); thick sequence of andesite-dacite lava flows and breccias, dikes, and sills (ca. 30–29) | Dacite-andesite lavas and breccias up to 2 km thick fill caldera; ca. 29 | Highly extended; original east-west extent unknown | John, 1995 |

(continued)

TABLE 1. CORRELATION, AGE, SOURCE, AND VOLUME OF MAJOR ASH-FLOW TUFF UNITS IN THE WESTERN NEVADA VOLCANIC FIELD (continued)

| Tuff (names, west to east) | Age (Ma) | Source caldera (Figs. 10, 11) | Caldera size (km, km ²) | Tuff volume (km ³) | | Pre-caldera volcanism; age (Ma) | Post-caldera volcanism; age (Ma) | Comments | References |
|---|-----------|---|-------------------------------------|--------------------------------|---|---------------------------------|--|--|--|
| | | | | Minimum | Maximum | | | | |
| Tuff of Dogskin Mountain – tuff of McCoy Mine | 29.4 | Probably Desatoya Mountains | Overprinted | Large | Unknown | Unknown | Unknown | | Henry and Faulds, 2010 |
| Lower tuff of Corcoran Canyon | 30.0 | Early Toquima complex | 2.5 x 7 remnant | Large? | | | | | Boden, 1986, 1987, 1992, 1994; Shawe, 2002; Henry et al., 1996; Henry and Faulds, 2010 |
| Tuff of Mine Canyon – part of Edwards Creek Tuff | 30.1 | ?Stillwater – Desatoya – Clan Alpine area | Unknown | Large | Unknown | Unknown | Unknown | | Henry et al., 2004, 2009; Faulds et al., 2005a, b |
| Tuff of Cove Spring – part of Edwards Creek Tuff | 30.3 | ?Stillwater – Desatoya – Clan Alpine area | Unknown | Large | Unknown | Unknown | Unknown | | Henry et al., 2004, 2009; Faulds et al., 2005a, 2005b |
| Tuff of Sutcliffe – part of Edwards Creek Tuff – unit B, Bates Mountain Tuff | 30.4–30.6 | ?Stillwater – Desatoya – Clan Alpine area | Unknown | Large | Unknown | Unknown | Unknown | | Henry et al., 2004, 2009; Faulds et al., 2005a, 2005b |
| Tuff of Rattlesnake Canyon – part of Edwards Creek Tuff – unit A, Bates Mountain Tuff | 31.2 | Stillwater – Desatoya – Clan Alpine area? | Unknown | Very large | Unknown | Unknown | Unknown | | Henry et al., 2004, 2009; Faulds et al., 2005a, 2005b |
| Tuff of Hardscrabble Canyon | 31.4 | Stillwater – Desatoya – Clan Alpine area? | Unknown | Moderate | Unknown | Unknown | Unknown | | Henry et al., 2004, 2009; Faulds et al., 2005a, 2005b |
| Tuff of Axehandle Canyon | 31.5 | Stillwater – Desatoya – Clan Alpine area? | Unknown | Very large | Unknown | Unknown | Unknown | | Henry et al., 2004, 2009; Faulds et al., 2005a, 2005b |
| Tuffs of Dry Canyon and Hoodoo Canyon | 32.6 | Dry Canyon, early Toquima complex | 4 x 3 remnant | Moderate to large | Unknown | Unknown | Unknown | | Boden, 1986, 1992; Shawe, 1999a; McKee, 1976; Henry et al., 1996 |
| Northumberland Tuff | 32.9 | Northern Toquima Range | 15 x 15; 180 | 720 | | | | | McKee, 1974, 1976 |
| Caetano Tuff | 34.0 | Caetano | 20 x 10–18; 280 | 1100 | Abundant rhyolite dikes; 35.69 ± 0.06, n = 13. Minor andesite lavas; 35.21 ± 0.18 | | Several rhyolite intrusions compositionally related to tuff; 33.95 ± 0.05, n = 3 | Highly extended; ash-flow tuff dike exposed for ~3.5 km; caldera floor locally exposed | John et al., 2008; Colgan et al., 2008, 2011 |
| Tuff of Hall Creek | 34.0 | Hall Creek | 20 x 10; ~150 | ≤Large | | | Several rhyolite intrusions, 33.51 ± 0.03 | Moderately extended so area uncertain | Stewart and McKee, 1968; J.P. Colgan and C.D. Henry, unpub. mapping |
| Tuff of Cove Mine | 34.4 | Beneath Reese River Valley? | Unknown | Moderate | Unknown | Unknown | Unknown | | John et al., 2008; Colgan et al., 2008, 2011 |

Note: ?—uncertain correlation.

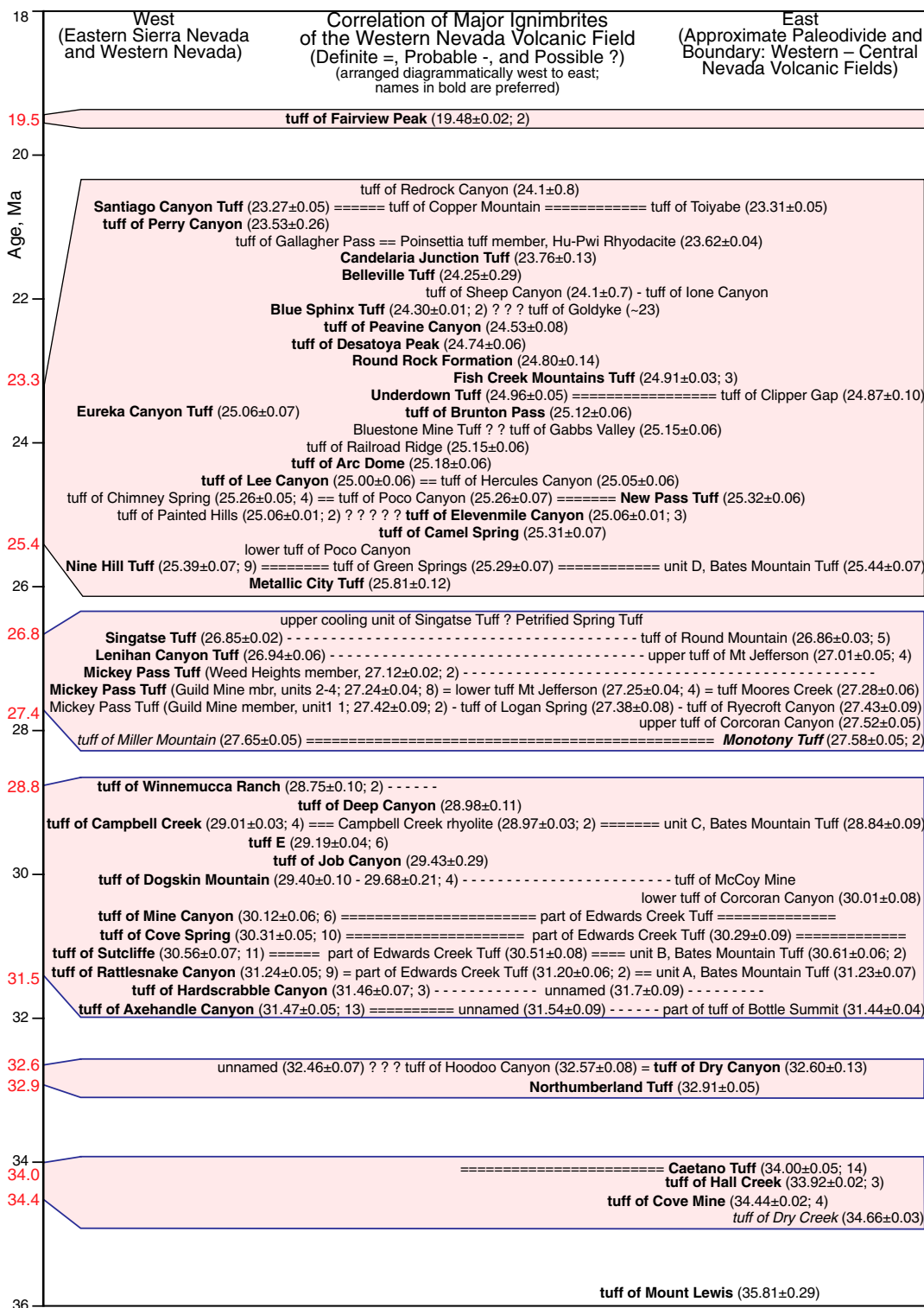


Figure 3. Definite (=), probable (-), or possible (?) correlations of major ash-flow tuffs of the western Nevada volcanic field. Position of names and extension of symbols beyond names (e.g., Caetano Tuff) diagrammatically indicate west-east distribution of tuffs. Names in bold are accepted formal names or best names based on dominant use. All but the tuffs of Perry Canyon, Gallagher Pass, and Mount Lewis are definite or probable caldera-forming units. The Monotony Tuff and probably the tuff of Dry Creek, which crops out only at the eastern edge of the western Nevada field, erupted from calderas in the central Nevada field (shown in italics). The Monotony, Nine Hill, and Underdown Tuffs and the tuff of Campbell Creek extend far into the central Nevada field. All ages (in Ma) are sanidine ⁴⁰Ar/³⁹Ar based on Fish Canyon Tuff sanidine at 28.201 Ma (Kuiper et al., 2008), except mostly plagioclase on tuffs of Dogskin Mountain and Mount Lewis and U-Pb zircon on tuff of Job Canyon, and are followed by the number of dated samples (n = 1 if not specified). Age of Underdown Tuff is from Best et al. (2013c).

TABLE 2. COMPOSITIONAL AND PETROGRAPHIC CHARACTERISTICS OF ASH-FLOW TUFFS IN THE WESTERN NEVADA VOLCANIC FIELD

| Tuff (names, west to east) | Other names and subunits | Age (Ma) | Rock name* | Composition | | | Petrographic characteristics | | | | | | | | | | Source of data | |
|--|---|--------------|------------|-------------------------------|--------------------------------|-------------------|--------------------------------|---------------------|--------------------------|-------------------------|----------------------|-------------------------|---------------------|---------------------------|-------------------------|---------------------------------------|--|---------------------------|
| | | | | SiO ₂ [†] | TiO ₂ [†] | Number of samples | Total phenocrysts [§] | Quartz [§] | Plagioclase [§] | K-feldspar [§] | Biotite [§] | Hornblende [§] | Opaque [§] | Other phenos [§] | Notable characteristics | | | |
| Tuff of Fairview Peak | | 19.5 | HSR | 75 to 76 | 0.12 to 0.13 | 2 | 10 to 15 | tr to 1 | 3 | tr to 1 | 3 | tr to 1 | 3 | tr to 1 | 0 | cpx 0 to tr | Henry (1996a, 1996b); this study | |
| Santiago Canyon Tuff – tuff of Copper Mountain – tuff of Toiyabe | Santiago Canyon Tuff Tuff of Copper Mountain Tuff of Toiyabe | 23.3 | HSR | 74.4 | 0.28 | 1 | 32 | 15 | 6 | 8 | 2 | 2 | 1 | 1 | sph | Abundant sphene phenocrysts | Bingler (1978a); Ekren et al. (1980); this study | |
| Tuff of Sheep Canyon | Tuff of Lone Canyon(?) | <24.3 | R-HSR | 72 to 75 | 0.25 to 0.34 | 21 | 28 to 44 (14.4 to 36.2) | 6.3 (2.1 to 11.3) | 9.8 (7.1 to 16.2) | 4.9 (1.8 to 7.8) | 1.9 (0.9 to 3.1) | 1.1 | 0.5 (0.1 to 1.0) | 0.5 | sph, cpx-tr to 1.3 | Abundant sphene phenocrysts | John (1992a) | |
| Blue Sphinx Tuff – tuff of Golddyke | Blue Sphinx Tuff | 24.3 | R-TD | 68 to 73 | 0.23 to 0.54 | 29 | 29.1 (13.9 to 44.3) | 3.1 (0.1 to 9.0) | 18.3 (6.8 to 32.1) | 4.0 (0.0 to 10.3) | 2.6 (0.5 to 5.8) | 0.3 | 0.8 (tr to 1.3) | 0.8 | z (all, ap, cpx, sph) | Abundant biotite-rich pumice | John (1992a); Whitebread et al. (1988) | |
| Tuff of Peavine Canyon | Tuff of Golddyke | 24.5 | HSR | 73.8 to 75 | 0.22 to 0.29 | 19 | 28.0 (9.4 to 36.4) | 5.1 (1.0 to 8.7) | 11.6 (3.3 to 28.1) | 9.7 (1.6 to 13.9) | 1.1 (0.2 to 2.1) | 0.1 | 0.4 (0.2 to 1.0) | 0.4 | px tr | Intensely resorbed quartz phenocrysts | Ekren and Byers (1985, 1986a, 1986b, 1986c) | |
| Tuff of Desatoya Peak | Lower cooling unit | 24.7 | R | 71 | | 2 | 24 to 25 | 3 | 2 | 5 | 2.5 to 3 | tr | 0.5 to 1 | 0.5 | z (all, ap, sph) | Vermicular quartz (sieve-textured) | John (1992a) | |
| Round Rock Formation | | 24.8 | R | zoned; 74 (base), 72 (top) | zoned; 0.16 (base), 0.34 (top) | 4 | 36 (26 to 50) | 14.5 (10 to 19) | 1 | 20.3 (15 to 29) | tr | | | | ap, z | Smoky quartz | McKee (1970); Cousens et al. (2011); this study | |
| Fish Creek Mountains Tuff | | 24.9 | HSR | 74 to 77 | 0.14 to 0.23 | 2 | 4 to 5 | 1 to 3 | 1 | 1 to 2 | 0 to tr | | | | | | Bomham (1970); this study | |
| Underdown Tuff – tuff of Clipper Gap | Underdown Tuff Tuff of Clipper Gap | 25.0 | HSR | 75.6 to 76.1 | 0.13 to 0.14 | 11 | 4 to 6 | 0.6 to 3 | 0.2 to 1.5 | 1.5 to 4.4 | 0 to 0.3 | 0 to tr | 0 to 0.3 | 0 to 0.3 | | | Rare altered fayalitic olivine | Bomham (1970); this study |
| Tuff of Arc Dome | Upper Lower | 25.2 | R | 73 to 74 | 0.18 to 0.25 | 14 | 33.0 (28.0 to 35.3) | 6.5 (5.2 to 8.7) | 11.3 (6.7 to 15.7) | 13.1 (11.2 to 15.4) | 1.3 (0.9 to 1.7) | 0.4 | 0.4 (0.1 to 0.8) | 0.4 | all, z | Smoky quartz | John (1992a) | |
| Eureka Canyon Tuff – Bluestone Mine Tuff(?) – tuff of Brunton Pass(?), tuff of Gabbs Valley(?), tuff of Camel Spring(?), tuff of Menter Canyon(?), tuff of Railroad Ridge(?) | Eureka Canyon Tuff to Brunton Pass Bluestone Mine Tuff Tuff of Brunton Pass | 25.4 to 25.1 | HSR | 77 | 0.09 to 0.12 | 7 | 30.6 (28.9 to 32.0) | 12.3 (9.5 to 15.3) | 4.5 (1.6 to 7.0) | 12.8 (11.8 to 14.4) | 0.5 (0.1 to 1.0) | 0.1 | 0.2 (0.1 to 0.2) | 0.2 | (z) | Smoky quartz, small black flame | John (1992a) | |
| | | | HSR | 76.9 to 77.1 | 0.08 to 0.11 | 6 | 6.5 (5.9 to 7.3) | 3.4 (2.7 to 4.3) | 0.4 (0.0 to 0.9) | 2.5 (2.2 to 3.1) | 0.0 (0.0 to 0.1) | 0.0 | 0.2 (0.1 to 0.3) | 0.2 | cpx tr | | Bingler (1978a); this study | |

(continued)

TABLE 2. COMPOSITIONAL AND PETROGRAPHIC CHARACTERISTICS OF ASH-FLOW TUFFS IN THE WESTERN NEVADA VOLCANIC FIELD (continued)

| Tuff (names, west to east) | Other names and subunits | Age (Ma) | Rock name* | Composition | | | | Petrographic characteristics | | | | | | | Other phenos ^a | Notable characteristics | Source of data |
|--|--|----------|------------|-------------------------------|-------------------------------|-------------------|--------------------------------|------------------------------|--------------------------------|-------------------------|----------------------|-------------------------|---------------------|----------|--|--|----------------|
| | | | | SiO ₂ [†] | TiO ₂ [†] | Number of samples | Total phenocrysts ^b | Quartz ^b | Plagioclase ^b | K-feldspar ^b | Biotite ^b | Hornblende ^b | Opaque ^b | | | | |
| Eureka Canyon Tuff – tuff of Brunton Pass(?), Gabbas Valley (GV3) | Upper tuff of Gabbas Valley (GV3) | | HSR | 74.8 to 76.1 | 0.09 to 0.13 | 8 | 8 | 2.3 | 2.3 | 3.3 | 0.1 | 0.1 | | | | Ekren and Byers (1976, 1985, 1986b); this study | |
| Bluestone Mine Tuff(?), tuff of Gabbas Valley(?), tuff of Camel Spring(?), tuff of Menter Canyon(?), tuff of Railroad Ridge(?) | Middle tuff of Gabbas Valley (GV2) | | HSR | 75.5 to 76.8 | 0.07 to 0.16 | 20 to 35 | 20 to 35 | 5 to 9 | 6 to 11 | 8 to 14 | 0.2 to 0.4 | | | | altered mafic 0.8 to 1.4 | Ekren and Byers (1985, 1986c); this study | |
| | Lower tuff of Gabbas Valley (GV1) | | R-HSR | 73.5 to 78 | 0.08 to 0.16 | 4 to 11 | 4 to 11 | 0.4 to 3.5 | 0.8 to 4.6 | 1.3 to 5.1 | tr to 1.0 | 0 to 1.0 | tr | | | Ekren and Byers (1985, 1986c); this study | |
| | Tuff of Gabbas Valley, Paradise Range (GV1?) | | R | 73.7 | 0.18 | 3 | 5.9 (4.7 to 6.5) | 0.8 (0.5 to 1.0) | 1.8 (1.4 to 2.4) | 3.0 (2.4 to 3.7) | 0.1 (tr to 0.1) | 0.0 (0.0 to 0.1) | tr (tr to 0.1) | | (cpx) | John (1992a) | |
| | Tuff of Menter Canyon | | HSR | 76.5 | 0.09 | 2 | 1.1 to 1.2 | 0.2 to 0.3 | 0.6 to 0.8 | 0.2 | 0 | 0 | tr | | | John (1992a) | |
| | Tuff of Camel Spring | | HSR | 74 to 75 | 0.13 to 0.18 | 8 | 6.3 (0.3 to 10.8) | 0.5 (0.0 to 2.0) | 5.6 (0.1 to 10.4) ^b | 2 | tr (0.0 to 0.2) | 0.0 (0.0 to tr) | 0.3 (tr to 0.8) | | (cpx, z) | John (1992a) | |
| | Tuff of Railroad Ridge | | R | 70.3 | 0.4 | 1 | 25 | 5 | 12 | 8 | 2 | ? | | (cpx tr) | Riehle et al. (1972); Hardyman et al. (1988); this study | | |
| | Tuff of Chimney Spring – tuff of POCO Canyon – New Pass Tuff | 25.2 | R-HSR | 71.8 to 77.4 | 0.10 to 0.28 | 1 | 34 | 15 | tr | 17 | 1 | tr | | | | This study | |
| | Tuff of POCO Canyon, upper unit | | HSR | 74 to 78 | 0.14 to 0.21 | 24 | 37.0 (22.6 to 49.1) | 13.6 (8.1 to 19.4) | 3.9 (0.4 to 9.1) | 18.6 (7.9 to 28.9) | 0.4 (0.0 to 1.3) | 0.0 (0.0 to 0.2) | 0.4 (0.1 to 0.8) | | all, z | John (1995); this study | |
| | Tuff of POCO Canyon, lower unit | | R-HSR | 71.2 to 78 | 0.06 to 0.35 | 15 | 41.3 (30.9 to 53.5) | 15.7 (9.5 to 19.7) | 4.4 (0.0 to 10.2) | 20.3 (10.2 to 30.6) | 0.4 (0.0 to 1.7) | 0.0 (0.0 to 0.2) | 0.5 (0.3 to 1.0) | | z (all) | John (1995); this study | |
| | New Pass Tuff | | HSR | 75.8 to 76.1 | 0.09 to 0.11 | 1 | 39.6 | 18.3 | 0.0 | 21.0 | 0.0 | 0.0 | 0.4 | | all, z | This study | |
| | Tuff of Elevenmile Canyon | 25.1 | R | 69 to 70 | 0.35 to 0.44 | 10 | 34.4 (26.8 to 40.3) | 3.7 (2.0 to 8.9) | 19.0 (15.4 to 22.5) | 8.2 (3.1 to 12.2) | 2.5 (1.4 to 3.1) | 0.4 (0.0 to 0.8) | 0.7 (0.1 to 1.3) | | ap, z | John (1995); this study | |
| | Tuff of Hercules Canyon | | R-TD | 65 to 71 | 0.32 to 0.67 | 48 | 36.4 (11.1 to 55.5) | 2.3 (0.0 to 9.2) | 23.2 (5.5 to 36.8) | 6.1 (1.2 to 14.0) | 3.4 (0.4 to 6.4) | 0.4 (0.0 to 1.3) | 0.7 (0.1 to 1.5) | | ap, z, (cpx) | John (1995, 1997); this study | |
| | Tuff of Elevenmile Canyon | | R-TD | 64 to 74 | 0.27 to 0.67 | 114 | 32.3 (14.5 to 58.1) | 4.0 (0.0 to 13.7) | 17.9 (3.9 to 47.3) | 7.8 (0.7 to 15.1) | 2.0 (0.1 to 5.3) | 0.3 (0.0 to 1.8) | 0.5 (0.1 to 1.2) | | ap, z, (all, cpx) | John (1995); this study | |
| | Nine Hill Tuff – unit D, Bates Mountain Tuff | 25.4 | R | 71 to 72 | 0.21 to 0.44 | 10 to 18 | tr to 0.5? | tr to 0.5? | tr to 2 | 8 to 15 | tr to 0.2 | | | | 2 to 6 | Three feldspars: Bingle (1978a); Deno (1985); this study | |
| | Nine Hill Tuff, lower unit | | HSR | 75 to 76 | 0.13 to 0.21 | 2 to 6 | tr to 0.5? | tr | tr | 1 to 5 | tr to 0.1 | | | | tr to 1 | Deno (1985); this study | |
| | Tuff of Green Springs | | HSR | 75 to 78 | 0.16 to 0.21 | 5 | 3.2 (1.3 to 6.4) | 0.2 (tr to 0.6) | 0.4 (0.0 to 0.7) | 2.5 (1.2 to 5.3) | 0.0 (0.0 to 0.1) | 0 | 0.2 (tr to 0.4) | | (z) | John (1992a) | |
| | Petrified Spring Tuff | | | | | 22 to 33 | 0 to 2.7 | 10 to 26 | 0 to 5 | 1.8 to 4.6 | 0 to 1.0 | 0 to 1.0 | tr to 0.7 | | cpx 0 to 0.5 | Ekren et al. (1980); Ekren and Byers (1985) | |

(continued)

TABLE 2. COMPOSITIONAL AND PETROGRAPHIC CHARACTERISTICS OF ASH-FLOW TUFFS IN THE WESTERN NEVADA VOLCANIC FIELD (continued)

| Tuff (names, west to east) | Other names and subunits | Age (Ma) | Rock name* | Composition | | | | Petrographic characteristics | | | | | | | Notable characteristics | Source of data |
|--|--|----------|------------|----------------------------------|--------------------------------|-------------------|--------------------------------|------------------------------|--------------------------|-------------------------|----------------------|-------------------------|----------------------|---|---|----------------|
| | | | | SiO ₂ [†] | TiO ₂ [†] | Number of samples | Total phenocrysts [‡] | Quartz [‡] | Plagioclase [§] | K-feldspar [§] | Biotite [§] | Hornblende [§] | Opaques [§] | Other phenos [§] | | |
| Singatase Tuff – tuff of Round Mountain | Tuff of Round Mountain | 26.9 | HSR-R | zoned; 77 (base), 74 (top) | zoned; 0.15 (base), 0.26 (top) | 10 | 21 (9 to 41) | 5 (1 to 9) | 8 (2 to 17) | 6 (3 to 10) | 1.5 (0.6 to 2.9) | 0.4 (0 to 1.3) | | | Boden (1986, 1992); Shaive (1995); this study | |
| Singatase Tuff | | | | 69 | 0.47 | 1 | 38 | 5 | 23 | 7 | 2.5 | | | Proffett and Proffett (1976) | | |
| Lenihan Canyon Tuff – upper tuff of Mount Jefferson | | 27.0 | R-TD | zoned; 75 (base), 67 (top) | zoned; 0.18 (base), 0.47 (top) | 15 | 32 (21 to 45) | 4 (1.5 to 8) | 17 (11 to 23) | 6 (2 to 10) | 3 (2 to 5) | 1 (0.6 to 1.6) | | Proffett and Proffett (1976); Ekren et al. (1980); this study | | |
| Mickey Pass Tuff (Weed Heights Member) | | 27.1 | R-HSR | zoned; 68 to 72 (base), 76 (top) | zoned; 0.50 (base), 0.20 (top) | | 20–25 (base) to 12–17 (top) | 1–2 (base) to 3–4 (top) | 5 (base) to 7 (top) | 1 (base) to 0.5 (top) | | sparse | | Proffett and Proffett (1976); Ekren et al. (1980); this study | | |
| Mickey Pass Tuff (main part of Guild Mine Member) – lower tuff of Mount Jefferson | | 27.3 | HSR | 75 to 78 | 0.12 to 0.21 | 18 | 24 (15 to 28) | 10 (7 to 13) | 3 (1 to 5) | 9 (4 to 12) | 1 (tr to 3) | 0 to tr | all | Boden (1986, 1992, 1994); Henry and Faulds (2010); this study | | |
| Mickey Pass Tuff (basal part of Guild Mine Member) – tuff of Loggan Spring – upper tuff of Corcoran Canyon | Tuff of Loggan Spring Upper tuff of Corcoran Canyon | 27.4 | R-TD? | 73 | 0.27 | 3 | 15 to 30 | 1 to 3 | 8 to 25 | 2 to 8 | 2 to 3 | tr to 0.5 | cpx, tr to 0.5 | Boden (1986, 1992, 1994); Henry and Faulds (2010); this study | | |
| Mickey Pass Tuff | | | TD | 66 | 0.60 | 2 | 30, 38 | 0, 1 | 25, 22 | 2, 1 | 2, 5 | | cpx, 1, 3 | Proffett and Proffett (1976); this study | | |
| Tuff of Loggan Spring | | | R-TD | 68 to 69 | 0.63 to 0.50 | 3 | 28 to 35 | 1 to 3 | 17 to 24 | 3 to 3 | 3 to 3.5 | 1 to 1.5 | cpx 0.2, opx 0.5 | Boden (1986, 1992, 1994); this study | | |
| Tuff of Deep Canyon | | | TD | 64 | 0.63 | 1 | 43 | 1.7 | 29 | 5.6 | 4.5 | 0.4 | 1.7 | Boden (1986) | | |
| Tuff of Campbell Creek – unit C, Bates Mountain Tuff | | 28.9 | HSR | 75 to 78 | 0.09 to 0.14 | 15 | 7.5 (4 to 10) | 2 (1 to 4) | 2.8 (1.5 to 4) | 2.1 (1 to 4) | 0.5 (tr to 1) | tr (0.0 to 0.2) | (opx) | Brooks et al. (2003, 2008); Henry et al. (2012a) | | |
| Tuff of Deep Canyon | | 29.0 | R | 71.5 | 0.29 | | | | | | | | | Riehle et al. (1972); Hardyman et al. (1988) | | |
| Tuff E | | 29.1 | R | 73.7 | 0.1 | 5 | 13 (7 to 22) | 4 (2 to 8) | 4 (3 to 7) | 4 (1 to 7) | 1 (tr to 2) | | | Brooks et al. (2003, 2008); this study | | |
| Tuff of Job Canyon | Extracaldera | ca. 29.4 | HSR-R | 72 to 76 | 0.31 to 0.44 | 25 | 21.8 (7.8 to 33.9) | 0.5 (0.0 to 1.6) | 13.6 (4.7 to 27.0) | 6.8 (2.8 to 15.7) | 0.3 (0.0 to 1.0) | 0.4 (0.0 to 1.3) | z. (ap) | John (1995); this study | | |
| Intracaldera | | | HSR-D | 68 to 77 | 0.14 to 0.58 | 29 | 13.9 (4.0 to 26.9) | 0.6 (0.0 to 5.1) | 5.3 (0.9 to 13.3) | 7.5 (1.8 to 15.0) | 0.2 (0.0 to 1.3) | 0.0 (0.0 to 0.4) | z. (ap) | John (1995); D.A. John, unpub. data | | |
| Tuff of Dogskin Mountain – tuff of McCoy Mine | | 29.4 | R-TD | 66.6 to 69.6 | 0.39 to 0.52 | 5 | 23.2 (15 to 30) | trace | 20 (12 to 25) | 0 to 1 | 2.5 (2 to 3) | | | McKee and Stewart (1971); Henry et al. (2004); this study | | |
| Lower tuff of Corcoran Canyon | | 30.0 | | | | 1 | 26 | 9 | 10 | 4 | 3 | | | Boden (1986, 1992); this study | | |

(continued)

TABLE 2. COMPOSITIONAL AND PETROGRAPHIC CHARACTERISTICS OF ASH-FLOW TUFFS IN THE WESTERN NEVADA VOLCANIC FIELD (continued)

| Tuff (names, west to east) | Other names and subunits (Ma) | Rock name* | Composition | | | Petrographic characteristics | | | | | | | | Source of data | |
|---|-------------------------------|------------|-------------|-------------------------------|-------------------------------|------------------------------|--------------------------------|---------------------|--------------------------|-------------------------|----------------------|-------------------------|----------------------|----------------|---|
| | | | Age (Ma) | SiO ₂ [†] | TiO ₂ [†] | Number of samples | Total phenocrysts [‡] | Quartz [‡] | Plagioclase [‡] | K-feldspar [‡] | Biotite [‡] | Hornblende [‡] | Opaques [‡] | | Other phenos [‡] |
| Tuff of Mine Canyon | 30.1 | | | | | 7 | 12.7 (8 to 15) | (0 to trace) | 7 (2 to 10) | 3.9 (1 to 6) | 1.8 (0.5 to 2.5) | | | | Henry et al. (2004); this study |
| Upper | | | | | | | 11 to 15 | 0 to tr | 8 to 10 | 3 to 6 | 1 to 2.5 | | | | |
| Lower | | | | | | | 8 to 11 | 0 to tr | 2 to 4 | 1 to 4 | 0.5 to 1 | | | | |
| Tuff of Cove Spring | 30.3 | | | | | 2 | 10 to 12 | tr | 6 to 8 | 1 to 3 | 0.5 to 1 | | | | Henry et al. (2004); this study |
| Lower | | | | | | 2 | 8 | tr | 3 | 5 | tr | | | | |
| Tuff of Sutcliffe – part of Edwards Creek Tuff – unit B, Bates Mountain Tuff | 30.4 to 30.6 | HSR-R | 74 to 75 | 0.20 to 0.23 | | 12 | 13.7 (5 to 18) | tr (0 to 1.5) | 6.8 (1 to 15) | 5.4 (tr to 8) | 1.6 (0 to 3) | 0 to tr | | | Sargent and McKee (1969); Henry et al. (2004); this study |
| Tuff of Rattlesnake Canyon – part of Edwards Creek Tuff – unit A, Bates Mountain Tuff | 31.2 | HSR-R | 73 and 75 | 0.16 and 0.18 | | 7 | 8.4 (5 to 12) | tr | 4.3 (3 to 7) | 4.1 (2 to 7) | 0 to tr | | | | Henry et al. (2004); this study |
| tuff of Hardscrabble Canyon | 31.4 | | | | | 2 | 6 to 7 | 1 to 2 | 2 to 2.5 | 2.5 to 3 | tr | | | | Henry et al. (2004); this study |
| Tuff of Axehandle Canyon | 31.5 | R | 68.5 to 73 | 0.28 to 0.41 | | 16 | 13.4 (5 to 21) | tr | 7.2 (2 to 15) | 4.4 (2 to 8) | 1.8 (tr to 2.5) | 0 to tr | | | Henry et al. (2004); this study |
| Upper | | | | | | 3 | 9 to 21 | tr | 4 to 13 | 3 to 6 | 1.5 to 2.5 | | | | |
| Lower | | | | | | 3 | 5 to 8 | tr | 2 to 3 | 2 to 4 | <1 to 1 | | | | |
| tuffs of Dry Canyon and Hoodoo Canyon | 32.6 | TD-R | 67 and 71 | 0.48 and 0.33 | | 3 | 25 | 1 to 2 | 16 to 20 | 1 to 3 | 2 to 4 | 0.1 to 0.2 | | | McKee (1976); Boden (1992); this study |
| Northumberland Tuff | 32.9 | HSR-R | 74 to 76 | 0.20 to 0.25 | | 1 | 25 | 13 | tr | 12 | tr | | | | McKee (1974, 1976); this study |
| Caetano Tuff | 34.0 | HSR-R | 71 to 77 | 0.10 to 0.26 | | 16 | 38.2 (31.7 to 43.0) | 10.3 (4.9 to 16.7) | 13.6 (8.0 to 19.4) | 11.7 (7.6 to 14.5) | 2.4 (0.6 to 4.2) | 0.1 (0.0 to 0.3) | 0.2 (0.0 to 0.4) | | Smoky quartz |
| Upper unit | | R | 72 to 75 | 0.14 to 0.27 | | 10 | 40.3 (24.9 to 47.2) | 9.6 (4.1 to 15.7) | 16.9 (12.3 to 20.9) | 10.5 (6.6 to 14.2) | 2.4 (1.0 to 3.5) | 0.3 (0.0 to 2.0) | 0.3 (0.0 to 2.0) | | Smoky quartz |
| Upper lower unit | | R | 72 to 75 | 0.18 to 0.30 | | 15 | 41.7 (36.3 to 49.1) | 10.4 (6.4 to 17.9) | 16.8 (14.3 to 23.3) | 10.4 (6.2 to 15.7) | 2.1 (1.2 to 3.5) | 0 to 0.4 | 0.4 (tr to 1.1) | | Smoky quartz |
| Lower lower unit | | HSR | 76 to 78 | 0.05 to 0.15 | | 18 | 39.9 (33.2 to 48.0) | 12.9 (5.7 to 18.9) | 12.7 (5.5 to 20.8) | 13.3 (9.2 to 14.9) | 1.6 (0.6 to 3.3) | 0 to tr | 0.2 (tr to 0.5) | | Smoky quartz |
| Tuff of Hall Creek | 34.0 | HSR-R | 72 to 76 | 0.10 to 0.17 | | 6 | 10 to 15 | 1 to 4 | 4 to 7 | 4 to 7 | 0.5 | tr | | | John et al. (2008); this study |
| Tuff of Cove Mine | 34.4 | R | 69 to 73 | 0.25 to 0.45 | | 18 | 43.1 (30.9 to 50.0) | 7.8 (2.8 to 9.4) | 21.7 (13.6 to 32.7) | 6.1 (2.5 to 14.9) | 6.2 (2.7 to 7.8) | 0.8 (0.0 to 2.3) | 0.4 (0.1 to 0.7) | | Smoky quartz |

Note: Numbers in italics are visual estimates from thin sections.
[†]D, dacite; HSR, high-silica rhyolite; R, rhyolite; TD, trachydacite.
[‡]Weight percent, major elements normalized to 100% volatile free.
[§]Mean and (range) in volume percent; tr, trace.
[¶]all—allantite; ap—apatite; cpx—clinopyroxene; opx—orthopyroxene; sph—sphene; tr—trace; z—Zircon.
^{**}Feldspar phenocrysts are mostly or entirely anorthoclase.

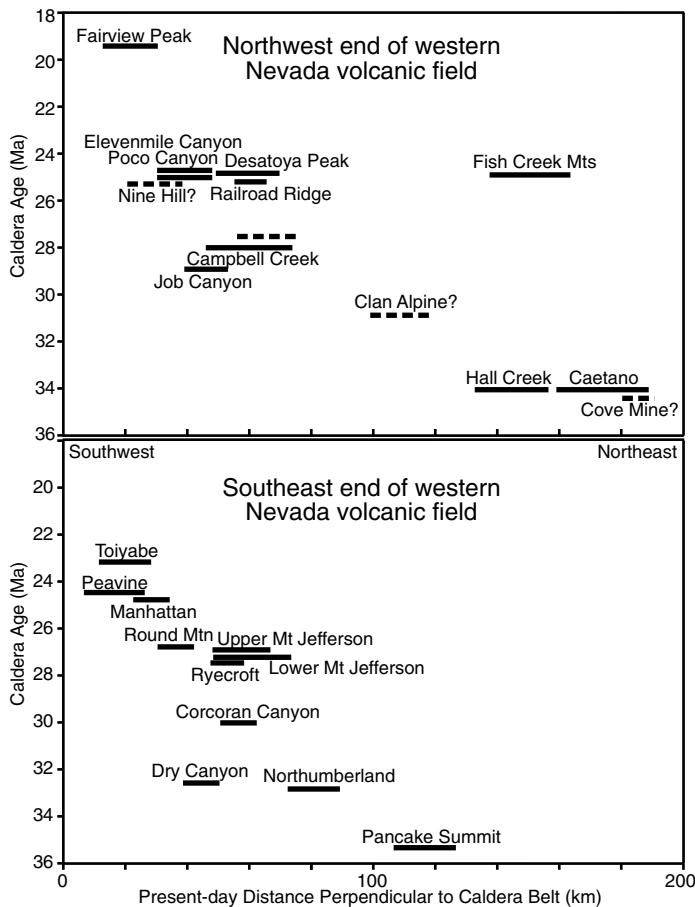


Figure 4. Ages of calderas plotted against distance along two northeast-striking transects perpendicular to the caldera belt (see Fig. 1 for location). Caldera magmatism migrated to the southwest through time along both transects but over a greater total distance along the northwest transect. Dashed lines show uncertainly located calderas.

continued, eruption of voluminous silic ash-flow tuffs resumed at ca. 37 Ma (Henry et al., 2010). Ignimbrite eruptions either initiated or increased greatly at ca. 37 Ma in the southern Rocky Mountain, Mogollon-Datil, Trans-Pecos Texas, and Sierra Madre Occidental fields (Lipman, 1992; Lipman and McIntosh, 2008; Chapin et al., 2004; McIntosh and Chapin, 2004; McDowell and McIntosh, 2012; Henry et al., 2012b). However, major ash-flow volcanism began earlier, as early as 49 Ma in the Challis volcanic field of Idaho (Fisher et al., 1992; Janecke et al., 1997) and 45 Ma in the Sierra Madre Occidental (McDowell and McIntosh, 2012), and both the rhyolitic and more intermediate magmatism in the Great Basin are probably related to slab rollback (Henry et al., 2009a, 2010). Ash-flow magmatism ended or paused at least by 22 Ma in all fields and resumed only in the Great Basin and Sierra Madre Occidental (Henry et al., 2010; McDowell and McIntosh, 2012). Southwestward to westward migration of the front of magmatism continued to the present in the Great Basin, with establishment of the ancestral Cascades arc in western Nevada and the Sierra Nevada at least by ca. 20 Ma (Cousens et al., 2008; Henry et al., 2009a; Vikre and Henry, 2011; John et al., 2012) and as early as 40 Ma in the Warner Range of northeastern California (Colgan et al., 2011a).

Despite the continuity of magmatism, major caldera-forming eruptions were restricted to areas northeast of the Walker Lane, a northwest-striking zone of late Cenozoic dextral slip (Fig. 1; Stewart, 1988; Faulds and Henry, 2008). Late Cenozoic faulting obviously could not have influenced the location of middle Cenozoic calderas. However, the Walker Lane is interpreted to be inherited from a zone of Mesozoic strike-slip faults (Albers, 1967; Oldow, 1984, 1992; Stewart, 1988; Schweickert and Lahren, 1990; Hardyman and Oldow, 1991), although this interpretation is also contested (Proffett and Dilles, 2008). These older structures may have precluded caldera formation in the Walker Lane by weakening or otherwise modifying the crust so that mantle-derived magmas traversed or accumulated in it differently. Although large-volume rhyolite magmatism and caldera collapse stopped at the future northeastern boundary of the Walker Lane, predominantly intermediate magmatism characterized by stratovolcanoes and lava domes continued to migrate to the southwest, across the Walker Lane and into the Sierra Nevada (Dilles and Gans, 1995; Busby et al., 2008; Cousens et al., 2008; Busby and Putirka, 2009; Colgan et al., 2011a; Vikre and Henry, 2011; John et al., 2012).

Although the greatest number and volume of rhyolite ignimbrite eruptions occurred between

Magmatism during shallowing of the slab is represented by scattered Late Cretaceous plutons in northern Nevada (Barton, 1996). During this “flat-slab subduction,” the slab is interpreted to have followed the base of the lithosphere instead of dipping into the asthenosphere. The relatively cool temperature of the shallow slab and the absence of an asthenospheric wedge between slab and overlying mantle lithosphere precluded melting. However, slab dewatering is interpreted to have intensely hydrated (metasomatized) the mantle lithosphere, “priming” it for future magma generation (Humphreys et al., 2003) and mineralization (Richards, 2009; Muntean et al., 2011). As the shallow slab rolled back and sank, asthenosphere welled up behind it, heated the overlying, fertile, hydrated lithosphere, and generated voluminous magmas. Questions about this overall mechanism include the degree to which the cold, shallow slab would

have dehydrated (English et al., 2003) and whether the great volumes of magma erupted during rollback would require sources in addition to mantle lithosphere, i.e., asthenospheric mantle (Farmer et al., 2008).

Despite the relatively regular progression of magmatism, composition and eruptive style varied considerably across the Great Basin in space and time (Henry et al., 2009a, 2010). Cenozoic magmatism began in northeastern Nevada ca. 46 Ma and was dominated by andesite and dacite lavas and their intrusive equivalents (Brooks et al., 1995; Henry and Ressel, 2000; Theodore, 2000; Ressel and Henry, 2006). Only about five voluminous, rhyolitic ash-flow tuffs erupted between 45 and 40 Ma in northeastern Nevada (Castor et al., 2003; Henry, 2008).

Following a hiatus of ~3 Ma during which southwestward migration and intermediate-composition lava flow and dome magmatism

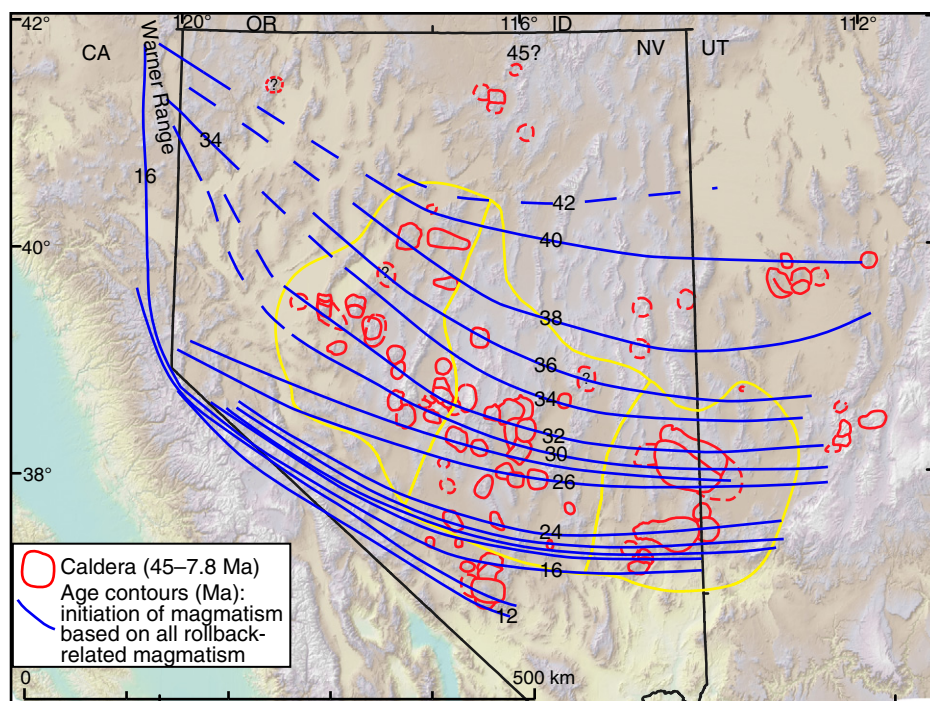


Figure 5. Contours of age of initial, locally sourced magmatism showing southwesterly migration of magmatism. Intermediate to silicic intrusive and effusive activity generally preceded caldera-forming ash-flow eruptions by 2–6 Ma in any area. For example, formation of the 34 Ma Caetano caldera followed ~6 Ma of nearby, semi-continuous andesitic to non-explosive rhyolitic activity (John et al., 2009). Development of the 32.9 Ma Northumberland caldera was preceded by andesitic and rhyolitic intrusions at 35.4 Ma (McKee, 1974, 1976; our unpublished data).

37 and 22 Ma, major caldera-forming eruptions continued to ca. 14 Ma in the Caliente caldera complex (Snee and Rowley, 2000) and to 8 Ma in the southwestern Nevada volcanic field (Fig. 1; Sawyer et al., 1994).

OVERVIEW OF IGIMBRITES

Composition and Phenocryst Assemblage

Igimbrites of the western Nevada volcanic field are characterized by 460 whole-rock chemical analyses, including 98 new analyses and additional trace element data for 87 published analyses (Supplemental Table 3³), and by petrographic and modal data summarized in Table 2. An additional ~100 chemical analyses were discarded due to hydrothermal alteration, most commonly manifested by alkali gains (K_2O) and losses (Na_2O) and/or addition of

³Supplemental Table 3. Chemical analyses of ash-flow tuffs in the western Nevada volcanic field. If you are viewing the PDF of this paper or reading it offline, please visit <http://dx.doi.org/10.1130/GES00867.S3> or the full-text article on www.gsapubs.org to view Supplemental Table 3.

CO_2 . In this section, we describe the general characteristics of the ignimbrites and briefly compare their characteristics to ignimbrites in the central Nevada and Indian Peak–Caliente fields (Best et al., 2013b, 2013c). Geochemical and petrographic features of ignimbrites related to the Stillwater and Caetano calderas are discussed in more detail in the section about these complexes.

Chemical analyses and petrographic data for most major ignimbrites in the western Nevada volcanic field define two major rock types: phenocryst-rich (>20 vol%) and phenocryst-poor (<15 vol%) rhyolites (Figs. 6A and 7; Table 2). Both types of rhyolitic tuff may be normally zoned upward from rhyolite or high-silica rhyolite to trachydacite, although a distinct subtype of phenocryst-rich rhyolite, which includes the Caetano Tuff and the tuff of Poco Canyon, is characterized by great intracaldera thicknesses (>2000 m) of relatively homogeneous crystal-rich (30–50 vol% phenocrysts) high-silica rhyolite (see discussion of the Caetano and Stillwater calderas). Phenocryst-poor rhyolites form several exceptionally widespread units, including the tuff of Campbell Creek (Henry et al.,

2012a) and the Nine Hill Tuff (Deino, 1989; Best et al., 1989). The Weed Heights Member of the Mickey Pass Tuff apparently is reversely zoned from silicic dacite to high-silica rhyolite (Table 2; Proffett and Proffett, 1976; Ekren et al., 1980).

Phenocryst assemblages in rhyolitic ignimbrites of the western Nevada volcanic field vary little (Table 2); quartz, plagioclase, and sanidine in variable proportions generally form >90% of the total phenocryst populations. Biotite and lesser hornblende are the principal mafic silicate minerals, with minor clino- and orthopyroxene present in some tuffs. Fe-Ti oxide minerals are ubiquitous. Anorthoclase is present in a few tuffs, most notably the Nine Hill Tuff, which has three feldspar phenocrysts, sanidine, anorthoclase, and plagioclase. Sphene is a notable accessory phase in the Santiago Canyon Tuff and its correlative units, the tuffs of Toiyabe and Copper Mountain. Zircon, apatite, and allanite are other common accessory phases.

Notably sparse in the western Nevada volcanic field are “Isom-type” trachydacite ignimbrites that contain a higher-temperature, anhydrous phenocryst assemblage of plagioclase, pyroxene, and Fe-Ti oxides. Tuffs with Isom-type characteristics but unknown sources that are probably in the western Nevada field include (1) the Belleville Tuff in the Candelaria area (Speed and Cogbill, 1979) and (2) the tuff of Crow Springs in the Royston Hills (Whitebread and Hardyman, 1987). In contrast, ignimbrites of this type are widespread and voluminous in the volcanic fields to the east and include tuffs of the Isom Formation that were erupted from the Indian Peak–Caliente field (Best et al., 2013b, 2013c).

Igimbrites in the western Nevada field are overwhelmingly rhyolite using the IUGS (International Union of Geological Sciences) classification system (Le Maitre, 1989), although a few tuffs are zoned to trachydacite or dacite (Fig. 6A). Most notable of the zoned tuffs is the tuff of Elevenmile Canyon, whose silica content ranges from 64.3 to 74.5 wt%. Other zoned tuffs and those that consist of dacite or trachydacite include the Nine Hill Tuff and tuffs of Sheep Canyon, Ryecroft Canyon, Logan Spring, Corcoran Canyon, Cove Mine, and Job Canyon and the upper unit of the tuff of Mount Jefferson (Fig. 6A). Five analyses of the correlative tuffs of Dogskin Mountain and McCoy Mine, a widespread, large-volume, plagioclase- and biotite-rich unit, extend from trachydacite and dacite to low-silica rhyolite with 66.6%–69.6% SiO_2 (Fig. 6). The tuff of Dogskin Mountain–McCoy Mine is the only ignimbrite of the western Nevada field that could be a “monotonous intermediate” (Hildreth, 1981). The tuff of Perry Canyon and

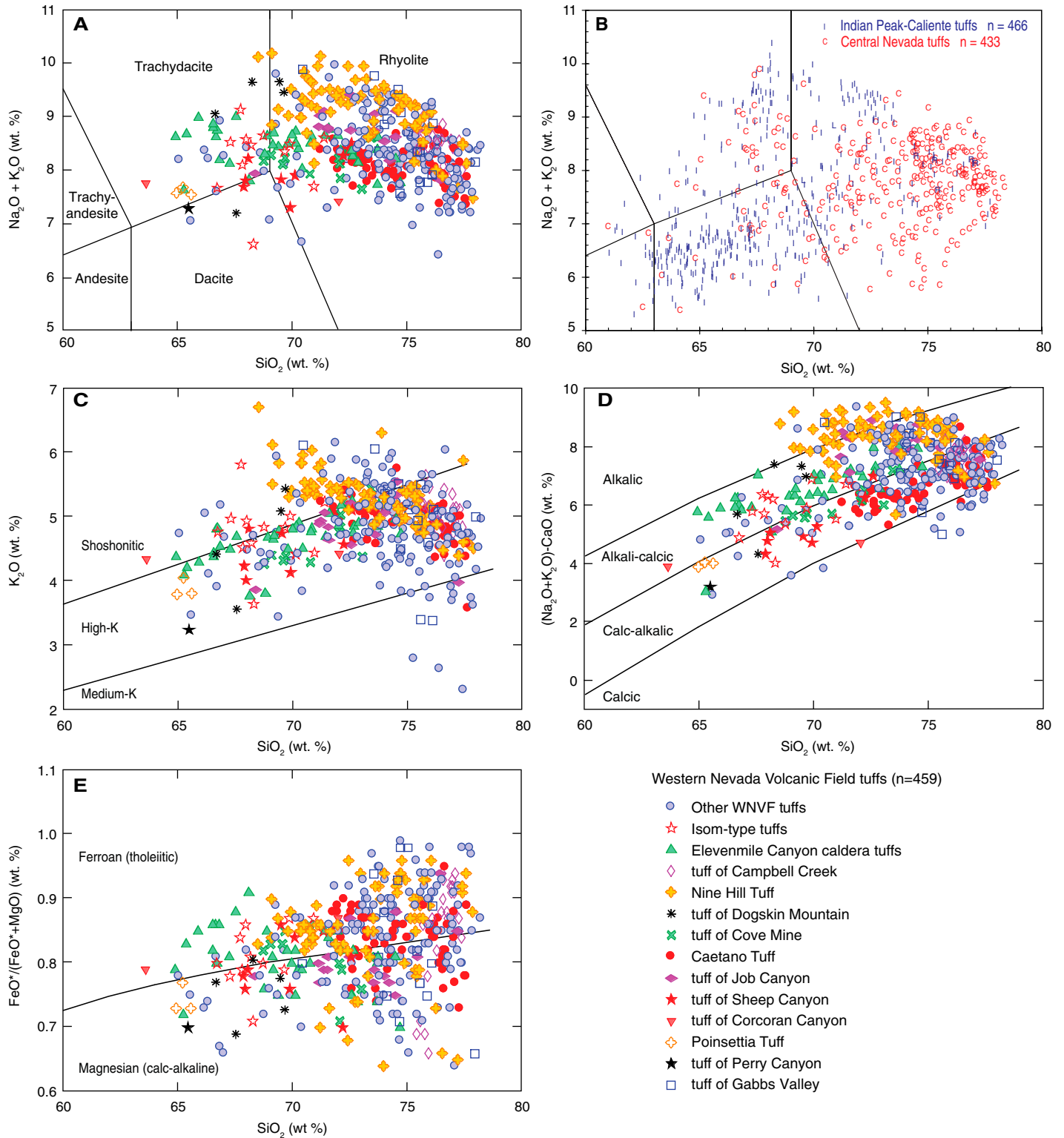


Figure 6. Chemical variation diagrams for whole-rock samples of ignimbrites in the western Nevada volcanic field (WNVF). All analyses normalized to 100% volatile free. Plotted data are tabulated in Supplemental Table 3 [see footnote 3]. (A) Total alkali-silica diagram. Field boundaries from Le Maitre (1989). (B) Total alkali-silica diagram for ignimbrites in the central Nevada and Indian Peak-Caliente fields from Best et al. (2013b, 2013c). (C) K₂O-SiO₂ diagram using classification of Le Maitre (1989) and Ewart (1982). (D) Modified alkali-lime index diagram using classification of Frost et al. (2001). (E) Ferroan-magnesian classification of Frost et al. (2001). FeO* is total Fe as FeO.

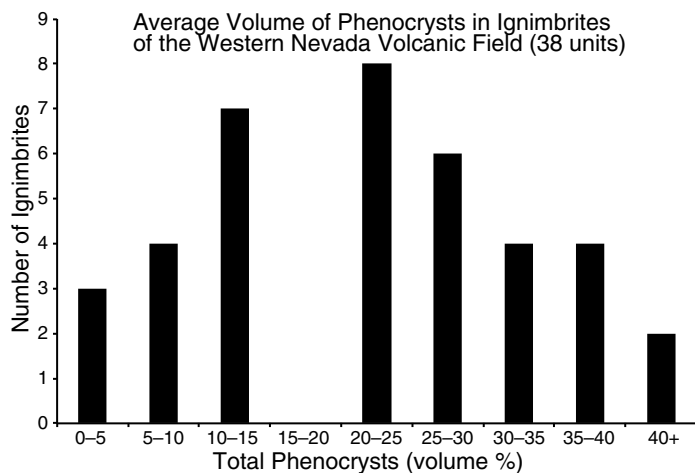


Figure 7. Histogram showing average volume percent of total phenocrysts in major ignimbrite units in the western Nevada volcanic field. Diagram constructed using modal data in Table 2.

the Poinsettia Tuff Member of the Hu-pwi Rhyodacite are small-volume dacitic ignimbrites probably related to intermediate lava domes.

The western Nevada ignimbrites mostly form a high-K series based on their K_2O and SiO_2 contents (Fig. 6C). Several units, including the trachydacitic parts of tuffs related to the Elevenmile Canyon caldera and the Nine Hill Tuff, are shoshonitic. On the basis of Na, K, and Ca contents, most western Nevada volcanic field ignimbrites are calc-alkalic to alkali-calcic (Fig. 6D). The Job Canyon and Elevenmile Canyon caldera tuffs and the Nine Hill Tuff are alkali-calcic. The Nine Hill Tuff is Na_2O rich and CaO poor, whereas the Caetano Tuff is CaO rich (Fig. 8). Other tuffs with trachydacite or dacite compositions are calc-alkalic. The western Nevada volcanic field ignimbrites are nearly evenly split between ferroan (tholeiitic) and magnesian (calc-alkaline) compositions using their Fe-Mg contents (Fig. 6E). Several units, including the Elevenmile Canyon caldera tuffs, have trachydacitic compositions with ferroan characteristics and rhyolitic compositions with magnesian characteristics.

Primitive mantle-normalized trace element patterns for western Nevada tuffs generally show notable depletions in high-field-strength elements (HFSE, Ti, Nb, Zr) and P suggestive of fractionation of Fe-Ti oxides, zircon, and apatite that are characteristic of subduction-related magmas (Fig. 9; e.g., Wood et al., 1979; Gill, 1981; Pearce et al., 1984; Hildreth and Moorbath, 1988). Large-ion lithophilic elements (LIL, Rb, Ba, K, Pb, U, Th, Sr) are variably enriched, although trace element contents vary significantly within zoned ignimbrites, such as the Caetano Tuff and tuffs related to the Elevenmile

and Poco Canyon calderas; this variation likely reflects crystal fractionation. The tuffs have a large range in Ba/Nb from ~1–180 that generally decreases with increasing silica content and suggests fractionation of alkali feldspar in the more silicic rocks, which also are notably depleted in Ba and Sr. The samples with lowest Ba/Nb values (<5) include the lower unit of the Nine Hill Tuff and the most highly evolved parts of the Caetano Tuff and tuffs related to the Poco Canyon caldera (Supplemental Table 3 [see footnote 3]).

The western Nevada ignimbrites are generally similar compositionally to rhyolite ignimbrites in the central Nevada and Indian Peak–Caliente fields to the east. All have high concentrations of K_2O and are calc-alkalic to alkali-calcic. However, ignimbrites in the western Nevada field have relatively high total alkali contents compared to contemporary ash-flow tuffs elsewhere in the Great Basin (Figs. 6A and B). Because overall concentrations of K_2O vary little throughout the Great Basin ignimbrite province with the exception of K-rich Isom-type ignimbrites, the increased alkalinity of the western Nevada volcanic field ignimbrites relative to their eastern counterparts is due to their generally higher Na_2O contents (Fig. 8), a trend also mimicked in lava flows in the western Nevada field (see section below on lava flows and domes), and may reflect different types of crust (accreted oceanic and transitional crust) underlying the western Nevada field. With the exception of highly evolved, high-silica rhyolite parts of several ignimbrites in the western Nevada volcanic field, most of the rhyolites in all 3 fields have trace element patterns and ratios of LIL/HFSE (e.g., Ba/Nb > 15) similar to those characteristic of subduction-related rhyolites elsewhere.

Western Nevada ignimbrites have several distinct characteristics relative to their eastern contemporaries, most notably the aforementioned dearth of dacite and Isom-type compositions that characterize large parts of the volcanic fields to the east (Fig. 6B; see Best et al., 2013b, 2013c). Western Nevada volcanic field ignimbrites are relatively enriched in alkalis, whereas calcic compositions are common in tuffs in the other two fields (K-rich Isom-type tuffs in the eastern fields, which have alkali-calcic to shoshonitic compositions, are notable exceptions). In addition, a larger proportion of the western Nevada ignimbrites are ferroan with higher $FeO^*/(FeO^* + MgO)$ than typical rhyolites in the eastern fields (Fig. 6E). In contrast, in the central Nevada field more than half of the ignimbrites are magnesian and in the Indian Peak–Caliente field virtually all but the Isom Formation tuffs are magnesian. The western Nevada ignimbrites, including sparse Isom-type units, are notably enriched in Pb and U and have lower Th and Th/U than the eastern ignimbrites (Fig. 9). These characteristics also likely reflect differences in underlying types of crust.

Volumes

At least four ignimbrites in the western Nevada volcanic field (Caetano Tuff, tuffs of Campbell Creek and Elevenmile Canyon, and the correlative Mickey Pass Tuff [main part of the Guild Mine Member] and lower tuff of Mount Jefferson), have estimated volumes >1000 km³ and qualify as resulting from supereruptions (de Silva, 2008; Miller and Wark, 2008) (Tables 1 and 2). However, based on their wide distributions, several other tuffs, including the Nine Hill Tuff, the tuffs of Axehandle and Rattlesnake Canyons, and the correlative tuffs of Dogskin Mountain and McCoy Mine, likely also have volumes >1000 km³. The volumes of many ignimbrites and the total volume of the ignimbrites in the western Nevada volcanic field cannot be estimated with available data due to large uncertainties in caldera dimensions, amount of collapse, and thickness of intracaldera tuff; channelization of outflow tuff by paleotopography; and uncertainty in regional correlations.

DESCRIPTION OF MAJOR IGIMBRITES AND THEIR CALDERAS

This section briefly summarizes characteristics of individual tuffs, known and suspected correlations, and what is known about calderas, organized by age groups beginning with the oldest. Many published geologic maps assigned local stratigraphic names to tuffs before correlation was possible between different areas.

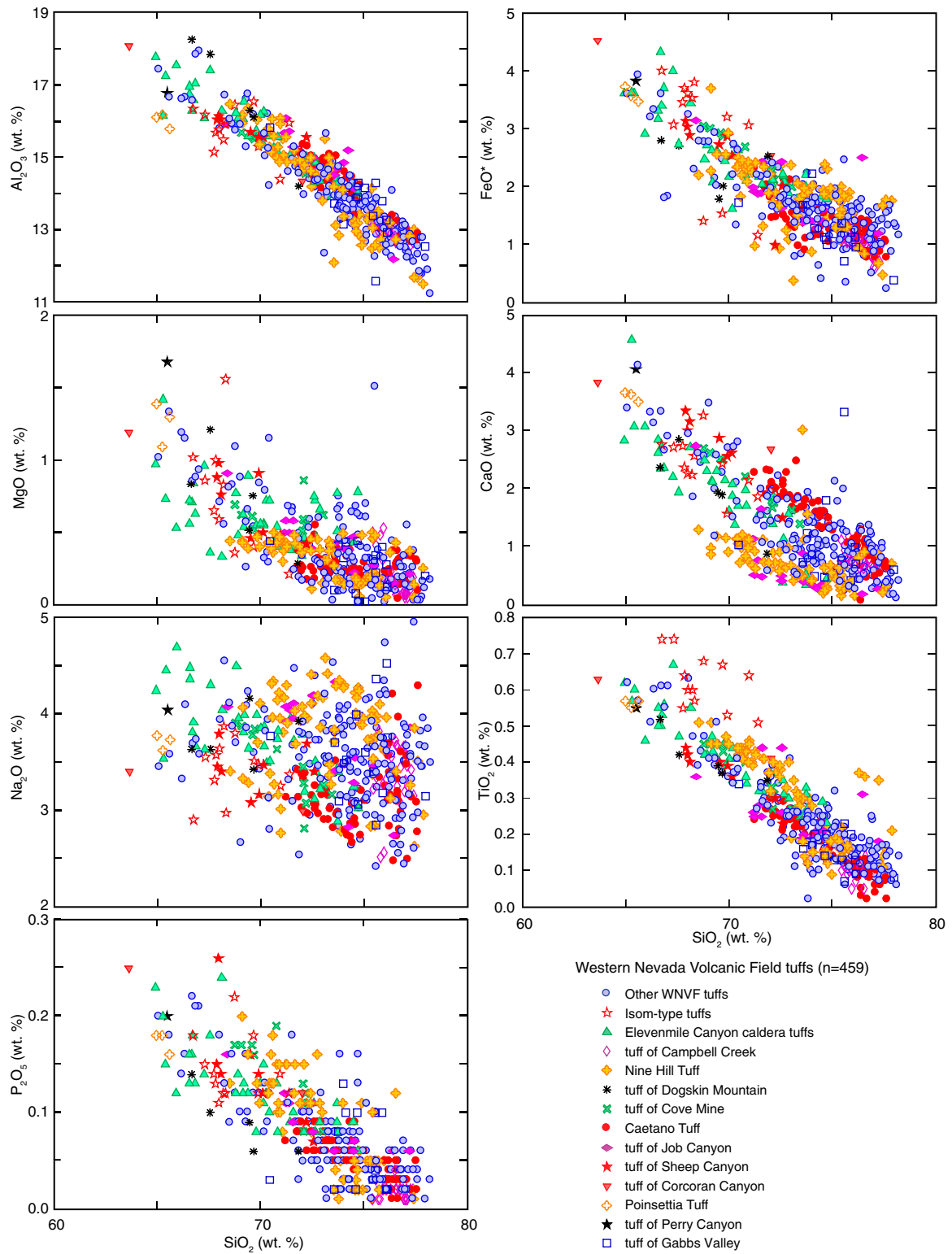


Figure 8. Harker variation diagrams for whole-rock samples of western Nevada volcanic field (WNVF) ignimbrites. All analyses normalized to 100% volatile free. Plotted data are tabulated in Supplemental Table 3 (see footnote 3). FeO* is total Fe as FeO.

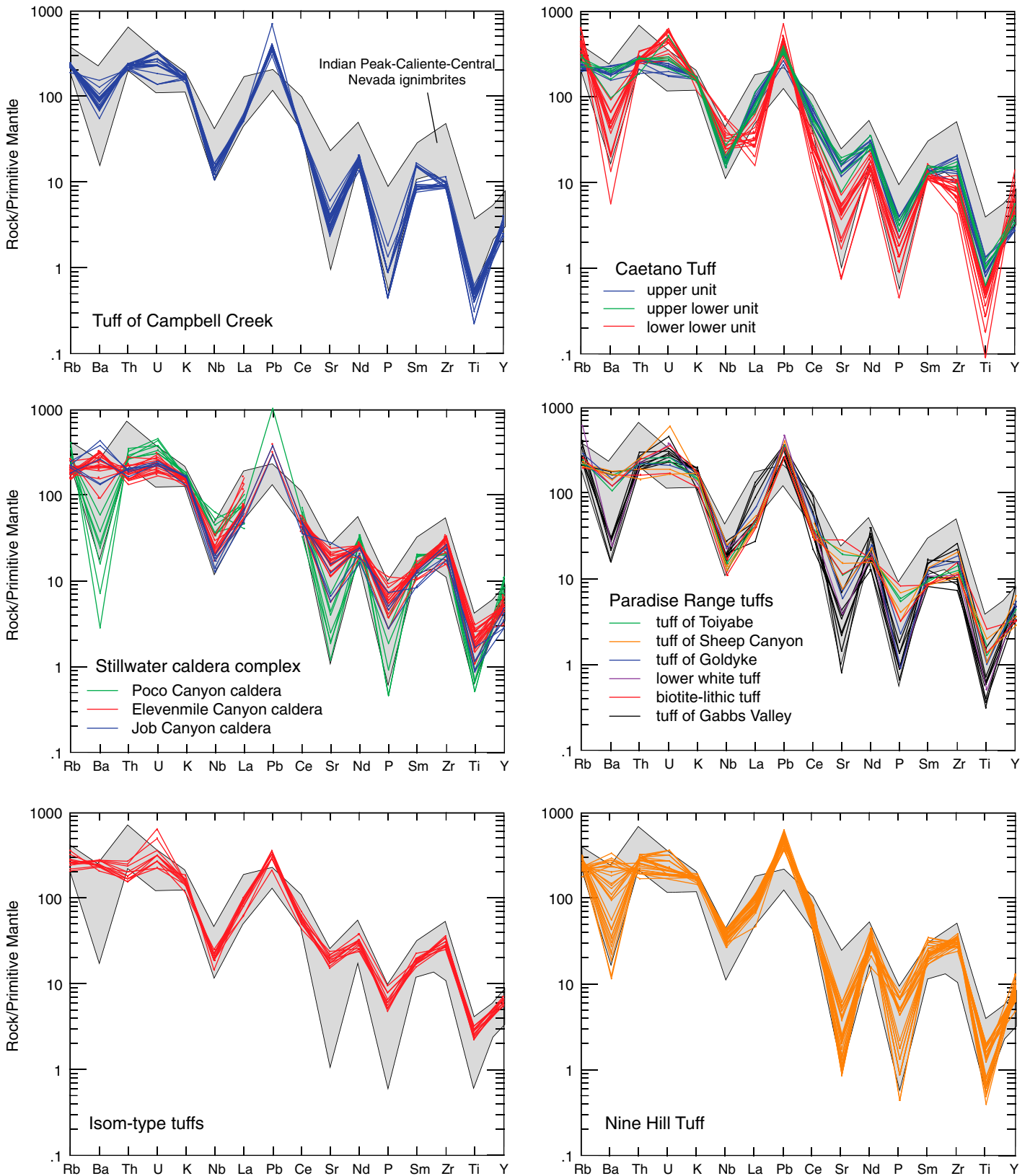


Figure 9. Primitive mantle-normalized trace element diagrams for whole-rock samples of tuffs in the western Nevada volcanic field. Shaded outline shows all tuffs in the central Nevada, Indian Peak, and Caliente fields from Best et al. (2013b, 2013c). Primitive mantle composition from Sun and McDonough (1989). Plotted data are tabulated in Supplemental Table 3 (see footnote 3).

For that reason, the number of names greatly exceeds the number of separate voluminous tuffs. Tables 1 and 2 summarize compositional, petrographic, age, and volume data. Figures 10 and 11 show tuff distributions and calderas divided into 34.4–26.8 Ma and 25.4–19.5 Ma segments, respectively. Supplemental Table 1 (see footnote 1) provides $^{40}\text{Ar}/^{39}\text{Ar}$ ages for most tuffs.

34.4–34.0 Ma

Caetano Tuff and Caldera (34.0 Ma)

The oldest ignimbrites of the western Nevada field are in its northeastern part (Fig. 10). The most

voluminous tuff of this episode is the ~1100 km³, 34.0 Ma Caetano Tuff (John et al., 2008). Most of this phenocryst-rich rhyolite accumulated within its caldera, where it is as much as 4 km thick and normally zoned from high- to low-silica rhyolite (see section on the Caetano and Stillwater calderas). Outflow tuff is found ~5 km to the east in the Cortez Mountains, 50 km to the west in the Tobin Range, and 40 km to the south inside the Hall Creek caldera (Gonsior, 2006; Gonsior and Dilles, 2008; John et al., 2008; this study).

Tuff of Cove Mine (34.4 Ma)

The 34.4 Ma, low-silica rhyolite tuff of Cove Mine, which is petrographically and

chemically similar to low-silica parts of the Caetano Tuff, is exposed locally north and west of the Caetano caldera. The possible location of the Cove Mine caldera is based on the distribution of the tuff and an exploration drill hole that penetrated more than 300 m of Caetano-like tuff in Reese River Valley north of the Caetano caldera (P.J. Dobak, Barrick Exploration Inc., 2012, personal commun.). This thick tuff may lie inside a Cove Mine caldera because (1) the tuff of Cove Mine was originally considered to be the Caetano Tuff, (2) the drill hole is in the middle of the distribution of the tuff of Cove Mine, and (3) outflow Caetano Tuff is generally no more than

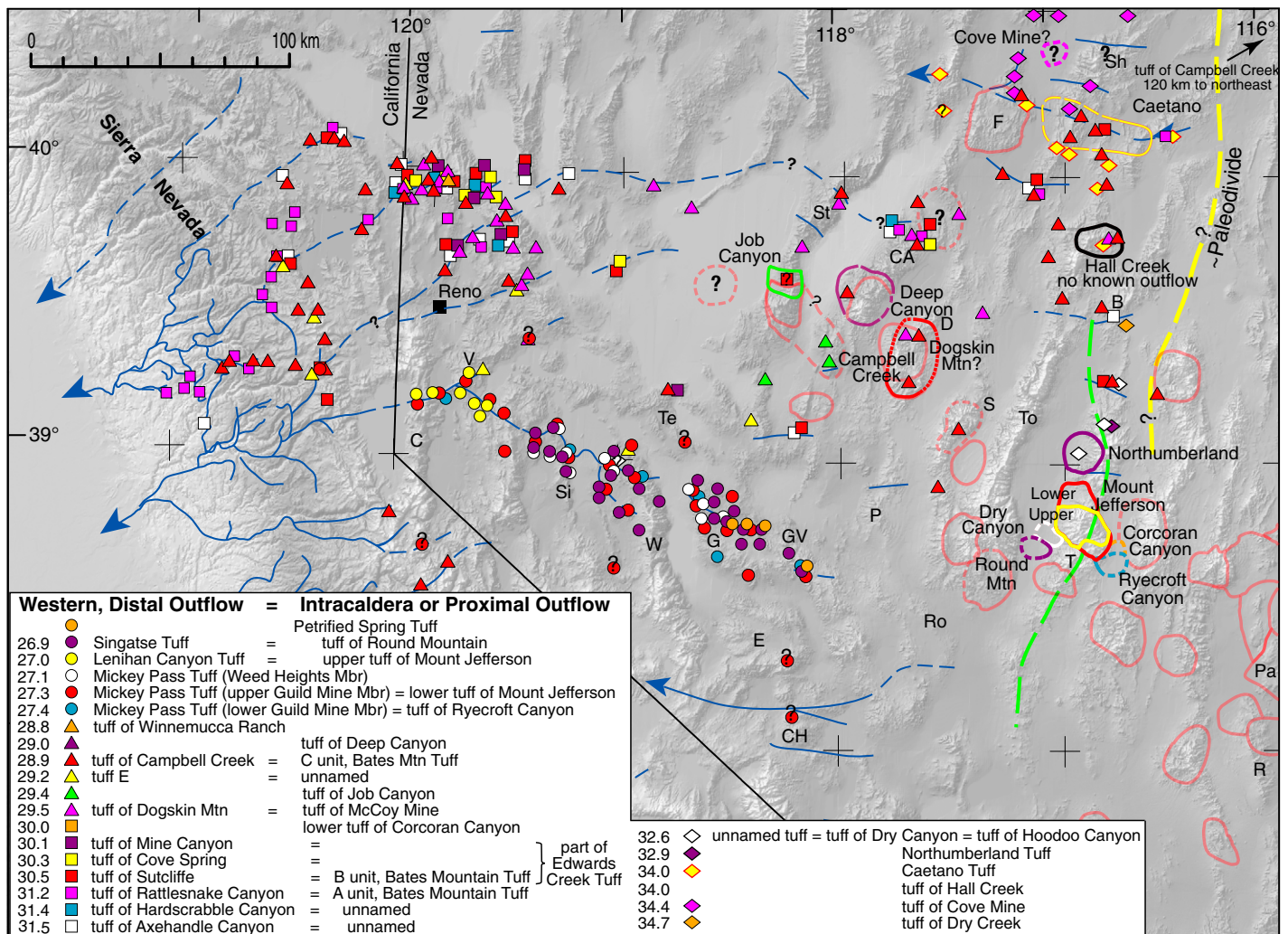


Figure 10. Calderas and tuffs formed during the 34.4–34.0, 32.9–32.6, 31.5–28.8, and 27.4–26.8 Ma intervals. Known locations of outflow deposits are shown as colored symbols, and approximate margins of associated calderas use the same color. Intracaldera tuff crops out extensively in all calderas. Calderas not erupting during the time intervals covered by this figure are in light red. Blue lines and arrowheads are paleovalleys as in Figure 1. B—Bates Mountain; C—Carson Range; CA—Clan Alpine Mountains; CH—Candelaria Hills; D—Desatoya Mountains; E—Excelsior Mountains; F—Fish Creek Mountains; G—Gillis Range; GV—Gabbs Valley Range; P—Paradise Range; Pa—Pancake Range; R—Reveille Range; Ro—Royston Hills; S—Shoshone Mountains; Sh—Shoshone Range; Si—Singatse Range; St—Stillwater Range; T—Toquima Range; Te—Terrill Mountains; To—Toiyabe Range; V—Virginia Range; W—Wassuk Range.

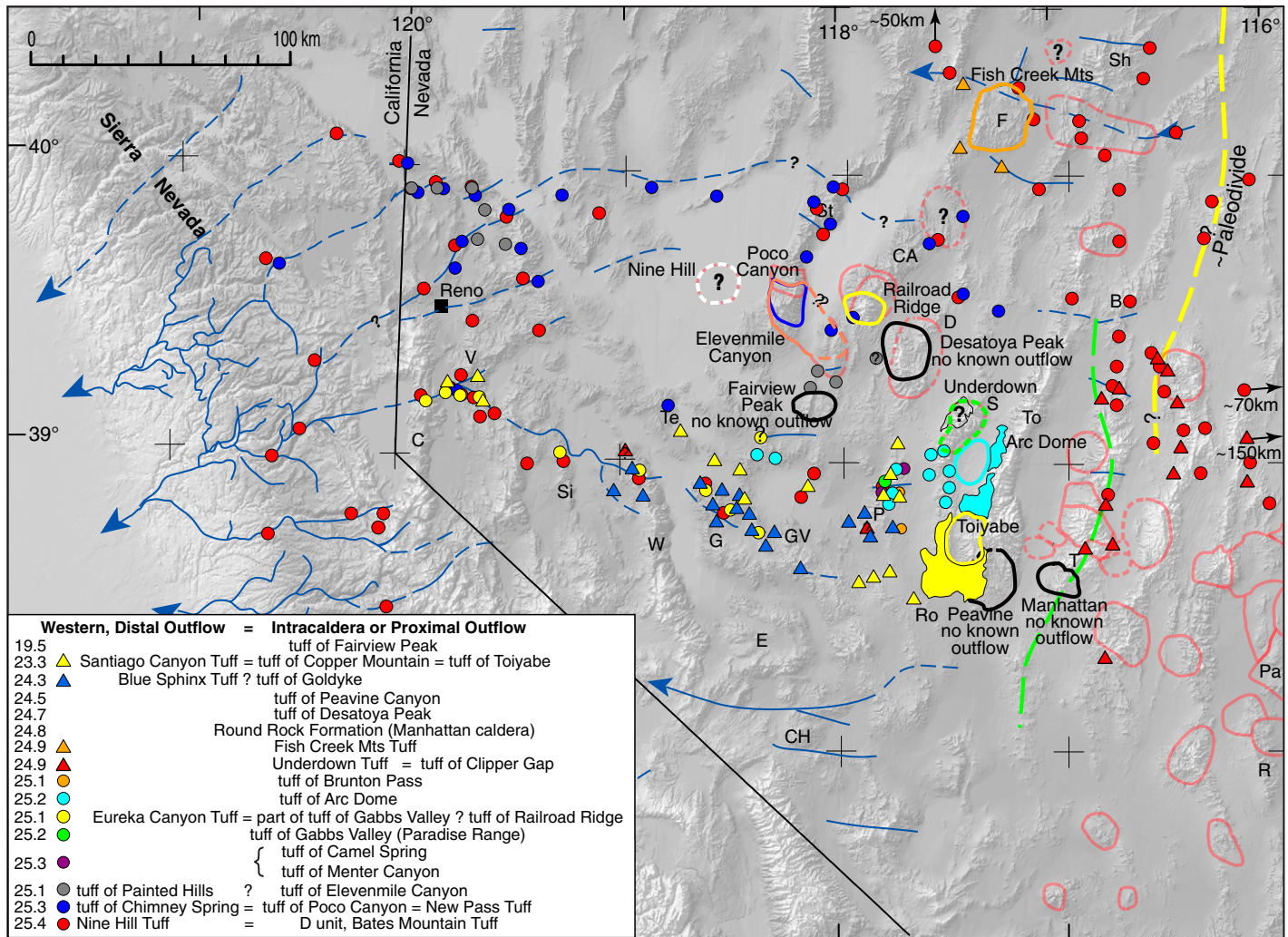


Figure 11. Calderas and tuffs formed during the 25.4–23.3 and 19.5 Ma intervals. Explanation and location abbreviations are given in Figure 10.

250 m thick (in Golconda Canyon) and virtually unknown north of the Caetano caldera (John et al., 2008).

Tuff of Hall Creek and Caldera (34.0 Ma)

The other definitely caldera-forming unit in this group is the sparsely porphyritic, 34.0 Ma tuff of Hall Creek. The tuff of Hall Creek is currently only known within its caldera (Stewart et al., 1977), where it is at least 600 m thick, contains megabreccia along its southern and northern caldera margins, and is cut by two sets of rhyolite domes of distinctly different age: one is ca. 33.5 Ma and possibly related to the magma erupted as tuff, and the other ca. 25 Ma and clearly unrelated (J.P. Colgan and C.D. Henry, unpub. reconnaissance mapping). The 10 × 20 km size of the caldera and ≥600 m thickness of intracaldera tuff suggest a minimum volume of 100 km³.

32.9 and 32.6 Ma

Northumberland Tuff and Caldera (32.9 Ma)

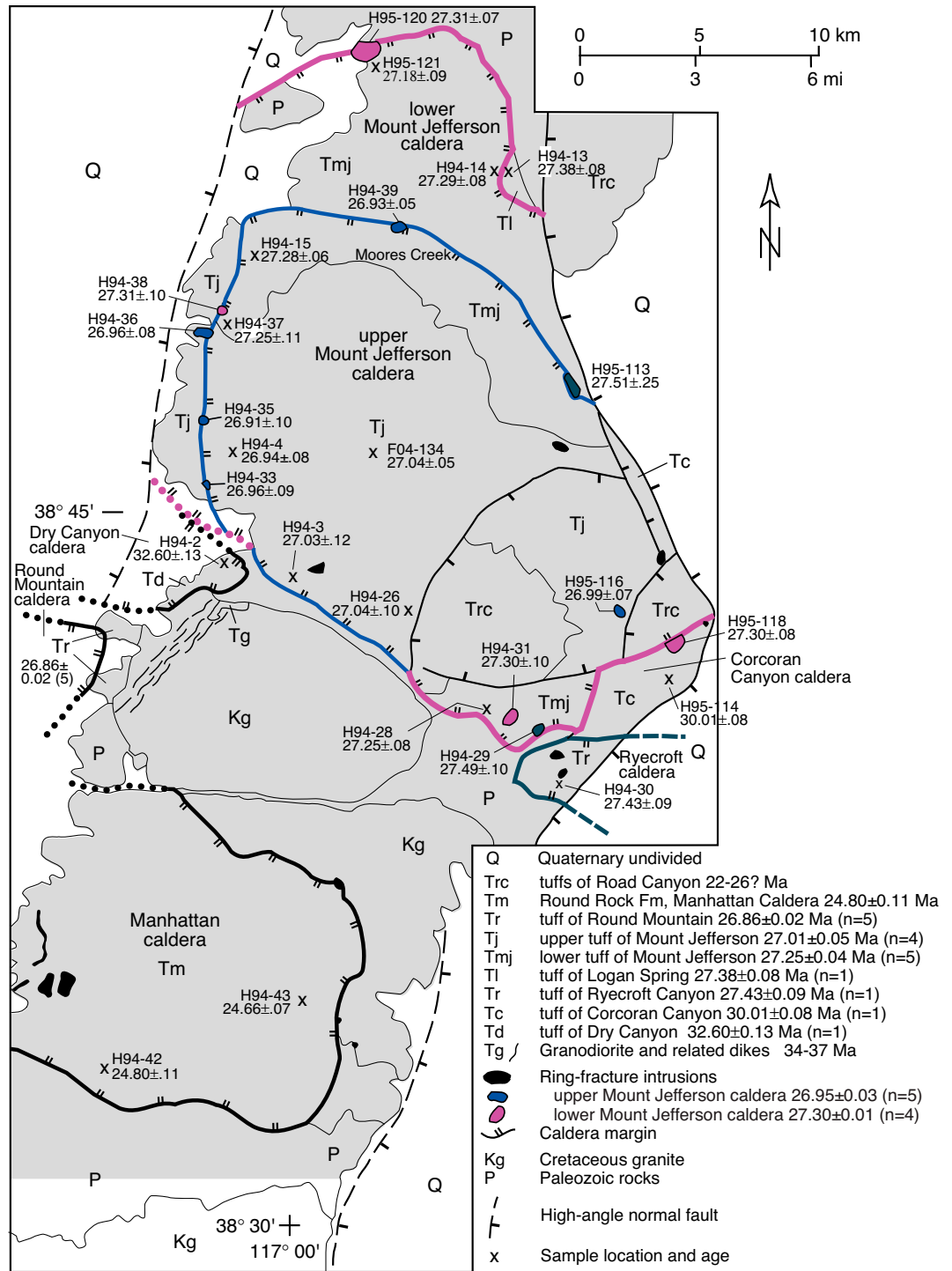
Two rhyolitic tuffs erupted at 32.9 and 32.6 Ma from calderas in the Toquima Range (Fig. 10). The 32.91 ± 0.05 Ma Northumberland Tuff is phenocryst-rich, high-silica rhyolite containing abundant quartz and sanidine and minor biotite and plagioclase (McKee, 1974, 1976). Although McKee (1976) did not map any outflow Northumberland Tuff, the tuff of Stoneberger Canyon, which he mapped as occupying a >250-m-deep paleovalley 7–15 km northeast of the caldera, is probably correlative based on its similar coarse-grained phenocryst assemblage, both tuffs being high-silica rhyolites, close spatial association, and its coarsely lithic character that suggests a proximal deposit. Based on the exposed eastern half of the Northumberland caldera—the western half is buried beneath Big

Smoky Valley—we estimate the caldera had an area of ~180 km². Neither the amount of collapse nor thickness of intracaldera tuff are known, but collapse of at least 1 km seems a minimum, so the erupted volume is probably at least 180 km³.

Tuff of Dry Canyon and Caldera (32.6 Ma)

The 32.60 ± 0.13 Ma tuff of Dry Canyon erupted from a caldera that is largely overprinted by younger calderas of the Toquima caldera complex (Figs. 10 and 12; Boden, 1992; Shawe, 2002). Phenocrysts in the fine-grained, moderately porphyritic tuff are mostly plagioclase and biotite with lesser quartz, sanidine, hornblende, and clino- and orthopyroxene. The exposed intracaldera tuff contains abundant blocks up to 300 m in diameter, mostly of Paleozoic sedimentary and Cretaceous granitic rocks, and pinches out abruptly against Paleozoic rocks at its eastern caldera wall (Boden, 1992; Henry, 1997).

Figure 12. Simplified geologic map of the Toquima caldera complex showing recognized and probable calderas (from Boden, 1986, 1992; Shawe, 1995, 1998, 1999a, 1999b, 2002; Shawe and Byers, 1999; Shawe et al., 2000; Henry et al., 1996; this study). Note that (1) Cenozoic magmatism started with intrusion of granodiorite ~4 Ma before any caldera-forming ash-flow eruptions, (2) caldera-forming magmatism lasted ~8 Ma, and (3) ring-fracture intrusions of both the lower and upper Mount Jefferson calderas are indistinguishable in age from their intracaldera tuffs.



The tuff of Hoodoo Canyon of McKee (1976), which crops out in and around the Northumberland caldera to the north, is probable outflow tuff correlative with the tuff of Dry Canyon based on indistinguishable $^{40}\text{Ar}/^{39}\text{Ar}$ age (32.57 ± 0.08 Ma), sanidine K/Ca, and phenocryst assemblage. A petrographically similar tuff also with an indistinguishable age (32.46 ± 0.07 Ma; H01-70) and sanidine K/Ca in the Wassuk Range

to the west may also be outflow (Figs. 3 and 10). Given the incomplete exposure of the caldera, no estimates of area or volume are possible.

31.5–28.8 Ma

At least eleven major rhyolitic tuffs and one trachydacitic tuff erupted between 31.5 and 28.8 Ma, probably all from calderas in

the Stillwater–Clan Alpine–Desatoya region in the western part of the volcanic field (Fig. 10). These are the tuffs of Axehandle Canyon, Hardscrabble Canyon, Rattlesnake Canyon, Sutcliffe, Cove Spring, Mine Canyon, Dogskin Mountain, Job Canyon, tuff E, Campbell Creek, Deep Canyon, and Winnemucca Ranch. All except the tuffs of Job Canyon and Deep Canyon were first mapped as outflow deposits in

western Nevada and the eastern Sierra Nevada (Brooks et al., 2003, 2008; Henry et al., 2004, 2009b; Faulds et al., 2005a, 2005b; Hinz et al., 2009) and then correlated eastward into the caldera belt on the basis of stratigraphy, phenocryst assemblage, age, sanidine K/Ca, and for some of the units, composition and paleomagnetic data (Gromme et al., 1972; Hinz et al., 2009; C.S. Gromme, 2010, written commun.). Compositions of glass shards further supported the correlations and extended the range of the tuffs of Axehandle Canyon, Rattlesnake Canyon, and Campbell Creek farther west in the Sierra Nevada (Cassel et al., 2009a). The tuff of Job Canyon is the oldest recognized caldera-forming tuff in the Stillwater caldera complex (John, 1995; Hudson et al., 2000), and the tuff of Deep Canyon is an intracaldera deposit in the Clan Alpine Mountains (Hardyman et al., 1988).

Important correlations include those of the tuffs of Rattlesnake Canyon, Sutcliffe, and Campbell Creek with the A, B, and C units, respectively, of the Bates Mountain Tuff (Fig. 3; Gromme et al., 1972; John et al., 2008; Henry et al., 2012a). Additionally, the tuffs of Sutcliffe, Cove Spring, and Mine Canyon probably comprise at least parts of the undivided Edwards Creek Tuff (McKee and Stewart, 1971; Stewart et al., 1977). McKee and Stewart (1971) noted similarities between parts of the Edwards Creek and Bates Mountain Tuffs. The tuff of Bottle Summit mapped at the south end of Bates Mountain (McKee, 1968) is the easternmost known outcrop of the tuff of Axehandle Canyon. Based on their distinctive phenocryst assemblages of abundant plagioclase and biotite, the tuffs of Dogskin Mountain and McCoy Mine also correlate (McKee and Stewart, 1971; Mark Hudson, 2002, personal commun.; our data).

Notable about most of these tuffs is their wide distribution from known or probable sources ~200 km westward to the Sierra Nevada and variably eastward toward and, at least for the tuff of Campbell Creek, across the interpreted paleodivide (Fig. 10; Supplemental Fig. 1⁴; Henry et al., 2012a). The tuffs flowed, were deposited, and are preserved in paleovalleys as much as 1.2 km deep in western Nevada

(Garside et al., 2005; Henry and Faulds, 2010; Henry et al., 2012a).

The ca. 30.5 Ma tuff of Sutcliffe and ca. 29.5 Ma tuff of Dogskin Mountain–McCoy Mine are composite units consisting of petrographically similar tuffs that were initially mapped as single tuffs in western Nevada (Henry et al., 2004). The large age range, variation in phenocryst assemblage and abundance in the tuff of Sutcliffe (although some variation is a result of zoning), and, most importantly, separation of cooling units by sedimentary sequences up to 10 m thick in western Nevada (Henry et al., 2009b) indicate that the tuffs of Sutcliffe and Dogskin Mountain were multiple eruptions possibly over longer time spans than most caldera-forming units.

Sources in the Clan Alpine–Desatoya region are interpreted mostly on the basis of recognition of several definite calderas (for the tuffs of Job Canyon, Campbell Creek, and Deep Canyon), recognition of probable calderas (for the tuff of Dogskin Mountain and tuff E), and tuff distribution. The tuffs are restricted to the east-west belt of paleovalleys, which indicates they must have erupted from sources in that belt, which can only be in the western part of the western Nevada volcanic field.

Volumes of most of these tuffs are poorly known but probably all are large. Based on the area of its caldera and probable amount of collapse, the tuff of Campbell Creek probably had a volume of at least 1200 km³ and possibly as much as 3000 km³ (Henry et al., 2012a). The tuff of Job Canyon has a minimum volume of 240 km³ but could be much larger because the estimate is based on only one slice through the caldera.

The only tuff in the 31.5–28.8 Ma age range not erupted from the Clan Alpine region is the 30.01 ± 0.08 Ma lower tuff of Corcoran Canyon in the Toquima Range (Boden, 1992; Shawe et al., 2000; Shawe, 2002). This tuff is as much as 2 km thick and contains megabreccia in its ~15 km² outcrop area, which is probably a remnant of its source caldera mostly overprinted by younger calderas of the Toquima caldera complex (Boden, 1992; Shawe et al., 2000). The tuff has not been recognized outside this area, and no outflow tuff is known. Given the very incomplete exposure of the caldera, no estimates of area or volume are possible.

Monotony Tuff from Central Nevada (27.6 Ma)

The 27.58 ± 0.05 Ma (n = 2, Supplemental Table 1 [see footnote 1]; 27.57 ± 0.04 Ma [n = 9], Best et al., 2013c) Monotony Tuff of the central Nevada volcanic field erupted from the

Monotony caldera in the Pancake Range (Supplemental Fig. 1 [see footnote 4]; Ekren et al., 1974; Best et al., 1989) and spread mostly eastward but also westward at least to the Candelaria Hills, where it was mapped as the 27.65 ± 0.05 Ma tuff of Miller Mountain (Petronis et al., 2009; Page, 1959; Speed and Cogbill, 1979; Stewart, 1979). Correlation is based on similarity in phenocryst assemblage and abundance, and compositions of biotite and hornblende phenocrysts (Best et al., 2013c, their figure 33). Ash flows of the Monotony Tuff also spread southwestward into what is now a restricted military range (Ekren et al., 1971). Fridrich and Thompson (2011) identified a tuff in the Funeral Mountains in California at about 36°37'N, 116°45'W, as Monotony based on petrographic characteristics. The voluminous (~4700 km³; Best et al., 2013c), crystal-rich Monotony Tuff is dacite to low-silica rhyolite that contains mostly plagioclase and biotite, with minor sanidine and quartz. It is the only tuff erupted from the central Nevada volcanic field known to have spread widely into western Nevada.

27.4–26.85 Ma

Numerous voluminous rhyolitic to trachydacitic ash-flow tuffs erupted from the Toquima caldera complex between 27.4 and 26.85 Ma in one of the most intense outpourings of tuff in the western Nevada volcanic field (Figs. 3, 10, and 12). The younger calderas are well exposed, but these calderas and Basin and Range faulting largely overprint the older calderas. As with the tuffs of the 31.5–28.8 Ma episode, these tuffs crop out far to the west. However, because the caldera sources were farther south, their outflow tuffs are similarly to the south. Outflow tuffs of the two episodes mostly do not overlap.

Figure 13 shows our proposed correlation of tuffs from the Toquima caldera complex to the westernmost outflow. Uncertainties remain, and some proposed correlations undoubtedly will prove incorrect. Many of the tuffs of this episode were mapped in considerable detail in the Toquima caldera complex (Boden, 1992; Shawe et al., 2002), in the Gillis and Gabbs Valley Ranges (Hardyman, 1980; Ekren and Byers, 1985, 1986a, 1986b, 1986c; Ekren et al., 1980), and from the Wassuk Range westward to the Singatse Range (Proffett and Proffett, 1976; Bingler, 1977; Proffett, 1977; Proffett and Dilles, 1984; Dilles and Gans, 1995) to westernmost Nevada (Bingler, 1978a, 1978b). The outflow tuffs were relatively well correlated in these early works, but their sources in the Toquima caldera complex have only recently been recognized and understanding of the complex is still evolving (Henry and Faulds, 2010). The

⁴Supplemental Figure 1. Distributions of the tuff of Campbell Creek, Monotony Tuff (only its western extent; see Best et al. [2013c] for full extent), Nine Hill Tuff, and Underdown Tuff—tuff of Clipper Gap and their known or suspected calderas. These tuffs have wider distributions than other tuffs, and all crossed the paleodivides proposed by Best et al. (2013a) and Henry et al. (2012a). If you are viewing the PDF of this paper or reading it offline, please visit <http://dx.doi.org/10.1130/GES00867.S4> or the full-text article on www.gsapubs.org to view Supplemental Figure 1.

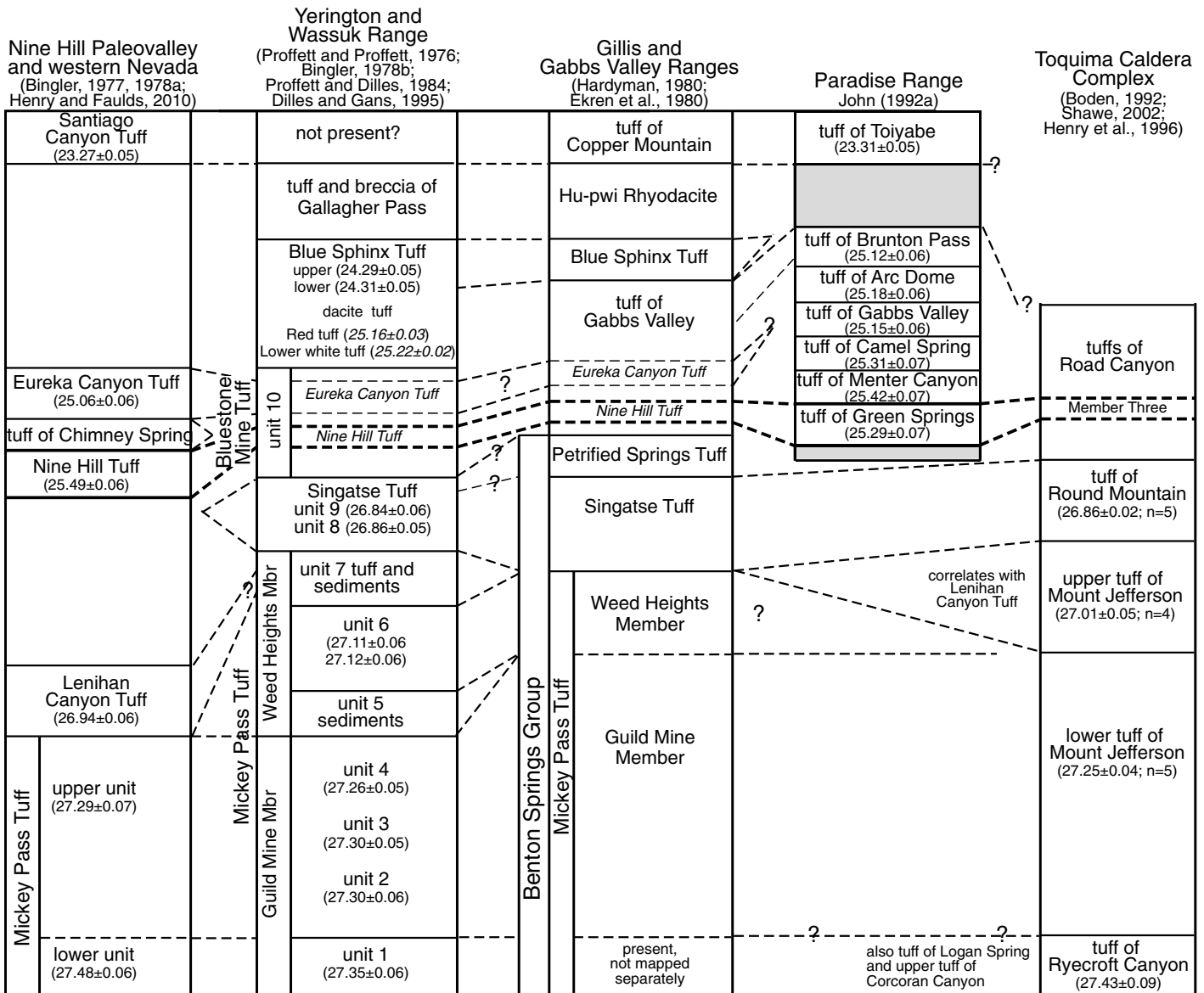


Figure 13. Proposed and possible correlations of ash-flow tuffs between western Nevada and the Toquima caldera complex. See text for comprehensive discussion. ⁴⁰Ar/³⁹Ar dates in regular type from Henry et al. (1996), Henry and Faulds (2010), and this study. ⁴⁰Ar/³⁹Ar dates in italics from Dilles and Gans (1995).

number, distribution, and sources of tuffs in the Toquima caldera complex remain incompletely understood. Further mapping, chemical analyses, and paleomagnetic data to test correlation are needed.

The Toquima caldera complex was independently mapped in two monumental efforts by Boden (1986, 1992) and D.R. Shawe and colleagues (Shawe, 1995, 1998, 1999a, 1999b, 2002; Shawe and Byers, 1999; Shawe et al., 2000). Named units are almost identical between the two projects, but their maps differ considerably in many areas. Our work (Henry et al., 1996, 1997; Henry, 1997; this study) builds upon

and variably revises their studies. However, volcanic stratigraphy, structure, and evolution in the Toquima Range are complex, and this summary is far from the last word.

Tuffs of Ryecroft Canyon, Corcoran Canyon, and Logan Spring (27.4 Ma)

The oldest tuffs of the Toquima caldera complex crop out in small areas that are mostly remnants of their older calderas (Fig. 12). The lower tuff of Ryecroft Canyon (Boden, 1992; Shawe and Byers, 1999) has sanidine ⁴⁰Ar/³⁹Ar ages of 27.43 ± 0.09 Ma (this study) and 27.48 ± 0.03 Ma (Shawe and Byers, 1999). Shawe et al.

(2000) reported an age of 27.52 ± 0.05 Ma on the upper tuff of Corcoran Canyon. Both tuffs are thick, megabreccia-containing units in the southeastern part of the complex, almost certainly in caldera remnants. Shawe and Byers (1999) and Shawe et al. (2000) reasonably interpreted a Ryecroft Canyon caldera to extend beneath Monitor Valley immediately east of these outcrops (Fig. 12). The 27.38 ± 0.08 Ma tuff of Logan Spring is a thin, presumably outflow deposit that underlies the lower tuff of Mount Jefferson in the northeastern part of the complex (Boden, 1992; this study). All three tuffs are petrographically similar, crystal-rich, low-silica

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rhyolites or trachydacites containing abundant plagioclase and biotite, with lesser sanidine, quartz, and other mafic minerals (Table 2). None of the three tuffs occur together in a stratigraphic section, so all could be the same tuff.

We interpret the lowermost part of the Mickey Pass Tuff of western Nevada (Gillis, Gabbs Valley, Singatse, and Virginia Ranges) to be outflow tuff correlative with one or more of the Toquima plagioclase-biotite tuffs (Figs. 10 and 13). This correlation is based on the indistinguishable age, sanidine K/Ca, and plagioclase- and biotite-rich, quartz- and sanidine-poor phenocryst assemblage of all the tuffs. Given the very incomplete exposure of the caldera(s), no estimate of area is possible. However, the great outflow distribution suggests at least one eruption was voluminous.

Lower Tuff of Mount Jefferson and Caldera (27.3 Ma)

What we term the lower tuff of Mount Jefferson and its lower Mount Jefferson caldera are the most voluminous tuff and largest caldera of the Toquima caldera complex (Fig. 12; Table 1). Our lower tuff of Mount Jefferson combines the tuff of Moores Creek, the lower tuff of Mount Jefferson, the lower tuff of Trail Canyon, and the upper tuff of Rycroft Canyon of Boden (1992) and parts of the tuffs of Rycroft Canyon and Antone Canyon and megabreccia of Meadow Canyon and possibly part of the principal member of the tuff of Mount Jefferson of Shawe (1999b; Shawe et al., 2000). This correlation is based on distribution of thick (≥ 1300 m), petrographically similar, breccia-rich, intracaldera tuff with indistinguishable ages (27.25 ± 0.04 Ma, $n = 5$) and association with rhyolite domes, also with indistinguishable ages (27.30 ± 0.01 Ma, $n = 4$), along the caldera margin (Fig. 12). The domes were emplaced almost immediately after ash-flow eruption and caldera collapse and presumably tapped the same magma chamber that erupted tuff.

The lower tuff of Mount Jefferson contains ~25% phenocrysts dominated by quartz and sanidine with lesser plagioclase and biotite (Table 2). It is high-silica rhyolite that is not obviously compositionally zoned but basal parts are not exposed and exposed sections have not been sufficiently examined to evaluate zoning. The lower tuff of Mount Jefferson is distinct in phenocryst assemblage, composition, and age from the older plagioclase-biotite tuffs and the younger upper tuff of Mount Jefferson.

The main part of the Guild Mine Member of the Mickey Pass Tuff of western Nevada is probably the outflow equivalent of the lower tuff of Mount Jefferson (Figs. 10 and 13). Correlation is again based on indistinguishable age ($27.28 \pm$

0.02 Ma, $n = 5$), sanidine K/Ca, and phenocryst assemblage of the quartz- and sanidine-rich outflow tuff. Chemical analyses and paleomagnetic data to test the correlation would be useful. The caldera has an area of ~ 400 km². With an exposed thickness of 1300 m of intracaldera tuff, the erupted volume is at least 520 km³. Total caldera collapse of 3–4 km, typical of calderas in Nevada and the western U.S. (Lipman, 1984, 2007; John, 1995; John et al., 2008), suggests a volume of 1200–1600 km³. This large volume is consistent with the great extent and thickness of outflow tuff (Fig. 10).

Weed Heights Member of the Mickey Pass Tuff (27.1 Ma)

The Weed Heights Member is outflow tuff recognized only in the Singatse, Wassuk, Gillis, and Gabbs Valley Ranges (Proffett and Proffett, 1976; Ekren et al., 1980; Proffett and Dilles, 1984). Ages from near the base and near the top in the Singatse Range are 27.12 ± 0.06 and 27.11 ± 0.06 Ma, respectively (this study). The Weed Heights Member is as much as 180 m thick in the Singatse Range and 120 m thick in the Gillis Range. The tuff is apparently reversely zoned, as indicated by both phenocryst assemblage and three chemical analyses, from low-silica rhyolite at the base to high-silica rhyolite near the top (Proffett and Proffett, 1976; Ekren et al., 1980). The source is unknown but could be in the Toquima Range as indicated by its temporal tie to other Toquima units (Fig. 13).

Upper Tuff of Mount Jefferson and Caldera (27.0 Ma)

The upper tuff of Mount Jefferson encompasses the upper and capping members of the tuff of Mount Jefferson of Boden (1986, 1992) and the principal member of the tuff of Mount Jefferson of Shawe (1999b, 2002) and Shawe et al. (2000). Correlation is based on similar zoned phenocryst assemblage, composition, age (27.01 ± 0.05 Ma, $n = 4$), and association with rhyolite domes of indistinguishable age (26.95 ± 0.03 Ma, $n = 5$) along its caldera margin (Fig. 12). The upper tuff of Mount Jefferson is irregularly zoned upward from less to more abundantly porphyritic, with increases in plagioclase, biotite, hornblende, and clinopyroxene and decreases in quartz and sanidine, and from rhyolite with 75% SiO₂ to trachydacite with 67% SiO₂ (Table 2; Boden, 1986, 1987, 1994). The uppermost parts of the upper tuff of Mount Jefferson (capping members of Boden, 1992) consist of several cooling units separated by tuffaceous sedimentary deposits. The implied change from continuous to apparently small, discontinuous eruptions is characteristic of

many calderas in the western Nevada volcanic field (Henry, 1996a, 1996b; John et al., 2008).

We interpret the Lenihan Canyon Tuff of westernmost Nevada to be the correlative outflow tuff (Fig. 13). Correlation is based on a similar phenocryst assemblage, age (26.94 ± 0.06 Ma, $n = 1$), and sanidine K/Ca. However, Lenihan Canyon Tuff has not been recognized elsewhere over a distance of ~ 200 km between western Nevada and the Toquima Range (Fig. 10). About half of the intervening area has been mapped in sufficient detail that the tuff might be recognized, but much of the mapping combined many tuffs into units with local names (Fig. 12). For example, the tuffs of Royston Hills and Cedar Mountains in the Royston Hills are phenocryst-rich rhyolites with K-Ar ages between 29 and 26 Ma (Whitebread and Hardyman, 1987). These tuffs could correlate with almost any of the tuffs of the Toquima caldera complex. In the Paradise Range, the 25.4 Ma Nine Hill Tuff, there called the tuff of Green Springs, is the oldest exposed Tertiary volcanic unit in the paleovalley that crosses the central part of the range (Fig. 13; John, 1992a).

The caldera for the upper tuff of Mount Jefferson is irregularly nested inside the lower Mount Jefferson caldera (Fig. 12). Total area is ~ 200 km². Cross sections by Boden (1992) indicate a probable intracaldera thickness of 1.5 km, so a minimum erupted volume of 300 km³.

Tuff of Round Mountain, Correlative Singatse Tuff, and Caldera (26.9 Ma)

The tuff of Round Mountain fills the eastern part of its caldera in the southwestern part of the Toquima caldera complex (Fig. 12). The tuff hosts one of the world's largest epithermal gold-silver deposits (Ferguson, 1921; Tingley and Berger, 1985; Shawe et al., 1986; Mills et al., 1988; Sander and Einaudi, 1990; Henry et al., 1996). Mining and drilling reveal two cooling units, a lower one more than 600 m thick that wedges out abruptly against pre-Cenozoic rocks against the eastern caldera topographic wall, and an upper one ~ 390 m thick that laps onto the caldera wall; both contain megabreccia (Boden, 1992; Henry et al., 1996). Phenocrysts in both units consist of subequal amounts of plagioclase, sanidine, and quartz, with lesser biotite, Fe-Ti oxides, zircon, apatite, and, locally, hornblende. The upper unit is zoned upward in phenocryst abundance, which increases from $\sim 10\%$ to 40%, and from high-silica rhyolite (77 wt% SiO₂) at its base to rhyolite (73 wt% SiO₂) at its top (Boden, 1986). ⁴⁰Ar/³⁹Ar dates on sanidine from both units are indistinguishable and average 26.86 ± 0.02 Ma ($n = 5$).

We interpret the Singatse Tuff, which is widespread in western Nevada (Singatse to

Gabbs Valley Ranges), to be correlative outflow tuff (Figs. 10 and 13). Correlation is based on similar phenocryst assemblage and identical $^{40}\text{Ar}/^{39}\text{Ar}$ ages on two samples from the northern Wassuk Range (26.84 ± 0.06 Ma on a lower cooling unit, and 26.86 ± 0.05 Ma on the main upper cooling unit). The only chemical analysis of Singatse Tuff, from the Singatse Range, has 69% SiO_2 (Proffett and Proffett, 1976), less silicic than analyses of the upper unit at Round Mountain, which may indicate that less-evolved parts were not preserved or sampled in the Toquima Range. The Petrified Spring Tuff, a local unit in the Gabbs Valley and Gillis Ranges (Ekren et al., 1980), may be an upper cooling unit of the Singatse Tuff, equivalent to the upper cooling unit east of Yerington. The Petrified Spring Tuff everywhere occurs with and almost everywhere rests on Singatse Tuff (Fig. 10), but intermediate lavas separate the two in a few places. The Petrified Spring Tuff has a similar phenocryst assemblage as the Singatse Tuff, but neither an age nor a chemical analysis is available.

So little of the Round Mountain caldera is exposed that no estimates of size or volume are possible. Proffett and Proffett (1976) reported a volume of 3500 km^3 for the Singatse Tuff but provided no supporting data. Regardless, the great outflow distance suggests a moderately voluminous tuff.

25.4–23.3 Ma

Ash-flow eruption and caldera collapse resumed in the western Nevada volcanic field at 25.4 Ma, following a hiatus of ~ 1.4 Ma, and proceeded essentially continuously to 23.3 Ma (Fig. 3). Calderas formed along the entire northwest-southeast length of the belt and, with one exception, are along its southwestern edge (Fig. 11).

Nine Hill Tuff (25.4 Ma)

The 25.38 ± 0.08 Ma ($n = 12$; Supplemental Table 1 [see footnote 1]) Nine Hill Tuff, the most widespread tuff of the Great Basin and one of the world's most widely distributed tuffs, led off this episode. The Nine Hill Tuff crops out from near Ely on the east to the western foothills of the Sierra Nevada on the west (Deino, 1989; Best et al., 1989; Fig. 11; Supplemental Fig. 1 [see footnote 4]). The Nine Hill Tuff consists of two petrographic and chemical phases, a lower, sparsely porphyritic high-silica rhyolite that defines its full distribution and an upper, moderately porphyritic low-silica rhyolite that is restricted to the central part (Table 2). The two phases were probably emplaced almost simulta-

neously because they form a single cooling unit in most places. Also, the lower, sparsely porphyritic phase commonly contains abundantly porphyritic pumice. However, both phases have a basal vitrophyre, indicating a slight cooling break, at Haskell Peak, a distal location in the eastern Sierra Nevada (Brooks et al., 2003, 2008). The Nine Hill Tuff is unusual among middle Cenozoic (36–18 Ma) cooling units in the Great Basin, because it contains phenocrysts of three feldspars—plagioclase, sanidine, and anorthoclase (Deino, 1985). Its high concentrations of both Zr and Nb are also unusual for tuffs in the Great Basin ignimbrite province. Deino (1985, 1989) demonstrated that the Nine Hill Tuff and the D unit of the Bates Mountain Tuff (Sargent and McKee, 1969; Gromme et al., 1972) in central Nevada are the same, based on distinctive composition, age, and paleomagnetic direction. Because further data have strengthened the correlation of these two units, the formal Nine Hill Tuff name (Bingler, 1978a) should be used. Its source has not been verified but is interpreted to be concealed beneath the Carson Sink based on its distribution, thickness, and facies variations (Deino, 1985; Best et al., 1989).

New Pass Tuff and Correlative Tuffs of Poco Canyon and Chimney Spring, and Poco Canyon Caldera (25.3 Ma)

The 25.27 ± 0.05 Ma ($n = 7$) New Pass Tuff is a petrographically distinctive, high-silica rhyolite that is widespread in western Nevada (Fig. 11). The tuff was recognized and named from New Pass in the northern part of the Desatoya Mountains (McKee and Stewart, 1971; Stewart et al., 1977). Deino (1989) correlated the tuff of Chimney Spring of westernmost Nevada with the New Pass Tuff, and John (1992b, 1995) correlated both with the upper tuff of Poco Canyon, the intracaldera tuff in the Poco Canyon caldera in the Stillwater Range. Intracaldera tuff of Poco Canyon is as much as 2500 m thick and is underlain by a megabreccia unit of high-silica rhyolite (the megabreccia of Government Trail Canyon) up to 1800 m thick that is inferred to be genetically related to the tuff of Poco Canyon (John, 1995; see section below on Caetano and Stillwater calderas). Correlation of the three named tuff units is based on distinctive phenocryst assemblage (abundant adularose sanidine and smoky quartz), composition (74–78% SiO_2), $^{40}\text{Ar}/^{39}\text{Ar}$ age, sanidine K/Ca, and paleomagnetic direction (Deino, 1989; John, 1995; Hudson et al., 2000; Henry and Faulds, 2010). Five $^{40}\text{Ar}/^{39}\text{Ar}$ dates in western Nevada and the eastern Sierra Nevada average 25.26 ± 0.05 Ma (Brooks et al., 2008; Henry and Faulds, 2010).

A date on the upper tuff of Poco Canyon is 25.26 ± 0.07 Ma, and on the New Pass Tuff at New Pass is 25.32 ± 0.06 Ma. The New Pass Tuff nicely illustrates the distinctly elliptical distribution, preferentially west of the caldera, of many western Nevada tuffs (Fig. 11).

Tuff of Elevenmile Canyon and Caldera (25.1 Ma)

The phenocryst-rich tuff of Elevenmile Canyon erupted from the Elevenmile Canyon caldera in the southern Stillwater Range (John, 1995). Additional mapping since 1995 indicates that the Elevenmile Canyon caldera extends eastward into the southern Clan Alpine Mountains, thus making it one of the largest calderas in the western Nevada volcanic field (Fig. 11; John, 1997; this study). Intracaldera tuff is more than 3400 m thick in the Stillwater Range and as much as 2000 m thick with abundant megabreccia in the southwestern Clan Alpine Mountains. The tuff of Lee Canyon, which overlies the tuff of Elevenmile Canyon in the Stillwater Range, and the tuff of Hercules Canyon, which overlies Elevenmile Canyon in the Clan Alpine Mountains, are late eruptions probably of the Elevenmile Canyon magma system (John, 1995, 1997). All three tuffs are petrographically similar (Table 2). The volume of erupted tuff from the Elevenmile Canyon caldera is probably $\geq 1050\text{--}1400 \text{ km}^3$.

Intracaldera tuff of Elevenmile Canyon in the southern Clan Alpine Mountains is 25.07 ± 0.06 Ma, and the tuff of Hercules Canyon there is 25.05 ± 0.06 Ma. The tuff of Elevenmile Canyon in the Stillwater Range is altered and undated, but the overlying tuff of Lee Canyon is 25.00 ± 0.06 Ma.

A major stratigraphic and $^{40}\text{Ar}/^{39}\text{Ar}$ age conundrum remains about the New Pass Tuff and tuff of Elevenmile Canyon. In the Stillwater Range, the sequence consists of lower tuff of Poco Canyon, tuff of Elevenmile Canyon, megabreccia of Government Trail Canyon, upper tuff of Poco Canyon (New Pass Tuff), and tuff of Lee Canyon (John, 1995). At New Pass, the New Pass Tuff consists of a single cooling unit that is overlain by an undated tuff that is petrographically similar to the tuffs of Elevenmile Canyon, Lee Canyon, and Hercules Canyon. In western Nevada, the tuff of Chimney Spring, correlative with the New Pass Tuff, is also a single cooling unit and is overlain by the tuff of Painted Hills. Henry et al. (2004) interpreted the tuff of Painted Hills to correlate with the tuff of Elevenmile Canyon on the basis of similar phenocryst assemblage, $^{40}\text{Ar}/^{39}\text{Ar}$ dates (25.05 ± 0.06 and 25.06 ± 0.06 Ma, on two samples of tuff of Painted Hills), and sanidine K/Ca. Alternatively, the tuff of Painted Hills could

correlate with either of the tuffs of Lee Canyon or Hercules Canyon. The apparent contradiction between field relations and $^{40}\text{Ar}/^{39}\text{Ar}$ ages indicates the need for further study.

Eureka Canyon Tuff, Bluestone Mine Tuff, Tuff of Gabbs Valley, Paradise Range Tuffs, and Railroad Ridge Caldera (25.4–25.1 Ma)

A group of sparsely porphyritic rhyolite tuffs with similar ages (ca. 25.4–25.1 Ma) and phenocryst assemblages crops out extensively from western Nevada to the Paradise Range and Shoshone Mountains (Figs. 3 and 11). From west to east, these include the Eureka Canyon Tuff (Bingler, 1978a; Henry and Faulds, 2010), tuffs in the heterogeneous Bluestone Mine Tuff near Yerington (Proffett and Proffett, 1976; Proffett and Dilles, 1984), tuffs in the heterogeneous tuff of Gabbs Valley (Ekren et al., 1980), the tuffs of Menter Canyon, Camel Spring, Gabbs Valley, and Brunton Pass in the Paradise Range (John, 1992a; Fig. 13), and the tuff of Union Canyon in the Shoshone Mountains (Whitebread et al., 1988). Specific correlations between these tuffs are incompletely resolved. The Nine Hill Tuff is part of this suite, and its presence in the Bluestone Mine Tuff, in the tuff of Gabbs Valley, and as the tuff of Green Springs in the Paradise Range is now certain. The Eureka Canyon Tuff is definitely present throughout much of the region, but its greater phenocryst abundance makes correlation with any of the tuffs in the Paradise Range uncertain. Henry and Faulds (2010) suggested correlation of the Eureka Canyon Tuff with the Railroad Ridge caldera in the Clan Alpine Mountains, but the differences in distribution and phenocryst abundance make this correlation unlikely, despite the age match. If none of these tuffs correlate with the tuff of Railroad Ridge, then that caldera has no recognized outflow tuff. The distribution of this group of tuffs suggests source calderas in or around the Toiyabe Range.

Tuff of Arc Dome and Caldera (25.2 Ma)

The 25.18 ± 0.06 Ma, crystal-rich, rhyolitic tuff of Arc Dome is characterized by distinctive bipyramidal smoky quartz phenocrysts. The tuff is normally zoned from high- to low-silica rhyolite (Table 2; John, 1992a). The tuff of Arc Dome blankets much of the area around an interpreted caldera buried beneath the south end of Reese River Valley between the Toiyabe Range and Shoshone Mountains (Fig. 11; Brem et al., 1991; John, 1992a). However, a sample petrographically and compositionally indistinguishable from the tuff of Arc Dome (John, 1992a) from the large outcrop area in the northern Shoshone Mountains gives distinctly younger $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 24.72 ± 0.05

and 24.70 ± 0.05 Ma, determined at the New Mexico Tech and Berkeley Geochronology labs (A.L. Deino, 2012, written commun.), respectively. The match in ages demonstrates the age discrepancy is not analytical and that the two tuffs are of different age. The ages are also consistent with stratigraphic relations in the northern Shoshone Mountains as the younger tuff overlies the 25.0 Ma Underdown Tuff. The Arc Dome source caldera is located on the basis of increasing unit thickness, lithologic diversity, hydrothermal alteration, a negative gravity anomaly, and the presence of nearby dikes and rhyolite domes (Brem et al., 1991). The tuff of Arc Dome was probably moderately voluminous, but its limited westward distribution suggests a smaller volume than many other tuffs.

Underdown Tuff and Caldera and Tuff of Clipper Gap (25.0 Ma)

The 24.96 ± 0.05 Ma (Best et al., 2013c), sparsely porphyritic Underdown Tuff is recognized only as an intracaldera deposit along the eastern edge of the Shoshone Mountains (Fig. 11; Bonham, 1970). However, several outflow tuffs are probable or possible equivalents. Based on analyses of three intracaldera samples (Supplemental Table 3 [see footnote 3]), the Underdown Tuff is high-silica rhyolite with distinctive high Y (36–42 ppm) and Nb (28–31 ppm) contents. The tuff of Clipper Gap, widely distributed east of the caldera (Fig. 11; Supplemental Fig. 1 [see footnote 4]), is probably correlative outflow tuff based on similar age (24.87 ± 0.10 Ma, Supplemental Table 1 [see footnote 1]; 24.95 ± 0.07 Ma, Best et al., 2013c), phenocryst assemblage, and chemical composition, including unusually high Nb (Best et al., 2013c; this study). The “red tuff” of John (1992a) in the Paradise Range may also be correlative based on similar phenocryst assemblage and chemical composition, especially high Y (37 ppm) and Nb (35 ppm). Two other tuffs have similar sparse phenocryst assemblages, have high Y and Nb, and are in reasonable stratigraphic positions but have reported $^{40}\text{Ar}/^{39}\text{Ar}$ ages that appear slightly too old. Dilles and Gans (1995) tentatively correlated a 25.15 ± 0.03 Ma tuff with high Y (34 and 45 ppm) and Nb (35 and 36 ppm) in the Wassuk Range with the “red tuff”. The 25.12 ± 0.03 Ma tuff of Brunton Pass in the Paradise Range also has high Y (42 and 63 ppm) and Nb (34 and 35 ppm). Further work is needed to fully evaluate possible correlations. The combined Underdown Tuff–tuff of Clipper Gap is younger than but appears to be a continuation of the 25.4–25.1 Ma pulse of sparsely porphyritic tuffs (Fig. 3).

Probable intracaldera Underdown Tuff is exposed for ~23 km along the eastern range

front of the Shoshone Mountains, suggesting the caldera is at least that wide. The Bonita Canyon Formation, a >500-m-thick sequence of epiclastic rocks interbedded with thin air-fall and ash-flow tuffs and local megabreccia blocks of pre-Tertiary rocks, overlies the Underdown Tuff, extends 7 km farther north, and lithologically resembles caldera-fill sequences elsewhere, possibly suggesting an even larger caldera.

Fish Creek Mountains Tuff and Caldera (24.9 Ma)

The 24.91 ± 0.03 Ma ($n = 3$) Fish Creek Mountains Tuff crops out mostly inside its caldera in the northeastern part of the western Nevada field, where it is anomalously young compared to surrounding magmatism (McKee, 1970; Cousens et al., 2011; Figs. 4, 5, and 11). Outflow tuff has been found in a paleovalley ~10 km west of the caldera (Gonsior and Dilles, 2008). The abundantly porphyritic, high-silica rhyolite contains abundant smoky quartz and sanidine, minor plagioclase, and a trace of biotite (Table 2). The caldera is distinctly non-resurgent; gently inward-dipping tuffaceous sediments occupy an intracaldera basin. The caldera is also remarkably unutilized and therefore unextended, occupying a domain of low extension separating two highly extended domains to the west and east (Colgan et al., 2008; Fosdick and Colgan, 2008; Gonsior and Dilles, 2008; Colgan and Henry, 2009; see section on the Caetano caldera). McKee (1970) estimated a total volume of 310 km^3 .

Round Rock Formation and Manhattan Caldera (24.8 Ma)

The 24.80 ± 0.14 Ma Round Rock Formation and its source Manhattan caldera are the youngest ignimbrite and caldera identified in the Toiyabe Range (Fig. 11; Shawe, 1999a). The tuff is normally zoned rhyolite containing sparse phenocrysts of plagioclase, quartz, sanidine, biotite, and hornblende that increase in abundance upward (Shawe, 1999a). No correlative outflow facies tuff has been identified. Geophysical modeling indicates the intracaldera ignimbrite is 1.0–1.4 km thick in the 11×17 km caldera (Shawe and Snyder, 1988), which indicates a minimum tuff volume of ~135 km³.

Tuff of Desatoya Peak and Caldera (24.7 Ma)

The tuff of Desatoya Peak is a crystal-rich rhyolite known only from its own caldera in the Desatoya Mountains (Fig. 11). The tuff consists of two, petrographically similar cooling units separated by up to 100 m of air-fall tuff, tuffaceous sedimentary rocks, and lava flows (McKee and Conrad, 1987). The upper unit is at least 600 m thick, and the lower unit greater than

400 m with no base exposed. McKee and Conrad (1987) reported a partial chemical analysis that indicates the tuff is rhyolite with ~71% SiO₂. The lower unit has a sanidine ⁴⁰Ar/³⁹Ar date of 24.74 ± 0.06 Ma. Reconnaissance observations suggest a ~12 × 17 km caldera that has an area of 160 km² (Henry et al., 2012a). Minimum volume based on the 1.2 km thickness of intracaldera deposits is ~190 km³.

Tuff of Peavine Canyon and Peavine Caldera (24.5 Ma)

The 24.53 ± 0.08 Ma tuff of Peavine Canyon crops out only(?) in the relatively unstudied Peavine caldera of the southeastern Toiyabe Range (Fig. 11; Brem and Snyder, 1983; Brem et al., 1991). The sparsely porphyritic tuff is distinctive in having anorthoclase rather than sanidine phenocrysts (Table 2). Tuff that crops out in the Manhattan caldera that was considered possible outflow tuff of Peavine Canyon is petrographically dissimilar to the tuff of Peavine Canyon and is therefore not correlative (Shawe, 1999a). The Peavine caldera is partly covered by the younger Toiyabe caldera but covers an area of at least 215 km² (Table 1).

Blue Sphinx Tuff (24.3 Ma) and Possibly Correlative Tuff of Goldyke

The 24.30 ± 0.02 Ma (n = 2) abundantly porphyritic Blue Sphinx Tuff crops out in the Wassuk, Gillis, and Gabbs Valley Ranges (Fig. 11; Bingler, 1978b; Ekren et al., 1980; Hardyman, 1980; Proffett and Dilles, 1984). The tuff is distinctive for its variety of phenocryst phases and especially for its intensely resorbed quartz and plagioclase phenocrysts (Table 2). The possibly correlative tuff of Goldyke, which crops out south of the Paradise Range, is petrographically similar and has 23.0 ± 0.9 and 23.6 ± 0.7 Ma K-Ar biotite ages (John, 1992a). However, the tuff of Goldyke underlies some minor tuffs that may correlate with two tuffs in the Wassuk Range that are dated at around 25.2 Ma (John, 1992a; Dilles and Gans, 1995). The tuff of Goldyke has 74–75% SiO₂; no analyses are available for the Blue Sphinx Tuff. Unlike the previous calderas with no known outflow, the Blue Sphinx and Goldyke are outflow tuffs with no known source caldera, which is likely to be near the Toiyabe Range on the basis of their distribution.

Tuffs of Sheep Canyon and Ione Canyon (ca. 24 Ma)

The tuffs of Sheep Canyon and Ione Canyon comprise multiple cooling units of crystal-rich rhyolite, dacite, and trachydacite in the southern Paradise Range and central Shoshone Mountains, respectively. The ca. 24 Ma tuff of Sheep

Canyon consists of four cooling units as much as 600 m thick, contains abundant, coarsely porphyritic biotite-rich pumice, and is noteworthy for its variable and relatively mafic composition (Fig. 6; John, 1992a). The possibly correlative tuff of Ione Canyon consists of two cooling units several hundred meters thick that are megascopically identical to the tuff of Sheep Canyon (Whitebread et al., 1988). The source(s) of these tuffs is unknown.

Santiago Canyon Tuff and Correlative Tuffs of Toiyabe and Copper Mountain, and Toiyabe Caldera (23.3 Ma)

The Santiago Canyon Tuff is the youngest widespread ash-flow tuff of the ignimbrite flareup (Fig. 11). The abundantly porphyritic tuff is distinctive in having abundant sphene in addition to the usual assemblage of quartz, sanidine, plagioclase, biotite, and hornblende (Table 2). The presence of sphene and its young age (23.27 ± 0.05 Ma) provide certain correlation with the tuff of Copper Mountain in the Gillis and Gabbs Valley Ranges (Ekren et al., 1980; Hardyman, 1980) and the tuff of Toiyabe (23.31 ± 0.05 Ma) in the Paradise and Toiyabe Ranges (Brem et al., 1991; John, 1992a; Henry and Faulds, 2010). Outflow tuff of Toiyabe is normally zoned from 75% to 72% SiO₂ (Table 2; John, 1992a). The source caldera is in the southern Toiyabe Range, where reconnaissance mapping identified thick intracaldera tuff of Toiyabe, part of a caldera wall, and several ash-flow tuff feeder vents (Whitebread et al., 1988; Brem et al., 1991; John, 1992a). The Santiago Canyon Tuff has not been recognized in the Singatse or Wassuk Ranges (Proffett and Dilles, 1984; Bingler, 1978b).

Tuff of Fairview Peak and Caldera (19.5 Ma)

The 11 × 18 km Fairview Peak caldera (Henry, 1996a, 1996b), source of the 19.48 ± 0.03 Ma tuff of Fairview Peak, is the youngest caldera in the western Nevada volcanic field (Figs. 3, 4, and 11) and is almost 4 Ma younger than the next-youngest caldera. The sparsely porphyritic, high-silica rhyolite contains phenocrysts of plagioclase, sanidine, quartz, and biotite. All tuff is either intracaldera, where it is greater than 700 m thick, or no more than ~5 km outside the caldera topographic wall. An east-striking dike of vertically foliated tuff of Fairview Peak that cuts through subhorizontal tuff near the approximate northeastern margin of the caldera is a probable ash-flow vent (39.2054°N, 118.0469°W). The dike is 200 m long, up to 20 m wide, and has a vitrophyric margin and a devitrified core. Eruption of the tuff of Fair-

view Peak was followed ~0.2–0.4 Ma later by emplacement of a series of rhyolite to dacite lava domes around the ring fracture of the caldera. The caldera has an area of ~160 km²; minimum erupted volume is ~115 km³ and could be several times that. Despite their young age, Fairview Peak caldera rocks have an arc geochemical signature and are petrographically and compositionally like older tuffs.

Previously Interpreted but Doubtful Calderas

Ash-Flow Tuffs of the Candelaria Hills

A sequence of twelve, ca. 27.6–23.3 Ma ash-flow cooling units, individually ranging from about 20 to 130 m thick, in the Candelaria Hills has been interpreted (1) to have been erupted from an east-trending, fault-bounded trough that was subsiding incrementally during the eruptions, possibly as a result of magma withdrawal, i.e., a “slow caldera” (*our term*) (Robinson and Stewart, 1984) or (2) to fill an east-north-east–elongated tectonic trough resulting from approximately north-south extension, i.e., a volcano-tectonic trough (Speed and Cogbill, 1979; Petronis et al., 2004; Petronis and Geissman, 2009). A caldera is inconsistent with the lack of any observable caldera-like structures, especially the thin deposits that lack megabreccia, and an east-northeast–elongated tectonic trough is inconsistent with the lack of a mappable structure of that orientation in the Candelaria Hills and of similarly oriented extensional basins anywhere else in the Great Basin (Fig. 14). On the basis of published data and our examination, we interpret the tuffs to fill approximately west-draining paleovalleys, similar to those to the north (Garside et al., 2002; Henry and Faulds, 2010; Henry et al., 2012a), and to have erupted from mostly unknown sources to the east.

Geologic maps of the area show that the tuffs filled a region with nearly 800 m of topographic relief and wedge out in depositional contact with pre-Cenozoic rocks (Fig. 14; see Stewart [1979, 1981] and Stewart et al. [1983] for wedge-out of individual tuffs). Tuff stratigraphy varies tremendously laterally, as individual tuffs pinch and swell. Most contacts between tuff and pre-Cenozoic rock were mapped as depositional. The distribution of tuffs and the geometry of contacts do not support either a caldera or an extensional basin. Attitudes of the tuffs vary tremendously; many dip away from basement rocks, probably reflecting primary dip where they compacted against slopes, e.g., north and east of Miller Mountain in the south-central part of Figure 14 (Henry and Faulds, 2010). Rock fragments in the tuffs are less than 2 cm in diameter, except in one unit where fragments

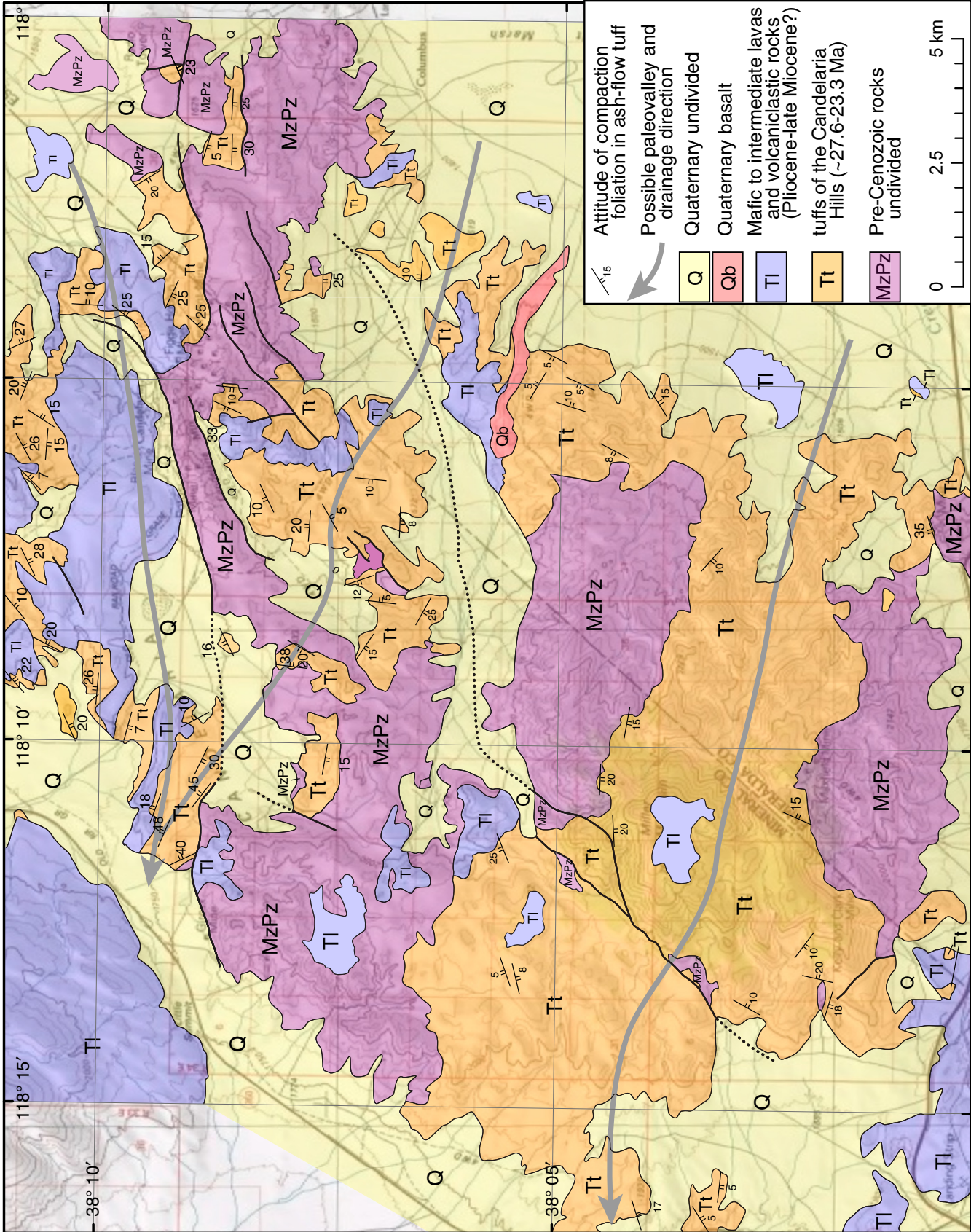


Figure 14. Geologic map of the Candelaria Hills, from Stewart (1979, 1981), Stewart et al. (1983), unpublished maps by R.C. Speed (Robert C. Speed Unpublished Geologic Maps Collection; <http://www.nbmj.unr.edu/Departments/GBSSRL/Collections.html>), and this study.

are up to 10 cm in diameter (Robinson and Stewart, 1984). Sedimentary deposits including coarse conglomerates with well-rounded pre-Cenozoic clasts up to at least 60 cm in diameter both underlie the oldest tuff and separate individual tuffs. All these characteristics seem more consistent with deposition of distal tuffs in paleovalleys, which has been widely recognized in the Great Basin (Eckberg et al., 2005; Henry and Faulds, 2010; Henry et al., 2012a).

Petronis and Geissman (2009) and Petronis et al. (2009) reported ten ⁴⁰Ar/³⁹Ar ages ranging from 27.65 ± 0.05 to 23.31 ± 0.17 Ma from the tuff sequence. Major tuffs are, from oldest to youngest, the Metallic City Tuff (25.73 ± 0.12 Ma), the Belleville Tuff (24.24 ± 0.29 Ma), and the Candelaria Junction Tuff (23.76 ± 0.13 Ma), the second youngest of the sequence.

As discussed above, the biotite-rich tuff of Miller Mountain, one of the oldest in the Candelaria Hills (27.65 ± 0.05 Ma; Petronis et al., 2009), is distal outflow Monotony Tuff from the central Nevada volcanic field. Other possible correlations are between two units of the tuff of Volcanic Hills (27.29 ± 0.16 and 27.07 ± 0.12 Ma; Petronis et al., 2009) and the 27.3 and 27.0 Ma lower and upper tuffs of Mount Jefferson in the Toquima caldera complex (Fig. 3).

Fissure Vent and Proposed Caldera in Gabbs Valley

Ekren and Byers (1976, 1986c) interpreted a caldera source for one or more of the tuffs of Gabbs Valley at Fissure Ridge in Gabbs Valley (Figs. 11, 15, and 16). Their interpretation seems to be based mostly on an interpreted ash-flow fissure vent in the western wall or boundary of a caldera. However, we find little or no evidence for a caldera and suggest the fissure vent is a narrow canyon eroded into older tuffs.

The fissure vent is an ~400-m-deep cut into two older units of the tuff of Gabbs Valley (Tgv1 and Tgv2 of Ekren and Byers, 1986c) filled with the youngest unit (Tgv3). White, poorly welded tuff up to ~3 m thick and overlying black, dense vitrophyre up to ~5 m thick are developed all along the walls of the fissure, and welding foliation parallels the walls at the walls (Figs. 15, 16A, and 16B). The vent widens upward from ~60 m to ~460 m at which point vent rock merges with initially horizontal Tgv3 outside the vent. Ekren and Byers (1986c) mapped the Cretaceous or Jurassic Granodiorite (KJg)-Tgv2 contact in the north and most tuff-tuff contacts except near the fissure as caldera-boundary faults, with down-to-the-east and/or left-lateral displacement. They further interpreted eastward thickening of Tgv1 and Tgv2 to ~300 m each ~8 km to the north along the interpreted caldera boundary to be into the caldera.

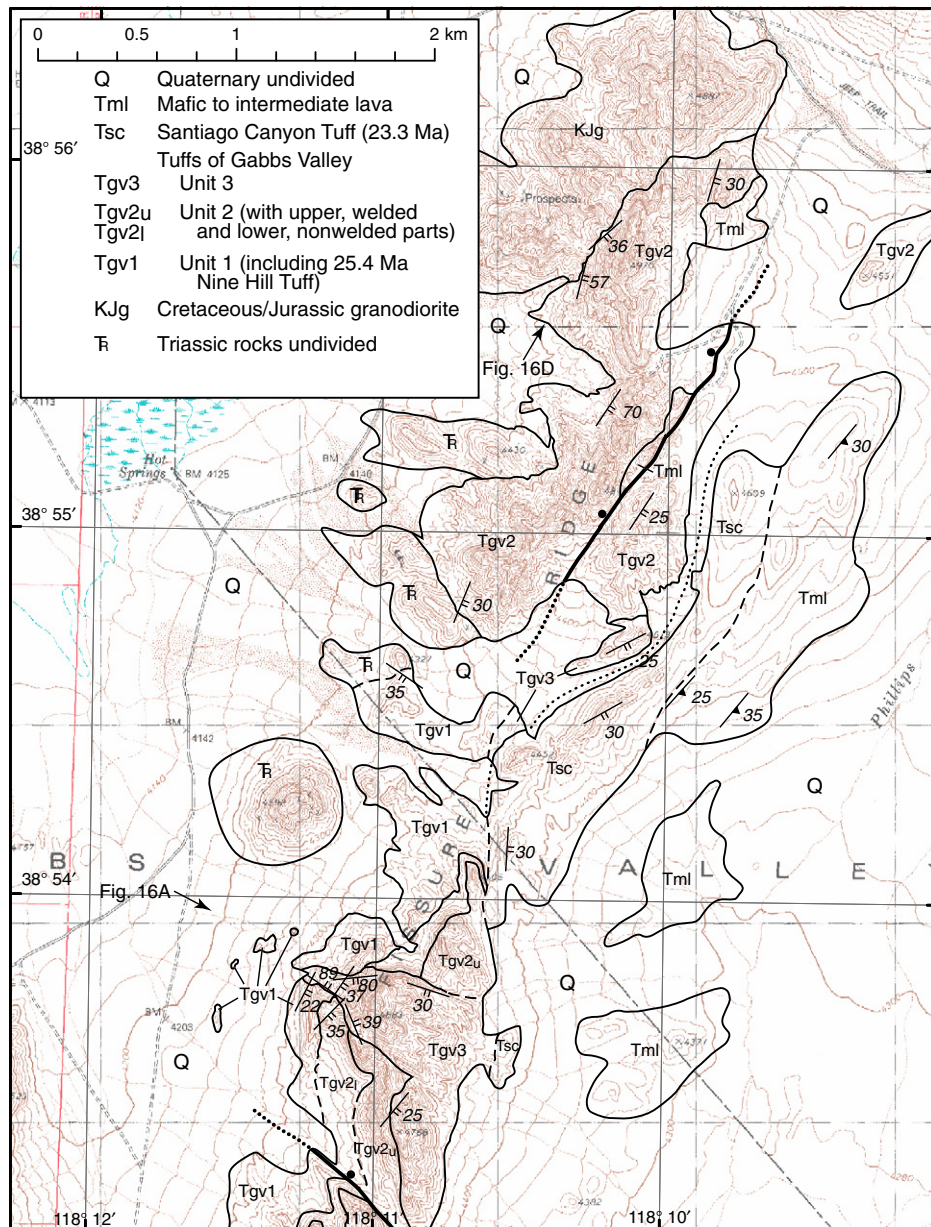


Figure 15. Geologic map of the Fissure Ridge area, Gabbs Valley, modified from Ekren and Byers (1976, 1986c) based on this study.

The interpreted fissure vent is the strongest evidence for a possible source at this location. Ekren and Byers (1976) discounted that it was a canyon filled by tuff in part because they found no evidence for significant erosion between the different tuffs. However, the area has no other evidence for a caldera. The tuffs of Gabbs Valley contain no megabreccia typical of intracaldera tuffs, and none of them, including Tgv3 in the fissure, are noticeably lithic or contain coarse lithic fragments (our observations). The geologic map of Ekren and Byers (1986c) shows Tgv1 depositionally on irregular

topography developed on pre-Cenozoic rocks north of the vent. Farther north, Tgv1 is absent, and Tgv2 fills a more than 300-m-deep valley in the pre-Cenozoic rocks. These relationships require that Tgv1 either not be deposited or be completely eroded before deposition of Tgv2. Similarly, all tuffs of Gabbs Valley and the overlying Santiago Canyon Tuff (their tuff of Copper Mountain; Fig. 3) thicken and thin along Fissure Ridge, which Ekren and Byers interpreted to be a result of caldera faulting. However, basal poorly welded zones and vitrophyres occur continuously along these contacts,

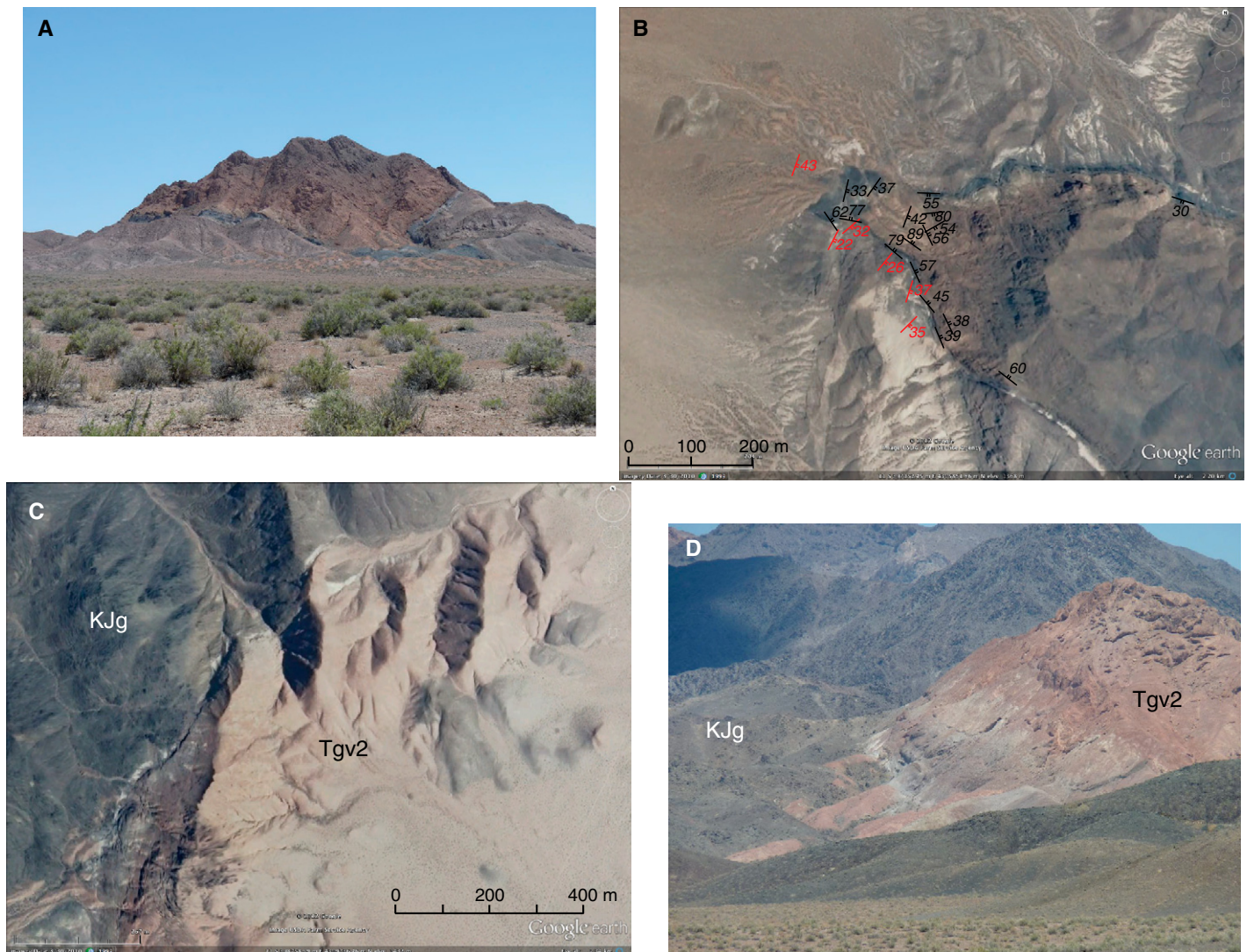


Figure 16. Photographs and Google Earth (www.google.com/earth/index.html) images of the Fissure Ridge area. (A) Photo looking south-east at the tuff-filled fissure. The fissure is ~400 m wide at its widest in this photo. (B) Google Earth image of the fissure showing strikes and dips in the wall rock (in red) and fissure-filling tuff (in black; Tgv3; Fig. 15). Steeply to moderately inward-dipping, white, poorly welded tuff and dense, black vitrophyre are developed all along the contact with wall rock. However, dips decrease inward and were flat in the middle of the fissure before extensional faulting that tilted rocks ~30° to the southeast. (C) Google Earth image of the contact between Tgv2 and Cretaceous or Jurassic granodiorite (KJg). Although interpreted as a down-to-the-east normal fault (Ekren and Byers, 1976, 1986c), we reinterpret the contact to be depositional. White, poorly welded tuff overlain by black, dense vitrophyre are developed all along the contact. Attitudes in the tuff parallel the east-dipping contact (Fig. 15). (D) Photo looking northeast along the contact between Tgv2 and Cretaceous or Jurassic granodiorite showing gentle to moderate east dips of the tuff and contact.

notably between Tgv2 and KJg, which we reinterpret to be a depositional contact (Figs. 16C and 16D). An alternative interpretation is that the tuffs were being alternately deposited and eroded in a paleovalley, typical of paleovalley relationships elsewhere in Nevada (Henry and Faulds, 2010). All tuffs, including the overlying, ~2 Ma-younger Santiago Canyon Tuff, are tilted 25° to 30° southeastward into Gabbs Valley.

Despite interpreting a caldera, Ekren and Byers (1986c) in their cross sections show all

tuffs maintaining constant thicknesses eastward into the caldera. However, all tuffs are buried beneath an extensional basin fill east of Fissure Ridge, so thicknesses to the east are unknown. Sedimentary sections encountered in two oil exploration wells in Gabbs Valley ~7 and 12 km east-northeast of the fissure and within the interpreted caldera are consistent with Gabbs Valley being a late Cenozoic basin-and-range graben but provide no evidence regarding a possible caldera. The western well went through middle

Miocene to Pliocene sedimentary rocks (Esmeralda Formation; Stewart et al., 1999) overlying probably older Miocene lavas to a total depth of 5060 feet (1540 m). The eastern well encountered a nearly 6000-ft-thick (1830 m) section of Esmeralda Formation to the bottom of the hole at a depth of 7707 ft (2349 m). Any caldera-related rocks would have to be at still greater depths.

An alternative explanation of the fissure is that it is a narrow canyon eroded into outflow

Tgv1 and Tgv2 in a paleovalley that was subsequently filled by outflow Tgv3. Tgv3 generally has a poorly welded zone and dense vitrophyre (Figs. 16A and 16B) along the wall, similar to depositional contacts of Tgv3 outside the fissure and of all other tuffs. The paucity of lithic fragments in Tgv3 in the fissure suggests that wall rock was not removed by eruption of tuff from the fissure and allows that it was removed by pre-fissure erosion. The fissure closes and vitrophyre dipping 35° eastward is continuous across the lowest exposed, western end, which would allow the western end to be the base of a paleovalley. Devitrified rock in the middle of the fissure also flattens inward, and both rock types in the middle of the fissure were approximately horizontal before the eastward, extensional tilting. Before extensional tilting, the slope of fissure walls ranged from steep, even vertical, to as little as ~35°. Moderate to steep slopes in the walls of canyons cut into modern, welded or even non-welded tuffs are common. A canyon as much as 33 m deep commonly with near-vertical walls was cut into the tuff of the Valley of Ten Thousand Smokes, Alaska, within 10 years of the tuff's deposition (Griggs, 1922); canyon walls remain near vertical today. Therefore, we conclude that no caldera is present in Gabbs Valley, and the fissure vent could be a narrow paleocanyon.

IMPLICATIONS OF TUFF DISTRIBUTION FOR EOCENE–OLIGOCENE TOPOGRAPHY

Geologic mapping of Cenozoic ash-flow tuffs in northern and western Nevada demonstrates that they erupted onto a surface with considerable relief—the paleovalleys of Figure 1 (Proffett and Proffett, 1976; Proffett and Dilles, 1984; Dilek and Moores, 1999; Faulds et al., 2005a, 2005b; Garside et al., 2005; Henry, 2008; Cassel et al., 2009a, 2009b; Henry and Faulds, 2010; Henry et al., 2012a). The region was an erosional highland, commonly referred to as the Nevadaplano, which was drained by major west- and east-trending rivers, with a north-south paleodivide through east-central Nevada. The highland developed at least by ca. 50 Ma, probably as a result of crustal thickening during Mesozoic–early Cenozoic shortening, Mesozoic batholith emplacement in the Sierra Nevada, and possibly a later contribution from slab removal (e.g., Humphreys et al., 2003; DeCelles, 2004; Dickinson, 2006; Best et al., 2009; Ernst, 2010; Mix et al., 2011). The absolute elevation and topographic-extensional evolution of this highland in the mid-Cenozoic remain highly controversial (Wakabayashi and Sawyer, 2001; Mulch et al., 2006, 2008; Cassel

et al., 2009a, 2009b, 2012; Druschke et al., 2009; Molnar, 2010; Henry et al., 2011, 2012a).

In northern and central to western Nevada and eastern California, the Eocene and Oligocene tuffs flowed, were deposited, and are preserved primarily in deep (as much as 1.2 km) but wide (8–10 km) paleovalleys (Proffett and Proffett, 1976; Henry, 2008; Henry and Faulds, 2010; Cassel et al., 2012; Henry et al., 2012a). The tuffs were able to flow great distances, as much as ~250 km corrected for extension (Henry and Faulds, 2010), down these valleys because they were channelized, did not disperse, and did not mix with air as much as would tuffs that spread more radially. The distribution of most of the tuffs in the western Nevada volcanic field suggests they spread radially only within a few tens of kilometers of their source.

Most tuffs of the western Nevada volcanic field are recognized in, and probably restricted to, areas west of the interpreted paleodivide. A topographic barrier along the divide may have restricted their flow to the east. However, the absolute elevation, location, and relief of a paleodivide are not fully resolved. The voluminous Nine Hill Tuff and tuff of Campbell Creek, and the Underdown Tuff–tuff of Clipper Gap of unknown volume, derived from sources at or north of 39°N, flowed far into eastern Nevada (Figs. 10 and 11; Supplemental Fig. 1 [see footnote 5]; Deino, 1985; Best et al., 1989; Henry et al., 2012a; Best et al., 2013c; A.L. Deino, 2012, written commun.). Whether the topographic barrier did not exist to the north, special eruption dynamics allowed these ash flows to travel especially far, or other factors were involved is unknown.

Our current view is that the divide shown by Henry et al. (2012a) south of ~40°N probably should be closer to that proposed by Best et al. (2009, 2013a) (Fig. 1). Uncertainty in location

and relief is especially high in the Toquima Range–Monitor Range area. On the basis of analogy to the modern Andes, Best et al. (2009; 2013a) proposed that a western side with a variable slope progressed up to a broad, gentle divide and an eastern plateau slightly less than 1 km below the divide (Fig. 17). Such a low divide would seem to be an insufficient topographic barrier to greatly limit eastward dispersion of ash flows. Eruptions from low on the west side would likely spread preferentially to the west, but eruptions from near the divide should have been able to spread both east and west. The abundance of calderas through central Nevada means that many of them must have been on or within a few tens of kilometers of the divide. That, with the few noted exceptions, tuffs apparently did not spread more symmetrically down both sides of the paleodivide requires further investigation.

LAVA FLOWS AND DOMES IN THE WESTERN NEVADA VOLCANIC FIELD

Lava flows and domes are a widespread but relatively small-volume component of the western Nevada and adjacent central Nevada fields (Fig. 18; Supplemental Text⁵). They commonly are the earliest manifestation of Tertiary magmatism within an area and often presage caldera-forming ignimbrite eruptions from the same area, but they commonly are hundreds of thousands to several million years older than associated calderas (Fig. 18 and Table 1). Lava flow sequences and domes range in composition

⁵Supplemental Text. Lava flows and domes in the western Nevada volcanic field. If you are viewing the PDF of this paper or reading it offline, please visit <http://dx.doi.org/10.1130/GES00867.S5> or the full-text article on www.gsapubs.org to view the Supplemental Text.

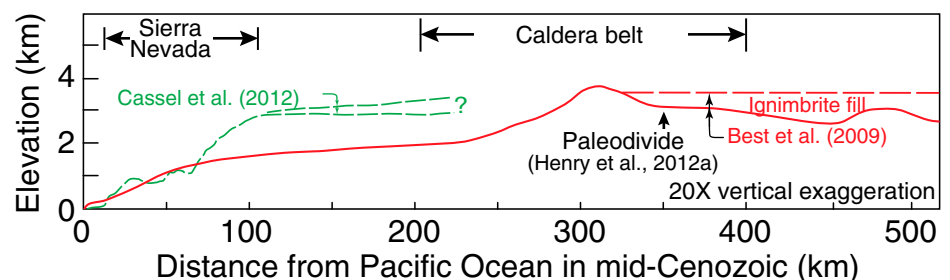


Figure 17. Interpreted paleotopography from the former Pacific Ocean eastward across the caldera belt and Nevadaplano, from Best et al. (2009, their figure 17) and Cassel et al. (2012, their figure 8). Profile from Best et al. also shows ignimbrite fill that would have smoothed pre-volcanic topography east of their paleodivide. Cassel et al. interpreted that the Sierra Nevada was at its present elevation in the late Eocene–early Oligocene and the area to the east a relatively flat plateau. Best et al. interpreted a slightly shallower gradient to between 200 and 250 km east of the Pacific Ocean but a similar maximum elevation.

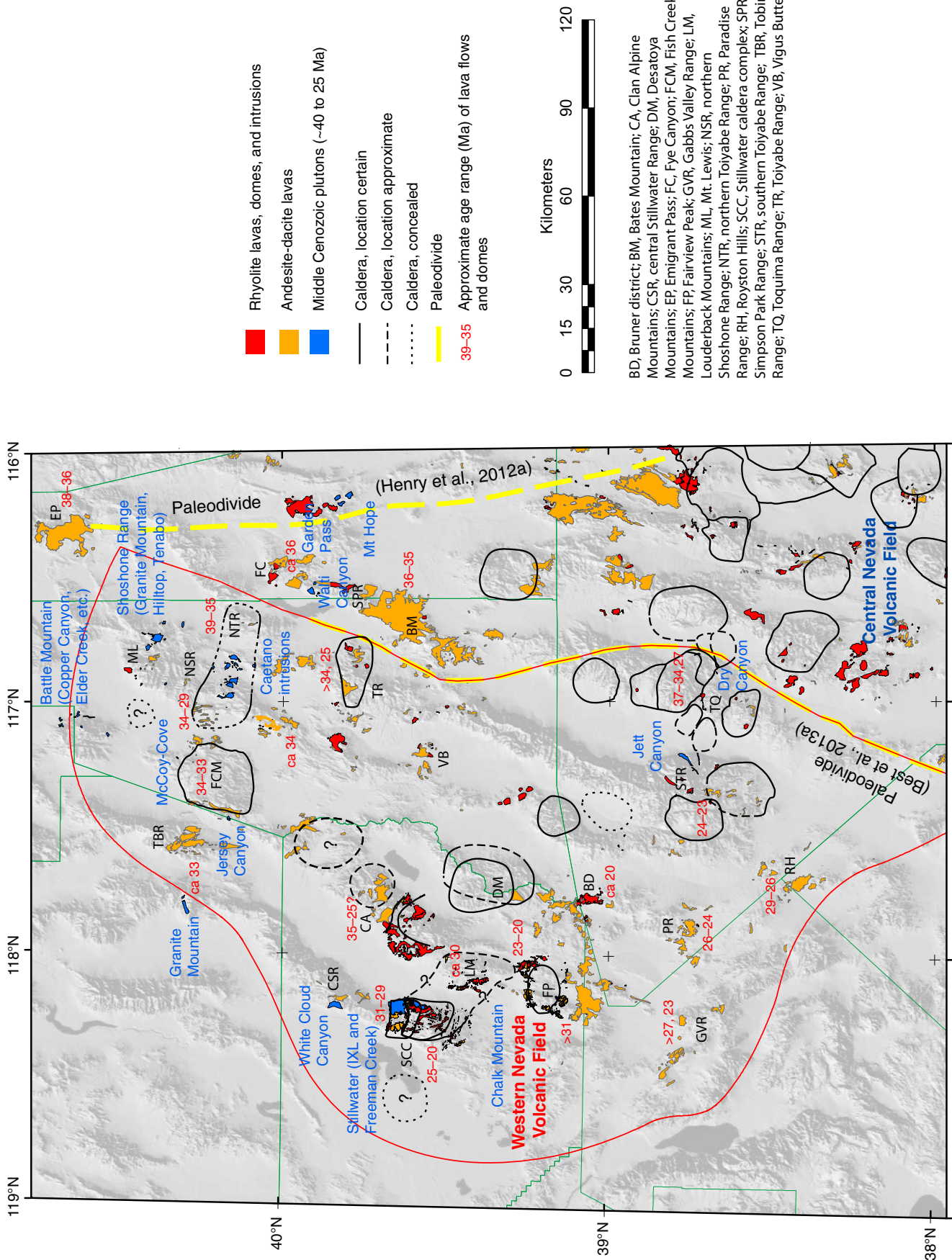


Figure 18. Map showing the distribution of lava flows and plutons in and immediately adjacent to the western Nevada volcanic field. Pluton distribution modified from John (1995), Crafford (2007), and Colgan et al. (2011b).

from high-silica rhyolite to mafic andesite, with rhyolites especially common in the central part of the field in the southern Stillwater Range and Clan Alpine Mountains (Figs. 18 and 19). Most larger volumes and areas underlain by lava flows and domes are >5–10 km distant from known calderas and appear unrelated to caldera-forming volcanism. Post-caldera, intermediate-composition lava flows are present in some calderas, notably in the Job Canyon caldera in the Stillwater Range and in the northwest part of the Caetano caldera. Silicic (rhyolite) domes and related lava flows are present in most calderas and represent resurgent magmas emplaced shortly after caldera formation.

Chemical analyses of 166 lava-flow and lava-dome samples scattered among these exposures indicate a wide range of compositions from high-silica rhyolite to basaltic andesite, although analyzed samples with <57% SiO₂ are scarce and restricted to the undated lavas north and south of Fairview Peak (Fig. 19). The rocks are K₂O rich and most form a high-K calc-alkaline series. Early rhyolites that underlie the Stillwater caldera complex and post-Caetano caldera andesites are notably K₂O rich. Samples from the western centers (Stillwater, Fairview Peak, Paradise Range) generally are more alkaline than samples collected farther east (Emigrant Pass, Fye Canyon, Bates Mountain), mostly as a consequence of higher Na₂O contents.

DACITE TUFFS RELATED TO LAVA DOMES

Several, small- to possibly moderate-volume dacitic ash-flow tuffs are more likely related to dacite lava domes, some possibly forming from dome collapse, than to caldera collapse. All known examples are outside of or at the edge of the caldera field. They are better considered part of the contemporaneous lavas than related to the calderas.

Dacite of Mount Lewis (35.8 Ma)

The 35.81 ± 0.29 Ma, small-volume, dacitic tuff of Mount Lewis erupted from an unknown but almost certainly non-caldera source probably in the northern Shoshone Range at 35.8 Ma. The tuff is spatially associated with a similar-age dacite intrusion and several aphyric rhyolite intrusions and is mostly confined to an east-west paleovalley across the range. The tuff is ~30 km north of the 34 Ma Caetano caldera. The tuff is moderately porphyritic, containing 20%–30%, fine-grained phenocrysts of plagioclase, hornblende, biotite, less-abundant quartz, and sparse K-feldspar and abundant small fragments of

Lava Flows and Domes in and adjacent to the Western Nevada Volcanic Field

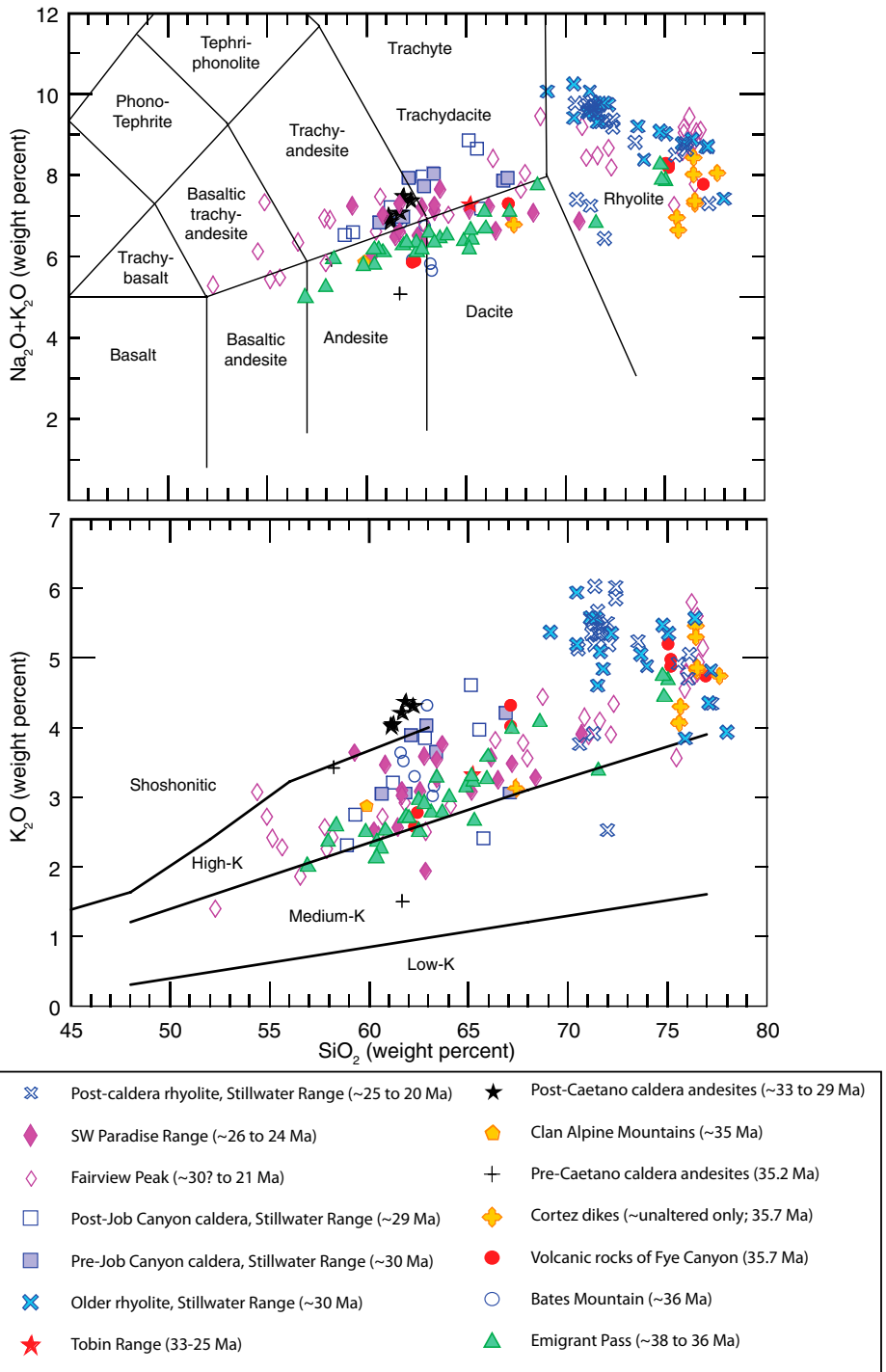


Figure 19. Total alkali-silica and K₂O-silica diagrams for lava flows of the western Nevada volcanic field. Sources of data: Stillwater Range, John (1995) and this study; Fairview Peak area, Henry (1996a, 1996b) and this study; Tobin Range, Gonsior and Dilles (2008); Clan Alpine Mountains, Riehle et al. (1972); Caetano, this study; Cortez, this study; Fye Canyon, this study; Bates Mountain, this study; Emigrant Pass, Ressel and Henry (2006) and this study. Classification from Le Maitre (1989) and Ewart (1982).

Paleozoic sedimentary rocks. It is generally propylitically altered and no chemical analyses are available.

Tuffs in the Hu-Pwi Rhyodacite (23.6 Ma)

The tuff and breccia of Gallagher Pass in the northern Wassuk Range and the time-correlative Hu-pwi Rhyodacite in the Gillis and Gabbs Valley Ranges consist predominantly of intermediate lavas with abundant phenocrysts of plagioclase, biotite, and pyroxene (Proffett and Proffett, 1976; Ekren et al., 1980; Hardyman, 1980; Proffett and Dilles, 1984; Dilles and Gans, 1995). Numerous cooling units of crystal-rich tuff having the same phenocryst assemblage are interbedded with the lavas and have been mapped as the lower Nugent Tuff Member and the trachydacitic upper Poinsettia Tuff Member in the Gillis and Gabbs Valley Ranges. Dilles and Gans (1995) reported a biotite $^{40}\text{Ar}/^{39}\text{Ar}$ age of 23.60 ± 0.04 Ma on what they interpreted to be the Poinsettia Tuff Member in the northern Wassuk Range. The combined group of tuffs crops out in a west-northwest-trending belt ~70 km long, which was probably shorter before strike-slip faulting of the Walker Lane, but the distribution of individual tuffs is unknown. Nevertheless, the tuffs are certainly more voluminous than the dacite of Mount Lewis or the tuff of Perry Canyon.

Tuff of Perry Canyon (23.5 Ma)

The 23.53 ± 0.26 Ma tuff of Perry Canyon in western Nevada north of Reno consists of two cooling units of crystal-rich, plagioclase > hornblende > biotite > quartz > opaque minerals dacite tuff (Garside et al., 2003). A single chemical analysis indicates 65% SiO_2 (Fig. 6). The tuff crops out over an area of ~50 km² and is spatially and temporally associated with dacite to rhyolite lava domes.

MIDDLE CENOZOIC PLUTONS IN THE WESTERN NEVADA VOLCANIC FIELD

Granitic plutons intruded approximately coeval with the ignimbrite flareup are scattered across the western Nevada volcanic field and the northwestern corner of the adjacent central Nevada volcanic field (Fig. 18; Table 3). Plutons are exposed both in the cores of highly extended calderas (Stillwater and Caetano calderas) and in areas that lack nearby exposures of coeval volcanic rocks (i.e., Eocene plutons near Battle Mountain). These plutons were emplaced between ca. 40 and 25 Ma, and their compositions range from granite to granodiorite with ~77–65 wt% SiO_2 (du Bray et al., 2007; this study).

TABLE 3. PLUTONS IN THE WESTERN NEVADA VOLCANIC FIELD

| Name | Location | Modal composition* | Composition (wt% SiO_2) | Age (Ma) | Mineralization | Comments | References |
|---------------------------|-----------------------------------|-------------------------------------|-----------------------------------|----------------|---|--|---|
| Long Canyon intrusions | Battle Mountain | Granodiorite-granite | 66.9 to 71.5 | 39.3 to 37.7 | None | Several small stocks and dikes | Theodore et al., 1992; du Bray et al., 2007 |
| Elder Creek stock | Battle Mountain | Monzonite-monzogranite | 67.2 to 71.4 | 38.2 | Porphyry Cu | Several small stocks and dikes | Theodore et al., 1973; du Bray et al., 2007 |
| Copper Canyon stock | Battle Mountain | Granodiorite-granite | 66.9 to 73.1 | 39.4 to 38.5 | Porphyry Cu, Au skarn, polymetallic veins | | Roberts, 1964; Theodore et al., 1973; du Bray et al., 2007 |
| Carico Lake pluton | Carico Lake valley | Granite | 71.3 to 73.2 | 34.0 | None | Resurgent intrusion in Caetano caldera | John et al., 2008, 2009; Colgan et al., 2011b; this study |
| Redrock Canyon intrusion | Carico Lake valley | Granite | 74.1 to 75.2 (altered) | 34.0 | None | Resurgent intrusion in Caetano caldera | John et al., 2008, 2009; Colgan et al., 2011b; this study |
| Granite Mountain pluton | Northern Shoshone Range | Granodiorite | 66.4 to 69.6 | ca. 39 to 38.5 | None | | Gilluly and Gates, 1965; Kelson et al., 2008 |
| Hilltop stock | Northern Shoshone Range | Granodiorite | 65.1 to 66.9 | 39.6 to 39.1 | Polymetallic veins | | Gilluly and Gates, 1965; Kelson et al., 2008; this study |
| Tenabo stock | Northern Shoshone Range | Granodiorite | 67.2 to 69.7 | 39.4 | Au skarn, Carlin-type Au | | du Bray et al., 2007; Kelson et al., 2008 |
| Granite Mountain stock | East Range | Granite | 68.5 to 69.7 | 31.4 | None | | Fosdick and Colgan, 2008; du Bray et al., 2007 |
| Brown stock | Northeastern Fish Creek Mountains | Tonalite-granodiorite | 65.0 to 67.7 | 40 to 39 | Au skarn, distal disseminated Au-Ag | | Brooks et al., 1991; Emmons and Eng, 1995; du Bray et al., 2007 |
| Jersey Canyon intrusion | Western Fish Creek Mountains | Granodiorite | | Undated | None | | du Bray et al., 2007 |
| White Cloud Canyon pluton | Stillwater Range | Granite | 74.2 to 77.0 | ca. 25 | None | Coarse K-feldspar phenocrysts with apitic groundmass | John, 1992c; du Bray et al., 2007; this study |
| IXL pluton | Southern Stillwater Range | Quartz monzodiorite to granodiorite | 61.4 to 68.5 | 29.1 | Polymetallic veins and skarn | Resurgent intrusion in Job Canyon caldera | John, 1992c, 1995; this study |
| Freeman Creek pluton | Southern Stillwater Range | Granite; granodiorite | 71.3 to 74.2; 66.2 to 67.1 | ca. 25 | None | Composite pluton underlying POCO Canyon and Elevenmile Canyon calderas | John, 1992c, 1995; this study |
| Chalk Mountain pluton | Chalk Mountain | Granodiorite-granite | 69.5 to 73.1 | ca. 25 | Polymetallic replacement | | John, 1992c; du Bray et al., 2007; this study |
| Jett Canyon dikes | Southern Toiyabe Range | Granite | 69.4 to 73.5 | 30.2 | None | | Whitebread and John, 1992; du Bray et al., 2007 |
| Dry Canyon pluton | Toiyama Range | Granodiorite | 62 to 65 | 37 to 34 | None | | Shawe et al., 1986; Shawe, 1995 |

*International Union of Geologic Sciences classification (Streckeisen, 1976).

The largest plutons include the 34.0 Ma Carico Lake and Redrock Canyon plutons in the center of the Caetano caldera and the 29.1 Ma IXL and ca. 25 Ma composite Freeman Creek plutons in the Stillwater caldera complex (discussed in detail in the section on the Caetano and Stillwater calderas). Other large plutons include the Granite Mountain, Hilltop, and Tenabo stocks in the northern Shoshone Range (all ca. 39 Ma), the Copper Canyon, Elder Creek, and Long Canyon stocks in the Battle Mountain area (ca. 39–38 Ma), the 40–39 Ma Brown stock in the northern Fish Creek Mountains, the ca. 31 Ma Granite Mountain pluton in the East Range, the ca. 25 Ma White Cloud Canyon pluton in the Stillwater Range, and the ca. 25 Ma Chalk Mountain pluton.

Of particular significance to the regional tectonic history, several plutons, notably ca. 40–38 Ma intrusions at Battle Mountain and in the northern Shoshone Range and northern Fish Creek Mountains, are not spatially associated with coeval volcanic rocks. However, these intrusions are locally unconformably overlain by late Eocene volcanic rocks (the 34.4 Ma tuff of Cove Mine or 35.8 Ma tuff of Mount Lewis; Seedorff, 1991; Doeblich, 1995; Johnston et al., 2008; this study), which indicates rapid and significant uplift (several kilometers) and erosion between ca. 38 and 35.8–34.4 Ma, just before major caldera-forming ignimbrite eruptions began in the region. Several of these intrusions also are noteworthy for significant genetically related base and precious metal deposits (e.g., ca. 39–38 Ma Copper Canyon and Elder Creek porphyry copper deposits and Fortitude gold skarn deposit at Battle Mountain and the ca. 39 Ma Cove-McCoy gold-silver skarn and distal disseminated deposits in the Fish Creek Mountains).

Middle Cenozoic plutons associated with significant mineral deposits in the western Nevada volcanic field are older than nearby calderas, although resurgent plutons formed large, but apparently unmineralized, hydrothermal systems in the Caetano and Job Canyon calderas (see section on the Caetano and Stillwater calderas). The most important mineral deposits in the western Nevada volcanic field are the 35.7 Ma Cortez and Cortez Hills Carlin-type gold deposits and the 26.1 Ma Round Mountain epithermal gold deposit. The Cortez and Cortez Hills gold deposits are spatially and temporally associated with rhyolite porphyry dikes, sills, and flow domes (the Cortez dikes; Wells et al., 1969; Muntean et al., 2011). The Round Mountain deposit is not associated with exposed intrusive rocks, but it formed along the structural margin of the Round Mountain caldera within 0.5 Ma of caldera formation (Henry et al., 1996, 1997).

IMPLICATIONS ABOUT DEVELOPMENT OF IGNIMBRITE CALDERAS AS INDICATED FROM THE CAETANO AND STILLWATER MAGMATIC SYSTEMS

The Caetano caldera and the three calderas of the Stillwater caldera complex in the western Nevada volcanic field are exceptionally well-exposed examples of single and multi-cycle (nested) calderas, respectively, formed by large-volume ignimbrite eruptions. Large-magnitude ($\geq 100\%$) crustal extension formed tilted fault blocks that expose extraordinary, 5–10-km-thick sections through the calderas and underlying granitic plutons. In this section, we review salient points of published studies of these calderas and add new data to establish the magmatic, structural, and hydrothermal nature of these unique calderas (Table 4). Special emphasis is given to their structure and how this contrasts markedly with current popular models of caldera subsidence, as invoked, for example, for the two largest calderas in the Indian Peak–Caliente field to the east (Best et al., 2013b) and the possible genetic relationship between large-volume ignimbrites and granitic plutons.

Caetano Caldera, North-Central Nevada

The 34.0 Ma Caetano caldera near the northeast end of the western Nevada volcanic field formed during eruption of $>1100 \text{ km}^3$ of Caetano Tuff (Figs. 1, 10, and 20; Table 4; John et al., 2008; Colgan et al., 2011b). Most of the tuff is a thick intracaldera sequence; little outflow tuff is preserved. Multiple resurgent granitic plutons were emplaced into the center of the caldera within 100 k.y. after caldera formation. Middle Miocene extensional faults cut and tilted the originally $20 \times 12\text{--}18 \text{ km}$ caldera into a set of fault blocks dipping $\sim 40^\circ$ east that expose paleodepths of as much as 5 km; the fault blocks include the entire caldera-fill sequence and the upper parts of resurgent plutons (Colgan et al., 2008).

Rocks related to the Caetano caldera include two units of intracaldera Caetano Tuff, two large granite porphyry plutons, and smaller rhyolite and rhyolite porphyry dikes and plugs (John et al., 2008; Colgan et al., 2011b). The Caetano Tuff is crystal rich and zoned upward from high-silica rhyolite to lower-silica rhyolite. It consists of the lower compound cooling unit as much as 3000 m thick separated by a complete cooling break from the upper unit, a thinner (500–2000 m thick) sequence of thin ash flows and interbedded lacustrine and fluvial sedimentary rocks. Zones and lenses of megabreccia and mesobreccia are abundant in the upper part of

the Caetano Tuff (John et al., 2008; Moore and Henry, 2010; Colgan et al., 2011b).

The crystal-rich Caetano Tuff (typically 35%–45% phenocrysts) is zoned upward from high-silica rhyolite (76%–78% SiO_2), which forms most of the lower unit, to rhyolite (72%–74% SiO_2) that forms the upper part of the lower unit and ash flows in the upper unit (John et al., 2008).

Intrusive rocks related to the Caetano caldera include the Redrock Canyon and Carico Lake plutons in the center of the caldera, smaller rhyolite porphyry ring-fracture plugs, and a composite tuff feeder dike along an outer ring fracture (Fig. 20; John et al., 2008; Colgan et al., 2011b). The Redrock Canyon intrusion is composed of medium-grained granite porphyry containing $\sim 30\%$ phenocrysts in a devitrified felsite to aplitic (0.01–0.10 mm) quartz-feldspar groundmass. The Carico Lake pluton consists of 55–65 vol%, medium-grained phenocrysts in a microcrystalline (0.05–0.10 mm) groundmass of quartz and feldspar. The Carico Lake and Redrock Canyon intrusions are compositionally similar to the upper unit of the Caetano Tuff.

Stillwater Caldera Complex, West-Central Nevada

The Stillwater caldera complex in the southern part of the Stillwater Range and the southwestern part of the Clan Alpine Mountains along the northwest margin of the western Nevada volcanic field comprises three partly overlapping, Oligocene ash-flow calderas and subjacent plutonic rocks (Figs. 1 and 21; Table 4; John, 1995). In the southern Stillwater Range, the north half of the caldera complex was tilted $70^\circ\text{--}90^\circ$ west and the south half tilted $55^\circ\text{--}75^\circ$ east during large-magnitude crustal extension in the latest Oligocene (John, 1992b, 1995; Hudson et al., 2000) and/or middle Miocene (J.P. Colgan, 2011, personal commun.). The calderas and underlying plutons are exposed in cross section over an $\sim 10 \text{ km}$ depth range. Our ongoing mapping indicates that the Elevenmile Canyon caldera extends east and southeast into the southern Clan Alpine Mountains (Fig. 11).

The ca. 29 Ma Job Canyon caldera consists of two structural blocks (Figs. 22 and 23). The 6-km-wide, $70^\circ\text{--}90^\circ$ -west-dipping north block is the least structurally deformed part of the Stillwater caldera complex. It consists of as much as 1500 m of pre-collapse intermediate-composition lava flows and breccias (older dacite and andesite unit) overlain by 2000 m of relatively crystal-poor rhyolite intracaldera ash-flow tuff (tuff of Job Canyon) locally interbedded with thick sequences of caldera-collapse breccia, overlain in turn by up to 2500 m of intermediate lava flows and minor lacustrine and fluvial

TABLE 4. SUMMARY OF MAJOR FEATURES OF THE CAETANO CALDERA AND CALDERAS IN THE STILLWATER CALDERA COMPLEX, NEVADA

| | Caetano caldera 34.0 Ma | Job Canyon caldera ca. 29 Ma | Poco Canyon caldera ca. 25 Ma | Elevenmile Canyon caldera ca. 25 Ma |
|---|---|---|---|--|
| Age | | | | |
| Caldera-related units | Caetano Tuff (lower compound cooling unit, upper unit of thin ash-flow tuffs and sedimentary rocks), Redrock Canyon intrusion, Carico Lake pluton, small ring-fracture intrusions | Younger dacite and andesite (lavas), tuff of Job Canyon, older dacite and andesite (lavas), IXL pluton | Tuff of Poco Canyon (2 cooling units), megabreccia of Government Trail Canyon, granite phase of Freeman Creek pluton, granite porphyry | Tuff of Lee Canyon, sedimentary tuff unit (?), tuff of Elevenmile Canyon, granodiorite phase of Freeman Creek pluton |
| Thickness of caldera fill | Up to 4 km | North block, ± 5 km; south block, ≤ 2.5 km | >4 km along north margin; >1.5 km at south end | >3 km in south and central parts, <1 km at north end |
| Volume of ignimbrites | 1100 km ³ | >200 km ³ | >200 km ³ | >600 km ³ |
| Megabreccia and mesobreccia | Lenses in intracaldera tuff near caldera walls especially in upper part of lower tuff unit and upper unit | Lenses in intracaldera tuff along north wall and south side of north block; blocks in intracaldera tuff near south end | Blocks in upper cooling unit of (intracaldera) tuff of Poco Canyon; 1.8-km-thick megabreccia unit beneath upper cooling unit | Blocks and lenses in tuff along north and south walls; slide blocks in tuff as much as 6 km from south wall |
| Intrusive rocks | Two large (25 km ²) intrusions: Redrock Canyon intrusion—medium-coarse grained biotite-hornblende granite porphyry (~30% phenocrysts); Carico Lake pluton—biotite-(hornblende) granite porphyry (50%–60% phenocrysts); porphyro-aphanitic rhyolite plugs along ring fractures | IXL pluton—fine- to medium-grained, biotite-hornblende granodiorite to quartz monzodiorite | Freeman Creek pluton—medium- to coarse-grained granite; granite porphyry ring-fracture dike—porphyroaphanitic rhyolite porphyry (shallow) to medium-grained granite porphyry (deep) | Freeman Creek pluton—medium-grained biotite-hornblende granodiorite porphyry |
| Composition of caldera-related rocks and compositional zoning | Caetano Tuff 78%–72% SiO ₂ , zoned upward from high-silica rhyolite to rhyolite; Carico Lake pluton 71%–73% SiO ₂ ; Redrock Canyon intrusion 74%–75% SiO ₂ ; ring-fracture plugs 70.5%–73% SiO ₂ | Trachyandesite to rhyolite (~55%–75% SiO ₂), rhyolite to silicic dacite tuff (68%–75% SiO ₂); IXL pluton—60%–68% SiO ₂ , felsic to mafic zoning from roof downward | Rhyolite to high-silica rhyolite tuff (73%–77% SiO ₂), weak normal zoning; Freeman Creek pluton—72%–74% SiO ₂ ; granite porphyry—71%–75% SiO ₂ | Trachydacite to rhyolite tuff (~61%–74% SiO ₂); compositional zoning uncertain in extrusive rocks; 66%–67% SiO ₂ in plutonic rocks |
| Outflow facies tuff and regional distribution | Outflow tuff filled paleovalley extending 40 km west; outflow tuff extends 35 km south | Regional distribution unknown; outflow(?) tuff extends 30 km southeast to Loderback and southwestern Clan Alpine Mountains | Outflow tuff regionally widespread—New Pass Tuff in central Nevada (60 km east) and tuff of Chimney Spring(?) in western Nevada (150 km west) | Regional distribution unknown; outflow tuff extends 25–30 km east and southeast to Loderback and southern Clan Alpine Mountains and north end of Fairview Peak |
| Caldera host rocks and bounding structures | Paleozoic rocks; vertical to steep inward dipping fault (60°–90°) extending >4 km deep; local topographic rim ≤ 1 km outside ring-fracture fault | Mesozoic rocks; sub-vertical fault penetrating at least 5 km deep | Tertiary volcanic rocks (Job Canyon caldera and older rhyolite unit); high-angle fault at north end | South wall—Mesozoic rocks, subvertical fault at least 3 km deep; north wall—Tertiary volcanic rocks, high-angle fault obscured by intrusive rocks |
| Caldera floor | Paleozoic rocks, Eocene gravel deposits; Eocene andesite; continuous for as much as 4 km along strike | Oligocene volcanic rocks where preserved; continuous for as much as 3 km along strike | Oligocene volcanic rocks; continuous for as much as 4 km along strike | South end—Mesozoic rocks; north end—Oligocene volcanic rocks; continuous for as much as 5 km along strike |
| Caldera resurgence | Carico Lake pluton forcibly emplaced and domed center of caldera | North block—uncertain, with possible piston-like uplift ≤ 1 km; south block— not evident | Not evident | Possible doming of center of caldera forming moat in southern third of caldera filled by sedimentary tuff unit |

sedimentary rocks (younger dacite and andesite unit). The 3-km-wide south block of the caldera consists of thinner sequences (total thickness ≤ 2500 m) of intermediate lava flows and overlying intracaldera tuff of Job Canyon with local interbedded collapse breccia; post-caldera collapse lava flows are absent in this structural block. The north part of the caldera is intruded by the compositionally zoned granodiorite-quartz monzodiorite IXL pluton. Outflow tuff of Job Canyon is widely distributed to the south and southeast (Fig. 10; John, 1995, 1997; this study).

The ca. 25 Ma Poco Canyon and Elevenmile Canyon calderas and underlying Freeman Creek pluton overlap in time and space. Poco Canyon caldera-related deposits comprise two cooling units of intracaldera crystal-rich rhyolite and high-silica rhyolite tuff (lower and upper units of the tuff of Poco Canyon) separated by a unit of crystal-poor high-silica rhyolite tuff and caldera-collapse breccia (megabreccia of Government Trail Canyon) and by a thick unit of crystal-rich rhyolite and trachydacite ash-flow tuff related to the Elevenmile Canyon caldera (tuff of Elevenmile Canyon) (Figs. 22 and 23). Coarse, locally tuffaceous breccias and sandstones containing abundant detritus derived from the lower tuff of Poco Canyon locally overlie this unit near the south margin of the caldera (sandstone and sedimentary breccia unit, Fig. 22; John, 1993). Total thickness of caldera-related deposits in the Poco Canyon caldera is locally >4500 m. The New Pass Tuff to the east and the tuff of Chimney Spring to the west are outflow tuffs that correlate with the intracaldera tuff of Poco Canyon (Fig. 3).

In the Stillwater Range, the Elevenmile Canyon caldera is filled by >3000 m of crystal-rich rhyolite to trachydacite ash-flow tuff (tuff of Elevenmile Canyon), locally overlain in the southern and central parts of the caldera by a thin sequence of probable caldera-lake deposits (sedimentary tuff unit), and in the northern and central parts of the caldera by the locally thick (as much as 1100 m) rhyolitic tuff of Lee Canyon. The total thickness of Elevenmile Canyon caldera deposits in the Stillwater Range locally exceeds 4000 m (Fig. 22). In the eastern part of the caldera in the southwestern Clan Alpine Mountains, intracaldera tuff of Elevenmile Canyon is as much as 2000 m thick and is overlain by as much as 1100 m of the trachydacitic to rhyolitic tuff of Hercules Canyon (Fig. 22; John, 1997; this study).

The composite Freeman Creek pluton intrudes the central and northern parts of the Poco Canyon and Elevenmile Canyon calderas (Figs. 21 and 23). It consists of an older granodiorite porphyry phase probably related to the Elevenmile Canyon caldera and a younger porphyritic granite phase probably related to the

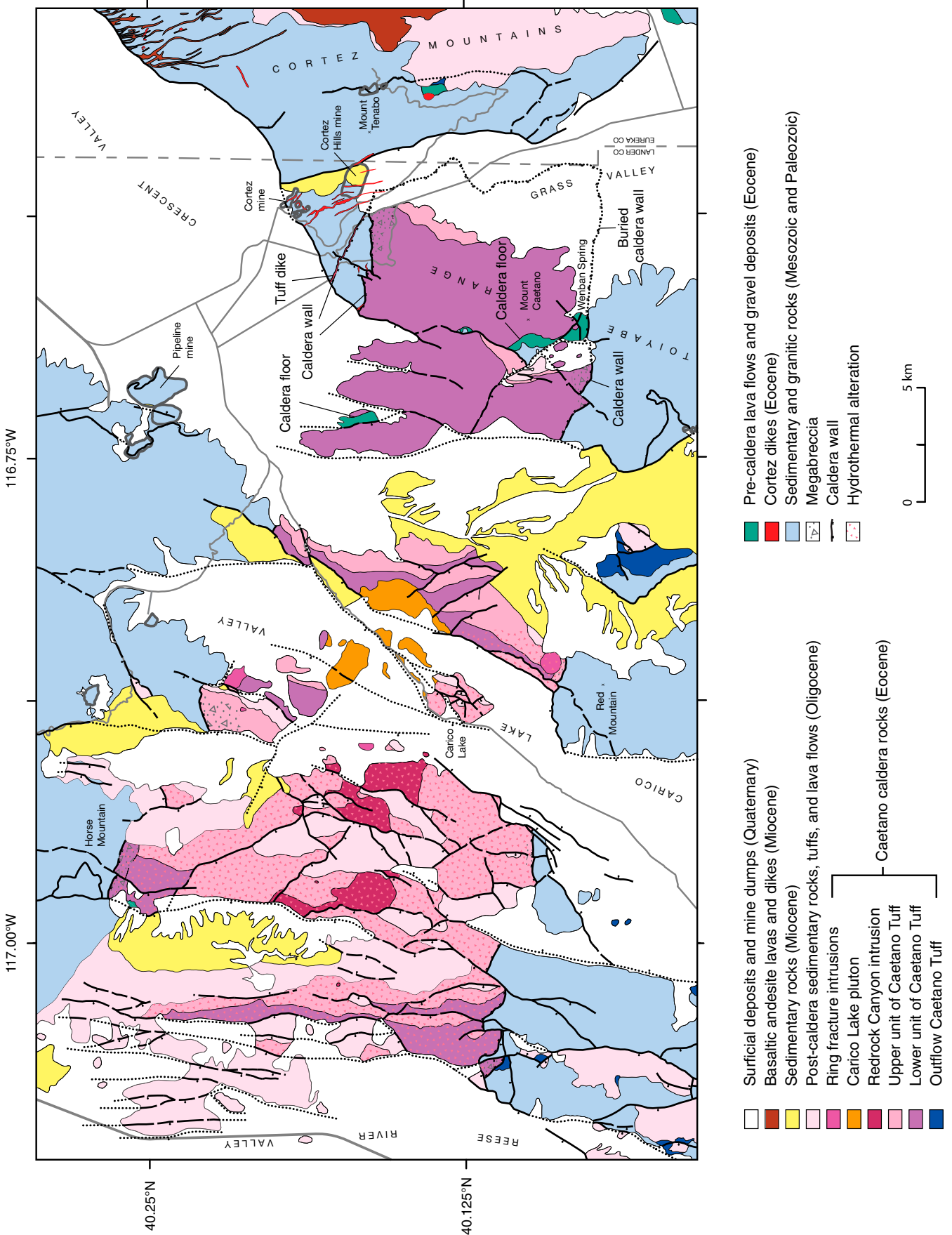


Figure 20. Simplified geologic map of the Caetano caldera. Modified from John et al. (2008) and Colgan et al. (2011b). See Figure 1 for location.

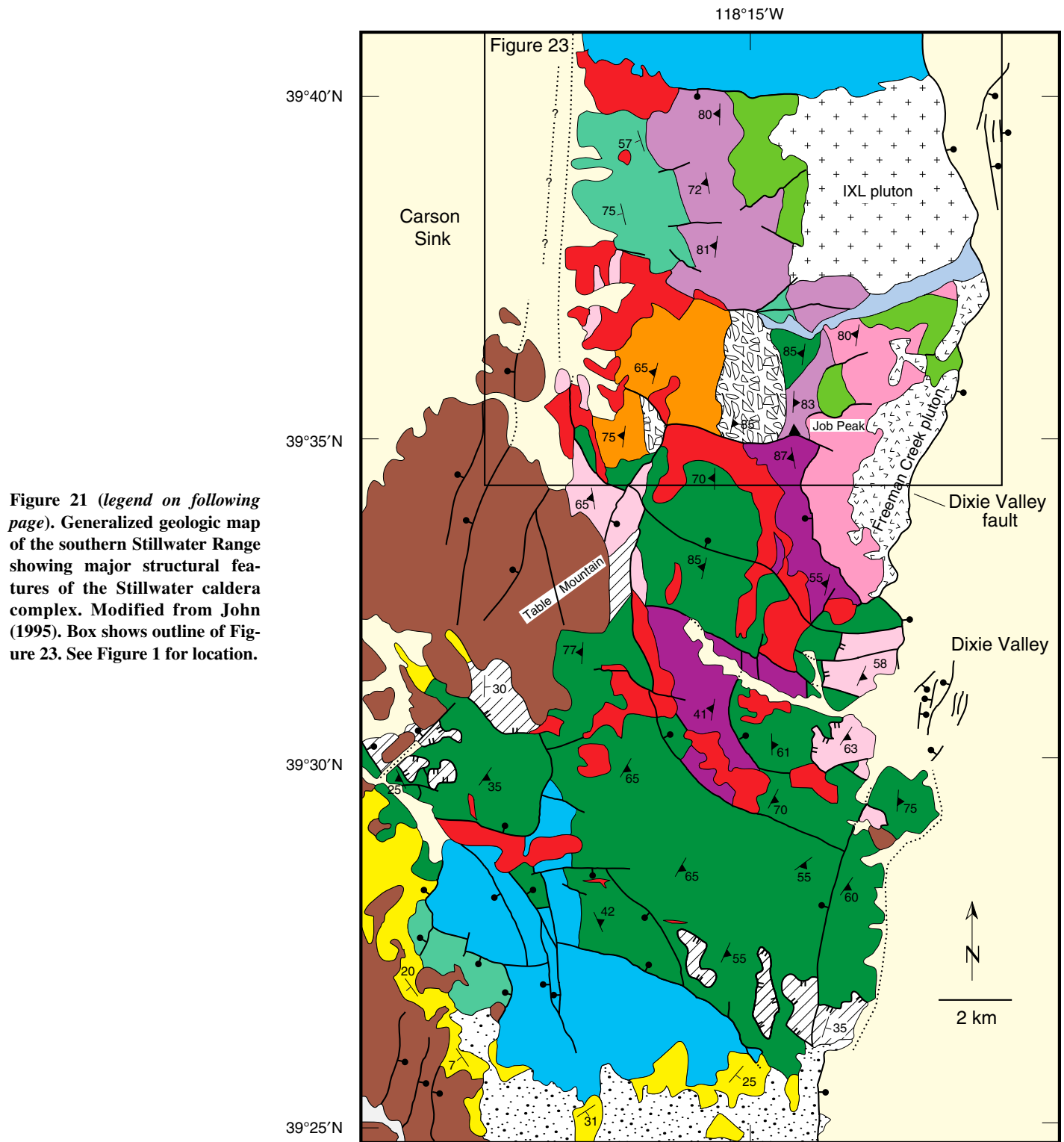


Figure 21 (legend on following page). Generalized geologic map of the southern Stillwater Range showing major structural features of the Stillwater caldera complex. Modified from John (1995). Box shows outline of Figure 23. See Figure 1 for location.

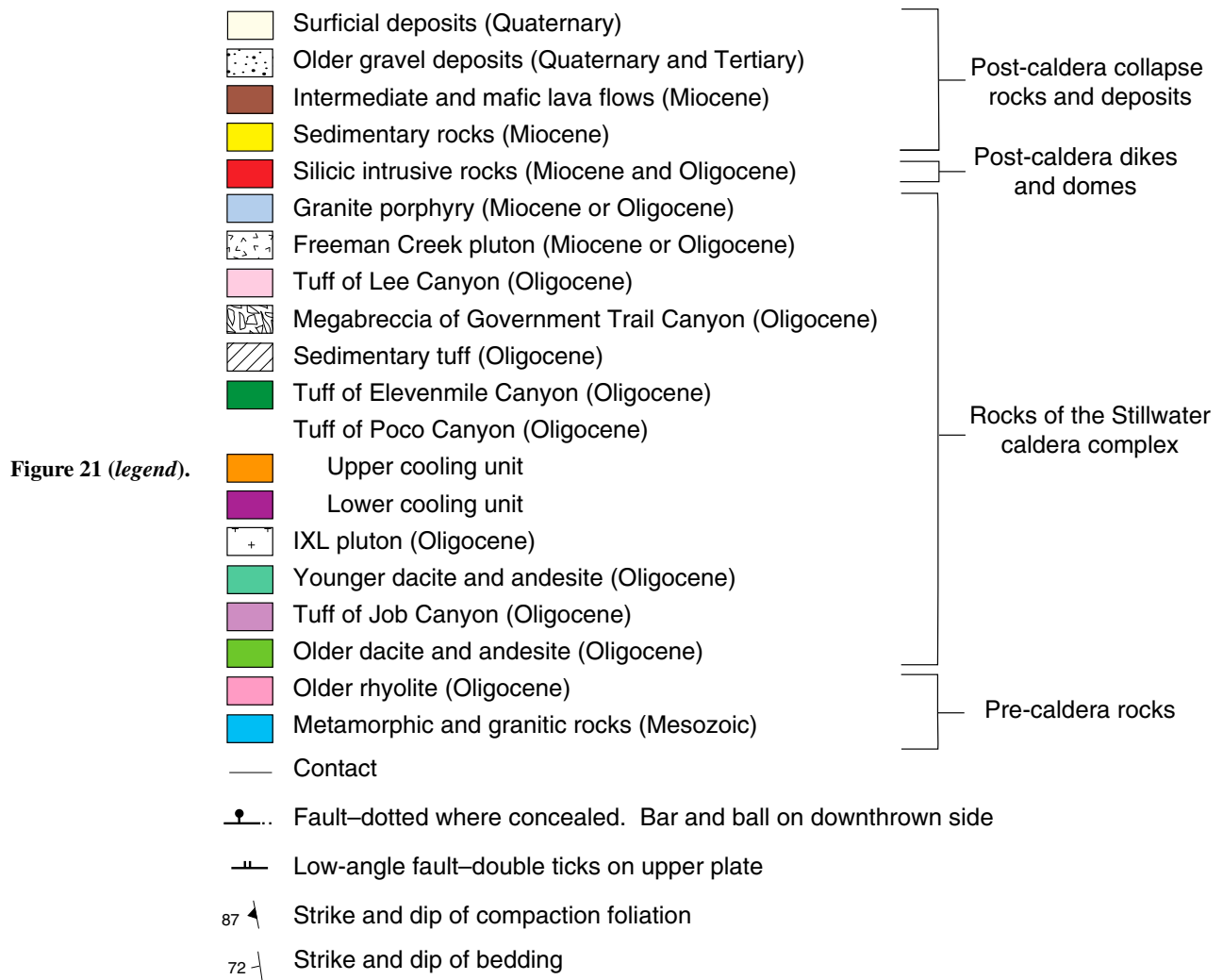
Poco Canyon caldera. A 7-km-long dike zoned from porphyro-aphanitic rhyolite on the west to granite porphyry on the east intruded the north edges of these calderas and may be a ring-fracture dike related to the Poco Canyon caldera (Figs. 21 and 23).

Comparison of the Caetano and Stillwater Calderas

The Caetano and Stillwater calderas provide unusual exposures of a broad spectrum of volcanic, subvolcanic, and plutonic rocks and

structural features that offer unique insights into caldera formation (Table 4). Many features are similar to those described in typical well-studied caldera systems, including thick intracaldera tuff, megabreccia, and resurgent intrusions, but other features, as described below, are unique.

EXPLANATION



Regional and Pre-Caldera Collapse Magmatism

Caldera-forming ignimbrite eruptions in both the Stillwater and northern Toiyabe Ranges were preceded by several million years of effusive and shallow intrusive silicic magmatic activity. Cenozoic magmatic activity in the southern Stillwater Range began ca. 31(?) Ma with eruption of lava flows and minor tuffs of the older rhyolite unit that probably formed a large lava dome complex that extended east to the west side of the Clan Alpine Mountains (Figs. 18, 21, 23, and 24; Reihle et al., 1972; Hardyman et al., 1988; John, 1995, 1997; C. Henry and D. John, unpub. mapping). These rhyolites are geochemically distinct from, and probably genetically unrelated to, the Job Canyon caldera system (John, 1995).

Magmatic activity related to the Job Canyon caldera system began with eruption of a thick

series of intermediate-composition lava flows and breccias on the north side of the rhyolite dome complex (older dacite and andesite unit, Fig. 23). Lava flows and a sill-and-dike complex underlying the south side of the caldera may have formed one or more small composite volcanoes. Following emplacement of the IXL pluton into the Job Canyon caldera a 3–4 Ma hiatus in magmatic activity preceded initial eruptions that formed the Poco Canyon and Elevenmile Canyon calderas. Pre-caldera collapse lava flows are not exposed beneath the intracaldera tuffs of either the Poco Canyon or the Elevenmile Canyon calderas; these calderas were developed in part on Mesozoic rocks, as well as on the older rhyolite sequence and the south block of the Job Canyon caldera (Figs. 21 and 23). Outflow tuff from the Job Canyon caldera, an uncorrelated tuff (reomorphically flowed tuff of John [1997]), and lava flows of

the older rhyolite sequence form the floor of the Elevenmile Canyon caldera in the southwestern Clan Alpine Mountains (Fig. 22).

The Caetano caldera was mostly built directly on Paleozoic sedimentary rocks, although it was preceded by ~6 Ma of regional, mostly silicic and mostly intrusive magmatism, including 35.7 Ma rhyolite dikes and flow domes (Cortez dikes) emplaced just outside the northeast margin of the caldera (John et al., 2008, 2009; Moore and Henry, 2010). Only a very small volume of andesitic lava erupted locally ~1.2 Ma prior to caldera formation. Major pre-collapse doming of surrounding Paleozoic rocks is not evident.

Caldera Subsidence and Caldera Floors

Collapse of the Caetano and Stillwater calderas occurred along subvertical to steeply inward-dipping faults that penetrated to depths of >5 km.

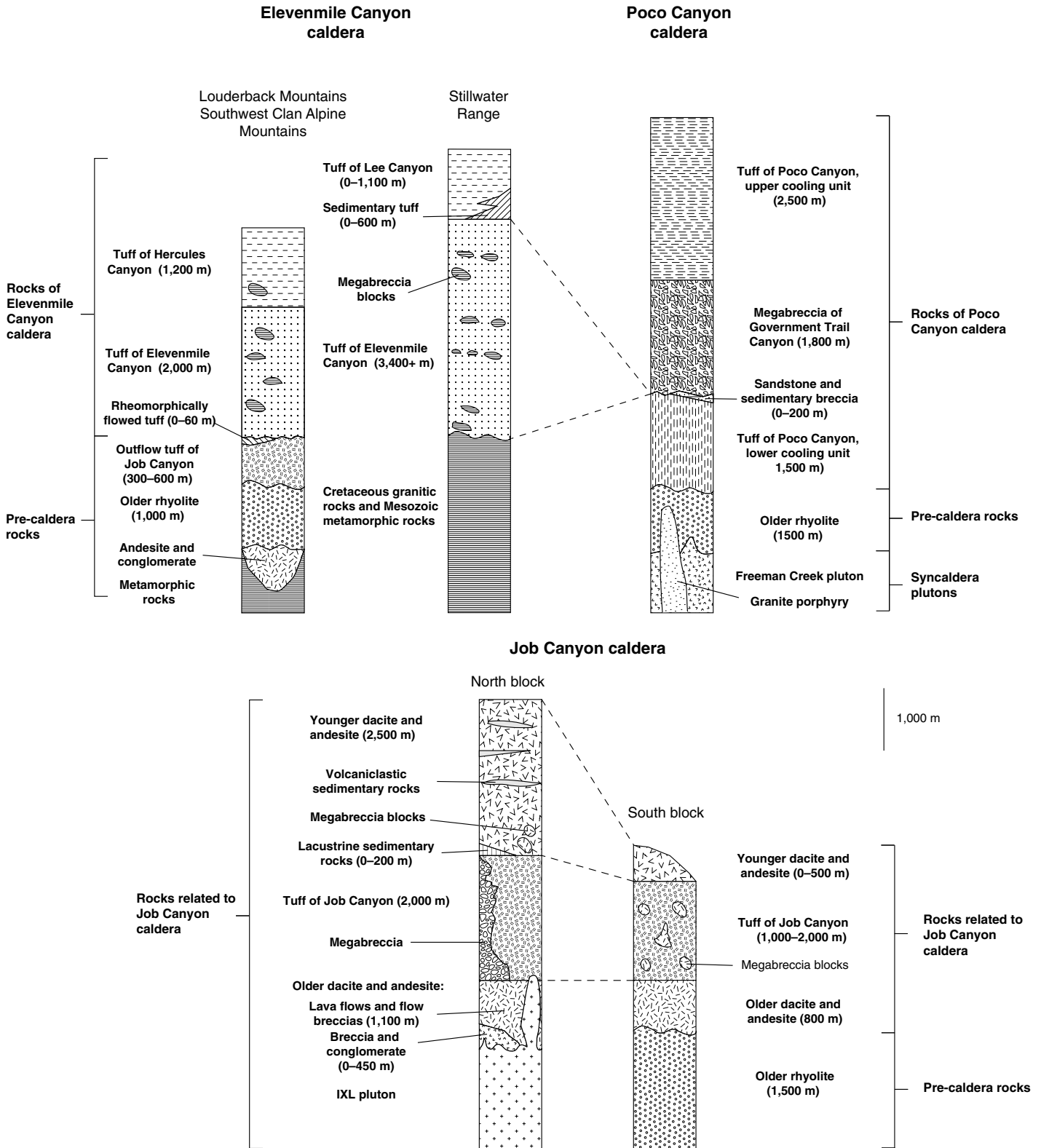
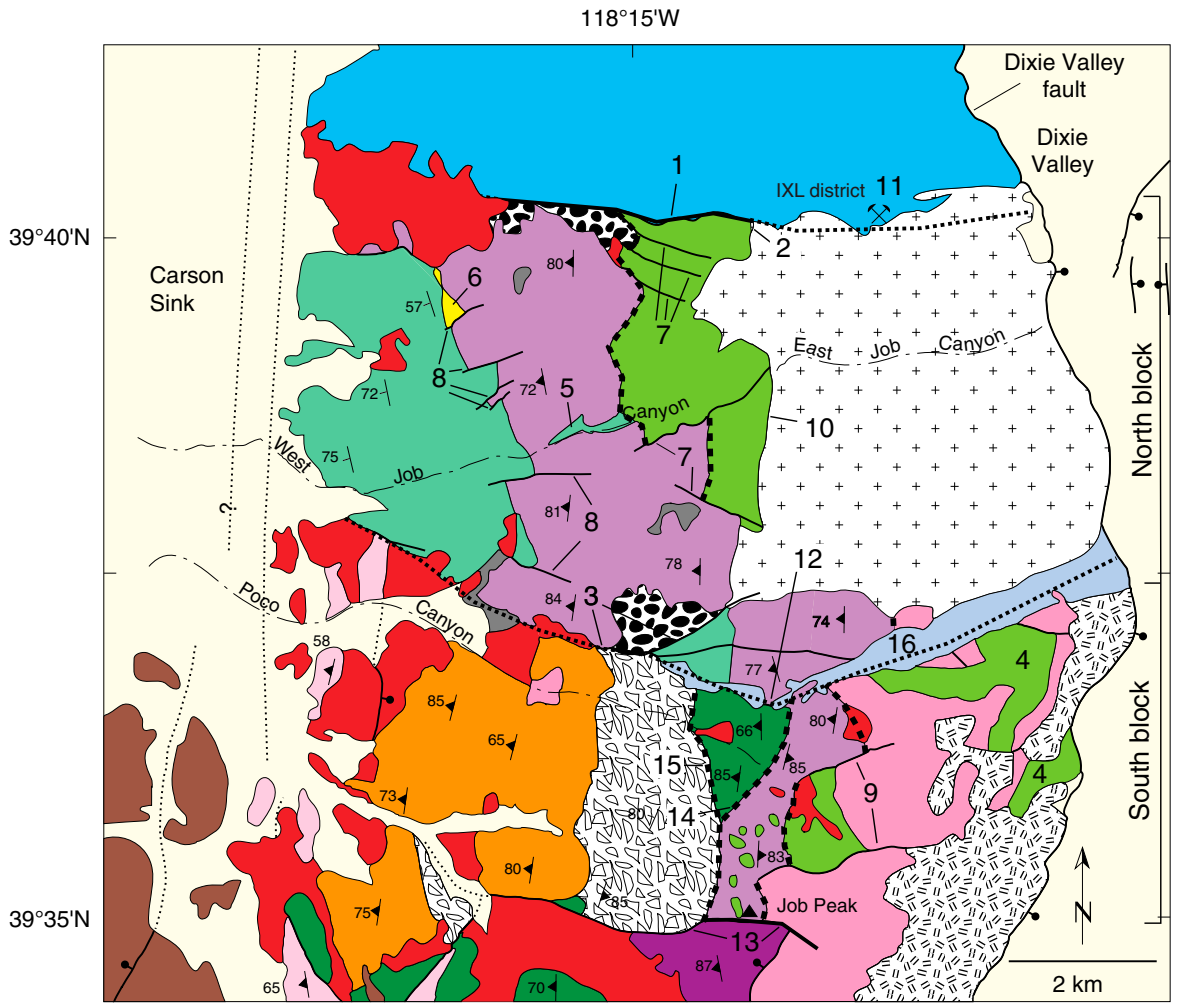


Figure 22. Summary composite stratigraphic sections for the Stillwater caldera complex in the southern Stillwater Range, Louderback Mountains, and southwestern Clan Alpine Mountains.

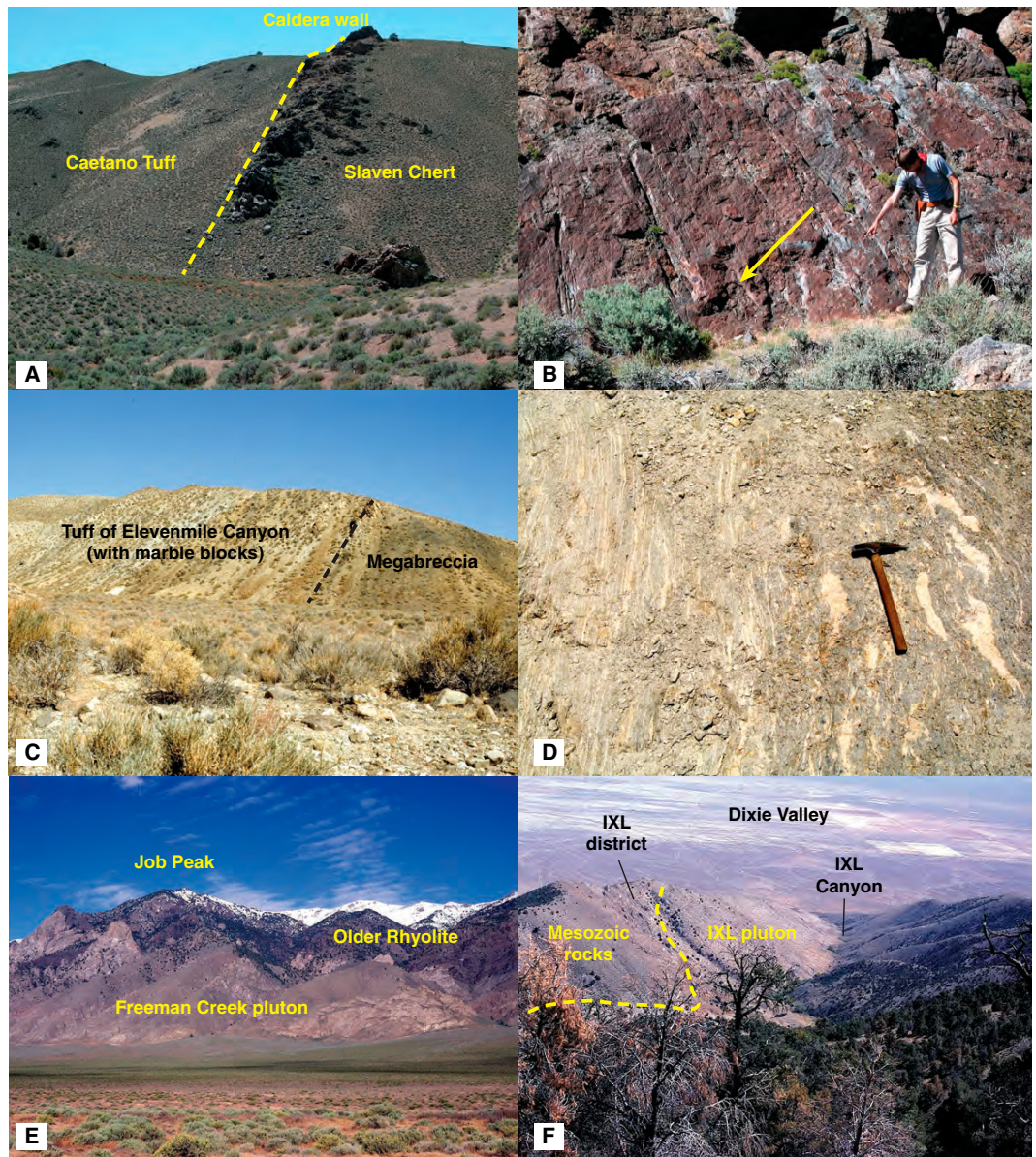


EXPLANATION

- | | | | |
|---|---|--|---|
| <ul style="list-style-type: none"> □ Surficial deposits ■ Basalt and basaltic andesite ■ Silicic intrusive rocks ■ Intermediate intrusive rocks ■ Granite porphyry ▨ Freeman Creek pluton ■ Tuff of Poco Canyon ■ Upper cooling unit ■ Lower cooling unit ▨ Megabreccia of Government Trail Canyon ■ Tuff of Lee Canyon ■ Tuff of Elevenmile Canyon | <ul style="list-style-type: none"> — Post-caldera rocks and deposits — Poco Canyon caldera — Elevenmile Canyon caldera | <ul style="list-style-type: none"> □ IXL pluton ■ Younger dacite and andesite ■ Lacustrine sedimentary rocks ■ Lava flows and flow breccias ▨ Megabreccia ■ Tuff of Job Canyon ■ Older dacite and andesite ■ Older rhyolite ■ Metamorphic rocks | <ul style="list-style-type: none"> — Job Canyon caldera — Pre-caldera rocks |
| <ul style="list-style-type: none"> — Ring fracture—dotted where concealed — Caldera floor 5 Structural features described in text | <ul style="list-style-type: none"> — Contact ⊥ Fault—dashed where concealed. Bar and ball on downthrown side 87 ↙ Strike and dip of compaction foliation 72 ↙ Strike and dip of bedding | | |

Figure 23. Geologic map of the Job Canyon and northern parts of the Elevenmile Canyon and Poco Canyon calderas (modified from John, 1995). Numbered features: 1—Job Canyon caldera wall; 2—Mesozoic rocks underlying floor of Job Canyon caldera; 3—structural boundary between north and south blocks in Job Canyon caldera; 4—andesite sill and dike complex underlying south block of Job Canyon caldera; 5—feeder dike for younger dacite and andesite unit; 6—lacustrine sedimentary rocks in Job Canyon caldera; 7 and 8—subsidiary growth faults in interior of Job Canyon caldera; 9—early faults in south block of Job Canyon caldera; 10—roof of IXL pluton; 11—polymetallic skarn occurrences around IXL pluton (IXL district); 12—north ring fracture of Elevenmile Canyon and Poco Canyon calderas; 13—fault separating structural blocks in Elevenmile Canyon and Poco Canyon calderas and forming south margin of Job Canyon caldera; 14—floor of Elevenmile Canyon caldera; 15—floor of Poco Canyon caldera; 16—granite porphyry ring-fracture dike along north margin of Poco Canyon caldera.

Figure 24. Photographs of features in the Caetano and Stillwater calderas. (A) Northeast structural wall of the Caetano caldera in the Toiyabe Range. View looking west along the Copper fault, which separates Devonian Slaven Chert from intracaldera Caetano Tuff. (B) Exposure of the northeast structural wall of Caetano caldera in the northern Toiyabe Range looking north (outcrop on ridgeline at the top of photo A). Wall rocks consist of brecciated, silicified Slaven Chert with oblique west-plunging slickensides (yellow arrow). Compaction foliation in Caetano Tuff dips $\sim 40^\circ$ east and restoration of tuff to horizontal results in approximately vertical-plunging slickensides. (C) View looking east at the southwest structural margin of the Elevenmile Canyon caldera in Elevenmile Canyon, southern Stillwater Range. The caldera margin consists of a silicified fault contact between intracaldera tuff of Elevenmile Canyon containing sparse, large blocks of Mesozoic marble and massive megabreccia consisting mostly of blocks of hornfelsed Mesozoic argillite with minor tuff matrix. The fault is presently oriented approximately $N60^\circ W, 85^\circ SW$ but untilting by rotation of compaction foliation in nearby tuff to horizontal restores fault to a steep north dip into the caldera. This fault apparently is a subsidiary fault subparallel to the caldera-bounding fault that lies ~ 150 m south in Elevenmile Canyon where the photograph was taken. (D) Caetano Tuff dike intruded into the outer ring fracture along northeast margin of the caldera. The dike in the left side of the photo is strongly flow banded with light-colored, semi-continuous, variably vesicular bands separated by darker bands. Fragments of Paleozoic rocks up to a few centimeters in diameter are common. Across a sharp transition zone in the middle of the photo, the light-colored bands have broken into individual pumice blocks. The rock on the right is a densely welded, pumice-rich ash-flow tuff in which the compaction foliation is irregularly parallel to the dike margin. Flow-banded rock and individual pumice blocks have compositions that are indistinguishable from the most highly evolved Caetano Tuff. Hammer handle is ~ 55 cm long. (E) View looking west at the west side of Dixie Valley and the east side of the Stillwater Range. The composite Freeman Creek pluton, which is inferred to represent unerupted residual magma from the Elevenmile and Poco Canyon calderas, forms light-colored outcrops in the lower part of the range. Most of the darker rocks in the upper part of the range are composed of an ~ 2 -km-thick sequence of rhyolite flow domes (older rhyolite unit) that underlie all three calderas of the Stillwater caldera complex. The roof of the pluton dips steeply (60° – 90°) west. On the left side of the photo, the total vertical relief through the pluton is ~ 800 m and Job Peak is ~ 1400 m above the range front. (F) View east down IXL Canyon from the top of the Stillwater Range showing the north margin of the IXL pluton. The pluton margin is mostly an approximately east-striking, subvertical intrusive contact into Mesozoic metasedimentary rocks that lies directly below (east of) the Job Canyon caldera wall. Mine workings in the IXL district are in small polymetallic skarns formed in limestone beds within the Mesozoic rocks.



Thicknesses of intracaldera tuffs and post-caldera collapse rocks indicate that caldera subsidence exceeded 4 km for all four calderas. Caldera-collapse fault zones (structural margins) are narrow in the deeper parts of the calderas and transition upward to wall slump zones (topographic margins) that contain large landslide blocks of wall rocks; the structural and topographic margins are separated by ≤ 1 km. Exposed caldera floor segments are coherent blocks as much as 4 km long and imply that the calderas collapsed as large piston-like blocks.

Caldera subsidence in the Stillwater Range dropped coherent wall-rock masses along subvertical ring-fracture faults, thereby juxtaposing thick sequences of intracaldera ash-flow tuff against Mesozoic wall rocks and large landslide blocks sloughed off caldera walls forming caldera-collapse breccia (Figs. 21 and 24; John and Silberling, 1994; John, 1995). Topographic walls outside the structural margins are not preserved, and along the north side of the Job Canyon caldera and the south side of the Elevenmile Canyon caldera, the topographic walls nearly coincided with the ring fractures. Each of the calderas collapsed as two or more internally coherent blocks that had greatly different amounts of subsidence (e.g., ~ 5 km and ~ 2 km in the Job Canyon caldera, Figs. 21, 22, and 23). Subsidiary growth faults with relatively minor displacement cut post-caldera collapse deposits within the Job Canyon caldera. A fault that separates the two structural blocks of the Job Canyon caldera later formed the north walls of both the Poco Canyon and Elevenmile Canyon calderas. A second, long-active fault formed the south margin of the Job Canyon caldera and separated the Poco Canyon and Elevenmile Canyon calderas into structural blocks with distinctly different thicknesses of caldera fill (Fig. 23).

The floors of the Elevenmile Canyon and Poco Canyon calderas and of the south block and part of the north block of the Job Canyon caldera are mostly preserved in the Stillwater Range (Figs. 21 and 23). In addition, much of the floor of the Elevenmile Canyon caldera is preserved in the southwestern Clan Alpine Mountains (John, 1997; this study). Topography of caldera floors is characterized by relatively flat to undulatory surfaces several kilometers long, locally offset by small-displacement (hundreds of meters), steeply dipping faults that were active during or shortly following caldera collapse (Fig. 23). Compaction foliation in tuffs and bedding in caldera-filling deposits are subparallel to caldera floors.

Collapse of the Caetano caldera occurred dominantly along a single ring fracture, although a composite ring fracture is preserved on the north-central and northeast caldera mar-

gins (Fig. 20; John et al., 2008; Moore and Henry, 2010; Colgan et al., 2011b). On the northeast margin, the greatest collapse (up to 4 km) occurred along an inner fracture that transitions upward from a discrete fault zone to a topographic wall. The topographic wall here (and elsewhere in the caldera) is only ~ 1 km outboard of the structural margin, and megabreccia constitutes no more than a few percent of intracaldera fill. A smaller amount of collapse occurred along an outer ring fracture, which contains a 10–40-m-wide, flow-banded to lithic and pumice-rich ash-flow tuff dike that fed a highly evolved phase of the eruption (Figs. 20 and 24; see section on vents and feeder dikes). Caldera structural-margin faults, including the outer ring fracture, are locally exceptionally well exposed and dip 60° – 70° inward to ~ 4 – 5 km depth (Fig. 20). Outward-dipping reverse faults required to make space for caldera collapse and prominently developed in analog models (Kennedy et al., 2004; Acocella, 2007; Howard, 2010) are not exposed but could be concealed beneath intracaldera tuff.

The Caetano caldera floor subsided asymmetrically as apparently coherent block(s) during eruption of the Caetano Tuff. Early subsidence was greater in the eastern part of the caldera, where lower intracaldera Caetano Tuff ponded and formed a single compound cooling unit as much as 3 km thick. Subsequent subsidence was greater in the western part of the caldera, where the upper intracaldera Caetano Tuff is as much as 2 km thick but thins to < 0.5 km in the eastern part of the caldera. Where exposed mostly in the eastern part of the caldera, the caldera floor is a planar surface subparallel to compaction foliation in overlying tuff (Fig. 20).

Vents and Feeder Dikes for the Caldera-Related Rocks

A 10–40-m-thick composite dike, composed of sparsely porphyritic rhyolite cut by abundantly porphyritic rhyolite tuff, is exposed along an outer ring fracture on the northeast margin of the Caetano caldera (Figs. 20 and 24; Moore and Henry, 2010; Colgan et al., 2011b). The sparsely porphyritic rhyolite is massive to locally flow banded, whereas the tuff is strongly flow banded near its margins; the flow bands disaggregate into individual flattened pumice blocks inward. Both the flow-banded and disaggregated parts of the tuff contain abundant to sparse fragments up to 25 cm in diameter of Paleozoic rocks similar to nearby wall rocks and of porphyritic rhyolite. Both sparsely porphyritic rhyolite and tuff contain the same phenocryst assemblage as Caetano Tuff, and the tuff has phenocrysts in the same abundance as the Caetano Tuff. Both phases have sanidine $^{40}\text{Ar}/^{39}\text{Ar}$ ages and

K/Ca indistinguishable from those of Caetano Tuff and chemical analyses of both rock types overlap with analyses of evolved Caetano Tuff. The composite dike is interpreted as a feeder for erupted Caetano Tuff that tapped sparsely and finely porphyritic rhyolite—the most compositionally evolved part of the Caetano magma chamber—and subsequently more abundantly porphyritic rhyolite that was highly evolved but more typical of Caetano Tuff. Similar highly evolved, finely porphyritic tuff locally forms the basal few tens of meters of intracaldera Caetano Tuff near Wenban Spring in the Toiyabe Range, where it is overlain by > 3000 m of crystal-rich (30%–45% phenocrysts) rhyolite tuff.

Ash-flow tuff feeder dikes are not exposed in the Stillwater Range. Several andesite dikes intrude intracaldera tuff of Job Canyon and apparently fed post-collapse lava flows of the Job Canyon caldera (John, 1995). These dikes are truncated by the underlying IXL pluton.

Late Intracaldera Magmatism

Both the Caetano caldera and calderas in the Stillwater Range are underlain by large composite granitic plutons. Extensive data for the Caetano system argue strongly that the granitic bodies were residual magma from the reservoir that was the source of the Caetano Tuff. Preliminary data from the Stillwater calderas allow a similar interpretation. These plutons have steep margins and flat roofs, are exposed over several-kilometer depth ranges, and have faulted lower contacts, which indicate that they are large stocks or the apical parts of batholiths (Figs. 20, 21, and 24). The shallowly emplaced (≤ 1 – 2 km) resurgent plutons in the Caetano caldera show evidence for both passive and forcible emplacement, whereas the more deeply emplaced (≥ 4 – 5 km) plutons in the southern Stillwater Range appear to have been passively emplaced. Stopped wall-rock blocks are rare to absent in all plutons.

Eruption of the upper Caetano Tuff was closely followed (≤ 0.1 Ma) by emplacement of two large resurgent granitic intrusions in the center of the caldera (Fig. 20; John et al., 2008; Colgan et al., 2011b). Both the 34.01 ± 0.05 Ma Carico Lake and the 33.90 ± 0.05 Ma Redrock Canyon plutons intruded the upper unit of the Caetano Tuff indicating that they ascended to within < 1 – 2 km of the surface. Sensitive high resolution ion microprobe (SHRIMP) U–Pb zircon and sanidine $^{40}\text{Ar}/^{39}\text{Ar}$ tuff eruption and pluton ages are indistinguishable, and the ages, whole-rock major, trace element, and rare earth element (REE) geochemistry, mineral chemistry, and oxygen isotope signatures (John et al., 2008, 2009; Watts et al., 2012a, 2012b) indicate a genetic connection between the Caetano Tuff and Carico Lake pluton (Fig. 25).

Caetano Caldera Rocks

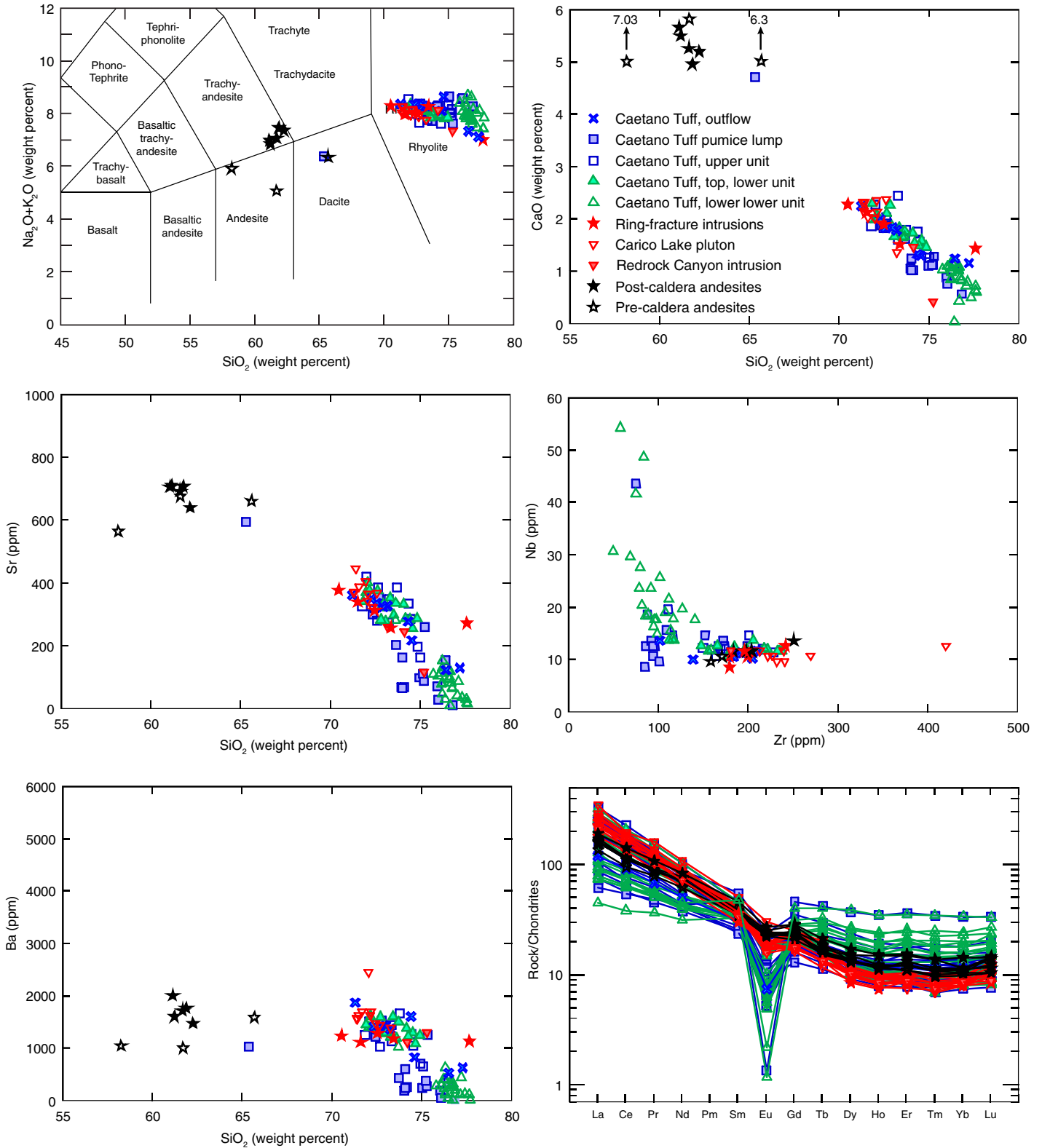


Figure 25 (on this and following three pages). Plots showing compositional variation of rocks related to the Caetano, Job Canyon, Poco Canyon, and Elevenmile Canyon calderas. For each caldera system, total alkali-silica, CaO-SiO₂, Sr-SiO₂, Ba-SiO₂, Zr-Nb, and rare earth elements are plotted for whole-rock samples. Major elements are normalized to 100% volatile free. Rare earth elements are normalized using chondrite values of Sun and McDonough (1989). (A) Caetano caldera, including pre- and post-caldera lava flows, and pumice lumps.

Job Canyon Caldera Rocks

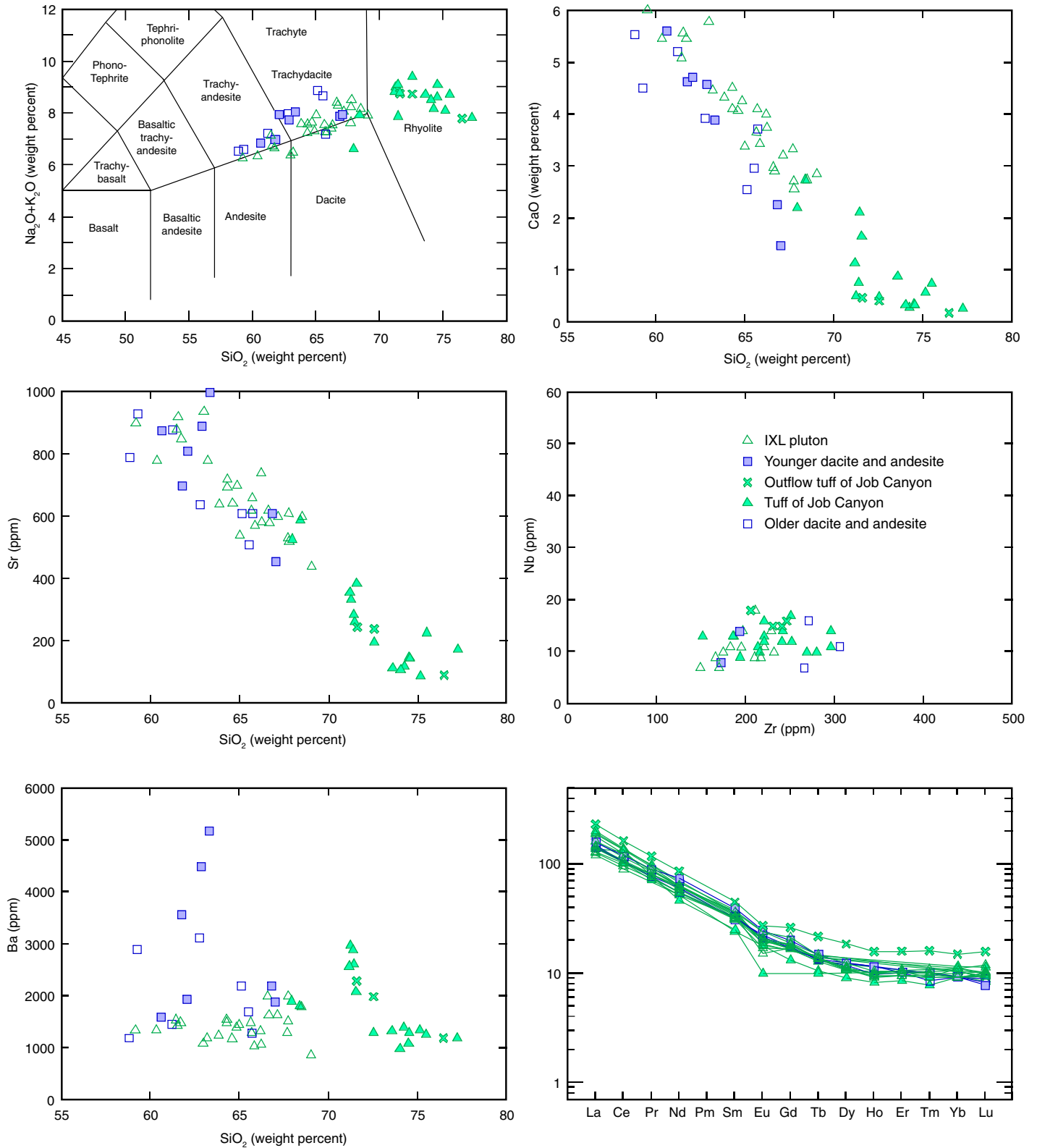


Figure 25 (continued). (B) Job Canyon caldera.

Elevenmile Canyon Caldera Rocks

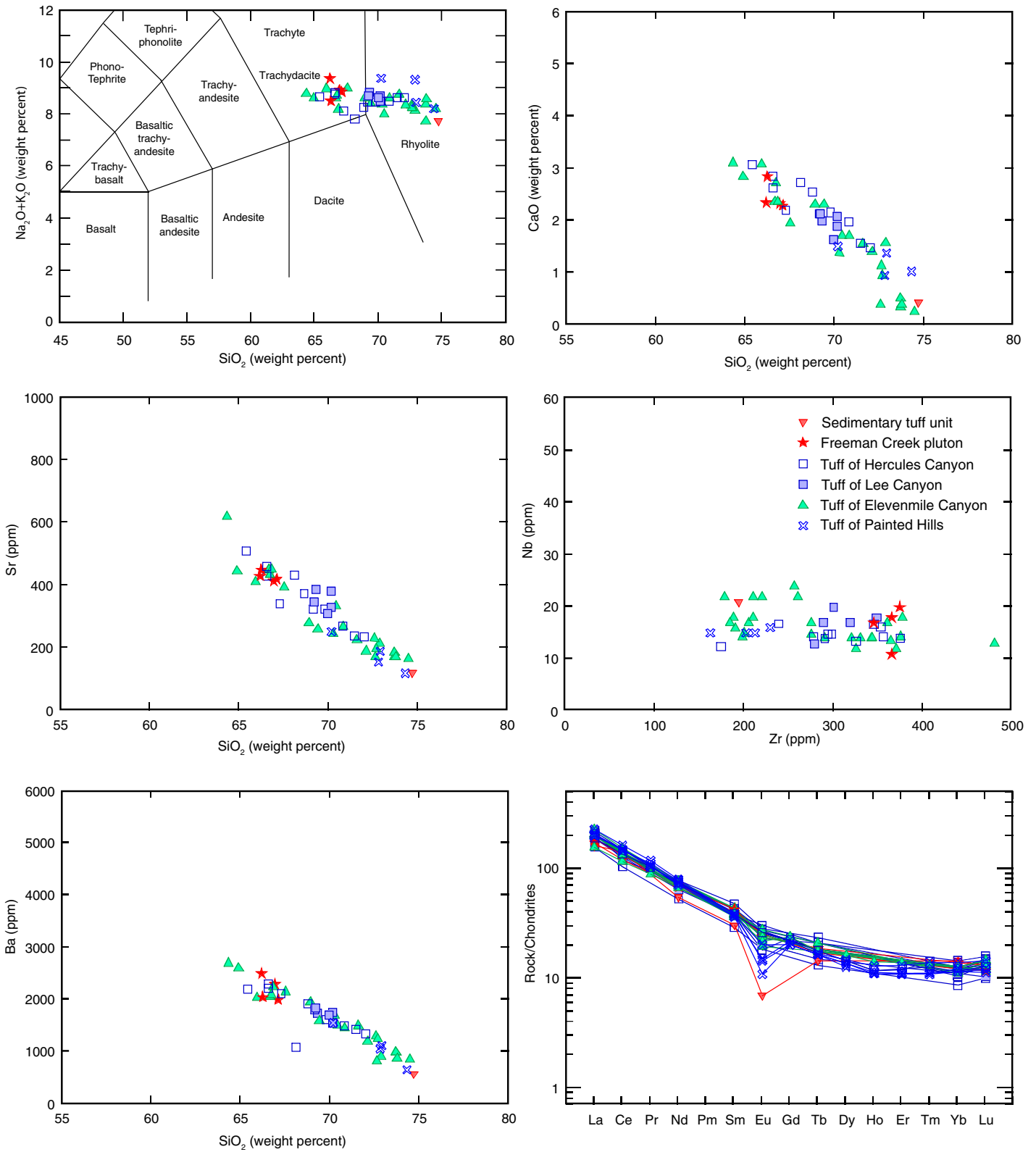


Figure 25 (continued). (C) Elevenmile Canyon caldera and tuff of Painted Hills.

Poco Canyon Caldera Rocks

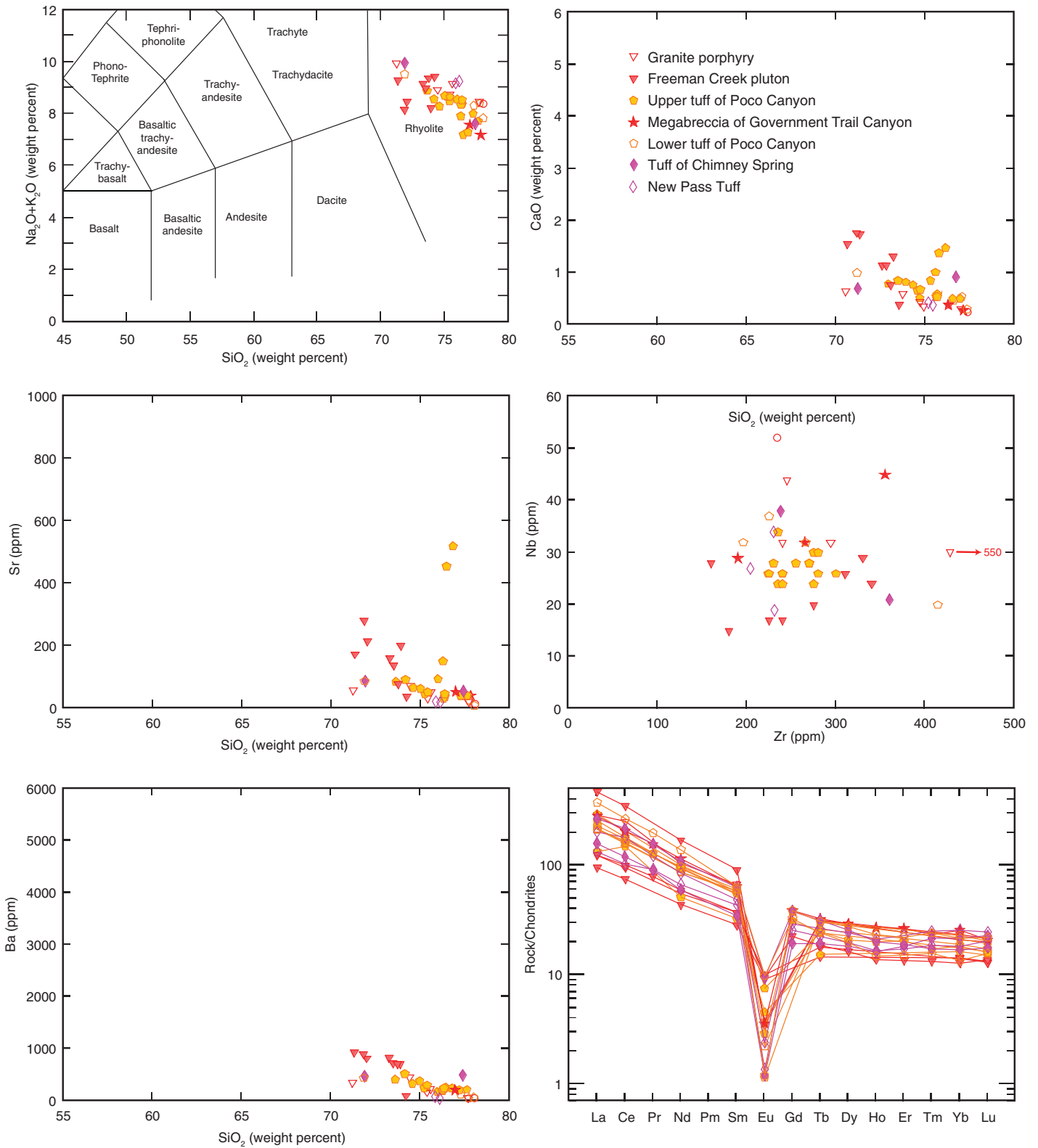


Figure 25 (continued). (D) Poco Canyon caldera, New Pass Tuff, and tuff of Chimney Spring. Data from John (1995), John et al. (2008), and this study.

To explain cessation of tuff eruption while similar magma remained in the chamber, we speculate that collapse along the exposed, upper, inward-dipping faults along with ponding of several kilometers of intracaldera Caetano Tuff self-sealed eruption conduits.

The early Redrock Canyon intrusion steeply cuts and did not deform the surrounding tuff. In contrast, Caetano Tuff wall rocks of the slightly younger Carico Lake pluton were rotated $\sim 90^\circ$ along both the north and south sides of the pluton indicating forcible emplacement of the pluton. The contrasting behavior of the two intrusions might be related to variations in their crystallinity; the Redrock Canyon intrusion contains $<30\text{--}40$ vol% phenocrysts in an altered felsite groundmass, whereas the Carico Lake pluton contains $\sim 60\%$ phenocrysts in an aplitic groundmass. Locally abundant xenoliths of high-grade metamorphosed pelitic rocks in the Carico Lake pluton are interpreted as mid-crustal wall rocks of the Caetano magma chamber that were incorporated in the pluton during its emplacement.

In the Stillwater Range, the large exposed thicknesses of the IXL pluton (4000–5000 m) and the Freeman Creek pluton (~ 2000 m), both with faulted lower contacts, suggest that they probably are the upper parts of batholiths.

The IXL pluton underlies and intrudes the lower part of the Job Canyon caldera (Fig. 23). It is zoned compositionally from a more felsic top with ~ 68 wt% SiO_2 to $\sim 59\text{--}60\%$ SiO_2 in its deepest exposures ~ 4.5 km below the pluton roof (John, 1995). The deeper parts of the pluton are compositionally similar to the post-collapse lava flow sequence in the Job Canyon caldera, whereas the top of the pluton is compositionally similar to the most mafic part of intracaldera tuff of Job Canyon (Fig. 25). Weighted mean SHRIMP zircon U-Pb ages for intracaldera tuff of Job Canyon and the IXL pluton are 29.43 ± 0.29 and 29.20 ± 0.29 Ma, respectively. Whole-rock geochemical data, U-Pb zircon ages, and field relations suggest a genetic relationship between caldera-filling rocks and the IXL pluton.

The composite Freeman Creek pluton underlies the Poco Canyon caldera and northern part of the Elevenmile Canyon caldera and is inferred to represent the plutonic roots of both caldera systems (John, 1995). The Freeman Creek pluton consists of two main phases, an earlier medium-grained hornblende biotite granodiorite (66–67 wt% SiO_2) and later coarse-grained porphyritic biotite granite (71–74 wt% SiO_2) that locally contains K-feldspar megacrysts as much as 1.5 cm long. On the basis of whole-rock geochemical data and distribution of the two phases, the granodiorite is inferred to be related to the Elevenmile Canyon caldera sys-

tem, whereas the granite phase likely is related to the Poco Canyon caldera (John, 1995). Available geochronologic data provide general support for this interpretation, but as noted previously, there are inconsistencies between stratigraphic relations and sanidine $^{40}\text{Ar}/^{39}\text{Ar}$ ages and between U-Pb and $^{40}\text{Ar}/^{39}\text{Ar}$ ages. Weighted mean SHRIMP zircon U-Pb ages of the granodiorite (25.01 ± 0.15 Ma) and granite (24.83 ± 0.20 Ma) phases of the Freeman Creek pluton are analytically indistinguishable, and only slightly younger than sanidine $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 25.07 ± 0.06 Ma, 25.05 ± 0.06 Ma, and 25.26 ± 0.07 Ma for the tuff of Elevenmile Canyon, tuff of Hercules Canyon, and upper unit of the tuff of Poco Canyon, respectively (Table 1). However, these tuffs have weighted mean SHRIMP zircon U-Pb ages of 25.27 ± 0.25 Ma, 25.43 ± 0.19 Ma, and 25.70 ± 0.19 Ma, respectively, which are somewhat older than their sanidine $^{40}\text{Ar}/^{39}\text{Ar}$ ages. Tuff U-Pb zircon ages being older than $^{40}\text{Ar}/^{39}\text{Ar}$ eruption ages is similar to age relations characteristic of some other large-volume silicic ignimbrites, such as in the Timber Mountain–Oasis Valley caldera complex, Nevada (Bindeman et al., 2006). The composite, porphyro-aphanitic–granitic ring-fracture dike along the north margin of the Freeman Creek pluton is the only exposed intrusion along a ring fracture in the Stillwater calderas (Fig. 21).

The absence of structural doming of the calderas in the Stillwater Range during magma resurgence and pluton emplacement may be due to relatively deep pluton emplacement and/or continued eruption of lava flows during magma resurgence. The top of the IXL pluton was emplaced at a depth of $\sim 4\text{--}5$ km. The depth of emplacement of the top of the Freeman Creek pluton is indeterminate due to structural uncertainties, but probably it was somewhat deeper ($>6\text{--}7$ km from the paleosurface) than the top of the IXL pluton (John, 1995). Eruption of the thick younger dacite and andesite unit after collapse of the Job Canyon caldera and prior to emplacement of the IXL pluton suggests that pressure caused by magma resurgence was released by magma eruption rather than by structural doming of caldera-related deposits; subsequent emplacement of the IXL pluton may have been at depths too great to uplift the caldera and the overlying lava flows.

Volumes of the Caetano and Stillwater Tuffs

The volume of erupted magma that formed the Caetano Tuff was estimated as ~ 1100 km³ by John et al. (2008). This estimate presumes a restored caldera area of ~ 280 km² after removing 100% post-caldera extension (Colgan et al., 2008), an average thickness of 3 km for intracaldera Caetano Tuff, and caldera collapse ~ 1 km

greater than the thickness of intracaldera tuff, which probably represents Caetano magma erupted as extracaldera fallout and ash-flow tuffs. Because of the extraordinary exposures of intracaldera rocks, this estimate is more robust than estimates for most other large-volume, caldera-forming tuffs, including those for the Stillwater calderas.

John (1995) estimated volumes of intracaldera rocks as ~ 200 km³ each for the Job Canyon and Poco Canyon calderas and ~ 600 km³ for the Elevenmile Canyon caldera. These estimates assumed circular calderas and used exposed widths of calderas and average thicknesses of caldera-related deposits to calculate volumes. The volume estimated for the Elevenmile Canyon caldera did not consider continuation of the caldera east of Dixie Valley into the southwestern Clan Alpine Mountains. Subsequent mapping indicates that the tuff of Elevenmile Canyon is >2000 m thick and contains megabreccia lenses in this area, which suggest that the Elevenmile Canyon caldera is much larger and the volume of erupted products (mostly the tuff of Elevenmile Canyon) was much greater than 600 km³, probably at least 1050–1400 km³ (Fig. 21; John, 1997; this study). The widespread distribution of outflow tuff from the Poco Canyon caldera (New Pass Tuff, tuff of Chimney Spring; Fig. 11) suggests that the previous volume estimate for this caldera also is likely too small.

Compositions of the Caetano and Stillwater Tuffs

Both the Caetano Tuff and the upper tuff of Poco Canyon have rhyolite to high-silica rhyolite compositions (72%–78% SiO_2) and generally contain 30–45 vol% phenocrysts mostly composed of quartz, sanidine, and plagioclase (Figs. 25 and 26, Supplemental Fig. 2⁶; Table 2; John, 1995; John et al., 2008). The Caetano Tuff has relatively uniform crystal content, is compositionally zoned upward from high-silica rhyolite to low-silica rhyolite, and except for sparse crystal-poor pumice clasts, lacks a crystal-poor eruptive phase. The earliest eruptive part of the Caetano magma is geochemically the most highly evolved, but these rocks still contain $\sim 25\%$ phenocrysts. The upper tuff of Poco Canyon is less well characterized than the Caetano Tuff, but available data suggest compositional

⁶Supplemental Figure 2. Plots showing whole-rock composition and phenocryst abundance in crystal-rich rhyolites of (A) the Caetano Tuff and (B) upper unit of the tuff of Poco Canyon. Data from John (1995), John et al. (2008), and this study. If you are viewing the PDF of this paper or reading it offline, please visit <http://dx.doi.org/10.1130/GES00867.S6> or the full-text article on www.gsapubs.org to view Supplemental Figure 2.

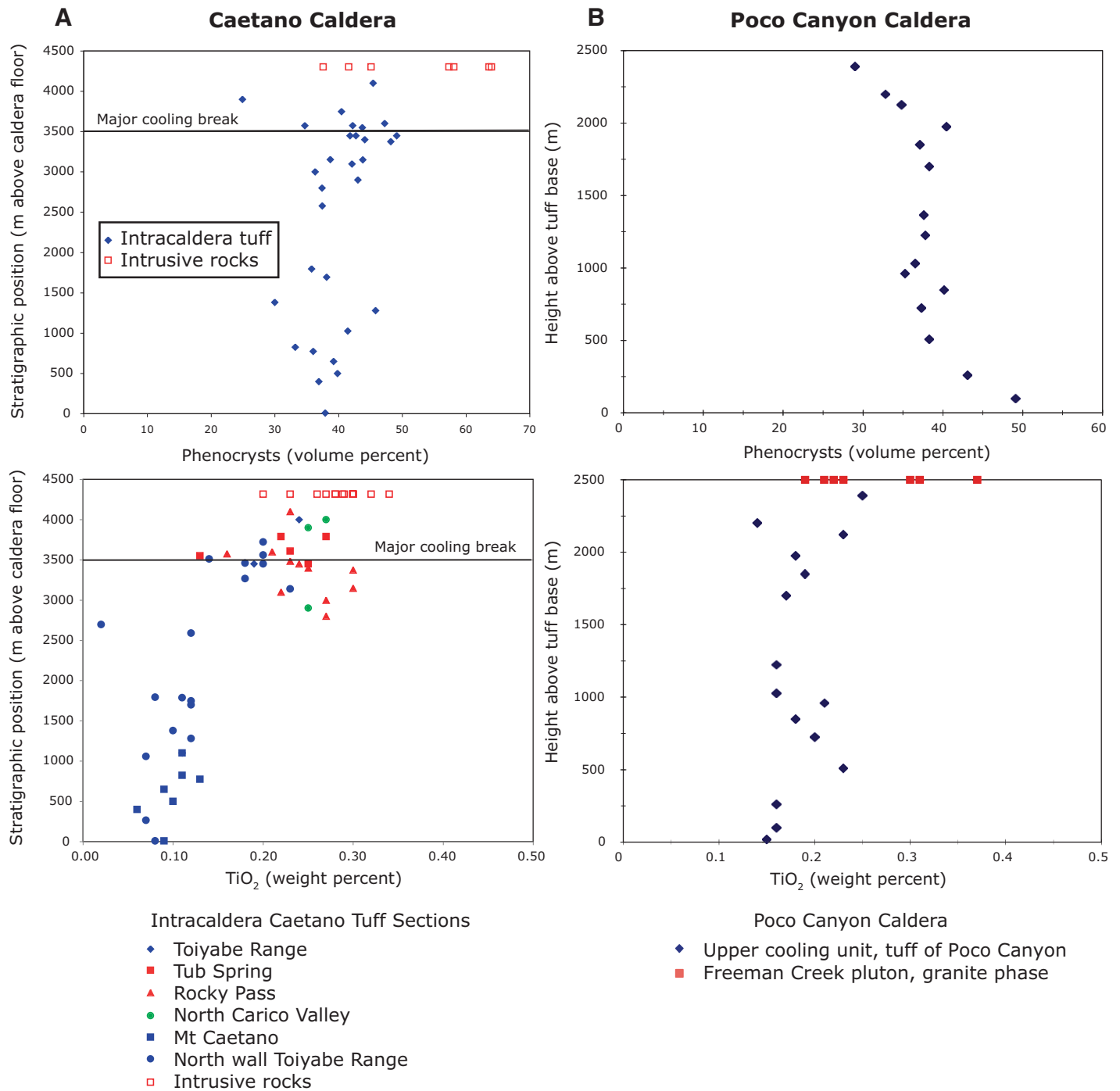


Figure 26. Plots showing whole-rock composition and phenocryst abundance in crystal-rich rhyolites of (A) the Caetano Tuff and (B) the upper unit of the tuff of Poco Canyon. Data from John (1995), John et al. (2008), and this study.

zoning is more irregular than in the Caetano Tuff. These data indicate that both the Caetano and Poco Canyon magma systems erupted large volumes (>200–1100 km³) of crystal-rich, mostly high-silica rhyolite.

The eruptive history of the Poco Canyon caldera is more complicated than that of the Caetano caldera (Fig. 22), but the earliest erupted mate-

rial, the crystal-rich, high-silica rhyolite (77%–78% SiO₂) lower unit of the tuff of Poco Canyon has distinctly highly evolved compositions. This small(?) volume tuff has not been identified outside of the caldera, and its eruption may have caused crustal sagging rather than caldera collapse (John, 1995). Alternatively, perhaps only the southern part of the caldera, where the tuff is

locally overlain by as much as 200 m of coarse collapse breccia, sedimentary rocks, and minor tuff (sandstone and sedimentary breccia unit, Fig. 22), collapsed in trapdoor fashion. Major caldera subsidence accompanied eruption of the overlying megabreccia of Government Trail Canyon, which is a highly evolved, crystal-poor (5%–10% phenocrysts) high-silica (77%–78%

SiO₂) rhyolite that contains numerous large wall-rock blocks including some composed of the lower tuff of Poco Canyon. The megabreccia is overlain by the volumetrically dominant, crystal-rich (25%–45%) upper unit of the tuff of Poco Canyon, which also is mostly high-silica rhyolite (74%–78% SiO₂) and which, based on ⁴⁰Ar/³⁹Ar ages, is correlative with the regionally widespread New Pass Tuff and tuff of Chimney Spring.

The Caetano Tuff and tuff of Poco Canyon are similar to several other large-volume ignimbrites in the western Nevada volcanic field, including the tuffs of Arc Dome and Toiyabe (John, 1992a), the tuff of Round Mountain (Shawe, 1995), the tuff of Moores Creek (Boden, 1994), and the Fish Creek Mountains Tuff (McKee, 1970). All are phenocryst rich (>20%–50%), composed of high-silica (73%–78% SiO₂) rhyolites, and have volumes of hundreds to more than one thousand cubic kilometers. None of these tuffs are known to have appreciable volumes of crystal-poor (<20 vol%) tuff. Crystal-rich rhyolite tuffs also constitute a substantial part of the central Nevada volcanic field (Best et al., 2013c). These tuffs are strikingly different from the two end members usually envisioned for large zoned caldera-forming ignimbrites: crystal-poor, high-silica rhyolite, which grades upward into lower-silica rhyolite, and crystal-rich dacite (monotonous intermediates of Hildreth [1981]) (Bachman and Bergantz, 2008; Huber et al., 2012), suggesting that some current models for large-volume ignimbrites are incomplete.

The compositional continuum from ash-flow tuff to post-collapse intrusions in the Caetano and Stillwater calderas and the overlap in their ages demonstrate that large volumes of residual, unerupted magma remained in their source magma chambers following ash-flow tuff eruption. Several other calderas in the western Nevada volcanic field, such as the lower and upper Mount Jefferson calderas of the Toquima caldera complex and the Fairview Peak caldera, erupted ring-fracture rhyolites that are indistinguishable in age from or slightly younger than their associated intracaldera tuffs (this study). The Campbell Creek caldera includes post-collapse rhyolite and dacite intrusions that are indistinguishable in age from and compositionally similar to intracaldera tuff (Henry et al., 2012a). Moreover, the dacite intrusions are petrographically and compositionally similar to sparse dacite pumice in the intracaldera tuff. Whether or not other calderas in the western Nevada volcanic field also retained significant volumes of unerupted residual magma is not known, because (1) they have not been studied sufficiently or (2) they are not as well

exposed as the tilted and extended Caetano and Stillwater calderas. Consequently, a significant group of western Nevada volcanic field calderas appear to have retained large volumes of residual magma after ash-flow eruption (see also Lipman, 2007; Zimmerer and McIntosh, 2013), which contrasts with interpretations that some calderas have small magma chambers that are completely evacuated during ash-flow eruption (Glazner et al., 2008; Tappa et al., 2011; Zimmerer and McIntosh, 2012).

Overlap of the Poco Canyon and Elevenmile Canyon Calderas

Overlap of the Elevenmile Canyon and Poco Canyon calderas both in time and space suggests that two nearby, shallow-level, silicic magma chambers were erupting nearly simultaneously. However, these calderas have discrete collapse features and different distributions of intracaldera rocks. In addition, lithic fragments in caldera-related pyroclastic rocks are notably different; pre-Tertiary lithic clasts are scarce to absent in rocks related to the Poco Canyon caldera but are nearly ubiquitous in the tuff of Elevenmile Canyon, suggesting that these rocks were erupted through different parts of the crust. Thicknesses of intracaldera rocks, locations of collapse features, and the location of the Freeman Creek pluton suggest that the magma chamber of the Elevenmile Canyon caldera was located somewhat south, west, or east of the magma chamber that formed the Poco Canyon caldera. Compositions of intracaldera rocks of the two calderas overlap at ~74 wt% SiO₂, but their trace element and REE contents have distinct trends and differences at similar silica contents (Fig. 25), which suggests that post-collapse caldera-filling rocks of both calderas did not tap different parts of a single, compositionally zoned magma chamber. Although the post-collapse sedimentary tuff unit in the southern and central parts of Elevenmile Canyon was inferred to be related to the Elevenmile Canyon caldera (John, 1995), its REE pattern more closely resembles REE patterns of units in the Poco Canyon caldera; consequently this unit may be a late, small-volume eruptive product of the Poco Canyon caldera. These data suggest the existence of two separate magma chambers beneath the Poco Canyon and Elevenmile Canyon calderas. Characteristics of their eruptive products preclude physical interaction of these magma chambers during caldera-forming activity, although eruption of one magma chamber may have triggered eruption of the other (e.g., Christiansen, 1979). However, multiple, compositionally varied intrusions within the Freeman Creek pluton suggest magma interaction during pluton emplacement.

Caldera Hydrothermal Systems

Large hydrothermal systems were active in Caetano and Job Canyon calderas shortly after their formation; a wide variety of hydrothermal features are exposed in caldera-filling rocks and in underlying plutons (Fig. 27; John and Pickthorn, 1996; John et al., 2011). However, the types of hydrothermal systems were significantly different in the two calderas, and no significant mineral deposits have been identified in either caldera.

In the Caetano caldera, hydrothermally altered rocks are exposed over >100 km² and represent depths from the paleosurface to >1–1.5 km throughout much of the west half of the caldera and farther east along the caldera's south wall (Figs. 27A and B; John et al., 2011). The hydrothermal system affected intracaldera Caetano Tuff—especially thin ash flows and interbedded lacustrine sediments in the upper unit of the tuff—the Redrock Canyon intrusion, and locally, Paleozoic wall rocks. The Carico Lake pluton postdates and truncates the hydrothermal system, which indicates (1) very rapid development of the hydrothermal system following caldera collapse and (2) that the hydrothermal system had a short lifespan (<<100 k.y.). The hydrothermal system formed in two stages. An early stage resulted in widespread, relatively homogeneous, advanced argillic alteration dominated by a quartz-kaolinite-pyrite assemblage that also includes local occurrences of dickite, alunite, and residual quartz alteration, mostly along the south caldera wall (Fig. 27A). Alteration intensity generally decreases to the north and locally to the west, where it grades into an intermediate argillic assemblage (montmorillonite, illite, and/or kaolinite after plagioclase phenocrysts), in which sanidine phenocrysts are unaltered, and eventually into unaltered tuff. The alteration assemblage apparently varies little vertically, which suggests relatively uniform temperatures ≤270 °C based on the absence of pyrophyllite. Early hydrothermal fluids are inferred to represent upflow of hot magmatic volatiles released from crystallizing residual Caetano magma at depth and condensation into the lake covering the caldera floor and the underlying water-saturated upper unit of Caetano Tuff. Magmatic volatile and heat fluxes likely were greatest along the south caldera wall where alunite, dickite, and residual quartz alteration was most strongly developed. The hydrothermal fluids flowed laterally, mostly to the north and east along the caldera wall, resulting in the pervasive advanced argillic alteration. The later stage of the hydrothermal system is characterized by oxidation of early-formed pyrite and cross-cutting hematite ± barite alteration (Fig. 27B). This stage probably represents decreased

fluxes of heat and magmatic volatiles, cooling, and downward collapse of the hydrothermal system, which resulted in narrow cross-cutting veins and hydrothermal breccias.

In the north block of the Job Canyon caldera, most caldera-filling rocks, including the pre-caldera lava flow sequence, and the upper parts of the underlying IXL pluton are pervasively hydrothermally altered (Fig. 27C; John and Pickthorn, 1996). Alteration is dominated by near-neutral pH, propylitic alteration assemblages of quartz–Na-plagioclase–sericite–chlorite and quartz–Na-plagioclase–sericite–chlorite–epidote (both \pm calcite, K-feldspar, and pyrite). Epidote-bearing rocks are present at paleodepths >3 –4 km except along the south margin of the caldera, where epidote at much shallower paleodepths (<1 km) likely indicates an upflow zone for hot hydrothermal fluids. Hydrothermal fluids were low salinity (mostly ≤ 1 wt% NaCl equivalent), neutral pH, and supercritical, and had low CO_2 contents. Fluid inclusions in quartz–calcite veins at 3–4 km paleodepth indicate fluid temperatures of ~ 290 – 315 °C. Oxygen and hydrogen isotope data indicate that the fluids were dominantly derived from heated ^{18}O -enriched meteoric groundwater and that large volumes of meteoric water reacted with caldera fill and with fractured upper parts of the IXL pluton to paleodepths greater than 7 km (John and Pickthorn, 1996).

The fluid chemistry of hydrothermal systems in the two calderas was fundamentally different with near-neutral pH fluids (K-feldspar stable) dominant in the Job Canyon caldera system and moderately acidic fluids (kaolinite stable) dominant in the Caetano caldera system. The Job Canyon caldera hydrothermal system was similar to alkali-chloride geothermal systems that form where magmatic sources of heat and volatiles are relatively deep, such as in the Taupo volcanic zone, New Zealand (Fig. 27C; Henley and Ellis, 1983; Simmons and Browne, 2000). In contrast, in the Caetano caldera, much shallower magmatic sources of heat and volatiles (represented by the Redrock Canyon intrusion and other residual magma) resulted in more intense and sustained flux of acid-forming magmatic volatiles into the Caetano hydrothermal system where it formed the acidic fluids typical of magmatic-hydrothermal systems (Figs. 27A and 27B; Henley and Ellis, 1983; Giggenbach, 1997; Hedenquist and Lowenstern, 1994). The relatively homogeneous alteration and lower alteration temperature in the Caetano caldera probably reflect a caldera-filling lake and underlying water-saturated, interbedded sedimentary rocks and poorly welded upper Caetano Tuff, which allowed widespread dispersion, dilution, and cooling of the acidic fluids.

SUMMARY AND COMPARISON WITH THE CENTRAL NEVADA AND INDIAN PEAK–CALIENTE CALDERA FIELDS

The western Nevada volcanic field, the westernmost part of the Great Basin ignimbrite province, covers $\sim 40,000$ km² in western and central Nevada. At least 23 calderas and their caldera-forming tuffs are reasonably well identified in the western Nevada volcanic field, and the presence of another 14 areally extensive, apparently voluminous ash-flow tuffs whose sources are unknown suggests a similar number of additional calderas. Most tuffs are rhyolites, with approximately equal numbers of sparsely porphyritic ($\leq 15\%$ phenocrysts) and abundantly porphyritic (~ 20 – 50% phenocrysts) tuffs. Both sparsely and abundantly porphyritic rhyolites commonly show chemical or petrographic evidence of normal compositional zoning from silicic bases to more mafic compositions upward. Several well-exposed, thick (2–4 km) intracaldera tuffs are almost entirely crystal-rich (30%–45% phenocrysts) rhyolite or high-silica rhyolite, an end member generally not included in current models for large zoned caldera-forming ignimbrites. The only possible “monotonous intermediate” tuff is the plagioclase-biotite-dominated, probably large-volume tuff of Dogskin Mountain–McCoy Mine, for which five

analyses range from 66.6% to 69.6% SiO_2 . This tuff is a composite unit consisting of as many as four cooling units possibly emplaced over several hundred thousand years, much longer than typical of caldera-forming ignimbrites. However, the Monotony Tuff, a definite monotonous intermediate of the central Nevada field, has a wider compositional range and also consists of multiple cooling units (Best et al., 2013c). Isom-type trachydacites are restricted to a few, small-volume, probably non-caldera forming tuffs. In contrast, monotonous intermediates and Isom-type trachydacites are common and particularly voluminous in the other two fields (Best et al., 2013b, 2013c). Volumes of individual tuffs are difficult to estimate in the western Nevada field, because many tuffs primarily filled their source calderas and/or flowed and were deposited in paleovalleys, so that they are irregularly distributed. Nevertheless, volumes of individual tuffs could range as large as ~ 3000 km³; at least four are probably greater than 1000 km³.

The tuffs and calderas range in age from 34.4 to 23.3 Ma, and one caldera formed at 19.5 Ma. Most caldera-forming eruptions clustered into five ~ 0.5 – 2.7 -Ma-long episodes separated by quiescent periods of ~ 1.4 Ma. Caldera-related as well as other magmatism migrated to the southwest with time, probably resulting from rollback of the shallow-dipping Farallon slab.

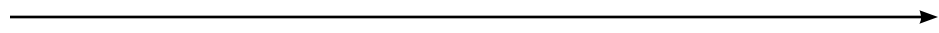


Figure 27 (on following page). Schematic north-south cross sections outlining the evolution of hydrothermal systems in the Caetano and Job Canyon calderas (modified from John et al. [2011] and John and Pickthorn [1996], respectively). (A) Early stage of Caetano hydrothermal system. Hot magmatic gases released from crystallizing residual Caetano magma rise up through the Redrock Canyon intrusion and along the south caldera wall and mix with cool lake water and with groundwater in interbedded poorly welded ash-flow tuffs, clastic and lacustrine sedimentary rocks, and breccias in the upper unit of Caetano Tuff. The resulting moderately acidic (pH < 4), moderate-temperature (< 270 °C) fluid flows laterally mostly through the upper unit of Caetano Tuff, hydrothermally altering the tuff and the upper parts of the Redrock Canyon intrusion mostly forming pervasive quartz-kaolinite-pyrite alteration that grades northward and westward to quartz-kaolinite or illite/montmorillonite alteration with relict sanidine. CCHS—Caetano caldera hydrothermal system. (B) Late stage of the Caetano hydrothermal system. Flux of hot magmatic gas has diminished or ended and hydrothermal fluids cool and oxidize as they continue to mix with cold oxygenated meteoric water. The hydrothermal system collapses with downflow of cooler denser fluids into previously altered and fractured upper Caetano Tuff and Redrock Canyon intrusion. These fluids oxidize pyrite formed during early alteration and precipitate hematite and barite in widespread fractures and breccias. (C) Job Canyon hydrothermal system showing fluid circulation paths in the Job Canyon caldera and IXL pluton. Recharge areas are in the north part of the caldera and outside the caldera in Tertiary volcanic rocks to the south and Triassic metasedimentary rocks to the north. Meteoric fluids circulate to depths of 7 km or more before rising along the sides and flowing into the uppermost part of IXL pluton and into the overlying caldera. An upwelling zone of high-temperature fluids was present along the south margin of the caldera that offsets the epidote isograd to higher structural levels, and a zone of mixing was present in the central part of the caldera between fluids flowing from the south that were rising buoyantly and cooler, more dense fluids with higher $\delta^{18}\text{O}$ that were flowing laterally from the north.

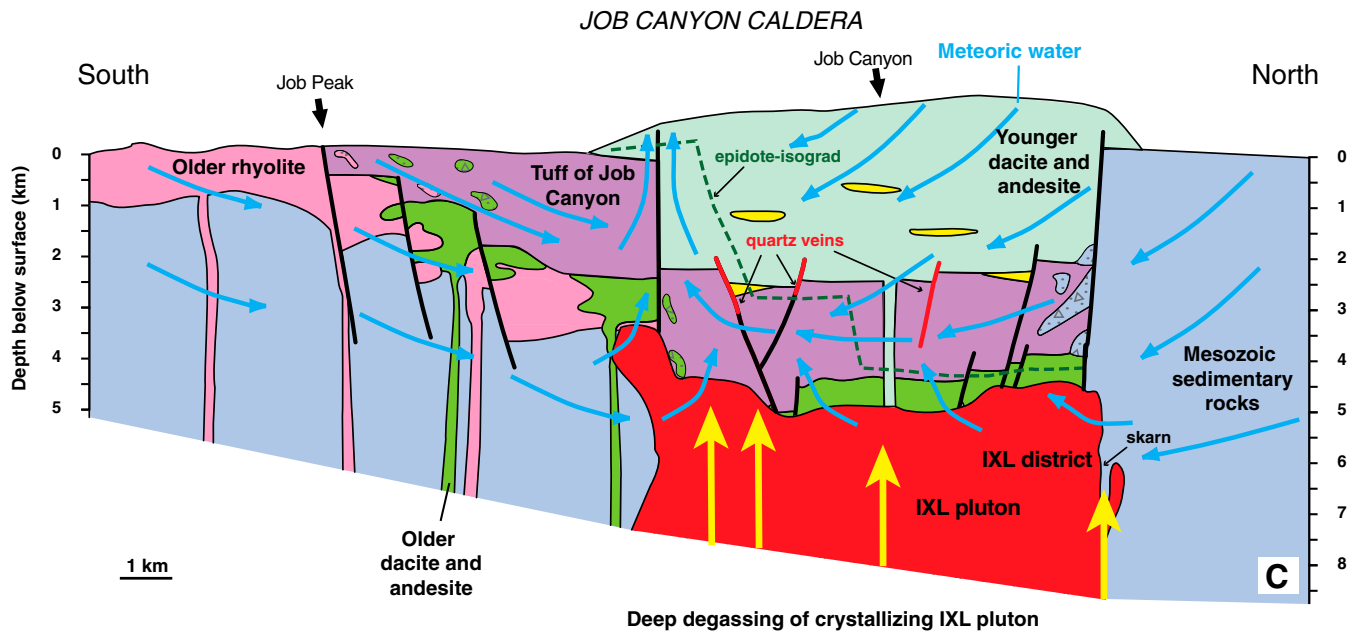
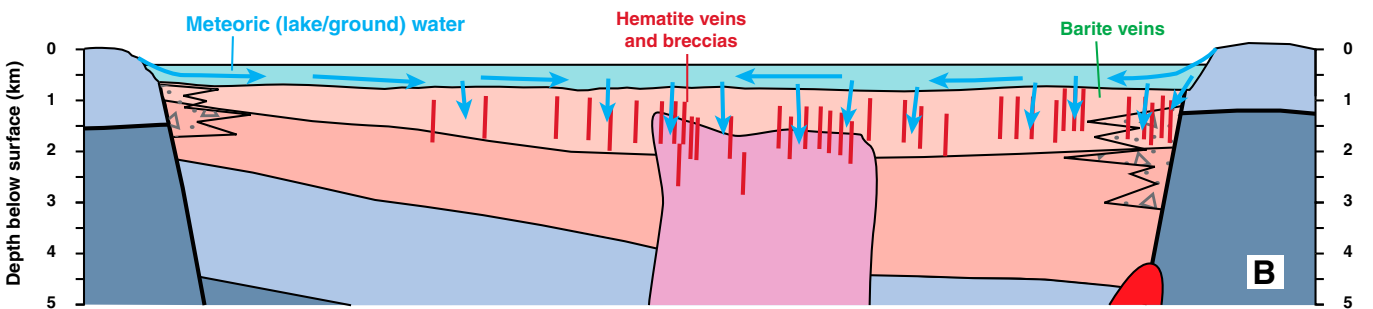
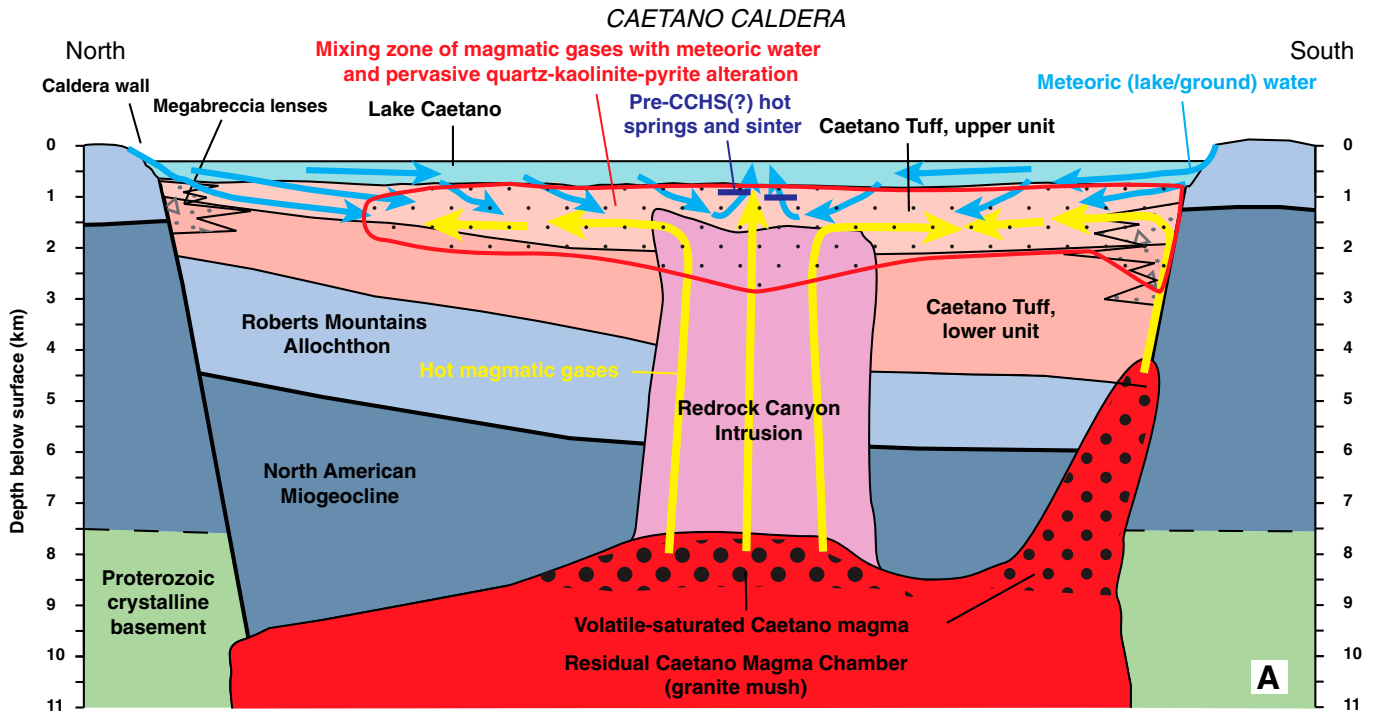


Figure 27.

Intermediate to silicic lavas or shallow intrusions commonly preceded caldera-forming eruptions by 1–6 Ma in any specific area. Calderas are restricted to the area northeast of what became the Walker Lane, although intermediate and silicic, effusive magmatism continued to migrate to the southwest across the Walker Lane, spatially and temporally transitioning into the ancestral Cascade arc (Cousens et al., 2008; John et al., 2012).

The caldera region of the western Nevada volcanic field extends across the boundary between accreted oceanic terrane on the west and rifted North American craton as defined by radiogenic isotopic data. The central Nevada field lies in the rifted craton, similar to the eastern part of the western field, whereas the Indian Peak–Caliente calderas are in the relatively stable craton.

Calderas are equant to slightly elongate, at least 12 km in diameter, and as much as 35 km in longest dimension. Exceptional three-dimensional exposure of the Stillwater caldera complex and Caetano caldera resulting from post-collapse extensional tilting show they subsided as much as 5 km as large piston-like blocks. Caldera walls were vertical to steeply inward dipping to depths ≥ 4 –5 km; topographic walls formed by slumping of wall rock into the caldera were only slightly outboard (≤ 1 km) of structural margins.

Most calderas had abundant post-collapse magmatism expressed as resurgent intrusions, ring-fracture intrusions, or intracaldera lavas that are closely related temporally (~0–0.5 Ma younger) to caldera formation. Granitoid intrusions exposed beneath several calderas were emplaced at paleodepths ranging from ≤ 1 to ~7 km and are compositionally similar to both intracaldera ash-flow tuffs and post-caldera lavas. For the tightly constrained Caetano caldera system, it seems inescapable that the caldera-forming magma chamber was large and not completely emptied during ash-flow eruptions.

Several calderas hosted large hydrothermal systems and caldera-related rocks underwent extensive hydrothermal alteration. Different types of hydrothermal systems (neutral-pH alkali-chloride and acid [low pH] magmatic-hydrothermal) may reflect proximity to (depth of) large resurgent intrusions. With the exception of the giant Round Mountain epithermal gold deposit, few caldera-related hydrothermal systems formed large mineral deposits. Major middle Cenozoic mineral deposits in and along the margins of the western Nevada volcanic field are mostly related to intrusions that preceded caldera-forming eruptions.

The western Nevada volcanic field has several additional apparent contrasts with the central Nevada and Indian Peak–Caliente complexes.

The most obvious and probably most certain are that the western Nevada field has a greater number of erupted ash-flow tuffs and resultant calderas, and mapped calderas are mostly smaller and more dispersed through the field. Only the Stillwater and Toquima caldera complexes cluster in a way somewhat similar to the Indian Peak–Caliente clusters. The differences between the volcanic fields could mostly reflect their different crustal settings, with magmas rising through and interacting with the crust in different ways. This is somewhat analogous to the presence and absence of calderas across the Walker Lane. Continued study is necessary to test these differences and their significance.

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