Eocene–Early Miocene paleotopography of the Sierra Nevada–Great Basin–Nevadaplano based on widespread ash-flow tuffs and paleovalleys

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ABSTRACT

The distribution of Cenozoic ash-flow tuffs in the Great Basin and the Sierra Nevada of eastern California (United States) demonstrates that the region, commonly referred to as the Nevadaplano, was an erosional highland that was drained by major west- and east-trending rivers, with a north-south paleodivide through eastern Nevada. The 28.9 Ma tuff of Campbell Creek is a voluminous (possibly as much as 3000 km³), petrographically and compositionally distinctive ash-flow tuff that erupted from a caldera in north-central Nevada and spread widely through paleovalleys across northern Nevada and the Sierra Nevada. The tuff can be correlated over a modern area of at least 55,000 km², from the western foothills of the Sierra Nevada to the Ruby Mountains in northeastern Nevada, present-day distances of ~280 km west and 300 km northeast of its source caldera. Corrected for later extension, the tuff flowed ~200 km to the west, downvalley and across what is now the Basin and Range–Sierra Nevada structural and topographic boundary, and ~215 km to the northeast, partly upvalley, across the inferred paleodivide, and downvalley to the east. The tuff also flowed as much as 100 km to the north and 60 km to the south, crossing several east-west divides between major paleovalleys. The tuff of Campbell Creek flowed through, and was deposited in, at least five major paleovalleys in western Nevada and the eastern Sierra Nevada. These characteristics are unusual compared to most other ash-flow tuffs in Nevada that also flowed great distances downvalley, but far less east and north-south; most tuffs were restricted to one or two major paleovalleys. Important factors in this greater distribution may be the great volume of erupted tuff and its eruption after ~3 Ma of nearly continuous, major pyroclastic eruptions near its caldera that probably filled in nearby topography.

Distribution of the tuff of Campbell Creek and other ash-flow tuffs and continuity of paleovalleys demonstrates that (1) the Basin and Range–Sierra Nevada structural and topographic boundary did not exist before 23 Ma; (2) the Sierra Nevada was a lower, western ramp to the Nevadaplano; and (3) any faulting before 23 Ma in western Nevada, including in what is now the Walker Lane, and before 29 Ma in northern Nevada as far east as what is now the Ruby Mountains metamorphic core complex, was insufficient to disrupt the paleodrainages. These data are further evidence that major extension in Nevada occurred predominantly in the late Cenozoic.

Characteristics of paleovalleys and tuff distributions suggest that the valleys resulted from prolonged erosion, probably aided by the warm, wet Eocene climate, but do not resolve the question of the absolute elevation of the Nevadaplano. Paleoasl valleys existed at least by ca. 50 Ma in the Sierra Nevada and by 46 Ma in northeastern Nevada, based on the age of the oldest paleovalley-filling sedimentary or tuff deposits. Paleovalleys were much wider (5–10 km) than they were deep (to 1.2 km; greatest in western Nevada and decreasing toward the paleo-Pacific Ocean) and typically had broad, flat bottoms and low-relief interfluves. Interfluves in Nevada had elevations of at least 1.2 km because paleovalleys were that deep. The gradient from the caldera eastward to the inferred paleodivide had to be sufficiently low so that the tuff could flow upstream more than 100 km. Two Quaternary ash-flow tuffs where topography is nearly unchanged since eruption flowed similar distances as the mid-Cenozoic tuffs at average gradients of ~2.5–8 m/km. Extrapolated 200–300 km (pre-extension) from the Pacific Ocean to the central Nevada caldera belt, the lower gradient would require elevations of only 0.5 km for valley floors and 1.5 km for interfluves. The great eastward, upvalley flow is consistent with recent stable isotope data that indicate low Oligocene topographic gradients in the Nevadaplano east of the Sierra Nevada, but the minimum elevations required for central Nevada are significantly less than indicated by the same stable isotope data.

Although best recognized in the northern and central Sierra Nevada, early to middle Cenozoic paleodrainages may have crossed the southern Sierra Nevada. Similar early to middle Cenozoic paleodrainages existed from central Idaho to northern Sonora, Mexico, and persisted over most of that region until disrupted by major Middle Miocene extension. Therefore, the Nevadaplano was the middle part of an erosional highland that extended along at least this length. The timing of origin and location of this more all-encompassing highland indicates that uplift was predominantly a result of Late Cretaceous (Sevier) contraction in the north and a combination of Late Cretaceous–early Cenozoic (Sevier and Laramide) contraction in the south.

INTRODUCTION

It is now well established that the Great Basin (western United States) was an erosional highland in the middle Cenozoic, following crustal thickening during Mesozoic–early Cenozoic shortening, batholith emplacement, and shallow slab subduction (e.g., Dilek and Moores, 1999; Humphreys et al., 2003; DeCelles, 2004;...
Dickinson, 2006; Best et al., 2009; Cassel et al., 2009a, 2009b; Colgan and Henry, 2009; Ernst, 2010; Henry and Faulds, 2010). Major river systems drained this highland both west to the Pacific Ocean and east to the Uinta Basin (Fig. 1; Henry, 2008; Cassel et al., 2009a, 2009b; Henry and Faulds, 2010). The absolute elevation and structural-topographic evolution of this highland in the mid-Cenozoic remain highly controversial, however, particularly the paleoelevation of what is now the Sierra Nevada (Wakabayashi and Sawyer, 2001; Mulch et al., 2006, 2008; Cassel et al., 2009a, 2009b; Molnar, 2010) and the timing of extension in northeastern Nevada, especially around the Ruby Mountains–East Humboldt Range metamorphic core complex (McGrew and Snee, 1994; Snake et al., 1997; McGrew et al., 2000; Howard, 2003; Colgan and Henry, 2009; Druschke et al., 2009; Colgan et al., 2010; Henry et al., 2011; Mix et al., 2011).

An ancestral Sierra Nevada probably formed during arc magmatism (Wernicke et al., 1996), and these batholithic rocks were exhumed during the Late Cretaceous–early Cenozoic (Dimitr, 1990; House et al., 1997; Cecil et al., 2006). Based on stable isotope and organic molecule paleothermometry and altimetry, together with detrital zircon geochronology, the Eocene–Oligocene Sierra Nevada is interpreted to have been at approximately the same elevation (~2.5–3 km at the latitude of Lake Tahoe) as it is today (Horton et al., 2004; Mulch et al., 2006, 2008; Cassel et al., 2009a, 2009b; Cecil et al., 2010; Hren et al., 2010). In contrast, Huber (1981), Unruh (1991), Wakabayashi and Sawyer (2001), Jones et al. (2004), Stock et al. (2004, 2005), and Clark et al. (2005) concluded from dated stream incision and gradients that the Sierra Nevada was much lower in the Eocene (≤1 km) and attained its present elevation following 1.5–2.5 km of uplift during the Late Miocene and Pliocene. Numerical models of bedrock channel erosion support two episodes of uplift, in the Late Cretaceous or middle Cenozoic, and the Late Miocene (Pelletier, 2007). Provenance studies and the distribution of ash-flow tuffs and paleovalleys unequivocally demonstrate that Eocene–Oligocene rivers drained from central Nevada across the northern and central Sierra Nevada to the Pacific Ocean, but these data do not constrain the absolute surface elevation of the Sierra Nevada at the time (Bate- man and Wahrhaftig, 1966; Yeend, 1972; Faulds et al., 2005; Garol et al., 2005; Henry, 2008; Henry and Faulds, 2010; Cassel et al., 2012b). In contrast, Cecil et al. (2010) and Lechler and Niemi (2011) interpreted the predominance of Mesozoic, i.e., Sierra Nevada batholith age, detrital zircons in Eocene sedimentary deposits of the Sierra Nevada to indicate that the Eocene bedwaters were in the modern Sierra Nevada, but is preserved in only a few locations. Only the most distal sections, Skillman Flat, Welches Canyon, and East Humboldt Range (Figs. 1 and 5C), are poorly welded and glassy throughout. Large pumice fragments, as much as 30 cm long, are common (Figs. 5D–5F). As with many other tuffs in western Nevada (Henry and Faulds, 2010), the tuff of Campbell Creek commonly shows primary dips where it compacted against moderate to steep topography in paleo-valleys (Fig. 5D). The tuff of Campbell Creek is slightly to rarely moderately rheomorphic within the caldera and in several proximal locations. Stretched pumice is the most common rheomorphic feature.

The tuff of Campbell Creek is a compositionally nearly homogeneous, high-SiO₂ rhyolite with 75–77% SiO₂. Major elements vary little, whereas trace elements show greater variation (Figs. 2 and 3). Whole-rock samples from this study and those of Brooks et al. (2003, 2008) have nearly identical compositions; minor differences probably reflect differences in analytical methods. Analytical bias is apparent in P₂O₅ and Nb, which are systematically higher in the Brooks et al. (2003, 2008) data set from Haskell Peak. Two whole-rock samples from Haskell Peak analyzed for this study have P₂O₅ and Nb concentrations similar to all other tuff of Campbell Creek samples.

Both major and trace elements of basal and upper samples of outflow tuff from individual sections are indistinguishable. Two individual pumice samples (H09-63, H09-65) from the Reese River Narrows section are indistinguishable from each other and from two bulk samples (H01-139, H01-140). Variations in the more mobile oxides (Na₂O, K₂O, CaO, MgO) most likely reflect hydration of nonwelded vitrophyres, especially in the westernmost (H07-230, Skillman) and easternmost (H07-95, H10-79, Welches Canyon, East Humboldt Range) distal, nonwelded, and strongly hydrated vitrophyres.

Despite the lack of observed compositional variation in individual sections, trace elements for the entire group of samples vary, and a few dacite pumice fragments are present in intra-caldera tuff. Total Zr concentration ranges from 136 to 83 ppm (mostly ≥98 ppm) in whole-rock samples and is as low as 76 ppm in glass shards (analyzed by Cassel et al., 2009a). This range probably results from crystallization and separation of zircon and allows use of Zr as an indicator of magma evolution. Samples with high Zr concentrations (H07-230 and H01-61) are least evolved, whereas samples with the lowest Zr concentrations (H08-35 and the shard samples of Cassel et al., 2009a) are most evolved. TiO₂, Sr, Ba, Eu, and Hf decrease linearly with decreasing Zr, whereas Rb and Th increase.

**TUFF OF CAMPBELL CREEK**

**Correlation: Petrography, Chemical Composition, Age, and Remanent Magnetization**

The 28.9 Ma tuff of Campbell Creek is a petrographically and compositionally distinctive ash-flow tuff that erupted from a caldera in north-central Nevada and spread widely in paleovalleys across northern Nevada and the eastern Sierra Nevada (Fig. 1). Correlation of the tuff, also called the C unit of the Bates Mountain Tuff in central Nevada (Gromme et al., 1972), from the Sierra Nevada to northeastern Nevada is based on its stratigraphic position, distinctive phenocryst assemblage, composition, ⁴⁰Ar/³⁹Ar age, and paleomagnetic direction (Figs. 2 and 3; Tables 1 and 2). The tuff contains ~5%–10% phenocrysts of sanidine, plagioclase, diopside, and biotite (Fig. 4). Large glass shards as much as 1 mm diameter and glass lumps to ~1 cm diameter are common.

All but the most distal sections, regardless of their thickness, show welding and crystallization zoning typical of ash-flow tuffs (Fig. 5). A 1–3-m-thick, poorly welded glassy base passes upward to a densely welded vitrophyre, which is black to white depending on the degree of hydration (Fig. 5B). Most of the tuff is densely welded and devitrified. An upper nonwelded zone was probably common before erosion,
Figure 1. Digital elevation map of western Nevada and the Sierra Nevada showing the distribution of known paleovalleys, including segments (from Lindgren, 1911; Jenkins, 1932; Faulds et al., 2005; Garside et al., 2005; Henry and Faulds, 2010), a proposed paleodivide (Henry, 2008), and known locations of the tuff of Campbell Creek and probably related underlying tuff E. The tuff of Campbell Creek erupted from the Campbell Creek caldera in the Desatoya Mountains and flowed as much as 200 km westward down paleovalleys. In contrast, the 760 ka Bishop Tuff erupted at the western margin of the Basin and Range and flowed a maximum of ~50 km into basins to the north, northeast, and southeast (Bailey, 1989). The queried caldera in the Carson Sink is the suggested possible source of the Nine Hill Tuff (Best et al., 1989).
Figure 2. Plots of major elements, rare earth elements (REEs; normalization values from Sun and McDonough, 1989), and Zr illustrating narrow compositional range of high-SiO₂ rhyolitic tuff of Campbell Creek. Major elements show essentially no zoning, whereas trace elements, especially Zr, show some zoning. TiO₂, Sr, Ba, Eu, and Hf decrease linearly with decreasing Zr, whereas Rb and Th increase (only TiO₂ and Rb shown). H10-79 is sample from easternmost occurrence in East Humboldt Range (Fig. 1).
Figure 3. Plots of Zr versus Nb. (A) Tuff of Campbell Creek and related tuff E have distinctive low Zr and Nb concentrations compared to other ash-flow tuffs of the ignimbrite flareup of Nevada. Comparison data are from Deino (1985), John (1992, 1995), Best et al. (1995), and Henry (2008). (B) Tuff of Campbell Creek data. Higher Nb concentrations in analyses from Brooks et al. (2003, 2008) suggest a systematic bias. Lower Zr concentrations in glass separates (Cassel et al., 2009a) probably reflect presence of zircon in the whole-rock samples.

These patterns probably indicate fractionation of titanomagnetite or ilmenite and plagioclase ± sanidine, all of which are present in the tuff of Campbell Creek.

Sparse, large dacite or low-SiO₂ rhyolite pumice fragments present in intracaldera tuff also indicate compositional variation (Fig. 5F; Table 1). These pumice fragments are more abundantly porphyritic than the host rhyolite, with as much as 40% phenocrysts, mostly of plagioclase, with lesser sanidine, biotite, minor vermicular quartz, and sparse hornblende. The dacite has significantly greater TiO₂, Al₂O₃, FeO, MgO, CaO, Sr, Zr, and Ba and lower SiO₂, Rb, Nb, Th, and U abundances than the main rhyolite (Table 1; H11-6).

We interpret the compositional variation of the entire group of samples, but the lack of variation in individual sections to indicate that different ash flows tapped different parts of a compositionally heterogeneous magma chamber. These different ash flows either did not all flow to the same locations, or mixed sufficiently to homogenize before deposition, or a combination of both. For example, the 4-km-thick intracaldera Caetano Tuff, whose caldera is 110 km northeast of the Campbell Creek caldera (Fig. 1), is zoned from high- to low-SiO₂ rhyolite and is even more strongly zoned in trace elements, whereas outflow sections show little zoning (John et al., 2008a, 2008b). More comprehensive study of the intracaldera tuff of Campbell Creek, which might preserve more complete zonation, would be worthwhile.

The tuff of Campbell Creek is distinct relative to other tuffs in the Great Basin in having low concentrations of Zr (~80–130 ppm) and Nb (~9–13 ppm) (Fig. 3; John, 1992; Deino, 1985; Best et al., 1995; Maughan et al., 2002; Brooks et al., 2003, 2008; Henry, 2008). Higher Zr concentrations in whole-rock samples compared to glass shards probably reflect presence of zircon in the whole-rock samples. This interpretation is supported by the linear relationship between Zr and Hf, which is incorporated in zircon by simple substitution for Zr (Hoskin and Schaltegger, 2003). Two Great Basin ash-flow tuffs with Zr and Nb concentrations similar to those of the tuff of Campbell Creek can be distinguished in other ways. The tuff of Toiyabe, the distribution of which partly overlaps that of the tuff of Campbell Creek, is abundantly porphyritic and 23.3 Ma (John, 1992; Henry and Faulds, 2010). The tuff of Big Cottonwood Canyon is petrographically similar to the tuff of Campbell Creek, but is 40.3 Ma and restricted to northeastern Nevada (Henry, 2008).

Over much of its distribution, the tuff of Campbell Creek is underlain by a slightly older but otherwise nearly indistinguishable tuff (Figs. 1, 2, and 3), informally referred to as tuff E (Brooks et al., 2003, 2008). Tuff E contains 10%–22% of the same phenocrysts as the tuff of Campbell Creek and is a high-SiO₂ rhyolite with similar major and trace element concentrations, including low Zr and Nb (Brooks et al., 2003, 2008). Tuff E can be distinguished from the tuff of Campbell Creek primarily by its stratigraphic position and slightly older ⁴⁰Ar/³⁹Ar age (29.2 Ma; Table 2), and to a lesser extent by its greater range in phenocryst abundance and less vermicular quartz phenocrysts that better
TABLE 1. CHEMICAL ANALYSES OF THE TUFF OF CAMPBELL CREEK AND RELATED ROCKS

<table>
<thead>
<tr>
<th>Sample</th>
<th>Location</th>
<th>Position</th>
<th>Sample Position</th>
<th>SiO₂</th>
<th>TiO₂</th>
<th>Al₂O₃</th>
<th>FeO*</th>
<th>MnO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na₂O</th>
<th>K₂O</th>
<th>P₂O₅</th>
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<td>lower</td>
<td>vitrophyre</td>
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(continued)
## TABLE 1. CHEMICAL ANALYSES OF THE TUFF OF CAMPBELL CREEK AND RELATED ROCKS (continued)

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<th>Longitude (W)</th>
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<td>0.76</td>
<td>41.05</td>
<td>15.14</td>
<td>159.147</td>
<td>77</td>
<td>85</td>
<td>15</td>
<td>3.17</td>
</tr>
<tr>
<td>Dogskin Mtn</td>
<td>H01-61</td>
<td>devitrified</td>
<td>39.922</td>
<td>119.735</td>
<td>75.57</td>
<td>0.118</td>
<td>13.12</td>
<td>1.20</td>
<td>0.03</td>
<td>0.76</td>
<td>41.05</td>
<td>15.14</td>
<td>159.147</td>
<td>77</td>
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<td>15</td>
<td>3.17</td>
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<tr>
<td>Dogskin Mtn</td>
<td>H02-24</td>
<td>devitrified</td>
<td>39.91787</td>
<td>118.03166</td>
<td>75.20</td>
<td>0.118</td>
<td>13.12</td>
<td>1.20</td>
<td>0.03</td>
<td>0.76</td>
<td>41.05</td>
<td>15.14</td>
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<td>15</td>
<td>3.17</td>
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<tr>
<td>Tule Peak</td>
<td>H00-10</td>
<td>devitrified</td>
<td>39.78746</td>
<td>117.63938</td>
<td>76.92</td>
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<td>13.12</td>
<td>1.20</td>
<td>0.03</td>
<td>0.76</td>
<td>41.05</td>
<td>15.14</td>
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<td>15</td>
<td>3.17</td>
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<tr>
<td>White Rock Canyon</td>
<td>H08-58</td>
<td>devitrified</td>
<td>39.78675</td>
<td>117.14153</td>
<td>76.77</td>
<td>0.118</td>
<td>13.12</td>
<td>1.20</td>
<td>0.03</td>
<td>0.76</td>
<td>41.05</td>
<td>15.14</td>
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<td>3.17</td>
</tr>
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<td>Shoshone Meadows section, Clan Alpine Mts</td>
<td>H09-75</td>
<td>devitrified</td>
<td>39.94387</td>
<td>117.14153</td>
<td>75.75</td>
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<td>13.12</td>
<td>1.20</td>
<td>0.03</td>
<td>0.76</td>
<td>41.05</td>
<td>15.14</td>
<td>159.147</td>
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<td>Shoshone Meadows section, Clan Alpine Mts</td>
<td>CA-31</td>
<td>devitrified</td>
<td>39.94387</td>
<td>117.14153</td>
<td>75.75</td>
<td>0.118</td>
<td>13.12</td>
<td>1.20</td>
<td>0.03</td>
<td>0.76</td>
<td>41.05</td>
<td>15.14</td>
<td>159.147</td>
<td>77</td>
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<td>15</td>
<td>3.17</td>
</tr>
<tr>
<td>Reese River Narrows</td>
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<td>39.94387</td>
<td>117.14153</td>
<td>75.75</td>
<td>0.118</td>
<td>13.12</td>
<td>1.20</td>
<td>0.03</td>
<td>0.76</td>
<td>41.05</td>
<td>15.14</td>
<td>159.147</td>
<td>77</td>
<td>85</td>
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<td>3.17</td>
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<tr>
<td>Reese River Narrows</td>
<td>H09-65</td>
<td>devitrified</td>
<td>39.94387</td>
<td>117.14153</td>
<td>75.75</td>
<td>0.118</td>
<td>13.12</td>
<td>1.20</td>
<td>0.03</td>
<td>0.76</td>
<td>41.05</td>
<td>15.14</td>
<td>159.147</td>
<td>77</td>
<td>85</td>
<td>15</td>
<td>3.17</td>
</tr>
</tbody>
</table>
TABLE 1. CHEMICAL ANALYSES OF THE TUFF OF CAMPBELL CREEK AND RELATED ROCKS (continued)

<table>
<thead>
<tr>
<th>Sample</th>
<th>Location</th>
<th>Position</th>
<th>Chemistry</th>
<th>XRF (ppm)</th>
<th>IC-PM (ppm)</th>
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<tr>
<td>H01-139</td>
<td>Reese River Narrows</td>
<td>lower</td>
<td>vitrophyre</td>
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<tr>
<td>H01-140</td>
<td>Reese River Narrows</td>
<td>upper</td>
<td>devitrified</td>
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<tr>
<td>H09-81</td>
<td>Clipper Gap</td>
<td>base</td>
<td>vitrophyre</td>
<td></td>
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<td>H09-82</td>
<td>Clipper Gap</td>
<td>top</td>
<td>devitrified</td>
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<td></td>
</tr>
<tr>
<td>06-DJ-26</td>
<td>Caetano caldera</td>
<td>lower</td>
<td>vitrophyre</td>
<td></td>
<td></td>
</tr>
<tr>
<td>H07-95</td>
<td>Welches Canyon</td>
<td>upper</td>
<td>devitrified</td>
<td></td>
<td></td>
</tr>
<tr>
<td>H10-79</td>
<td>E Humboldt Range</td>
<td>base</td>
<td>vitrophyre</td>
<td></td>
<td></td>
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<tr>
<td>H00-101</td>
<td>Intracaldera</td>
<td>top</td>
<td>devitrified</td>
<td></td>
<td></td>
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<tr>
<td>H08-33</td>
<td>Intracaldera</td>
<td>lower</td>
<td>clay alteration</td>
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<tr>
<td>H08-35</td>
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<td>upper</td>
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<td></td>
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<tr>
<td>H11-188</td>
<td>Intracaldera</td>
<td>in</td>
<td>megabreccia</td>
<td></td>
<td></td>
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</table>

**SiO₂** 76.23 76.77 76.55 76.52 76.89 75.12 77.76 77.36 76.78 77.40 75.96 76.80
**TiO₂** 0.103 0.103 0.094 0.101 0.113 0.112 0.113 0.096 0.119 0.101 0.138 0.08
**FeO** 1.06 1.01 1.01 1.07 0.79 1.08 1.07 1.02 1.09 1.08 1.26 1.52
**MnO** 0.051 0.051 0.069 0.042 0.017 0.017 0.017 0.012 0.015 0.015 0.039 0.073 0.07
**MgO** 0.26 0.07 0.08 0.07 0.07 1.55 1.78 0.19 0.17 0.24 0.21 0.09 0.0
**CaO** 0.79 0.71 0.67 0.63 0.69 1.22 2.24 0.63 0.80 1.49 1.06 0.65
**Na₂O** 3.19 3.69 3.26 3.68 3.67 2.05 0.33 4.85 4.81 4.45 4.14 4.84
**K₂O** 5.15 4.82 5.44 5.19 5.40 2.99 4.85 4.81 4.45 4.14 4.84
**P₂O₅** 0.015 0.012 0.015 0.018 0.023 0.022 0.022 0.018 0.024 0.031 0.029 <0.01
**LOI** 2.84 0.95 3.50
**Total** 93.86 97.67 96.57 98.58 98.77 90.28 89.39 95.77 96.81 96.51 94.68 98.70

**Sc** 2 3 2 2 3 3 3 3 3 3 3 3
**V** 3 1 3 7 13 7 5 2 4 8 10
**Cr** 1 1 3 3 1 2 4 1 2 2 5
**Ni** 0 0 1 2 5 0 3 0 0 2
**Cu** 1 1 1 0 2 1 0 0 0 3 18
**Zn** 78.35 44.34 44.27 42.74 42.39 39.43 46.44 56
**Ga** 14 14 15 15 15 15 15 15 15 15 15 16.2
**Rb** 145 147 154 152 158 128 48 154 150 147 202 202
**Sr** 80 67 68 64 78 99 40 292 60 90 105 116
**Y** 14 15 16 14 11 16 13 14 14 15 16
**Zr** 104 106 107 110 112 109 103 98 114 82 111
**Nb** 10.8 10.6 11.2 11.7 9.9 11.5 13.2 10.4 11.1 8.9 11.4
**La** 37 38 36 39 38 36 36 35 39 30 37
**Ce** 60 66 63 66 64 70 61 65 66 54 63
**Nd** 21 22 25 24 19 24 20 22 25 20 23
**Sm** 9 5 6 2 7 11 2 7 2 5 16
**Eu** 65 69 68 62 60 57 92 64 44 32 20 20
**Gd** 54 52 39 34 30 26 27 26 27 26 26 26
**Tb** 19 20 19 20 19 20 21 19 20 18 20
**U** 6 5 7 5 6 6 4 4 7 5 5

**Sc** 2 2 2 2 2 2 2.4 2.4 1.9 2.6
**Rb** 149 145.1 155 152.1 133.2 157.2 147.8 173
**Sr** 83 66 68 63 103 60 107 58.3
**Y** 14.5 13.8 14.7 12.63 15.39 13.93 12.97 14.3
**Zr** 108 106 107 110 112 109 103 83 106
**Nb** 11.7 10.35 11.6 11.38 12.2 11.49 9.76 12.2
**Cs** 7.02 5.63 7.52 4.92 6.97 5.29 4.3 7.46
**Ba** 659 685 662 573 693 442 810 521 644 932 557
**Pb** 34 29 30 29 29 26 27 27 26 22 18 28
**Th** 19 20 19 20 19 20 21 19 20 18 20
**U** 6 5 7 5 6 6 4 4 7 5 5

(continued)
### TABLE 1. CHEMICAL ANALYSES OF THE TUFF OF CAMPBELL CREEK AND RELATED ROCKS (continued)

<table>
<thead>
<tr>
<th>Sample</th>
<th>Tuff of Campbell Creek</th>
<th>Intrusive rocks of Campbell caldera</th>
<th>Tuff E</th>
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</thead>
<tbody>
<tr>
<td></td>
<td>Location</td>
<td>Position</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Devitrified</td>
<td>Devitrified</td>
</tr>
<tr>
<td></td>
<td></td>
<td>DACITE PUMICE</td>
<td>DACITE DOME</td>
</tr>
<tr>
<td>H11-189</td>
<td>Intracaldera</td>
<td>4</td>
<td>3</td>
</tr>
<tr>
<td>H11-190</td>
<td>Intracaldera</td>
<td>4</td>
<td>3</td>
</tr>
<tr>
<td>H11-6</td>
<td>Intracaldera</td>
<td>4</td>
<td>3</td>
</tr>
<tr>
<td>H10-97</td>
<td>Intracaldera</td>
<td>4</td>
<td>3</td>
</tr>
<tr>
<td>H10-93</td>
<td>Intracaldera</td>
<td>4</td>
<td>3</td>
</tr>
<tr>
<td>H10-104</td>
<td>Intracaldera</td>
<td>4</td>
<td>3</td>
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<tr>
<td>22-1B</td>
<td>Intracaldera</td>
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<td>3</td>
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<tr>
<td>HP-X</td>
<td>Intracaldera</td>
<td>4</td>
<td>3</td>
</tr>
<tr>
<td>H99-29</td>
<td>Intracaldera</td>
<td>4</td>
<td>3</td>
</tr>
<tr>
<td>H00-57</td>
<td>Intracaldera</td>
<td>4</td>
<td>3</td>
</tr>
<tr>
<td>H01-78</td>
<td>Intracaldera</td>
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<td>3</td>
</tr>
<tr>
<td>C05-695</td>
<td>Intracaldera</td>
<td>4</td>
<td>3</td>
</tr>
</tbody>
</table>

**Note:** LOI—loss on ignition. WSU—X-ray fluorescence (XRF) and inductively coupled plasma-mass spectrometry (ICP-MS) at Geoanalytical Lab, Washington State University; SNLL—X-ray fluorescence at Sandia National Laboratory, Livermore, California, USA. ALS—ICP-atomic emission spectrometry and ICP-MS at ALS Minerals. All analyses normalized to 100% anhydrous. FeO*—total Fe reported as FeO. Sum—total before normalization to 100% anhydrous.
<table>
<thead>
<tr>
<th>Sample</th>
<th>Tuff (or Dated Unit)</th>
<th>Location</th>
<th>State</th>
<th>Age (Ma) ± 2σ</th>
<th>Pheno. Analyses</th>
<th>Note</th>
</tr>
</thead>
<tbody>
<tr>
<td>H10-104</td>
<td>Anorthoclase rhyolite dike</td>
<td>Southern part of caldera</td>
<td>NV</td>
<td>25.26 ± 0.09</td>
<td>6.0 ± 2.0</td>
<td>6/6</td>
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<tr>
<td>H10-93</td>
<td>Rhyolite of Daatite-Rhyolite domes</td>
<td>Eastern part of caldera</td>
<td>NV</td>
<td>28.84 ± 0.05</td>
<td>54.7 ± 3.4</td>
<td>11/23</td>
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<tr>
<td>H10-97</td>
<td>Daatite of Daatite-Rhyolite domes</td>
<td>Eastern part of caldera</td>
<td>NV</td>
<td>28.83 ± 0.04</td>
<td>55.4 ± 6.3</td>
<td>15/21</td>
</tr>
</tbody>
</table>

## TABLE 2. 40Ar/39Ar SANIDINE AGES AND PHENOCRYST ABUNDANCES OF THE TUFF OF CAMPBELL CREEK AND RELATED ROCKS.

- Post-collapse rocks of Campbell Creek caldera
- **Sample**
  - **Tuff (or Dated Unit)**
  - **Location**
  - **State**
  - **Age (Ma) ± 2σ**
  - **Pheno. Analyses**
  - **Note**

### HP-3B
- **Sample**: distal welded, devitrified
- **Location**: Haskell Peak, eastern Sierra Nevada
- **State**: CA
- **Age (Ma) ± 2σ**: 28.86 ± 0.05
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

### H02-93
- **Sample**: Rhyolite of Dacite-Rhyolite domes
- **Location**: Eastern part of caldera NV
- **State**: CA
- **Age (Ma) ± 2σ**: 28.84 ± 0.05
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

### H03-115
- **Sample**: Rhyolite of Dacite-Rhyolite domes
- **Location**: Eastern part of caldera NV
- **State**: CA
- **Age (Ma) ± 2σ**: 28.83 ± 0.04
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

### H00-101
- **Sample**: Distal welded, devitrified
- **Location**: Campbell Creek, Desatoya Mts
- **State**: NV
- **Age (Ma) ± 2σ**: 28.99 ± 0.07
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

### H09-81
- **Sample**: Basal vitrophyre, lower Cooling unit
- **Location**: Clipper Gap, Toquima Range NV
- **State**: NV
- **Age (Ma) ± 2σ**: 28.92 ± 0.06
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

### H01-78
- **Sample**: Distal welded, devitrified
- **Location**: East Humboldt Range
- **State**: NV
- **Age (Ma) ± 2σ**: 29.17 ± 0.07
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

### H00-88
- **Sample**: Tuff of Railroad Ridge
- **Location**: South Fork War Canyon, NV
- **State**: NV
- **Age (Ma) ± 2σ**: 29.11 ± 0.07
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

### H00-57
- **Sample**: Basal vitrophyre
- **Location**: Sand Springs Range
- **State**: NV
- **Age (Ma) ± 2σ**: 29.11 ± 0.07
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

### GSCN 200
- **Sample**: Unnamed tuff
- **Location**: Cactus Range, Nellis Test Range
- **State**: NV
- **Age (Ma) ± 2σ**: 29.22 ± 0.04
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

### H00-139
- **Sample**: Basal vitrophyre
- **Location**: Reese River Narrows, NV
- **State**: NV
- **Age (Ma) ± 2σ**: 29.84 ± 0.09
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

### H03-695
- **Sample**: Distal welded, devitrified
- **Location**: Flowery Peak, Virginia Range
- **State**: NV
- **Age (Ma) ± 2σ**: 29.17 ± 0.07
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

### H04-544
- **Sample**: Distal welded, devitrified
- **Location**: Painted Rock, Virginia Range
- **State**: NV
- **Age (Ma) ± 2σ**: 29.21 ± 0.04
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

### H01-78
- **Sample**: Distal welded, devitrified
- **Location**: Eastern Humboldt Range
- **State**: NV
- **Age (Ma) ± 2σ**: 29.17 ± 0.07
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

### H00-57
- **Sample**: Basal vitrophyre
- **Location**: Sand Springs Range
- **State**: NV
- **Age (Ma) ± 2σ**: 29.11 ± 0.07
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

### GSCN 200
- **Sample**: Unnamed tuff
- **Location**: Cactus Range, Nellis Test Range
- **State**: NV
- **Age (Ma) ± 2σ**: 29.22 ± 0.04
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

### H00-136
- **Sample**: Basal vitrophyre
- **Location**: Shoshone Mtns, Gold Park
- **State**: NV
- **Age (Ma) ± 2σ**: 29.11 ± 0.07
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

### H00-88
- **Sample**: Tuff of Railroad Ridge
- **Location**: South Fork War Canyon, NV
- **State**: NV
- **Age (Ma) ± 2σ**: 29.15 ± 0.06
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

### H00-66
- **Sample**: Basal vitrophyre
- **Location**: South Block War Canyon, NV
- **State**: NV
- **Age (Ma) ± 2σ**: 29.15 ± 0.06
- **Pheno. Analyses**: NM ± 1
- **Note**: Total Plag San Qtz Biot Hbl Px

Note: Sanidine was separated from crushed, sieved samples by standard magnetic and density techniques, leached with dilute HF to remove matrix, and handpicked. All analyses except Wallace at New Mexico Geochronological Research Laboratory, New Mexico Institute of Mining and Technology. Analytical methods in McIntosh et al. (2003).

- Weighted mean 40Ar/39Ar ages of sanidine from individual samples calculated by the method of Samson and Alexander (1987). Samples were irradiated in Al discs for 7 hours in D-3 position, Nuclear Science Center, College Station, Texas, USA. Neutron flux monitor Fish Canyon Tuff sanidine (FC-1); assigned age = 28.201 Ma (Renne and others, 1998; Kuiper et al. 2008).
- Decay constants and isotopic abundances after Steiger and Jäger (1977); lb = 4.963 × 10–10 yr–1; le+e = 0.581 × 10–10 yr–1; *40K/K = 1.167 × 10–4
- *Number (n) of single grains used in age calculation/total number of sanidine grains analyzed.
preserve pyramidal shape. The similarity of the two tuffs means that tuff E may not have been recognized in areas where a cooling break between the two tuffs is poorly exposed. Tuff E probably represents an early eruption from the Campbell Creek caldera in the Desatoya Mountains (Brooks et al., 2008). It has not been positively identified in or near the caldera, although an older tuff that makes up part of the caldera wall may be tuff E (Fig. 6). A postcaldera rhyolite contains sparse, possibly xenocrystic sanidine grains that give ages indistinguishable from that of tuff E. 

Sanidine $^{40}$Ar/$^{39}$Ar ages of 12 samples of tuff of Campbell Creek that span its distribution vary between 29.04 ± 0.10 and 28.84 ± 0.09 Ma, with one anomalously low age from a hydrothermally altered sample at 28.73 ± 0.10 Ma (Table 2; all $^{40}$Ar/$^{39}$Ar ages discussed in this paper were calculated or recalculated to an age of 28.201 Ma for Fish Canyon Tuff sanidine; Kuiper et al., 2008). K/Ca clusters tightly ca. 55 Ma (range 49.1 ± 6.3–59.6 ± 6.6). Sanidine $^{40}$Ar/$^{39}$Ar ages of 6 samples of tuff E range from 29.22 ± 0.06–29.11 ± 0.07 Ma. K/Ca varies from 55.8 ± 20.0–88.2 ± 31.7, higher and more variable than in the tuff of Campbell Creek.

The tuff of Campbell Creek also has a relatively distinctive characteristic remanent magnetization (CHRM), which can be used to test possible correlations. For example, CHRMs from 10 paleomagnetic sites in the Diamond and Fort Sage Mountains in northeastern California and westernmost Nevada, respectively, yield a group mean of 206.1° (declination) and –43.0° (inclination) for the tuff of Campbell Creek (Fig. 7A; Hinz et al., 2009). Because vertical-axis rotation is negligible in this area, this mean can serve as a reference direction for the tuff of Campbell Creek. In central Nevada, the C unit of the Bates Mountain Tuff has a nearly identical CHRM (Fig. 7B). The group mean of the CHRM from six sites in the Bates Mountain C Tuff (203.6° declination and –44.1° inclination; Gromme et al., 1972; Hudson and Geissman, 1991) overlaps the reference direction of the tuff of Campbell Creek at the 95% confidence level (Fig. 7C), supporting correlation of the two tuffs.

Our age, composition, petrographic, and paleomagnetic data demonstrate that the tuff of Campbell Creek is the same as the C unit of the Bates Mountain Tuff of central Nevada. The Bates Mountain Tuff was defined to consist of several individual ash-flow tuffs at Bates Mountain (Fig. 1; McKee, 1968; Stewart and McKee, 1968; Sargent and McKee, 1969). Subsequent stratigraphic subdivision, based in part on paleomagnetic data, recognized four ash-flow cooling units, A–D, and showed the C unit to be present at Clipper Gap and Reese River Narrows (Gromme et al., 1972). Our work shows that the C units at those two locations and at the type locality at Bates Mountain are the same as the tuff of Campbell Creek. The informal name tuff of Campbell Creek is preferred over Bates Mountain Tuff, because the latter consists of four unrelated tuffs, all with informal letter designations.

Caldera Source

Although examined only in reconnaissance for this study, the caldera source for the tuff of Campbell Creek in the Desatoya Mountains is reasonably well established based on the presence of thick intracaldera tuff of Campbell Creek, megabreccia, probable southwestern, southern, and northwestern caldera walls, and post-tuff dacite-rhyolite intrusions (Fig. 6; Barrows, 1971; McKee and Stewart, 1971; Stewart and McKee, 1977; this study). The Desatoya Mountains and Campbell Creek calderas are at most slightly tilted (intracaldera tuff is flat-lying to gently east dipping), so exposure is limited to upper parts of the caldera. In the area of Campbell Creek, intracaldera tuff mostly consists of a single, petrographically homogeneous, densely welded cooling unit >600 m thick; the base is not exposed (Fig. 5G). The uppermost intracaldera tuff consists of poorly welded, lithic tuff interbedded with lenses of conglomerate.
Figure 5. Outcrop characteristics of tuff of Campbell Creek. (A) Typical exposure of outflow tuff in Caetano caldera (Fig. 1). Densely welded, devitrified, columnar-jointed tuff makes an ~8-m-thick resistant ledge with a thin welded basal vitrophyre. Upper, nonwelded tuff probably was deposited but stripped by erosion. (B) Typical section grading from poorly welded, glassy base to densely welded vitrophyre to densely welded devitrified tuff, Reese River Narrows (Fig. 1). Rock hammer (circled) in all photos is 42 cm long. (C) The most distal outflow tuff is poorly welded and vitrophyric throughout (Welches Canyon; Fig. 1). (D) Distal outflow tuff in canyon of Little Truckee River, Sierra Nevada (Fig. 1), showing ~40° primary dip where tuff compacted against steep slope developed in older ash-flow tuff in a paleovalley. (E) Densely welded intracaldera tuff with slightly stretched pumice, Desatoya Mountains (H00-101 location; Fig. 6). (F) Abundantly porphyritic dacite pumice fragments ~7 cm thick in intracaldera tuff, Desatoya Mountains (H00-101 location; Fig. 6). (G) Typical, thick, densely welded, columnar-jointed intracaldera tuff, Desatoya Mountains ~1.5 km west of location H00-101 (Fig. 6). (H) Megabreccia blocks of porphyritic andesite (left) and older ash-flow tuff (right) in moderately welded intracaldera tuff, southwest margin of caldera (Fig. 6).
Figure 6. Reconnaissance geologic map of the southern part of the Campbell Creek caldera in the Desatoya Mountains, central Nevada. Geology from Barrows (1971), Stewart and McKee (1977), McKee and Conrad (1987), and this study.
The most densely welded parts are slightly rheomorphic, with stretched pumice fragments as much as 60 cm long and 2 cm thick (Fig. 5E). Megabreccia consisting of blocks of the older ash-flow tuff or porphyritic andesite as much as 20 m across surrounded by vitrophyric tuff of Campbell Creek is present near the probable southwestern, southern, and northwestern margins of the caldera (Figs. 5H and 6).

Several intrusions are mapped in the caldera, although we have examined only the composite dacite-rhyolite intrusions (unit Tr; Fig. 6). Most of the observed dacite-rhyolite intrusive rock is coarsely and abundantly porphyritic (H10-97; 66% SiO₂; Table 1) that is related. Rhyolite bands are almost certainly residual (nonerupted) tuff of Campbell Creek magma. The two residual magma components are petrographically and compositionally indistinguishable from the dacite pumice. Dacite intrusions contain 40% phenocrysts, mostly plagioclase, with lesser sanidine, biotite, minor vermiculitic quartz, and sparse hornblende. The dacite intrusions are even less evolved than the dacite pumice, with higher TiO₂, Al₂O₃, Fe₂O₃, MgO, CaO, Sr, Zr, and Ba abundances and lower K₂O, Rb, Nb, Th, and U abundances than the tuff of Campbell Creek. Bands of rhyolite (H10-93; 77% SiO₂) in dacite in the northern outcrop area are petrographically and compositionally indistinguishable from the tuff. They contain <10% phenocrysts of plagioclase, sanidine, and highly vermiculitic quartz, and minor biotite. Major and trace elements show the same abundances as the tuff (Figs. 2 and 3). Sanidine ⁴⁰Ar/³⁹Ar ages from a rhyolite and a dacite are 28.84 ± 0.05 and 28.83 ± 0.04 Ma, overlapping with the age of the tuff of Campbell Creek (Table 2). The age and location of the dacite-rhyolite intrusions, the similarity of the rhyolite part to the main tuff of Campbell Creek, and the similarity of the dacite part to the dacite pumice are strong evidence that all are genetically related. Rhyolite bands are almost certainly residual (nonerupted) tuff of Campbell Creek magma. Dacite is probably additional residual but less silicic magma that provides further evidence for a compositionally variable magma chamber. The two residual magma components partly mixed during emplacement of the dacite-rhyolite intrusions. The sparse dacite pumice in intracaldera tuff indicates minor mixing during initial ash-flow eruption.

The younger, ca. 24.7 Ma Desatoya Peak caldera cuts out the western part of the Campbell Creek caldera (Fig. 6; Table 2). Probable outflow tuff of Desatoya Peak crops out in the Shoshone Mountains southeast of the Campbell Creek caldera. The tuff of Desatoya Peak contains 20%–25% phenocrysts of plagioclase, sanidine, quartz, biotite, and minor hornblende and clinopyroxene and orthopyroxene (Table 2). An anorthoclase-phryic dike that intrudes intracaldera tuff of Campbell Creek is significantly younger (25.26 ± 0.09 Ma; H10-104; Table 2), but older than the tuff of Desatoya Peak, and compositionally distinct from Campbell Creek rocks (Fig. 2).

We estimate the area of the Campbell Creek caldera to be ~600 km², and total volume of erupted tuff to be between 1200 and 3000 km³. Uncertainty about the location of the caldera margin in several places and total thickness of intracaldera tuff preclude more precise estimates (Figs. 1 and 6). The caldera is at least 35 km north-south from a well-located caldera margin near Buffalo Canyon to the northern part of the Desatoya Mountains, where intracaldera tuff is at least 400 m thick and contains abundant megabreccia. The west-east dimension is at least 14 km and probably closer to 20 km, but is poorly known because the younger Desatoya Peak caldera and basin-and-range faults cut off the Campbell Creek caldera to the west and east.

Estimating the volume of intracaldera outflow, or total erupted tuff is even more uncertain. Flow and deposition of the tuff of Campbell Creek in paleovalleys make estimates of outflow tuff especially difficult. We find that total erupted volume is best approximated from the volume of caldera collapse where it is known (John et al., 2008a; Henry and Faulds, 2010). Assuming 2–5 km of collapse, which is representative of calderas in Nevada, erupted volume could range from 1200 to 3000 km³. Even the lower value makes the tuff of Campbell Creek and caldera a supereruption and supervolcano (Mason et al., 2004; Sparks et al., 2005; Miller and Wark, 2008), which is consistent with the wide distribution of the tuff.
The tuff of Campbell Creek is distributed across an area of at least 55,000 km$^2$ from the western foothills of the Sierra Nevada east to the East Humboldt Range (northern Ruby Mountains). Notably, the tuff is found ~100 km east of our interpreted Oligocene drainage divide in east-central Nevada (Fig. 1; Henry, 2008), or even farther east of the alternative divide of Best et al. (2009). The present-day west-southwest–east-northeast extent, parallel to paleovalleys through that region, is ~530 km. The tuff reached a distance from its source caldera to the west of at least 200 km, corrected for later extension using the values of Henry and Faulds (2010), and ~215 km to the northeast, based on estimated extension of Colgan and Henry (2009). The actual flow distance would have been greater in each direction because the paleovalleys were not straight and because known outcrops probably do not represent the greatest original extent. As with other outflow tuffs in western and central Nevada, thickness correlates poorly with distance from source.

The distribution of the tuff of Campbell Creek is distinctive compared to that of other tuffs that erupted in the western part of the central Nevada caldera belt that also flowed long distances down the west-draining paleovalleys (Henry and Faulds, 2010). (1) It spread much farther east, upstream in the paleovalleys, even across the interpreted paleodivide, whereas other tuffs commonly traveled only ~60 km upstream. (2) Its north-northwest extent, perpendicular to the trend of the paleovalleys, is significantly greater, ~160 km, than the perpendicular extent of other tuffs, ~100 km. The tuff flowed ~100 km to the north and ~60 km to the south, crossing several east-west divides between major paleovalleys. (3) The tuff flowed in at least five major paleovalleys in western Nevada and the eastern Sierra Nevada, whereas most tuffs were restricted to one or two major paleovalleys. Only the Nine Hill Tuff, which spread similar distances to the west, almost as far to the northeast, and farther east to near Ely, Nevada, from a possible source in the Carson Sink, has a similar or greater distribution among tuffs of the Great Basin (Deino, 1985; Best et al., 1989). The Peach Springs Tuff in southern Nevada and northwestern Arizona also flowed northeastward across the probable paleodivide (see Regional Orogenic Highland discussion).

Why did the tuff of Campbell Creek travel so much farther than other tuffs from the central Nevada caldera belt? Particularly, how was it able to flow upstream across the inferred major north-south paleodivide and to cross several divides between west-draining paleovalleys near the source (Fig. 1)? The occurrences at Welches Canyon and the East Humboldt Range are particularly problematic because they are so far from the caldera, “upstream” from the source, and across several inferred east-west drainage divides. No single factor seems capable of generating such wide distribution, and each factor has its drawbacks. (1) Large eruptive volume seems essential, and the possible 3000 km$^3$ volume makes the tuff of Campbell Creek one of the most voluminous eruptions of the central Nevada caldera belt (Best et al., 1989, 1995; Mason et al., 2004). However, other voluminous tuffs are not so extensive. Both the ~1100 km$^3$ Caetano Tuff and ~1200–1600 km$^3$ lower tuff of Mount Jefferson are largely restricted to single paleovalleys (John et al., 2008a; Henry and Faulds, 2010). (2) The tuff of Campbell Creek may have had distinctive magmatic and eruptive characteristics that contributed to wider distribution. For example, it may have been hotter and had a higher eruption column than most other tuffs, although we have no direct evidence for either. The Nine Hill Tuff, which has a similar wide distribution, is sparsely porphyritic and moderately alkalic (but neither it nor the tuff of Campbell Creek are peralkaline); was hot, with magmatic temperatures of 850–930 °C (Deino, 1985); and is commonly highly rheomorphic. These features are contributors to or indicators of low viscosity and wide distribution. The tuff of Campbell Creek has the same features except for high magmatic temperature, which is unknown. A high eruption column would give the tuff the potential and then kinetic energy from column collapse to surmount significant topography. We have no information on eruption dynamics for either tuff, but suggest that the similarities between them indicate their wide distributions arise in part from similar causes. (3) Intervallief relief may have been especially low when the tuff of Campbell Creek erupted. The 28.9 Ma tuff of Campbell Creek erupted at the end of a nearly continuous, 2.5 Ma (31.4–28.9 Ma) period of voluminous ash-flow eruptions (Faulds et al., 2005) from calderas in the Stillwater Range–Clan Alpine Mountains–Desatoya Mountains region. Pyroclastic material from preceding eruptions at least partly filled topography close to the Campbell Creek caldera, which reduced intravalley relief and probably made it easier for the tuff to disperse more widely.

Factors external to the tuff of Campbell Creek, such as location of the paleodivide or gradients across the Nevadaplano, could also influence distribution but would apply to other tuffs erupted from the central Nevada caldera belt, which did not flow as far. Two interpretations depict paleotopography of the Nevadaplano very differently (Fig. 8). (1) Based on stable isotope studies, Cassel et al. (2010, 2012a) interpreted that the Oligocene Sierra Nevada had an elevation and steep gradient similar to its present-day elevation and gradient, and that the Nevadaplano east about to the Campbell Creek caldera had a much shallower gradient. A shallow gradient in the central Nevadaplano would have allowed the tuff of Campbell Creek, as well as any other tuff, to flow more easily upstream. In contrast, Best et al. (2009) used an analogy to the Andean Altiplano to interpret that the Nevadaplano had a nearly constant gradient to a paleodivide at ~4 km elevation slightly west of where we place it, and a nearly flat interior only slightly lower than the paleodivide to the
east. The steep western slope would have made it particularly difficult for any tuff to flow to the east and makes the distributions of the tuff of Campbell Creek, Nine Hill Tuff, and possibly Peach Springs Tuff even more problematic. (2) Intervallel relief may have been less near the paleodivide than to the west. Paleovalleys in western Nevada and eastern California were 380–1200 m deep (Proffett and Proffett, 1976; Brooks et al., 2003; Henry, 2008; Henry and Faulds, 2010). Paleovalleys are not as well exposed or examined in the central Nevada caldera belt, but, where exposed, they are >300 to >1000 m deep (this study; John et al., 2008a; Gonsior and Dilles, 2008). The Caetano Tuff, the caldera of which is much closer to our proposed paleodivide, spread almost entirely west of its source caldera (John et al., 2008a).

**IMPLICATIONS OF FAR-TRAVELED TUFFS**

**Structural Evolution of Northern Nevada and the Sierra Nevada**

Regardless of how they did so, the pyroclastic flows that deposited the tuff of Campbell Creek and several other 31.4–23.3 Ma ash-flow tuffs flowed great distances from sources in central Nevada through western Nevada and the Sierra Nevada, and other, mostly Eocene tuffs spread over great distances in northeastern Nevada (Fig. 1; Faulds et al., 2005; Henry, 2008; Hinz et al., 2009; Henry and Faulds, 2010). Their ability to flow far greater distances down paleovalleys across what are now highly faulted Basin and Range and Walker Lane structural provinces confirms the observation of Henry and Faulds (2010) that major faulting postdated ca. 23 Ma. That these tuffs and paleovalleys crossed what is now the Basin and Range–Sierra Nevada structural and topographic boundary further confirms that the Sierra Nevada was topographically lower than what is now the Basin and Range, regardless of the absolute elevations of either (Mulch et al., 2006; Cassel et al., 2009a; Henry and Faulds, 2010).

The fact that the tuff of Campbell Creek reached the East Humboldt Range places significant limits on the amount of pre–29 Ma extension across north-central Nevada to the Ruby Mountains–East Humboldt Range metamorphic core complex. Different studies have concluded markedly different amounts of pre–29 Ma extension in and around the core complex. Abundant thermochronologic and thermobarometric data indicate that the core complex underwent ~170 °C of cooling and 4 kbar of decompression between ca. 85 and ca. 50 Ma, and another 450 °C cooling and 4–5 kbar decompression between ca. 50 and ca. 21 Ma, which requires a total of ~30 km of exhumation (McGrew and Snee, 1994; Snee et al., 1997; McGrew et al., 2000; Henry et al., 2011). If accomplished by surface-breaking extension, that amount of exhumation would have greatly uplifted the Ruby Mountains and probably generated major basin-and-range topography around them. In contrast, studies of paleovalleys, mapped normal faults in northeast Nevada, and thermochronology of the southern Ruby Mountains indicate a minor episode of extension ca. 40 Ma and major extension after ca. 17 Ma (Henry, 2008; Colgan and Henry, 2009; Colgan et al., 2010).

The tuff of Campbell Creek crops out in the hanging wall of the west-dipping Ruby Mountains detachment fault, so the tuff would have been farther east relative to the core complex before fault displacement. However, ash flows could not have reached that far east if the Ruby Mountains had been a high range with adjacent low basins at the time. We conclude that either all exhumation and extension in the Ruby Mountains postdated 29 Ma (Colgan et al., 2010), or exhumation was accomplished by a process that did not affect the surface, such as diapirism balanced by adjacent downflow (Howard, 2003; Colgan, 2011; Henry et al., 2011).

The distribution of the 760 ka Bishop Tuff around the Long Valley caldera (Bailey, 1989; McConnell et al., 1995; Hildreth and Wilson, 2007) illustrates well the influence of Basin and Range topography on the flow distance and resulting distribution of a tuff (Fig. 1). With an estimated eruptive volume of ~600–750 km³, the Bishop Tuff is similar to many mid-Cenozoic tuffs in Nevada. The Bishop Tuff surmounted a ridge only a few hundred meters higher than the western topographic rim of the Long Valley caldera and flowed at least 14 km down the 3–4-km-wide Middle Fork of the San Joaquin River. The tuff spread more radially into three wide (≥20 km) basins to the north, east, and southeast. In each case, the Bishop ash flows spread out, made broad, sheet-like deposits, and traveled much shorter distances than did the channelized mid-Cenozoic tuffs. Maximum flow distance was ~50 km into the northern part of Owens Valley southeast of Long Valley (Bailey, 1989).

**Paleotopography of the Nevadaplano**

**Paleovalley Morphology**

Several characteristics of paleovalleys suggest that they resulted from prolonged erosion, possibly aided by the warm, wet Eocene climate (Zachos et al., 2001; Kelly et al., 2005).

(1) Although the oldest dated paleovalley-filling tuffs in central and western Nevada are ca. 34 and 31 Ma, the Eocene auriferous gravels that fill the bottoms of paleovalleys in the Sierra Nevada are ca. 52–50 Ma, based on paleoflora evidence (MacGinitie, 1941; Wing and Greenwood, 1993). In the Central Valley of California, the intertidal-deltaic equivalent of the auriferous gravels is the Middle Eocene (Lutetian) ca. 49–40 Ma Ione Formation (Creely and Force, 2007). Low-temperature thermochronology from the Sierra Nevada indicates rapid cooling and interpreted rapid exhumation between ca. 90 and 60 Ma, recording erosion following batholithic emplacement, and slower cooling and exhumation after 60 Ma (Cecil et al., 2006; see also House et al., 1997). Therefore, the Sierra Nevada and western Nevada paleovalleys existed by at least 50 Ma and possibly as early as 60 Ma. Paleovalley-filling tuffs in northeastern Nevada are as old as 45 Ma (Henry, 2008), and one underlying tuff is dated as 46 Ma (Haynes, 2003), so the northeastern paleovalleys could be as old as those in the Sierra Nevada. Wing and Greenwood (1993, p. 246) cited 50–52 Ma as the “peak of Cenozoic warmth,” and the paleovalleys were almost certainly being eroded at that time. Paleovalleys in Idaho existed by the Late Cretaceous (Janecke et al., 2000; Chetel et al., 2011), although they need not have formed at the same time as the drainages in Nevada. Preservation of deposits as old as 45–52 Ma in the bottoms of the paleovalleys indicates that they had been cut to their full depth by that time and did not deepen significantly before filling with Oligocene ash-flow tuffs.

(2) Paleovalleys were much wider (5–8 km) than they were deep (0.5–1.2 km), commonly with nearly flat bottoms and steep walls (Fig. 9), in strong contrast to the v-shaped canyons of the modern Sierra Nevada. These characteristics are exceptionally well illustrated by a three-dimensional view of the paleovalley at Black Mountain in the northern Sierra Nevada (Fig. 9B; see also Hinz et al., 2009), where intravalley relief was mostly developed on resistant rocks, e.g., metavolcanic rocks at Black Mountain. Interfluves were relatively flat (Fig. 9A; see also Slemmons, 1953). Paleovalleys in Idaho were also much wider than deep (Janecke et al., 2000).

(3) A paleoweathered zone as much as 30 m thick is common, especially in biotite-rich granitic rocks, throughout the distribution of paleovalleys. Weathering of the granitic rocks and of vitrophyric parts of ash-flow tuffs produced smectite. In the Sierra Nevada, vitrophyric parts of tuffs also have weathered to smectite (California Geological Survey, 2009) but, where more intensely weathered in the western Sierra Nevada, underlying granitic rocks are altered to kaolinite (Allen, 1929; Wood, 1994; Creely and Force, 2007). Kaolinite has not been found.
in the granitic rocks of Nevada. The different mineralogy could indicate either a spatial or temporal change in weathering characteristics. The absence of kaolinite, long recognized as forming by intense leaching in tropical climates (Wilson, 1999), could indicate higher elevation, lower temperature, and less precipitation in Nevada compared to the Sierra Nevada. However, weathering of granitic rocks to smectite can only be demonstrated to have occurred by ca. 31 Ma, the age of the oldest tuffs overlying smectite-weathered granitic rocks in Nevada. Therefore, the mineralogical change could indicate a change in climate over time, with less intense weathering during the cooler, dryer Oligocene compared to the Eocene. Berner and Kothavala (2001) showed a significant decrease in atmospheric CO₂ ca. 50 Ma that might have caused a drop in pCO₂ in soil moisture, higher pH, and less intense weathering. A temporal change might suggest that Eocene weathering in Nevada generated kaolinite, but none of it is preserved, which seems unlikely.

**Absolute Elevation**

Isotopic and paleobotanical data are interpreted to record a Nevadaplano and eastern, high part of what is now the Sierra Nevada as high as 3 km in the middle Cenozoic (Fig. 8; Wolfe et al., 1997; Horton and Chamberlain, 2006; Mulch et al., 2006; Crowley et al., 2008; Cassel et al., 2009a, 2009b, 2010; Hren et al., 2010), although Molnar (2010) contested the significance of the isotopic data. Based on estimates of crustal thickness and analogy to the Altiplano of the Andes, Best et al. (2009) interpreted a paleodivide at an elevation of ~4 km. In contrast, erosion and river incision data from the Sierra Nevada are interpreted to indicate that it underwent 1.5–2.5 km of uplift in the latest Cenozoic from much lower elevations in the middle Cenozoic (Wakabayashi and Sawyer, 2001; Jones et al., 2004; Stock et al., 2004). The distribution of mid-Cenozoic ash-flow tuffs demonstrates that the Sierra Nevada was a western ramp to the former Nevadaplano, but do not quantitatively indicate paleoelevation. That the tuffs commonly flowed as much as 200 km from source, much farther than most other documented ash-flow tuffs of the world (Cas and Wright, 1987), in channels carrying coarse boulders (Henry, 2008; Henry and Faulds, 2010; Cassel and Graham, 2011), suggests the rivers had moderately steep gradients.

However, the record of two far-traveled Quaternary ash-flow tuffs where the present-day topography is essentially unchanged since...
the time of eruption suggests that steep gradients are not required. The $>$60 km$^2$, 70% SiO$_2$, 90 ka Aso-411 tuff of Kyushu, Japan, flowed as much as 125 km to the sea from a caldera with a rim elevation of between 800 and 1200 m (Matumoto, 1943; Lipman, 1967; Kaneko et al., 2007), a topographic gradient of $\sim$8 m/km. The Morrinsville ignimbrite in New Zealand is interpreted to have flowed $\sim$200 km to the sea from a caldera near Taupo ($\sim$500 m elevation; Walker and Wilson, 1983), a gradient of only 2.5 m/km. Unfortunately, very little is published about the Morrinsville ignimbrite, including composition or tuff volume, and Walker and Wilson (1983, p. 131) stated that distal parts of the tuff “cease to have the characters normally regarded as typical of ignimbrites.”

Although neither the Aso tuff nor Morrinsville ignimbrite are ideal analogs for the mid-Cenozoic ash-flow tuffs, they demonstrate that an ignimbrite need not erupt from a caldera at high elevation to flow a long distance. A minimum gradient of 2.5 m/km would require paleovalley bottoms in the region of source calderas in central Nevada (200–300 km from the Oligocene coastline) to have been at altitudes of $\sim$500 m. With 300–1000 m-deep paleovalleys, interfluves would have reached elevations of $\sim$1500 m. The distribution of ash-flow tuffs in paleovalleys therefore allows but does not require high mid-Cenozoic elevations in central Nevada.

**DID PALEOVALLEYS CROSS THE SOUTHERN SIERRA NEVADA?**

Although well-known paleovalley systems crossed the northern and central Sierra Nevada, only one paleovalley is known to have crossed the southern Sierra Nevada south of Sonora Pass (Figs. 1 and 10). Huber (1981, p. 3) documented an ancestral San Joaquin River that “headed at least as far east as the present Mono Lake basin, possibly farther north or east in Nevada.” The valley of this ancestral San Joaquin River was much wider than deep (as much as 10 km by 450–750 m) and drained into the Eocene Ione Formation at the edge of the Great Valley. The oldest preserved deposits in the ancestral San Joaquin paleovalley are Late Miocene gravels containing 11 Ma pumice, which Huber (1981) interpreted to be derived from a similar-age tuff east of Mono Lake. The 10 Ma trachyandesite of Kennedy Table filled the paleovalley. Oligocene ash-flow tuffs or other pre–10 Ma deposits may be absent because they were never deposited in the ancestral San Joaquin River valley (source calderas in central Nevada were farther away), or because they were not preserved; few Miocene volcanic rocks were deposited that could have capped and preserved older deposits.

Neither paleovalley deposits nor the mid-Cenozoic landscape are preserved south of the ancestral San Joaquin River, but geologic data from the Death Valley region, southern Sierra Nevada, and San Joaquin Basin allow the existence of several trans-Sierra drainages. Eocene to Early Miocene tuff-bearing sedimentary sequences are common in the Death Valley, California, to Yucca Mountain, Nevada, region (Fig. 10). These deposits were termed Tertiary older sedimentary rocks (Barnes et al., 1982; Slate et al., 2000), the Amargosa Valley Formation (Cemen et al., 1999), and the Titus Canyon and Ubehebe Formations (Snow and Lux, 1999) in different areas. These sedimentary sequences are as much as 800 m thick and generally consist of a coarse basal conglomerate, overlain by mixed clastic rocks and freshwater limestone (Cemen et al., 1999; Snow and Lux, 1999). Dates determined for interbedded tuff in these deposits are (1) 34 Ma in the basal conglomerate and 30 Ma higher in the Titus Canyon Formation (Saylor and Hodges, 1994), (2) 30 Ma on pyroclastic-fall tuff interbedded with argillaceous limestone in the Tertiary older sedimentary rocks (Barnes et al., 1982; Slate et al., 2000), (3) 25–20 Ma on tuffs (mostly pyroclastic-fall deposits) in the Amargosa Valley Formation (Cemen et al., 1999), and (4) 24.0, 23.0, and 20.3 Ma on ash-flow tuffs in the Ubehebe Formation (Snow and Lux, 1999; their $^{40}$Ar/$^{39}$Ar ages recalculated to a monitor age equivalent to 28.201 Ma on Fish Canyon Tuff sanidine). Farther south in Nevada, west-flowing streams deposited gravels containing rounded quartzite boulders in the Late Oligocene or Early Miocene; the gravels are overlain by Middle Miocene ash-flow tuffs derived from the east (Fig. 10; Kohl, 1978; Hanson, 2008).

The tectonic environment of these mid-Cenozoic sedimentary and volcanic rocks is uncertain. Major extension in the Death Valley area began no earlier than ca. 16 Ma, but the existence of earlier extension is debated. Snow and Lux (1999), Snow and Wernicke (2000), and Niemi (2002) interpreted these deposits to have accumulated in ca. 36 and 25 Ma extensional basins, whereas Cemen et al. (1999) interpreted deposition in a broad floodplain but discounted any pre–Middle Miocene extension. Bedding in pre-Cenozoic rocks, Eocene to Early Miocene rocks, and basal parts of definitely Middle Miocene synextensional deposits is concordant (Cemen et al., 1999; Snow and Lux, 1999), which indicates that no measurable tilting, and presumably extension, occurred before the Middle Miocene. Pre–Middle Miocene sedimentary deposits in the Lake Mead area also preceded major extension beginning ca. 17 Ma (e.g., Lamb et al., 2011; Umhoefer et al., 2011).

We suggest that the deposits described here could have been deposited in paleovalleys that drained from the northeast and crossed what is now the southern Sierra Nevada. Barnes et al. (1982) specifically interpreted gravels to have been deposited by streams sourced to the north

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**Figure 10 (on following page).** Satellite image of southern Nevada and adjacent regions showing alternative interpretations of paleodrainages and paleodivides through time. The southernmost well-established paleodrainage across the Sierra Nevada is from Huber (1981). Lechler and Niemi (2011) interpreted a Paleocene–Eocene drainage system that flowed southeast along what is now Owens Valley east of the Sierra Nevada, then turned southwest, and did not drain from what is now Nevada. However, we suggest that Eocene–Oligocene deposits recognized in the Death Valley–Yucca Mountain area (Barnes et al., 1982; Saylor and Hodges, 1994; Cemen et al., 1999; Snow and Lux, 1999), although interpreted by some as deposited in extensional basins (Snow and Lux, 1999; Snow and Wernicke, 2000), could be parts of a paleodrainage system that crossed the southern Sierra Nevada and connected with the paleodrainage system of Lechler and Niemi (2011) and/or emptied into the San Joaquin Basin (Nilsen and Clarke, 1975; Bartow and McDougall, 1984; Graham and Olsen, 1988). A post-Sevier, pre-Laramide divide probably trended south into California (Goldstrand, 1992, 1994). Development of the Laramide Kingman uplift (ca.16 Ma or later; Beard et al., 2010) probably shifted the divide northeast through northwestern Arizona, where it stayed at least through the time of the 18.5 Ma Peach Springs Tuff (Young, 1979; Bohannon, 1984; Goldstrand, 1994; Beard, 1996; Spencer et al., 2008). Howard (1996, 2000) interpreted a large paleodrainage system to the southwest, to the Pacific Ocean, that would have been across the divide from these northeast-draining systems. In contrast, Davis et al. (2010) proposed that a Late Paleocene–Eocene Eocene system drained northeastward from southeastern California to Utah, perpendicular to the drainage of Howard (1996, 2000) and across the post-Laramide divide. A possible resolution is that the Davis et al. (2010) system is partly pre-Laramide. Distribution of Peach Springs Tuff is from Glazner et al. (1986) and Valentine et al. (1989) with interpreted calderas from Ferguson (2008). See text for further discussion.
Eocene–Early Miocene paleotopography

Kawich caldera
Cactus Range
Leach et al. (2009)

Late Cretaceous
Barnes et al. (1982)

Oligocene–Miocene
Davis et al. (2010)

Paleocene–Eocene
Howard (1996, 2000)

Paleocene–Eocene
Goldstrand (1992, 1994)

Paleocene–Miocene
Goldstrand (1992)

Late Cretaceous
Kohl (1978)

Hanson (2008)

Oligocene–Miocene
Young (1979)

Speicer et al. (2008)

Paleocene–Miocene
Lechler and Niemi (2011)

Paleocene–Oligocene
Young (1979)

Speicer et al. (2008)

Oligocene–Miocene
Barnes et al. (1982)

Valentine et al. (1989)

18.5 Ma Peach Springs Tuff

Paleodivide

Lechler and Niemi (2011)

Late Cretaceous
Leach et al. (2009)

Post-Laramide

Post-Laramide

Claroanus

Candelaria Hills

San Joaquin Basin

Mono Lake

Death Valley

Owens Valley

San Andreas Fault

Nevada

California

Arizona

Utah

Map legend:
- Caldera
- Mapped or interpreted individual paleovalley
- ? where existence or location are uncertain
- General trend of paleodrainage

Figure 10.
The idea that Eocene drainages in the northern and central Sierra Nevada were sourced no farther east than the modern drainages is incompatible with provenance studies and the distribution of ash-flow tuffs (Bateman and Wahrhaftig, 1966; Yeend, 1974; Faulds et al., 2005; Garside et al., 2005; Cassel et al., 2009a, 2009b, 2012b; Henry and Faulds, 2010; this study). A more extensive study of 1292 detrital zircons in Eocene–Oligocene fluvial sediments in the northern Sierra Nevada (Cassel et al., 2012b) found marked differences in populations (e.g., locally Jurassic-dominated versus mid-Cretaceous dominated) between different sites and 1%–17% Eocene zircons in the Jurassic-dominated populations. In Cassel et al. (2012b), Sierra Nevada drainage patterns are interpreted to have evolved complexly through time, possibly including eastward migration of the paleodrainage network. By Eocene time; this includes the headwaters of the northern and central Sierra drainages (Fig. 10).

In contrast, the zircon data of Cecil et al. (2010) and Lechler and Niemi (2011) are equally compatible with paleodrainages originating in central Nevada and crossing the Sierra Nevada. Paleogeologic maps of depositional basins and Tertiary rocks in the Sierra Nevada and Basin and Range north of 38°N indicate that Mesozoic plutonic rocks were widely exposed in California and western Nevada, in and east of the modern Sierra Nevada (Van Buer et al., 2009; Cassel et al., 2012b). These batholithic rocks are rich sources of zircons, which, being relatively young, would have been deposited in the upstream parts of the southern Sierra Nevada drainages (Fig. 10).

Eocene through Miocene sedimentary rocks are widespread in the surface and subsurface of the San Joaquin Basin, along and west of the western edge of the southern Sierra Nevada (Nilsen and Clarke, 1975; Bartow and McDougall, 1984; DeCelis, 1988; Graham and Olsen, 1988; see especially Bent, 1988). All studies agree upon an eroding Sierra Nevada as a source; Bartow and McDougall (1984, p. 7) noted that the “strong resemblance to the Eocene Ione Formation in its type area 350 km to the northwest,” and Nilsen and Clarke (1975) specifically suggested the Great Basin as a source area. Drainages were coming from the southern Sierra Nevada, the only question being the eastern location of their headwaters.

Combining the above data allows two end-member alternatives: (1) the southern Sierra Nevada was a high island in the early through middle Cenozoic and paleodrainages went around it, or (2) paleodrainages crossed the southern Sierra Nevada but both their deposits and geomorphic expression have been removed by erosion. The high island alternative requires that the southern Sierra Nevada underwent distinctly different Late Cretaceous or early Cenozoic evolution than did the northern and central Sierra Nevada. Batholithic development was similar along the entire range, especially in the most voluminous Cretaceous phase (Bateman, 1992; Ducea, 2001; Saleeby et al., 2008). However,
REGIONAL OROGENIC HIGHLAND

Combined with other interpreted paleodrainages in Idaho, Oregon, California, Arizona, and Sonora, Mexico, the data presented in this paper indicate that a late Mesozoic to mid-Cenozoic erosional highland extended from at least central Idaho to northern Sonora (Fig. 11). In Idaho, Janecke et al. (2000) and Chetel et al. (2011) mapped Middle Eocene (ca. 50–47 Ma) paleovalleys (Eocene Idaho River) that drained southeastward from the Sevier thrust belt into the Green River Basin in southwestern Wyoming. West of the paleovalve, several provenance studies indicate drainage from the Idaho batholith west to the Oregon Coast and southwest to northern California (Heller et al., 1985, 1987; Underwood and Bachman, 1986; Renne et al., 1990; Aalto et al., 1998). The southwestern drainage parallels the northwestern edge of the Cretaceous batholith belt, which turns northeast from the Sierra Nevada across northern Nevada toward the Idaho batholith (Barton et al., 1988; Lerch et al., 2007; Van Buer et al., 2009). The batholith belt probably was a topographic high in the early Cenozoic that had eastward drainages off its eastern flank (Van Buer et al., 2009). As pointed out by Chetel et al. (2011), the inferred paleodrainages in Nevada and Idaho are misaligned by ~200 km; they noted that this misalignment coincides with an interpreted Neoproterozoic transfer zone and major changes in Paleozoic facies (Lund, 2008), where the Roberts Mountains allochthon turns to the northeast in northeastern Nevada after trending north through the rest of Nevada (Stewart, 1980). Greater post-Eocene westward extension in Nevada relative to Idaho probably also contributes to the misalignment.

Considerable data show an orogenic highland through southern Nevada into Arizona (Figs. 10 and 11). Clasts in Late Cretaceous–Paleocene deposits of southern Utah were sourced from the Sevier thrust belt in southern Nevada and eastern California (Goldstrand, 1992; Young, 1979). The pre-Laramide divide probably followed the Sevier thrust belt into eastern California (Goldstrand, 1992). Development of the Laramide Kingman uplift (Bohannon, 1984; Faulds et al., 2001; Beard et al., 2010) probably shifted the divide to the northeast; this beheaded the older drainages (Young, 1979; Bohannon, 1984; Goldstrand, 1994; Beard, 1996). Apatite (U-Th)/He geothermometry data are consistent with major uplift of the southwestern Colorado Plateau following Sevier–Laramide contraction and erosion of 1-km-deep proto–Grand Canyon by the Early Eocene (Flowers et al., 2008). Wernicke (2011) placed this erosion as ca. 80–70 Ma, or following Sevier deformation and preceding the Kingman uplift. The post-Laramide drainage system persisted until eustational flooding began in the Middle Miocene (Bohannon, 1984; Beard, 1996).

The 18.5 Ma Peach Springs Tuff, the youngest major preextensional marker in its region, spread preferentially westward from its caldera source in northwestern Arizona near the borders with Nevada and California (Fig. 10; Glazner et al., 1986; Valentine et al., 1989; Ferguson, 2008). Throughout its distribution, the tuff flowed in paleovalleys, especially where it reached and overtopped the paleodivide to the northeast (Young and Brennan, 1974; Young, 1979; Glazner et al., 1986). The tuff’s distribution and flow are similar to those of the Oligocene–Miocene tuffs in western Nevada, although this southern region did not undergo the older volcanism. Notably, the Peach Springs Tuff also flowed upstream and crossed the paleodivide, although the divide was no more than ~50 km east of the caldera. The similarity in early Cenozoic (post-Laramide) and Middle Miocene drainages in southwestern Arizona suggests that this paleodivide maintained its location during that time (Young, 1979).

Interpretations of paleodrainages in southeastern California and northwestern Arizona partly conflict (Figs. 10 and 11). Davis et al. (2010) interpreted that a Late Paleocene–Early Eocene drainage system (California River) extended northeast from southeastern California to northeastern Utah. In contrast, Howard (1996, 2000) interpreted a Late Paleocene–Middle Miocene, ancestral Colorado River that drained southward across southeastern California, perpendicular to and across the drainage of Davis et al. (2010), and then west to the Pacific Ocean (Fig. 11). Lease et al. (2009) found two segments of a generally west-trending paleovalley offset ~24 km by right-lateral faulting in the eastern California shear zone where the two regional drainages would cross (Fig. 10), but did not identify which way the paleoriver flowed. A possible reconciliation is that detritus in Utah interpreted by Davis et al. (2010) to come from southeastern California did so before formation of the Kingman uplift. Beard et al. (2010) constrained the Kingman uplift to between 70 Ma and Paleocene (65.5–55.8 Ma; Walker and Geissman, 2009), because paleocanyons contain Paleocene deposits. Wernicke (2011) showed the northeastward-draining California River of Davis et al. (2010) being truncated by ca. 55 Ma and the upper reaches of that river becoming part of an “Arizona River” system that includes the ancestral Colorado River of Howard (1996, 2000) and flowed to the Pacific Ocean.

The southernmost recognized major paleodrainage systems are in southern Arizona and northern Sonora and connect across the San Andreas fault system to Eocene coastal deposits in southern California and northern Baja California (Fig. 11; Abbott and Smith, 1978, 1989; Howard, 1996, 2000). The distribution of these drainages indicates that the paleodivide turned far to the east, following the Laramide belt. Low-temperature thermochronology from the Peninsular Ranges in northern Baja California
Figure 11.
indicates rapid cooling and exhumation into the Paleogene, much slower cooling beginning by ca. 45 Ma, and exhumation related to rifting in the Gulf of California in the Late Miocene (Seiler et al., 2011). The cooling history and interpreted exhumation is similar to but displaced slightly younger than those for the Sierra Nevada (Cecil et al., 2006; Seiler et al., 2011). The thermochronology suggests that paleodrainages across the Peninsular Ranges were established at least by ca. 45 Ma, consistent with the Eocene age of deposits in the drainages, and could have been maintained into the Late Miocene, although upstream parts in Arizona and Sonora were probably disrupted by Oligocene–Early Miocene extension (Spencer et al., 1995; Gans, 1997).

These observations suggest that the location of the early to middle Cenozoic paleodivide was controlled to the north largely by uplift of the Sevier orogenic belt, whereas to the south it was largely controlled by the locus of Laramide deformation, which partly overprinted the Sevier belt and altered older drainage. Cenozoic magmatism everywhere postdates formation of the paleovalleys, so the processes that generated magmatism, including rollback of the shallow Farallon slab, did not generate the Nevada-plano. In particular, the caldera-batholith belt of the ignimbrite flareup (Best et al., 1989) long postdates the paleodrainages and does not correlate spatially with the paleodivide, so was postdates the paleodrainages and does not correlate spatially with the paleodivide, so was not a significant influence on regional paleo-topography. Establishment of paleodrainages along the entire length of the paleodivide by the Paleocene seems to contradict the interpretation that uplift migrated southward with magmatism (Mix et al., 2011).

From stable isotope data, Mix et al. (2011) interpret that uplift to 3–4 km elevation swept southward through the North American Cordilleran during the Eocene as a result of removal of the Farallon slab. Although Mix et al. (2011) conceded that a late Mesozoic Nevadaplano highland may have formed as a result of crustal thickening from contraction, they interpreted most uplift to be in the Eocene and that maximum elevations were reached in the Eocene–Oligocene. Ash-flow tuff and paleovalley distributions do not constrain absolute elevations, and so do not confirm or deny this interpretation. However, lack of significant paleovalley incision during magmatism seems more consistent with little surface uplift at that time.

CONCLUSIONS

1. The 28.9 Ma tuff of Campbell Creek erupted from a caldera in north-central Nevada and spread through paleovalleys across northern Nevada and the Sierra Nevada, over a modern area of at least 55,000 km². Corrected for later extension, the tuff flowed at least ~200 km to the west, downvalley and across what is now the Basin and Range–Sierra Nevada structural and topographic boundary to the western foothills of the Sierra Nevada, and ~215 km to the northeast, partly upvalley, across an inferred paleodivide, and downvalley to the east to the present East Humboldt Range in northeastern Nevada. With the Nine Hill and Peach Springs Tuffs, the tuff of Campbell Creek is one of the most extensive tuffs of western North America.

2. The distribution of the tuff of Campbell Creek and other middle Cenozoic ash-flow tuffs supports the concept that what is now the Great Basin was an erosional highland, commonly referred to as the Nevadaplano, with a north-south paleodivide through east-central Nevada. Major rivers drained westward to the Pacific Ocean and eastward, probably to the Uinta Basin.

3. The Sierra Nevada was the western flank of this erosional highland in the middle Cenozoic. Paleodrainages definitely crossed the northern and central Sierra Nevada and may have crossed the southern Sierra Nevada.

4. Based on comparison with Quaternary ash-flow tuffs, the great flow distances of middle Cenozoic tuffs do not require, but also do not preclude, that the erosional highland be much higher than ~1.5 km.

5. Major extension that dismembered the highland and generated Basin and Range structure and topography had to be mostly post–23 Ma in western Nevada and post–29 Ma in northeastern Nevada, including in the region of the Ruby Mountains metamorphic core complex.

6. The erosional highland extended at least from Idaho to northern Sonora, Mexico, and probably mostly resulted from Sevier contraction, overprinted by Laramide contraction in and south of southern Nevada.

REFERENCES CITED


Cassel, E.J., Calvert, A., and Graham, S.A., 2009a, Age, geo-


DeCelles, P.G., 1988, Middle *Cenozoic* depositional, tec-


Eocene–Early Miocene paleotopography


Wood, J.L., 1994, A re-evaluation of the origin of kaolinite in the Ione depositional system (Eocene), Sierra foothills, California [M.S. thesis]: Los Angeles, California State University, 85 p.


