Ash-flow tuffs and paleovalleys in northeastern Nevada: Implications for Eocene paleogeography and extension in the Sevier hinterland, northern Great Basin

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ABSTRACT

Northeastern Nevada is generally interpreted as an area of large-magnitude Eocene extension possibly due to gravitational collapse of crust thickened during the Sevier orogeny. The extensional interpretation is based in part on the presence of widespread Eocene conglomerates and lacustrine basins, as well as on thermochronology-based evidence of major Eocene cooling and uplift of the Ruby Mountains–East Humboldt Range core complex.

The distribution of 45–40 Ma ash-flow tuffs and interbedded coarse conglomerates and lacustrine deposits, however, indicates they were predominantly deposited in a system of east-draining paleovalleys incised into a plateau or moderate-relief upland. A large, contiguous sedimentary basin probably was never present. Paleovalleys were as much as 10 km wide and 500 m to possibly as much as 1.6 km deep, based on the thickness of intravalley deposits. Ash-flow tuffs are widely distributed near source calderas but are almost entirely confined to the paleovalleys as little as 20 km from their source. Basal, mostly pre-volcanic conglomerates contain clasts up to 6 m in diameter. The clasts are well rounded, indicating significant fluvial transport, not derivation from nearby fault scarps. Lacustrine deposits also are restricted to paleovalleys and accumulated during two periods that are interpreted to coincide with episodes of minor, northwest-directed extension, one before 41 Ma and possibly as old as 46 Ma, and another between 40 and 38 Ma. Extension formed small displacement, northeast-striking, mostly down-to-the-northwest faults that temporarily dammed the paleovalleys to form lakes. Lakes probably also formed where volcanic rocks or landslides dammed paleovalleys, a common process both in the Eocene and historically in the western United States. The absence of major Eocene extension suggests that gravitational collapse of overthickened crust, even assisted by thermal weakening of lithosphere by intense magmatism, was not sufficient to generate major extension.

Absolute elevation of the high plateau is uncertain, but it was high enough to have paleovalleys as much as 1.6 km deep. Based on published paleoflora data, interfluves could have been at elevations of ~4 km.

The Eocene paleovalleys in northeastern Nevada most likely drained eastward to remnants of the Uinta basin. An approximately north-south paleodivide through northeastern Nevada separated these east-draining paleovalleys from paleovalleys that drained westward to the Pacific Ocean.

Keywords: paleogeography, extension, Eocene, ash-flow tuff, Nevada.

INTRODUCTION

The Eocene tectonic history and paleotopography of northeastern Nevada are uncertain and debated (Axelrod, 1966b; Christiansen and Yeats, 1992; McGrew and Snee, 1994; Snoke et al., 1997; Chase et al., 1998; Wolfe et al., 1998; McGrew et al., 2000; Rahl et al., 2002; Howard, 2003; Horton et al., 2004; Cline et al., 2005; Hickey et al., 2005). The region underwent multiple episodes of contraction in the Paleozoic and Mesozoic followed by extension in the mid- to late Cenozoic (Armstrong, 1968; Coney and Harms, 1984; Christiansen and Yeats, 1992; Wernicke, 1992; DeCelles, 2004; Dickinson, 2006). Eastern Nevada was in the hinterland of the Late Cretaceous Sevier orogenic belt (Armstrong, 1968; Miller and Gans, 1989; Camilleri et al., 1997; Vandervoort and Schmitt, 1990; Wright and Snoke, 1993; Howard, 2003; DeCelles, 2004). That the crust was 50–60 km thick in a belt through eastern Nevada and western Utah following contraction has been inferred from restoration of Tertiary extension (Coney and Harms, 1984), estimates of shortening in the overthrust belt (Thorman et al., 1991; Camilleri et al., 1997), and metamorphic mineral assemblages (Camilleri et al., 1997; McGrew et al., 2000; Lee et al., 2003).

The region was probably an eroding highland at the beginning of the Cenozoic (Armstrong, 1968; Coney and Harms, 1984; Christiansen and Yeats, 1992; Dilek and Moores, 1999; DeCelles, 2004). Sediments probably were carried eastward to the Uinta and Green River Basins (Baars et al., 1988; Hintze, 1993; Goldstrand, 1994), with little sedimentary record in Nevada until the Eocene (Fouch et al., 1979; Solomon et al., 1979; Vandervoort and Schmitt, 1990). However, interpretations of the absolute elevation, timing of uplift, and paleotopography vary widely. Coney and Harms (1984), Dilek and Moores (1999), and DeCelles (2004), drawing analogies to present-day Tibet and the Andean Plateau, inferred elevations of >3 km. They interpreted the high elevation to have resulted from thickening of crust, in large part during the Sevier orogeny. Absolute Eocene elevations inferred from fossil leaves from Copper Basin in northeastern Nevada, which has a present-day elevation of 2.2 km, range from 1.1 km (Axelrod, 1966b; Christiansen and Yeats, 1992), to 2.0 ± 0.2 km (Wolfe et al., 1998), to 1.6 ± 1.6 km or 2.8 ± 1.8 km (Chase et al., 1998). Although not providing absolute elevations, Horton et al. (2004) used stable isotope data to interpret that northeastern Nevada rose ~2 km between the middle Eocene and early Oligocene, contemporaneous with the period of intense magmatism, then subsided ~1–2 km since the middle Miocene. A rise of 2 km would seem either to

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Geosphere; February 2008; v. 4; no. 1; p. 1–35; doi: 10.1130/GES00122.1; 13 figures; 3 tables; 1 supplementary file.

require that northeastern Nevada was relatively low in the middle Eocene, contradicting the tie between Mesozoic crustal thickening and high elevation, or that it rose to elevations of as much as 5 km.

Cenozoic extension is generally interpreted to have begun in the Eocene, but the timing and amount of early extension are poorly constrained. Thermochronologic data have been interpreted to record initial exhumation of the Ruby Mountains metamorphic core complex between 63 and 49 Ma (Hodges et al., 1992; McGrew and Snee, 1994; McGrew et al., 2000). These data are difficult to interpret, however, and Howard (2003) concluded that the timing of early unroofing of the Ruby Mountains is uncertain. A gradient in K-Ar and $^{40}Ar/^{39}Ar$ biotite ages from 36 to 20 Ma west-northwest across the range is generally interpreted to reflect cooling during prolonged uplift (Kistler, 1981; Dokka et al., 1986; McGrew and Snee, 1994). Apatite fission-track and U-Th/He dating indicate the Ruby Mountains underwent rapid cooling and uplift in the middle Miocene (ca. 15 Ma; Colgan and Metcalf, 2006). Sedimentary basins in northeastern Nevada began filling as early as 46 Ma (Solomon et al., 1979; Haynes, 2003; Cline et al., 2005; Hickey et al., 2005). Most basins have been interpreted to be extensional (Axelrod, 1966b; Solomon et al., 1979; Clark et al., 1985; Vandervoort and Schmitt, 1990; Satarugsa and Johnson, 2000; Haynes, 2003; Cline et al., 2005; Hickey et al., 2005), while others have not—for example, the White Sage basin in westernmost Utah (Fig. 1; Potter et al., 1995; Dubiel et al., 1996).

The goal of this study is to document the distribution of regional, Eocene ash-flow tuffs and character of the Eocene paleosurface in northeastern Nevada. The study grew out of an investigation of the magmatic-tectonic setting and origin of the Eocene Carlin-type gold deposits, whose relationship to magmatism or extension and the depth of formation are highly debated (Muntean et al., 2004; Cline et al., 2005; Hickey et al., 2005; Ressel and Henry, 2006). By analyzing the distribution of 45–40 Ma ash-flow tuffs and contemporaneous sedimentary deposits, I infer a system of paleovalleys that drained eastward to the remnants of the Uinta-Green River basins of Utah. Complementary paleovalleys drained westward across the central Great Basin and Sierra Nevada to the Pacific Ocean. These data also indicate that Eocene extension consisted of possibly two episodes of relatively minor extension, one before ca. 41 Ma and another between 40 and 38 Ma. Most extension

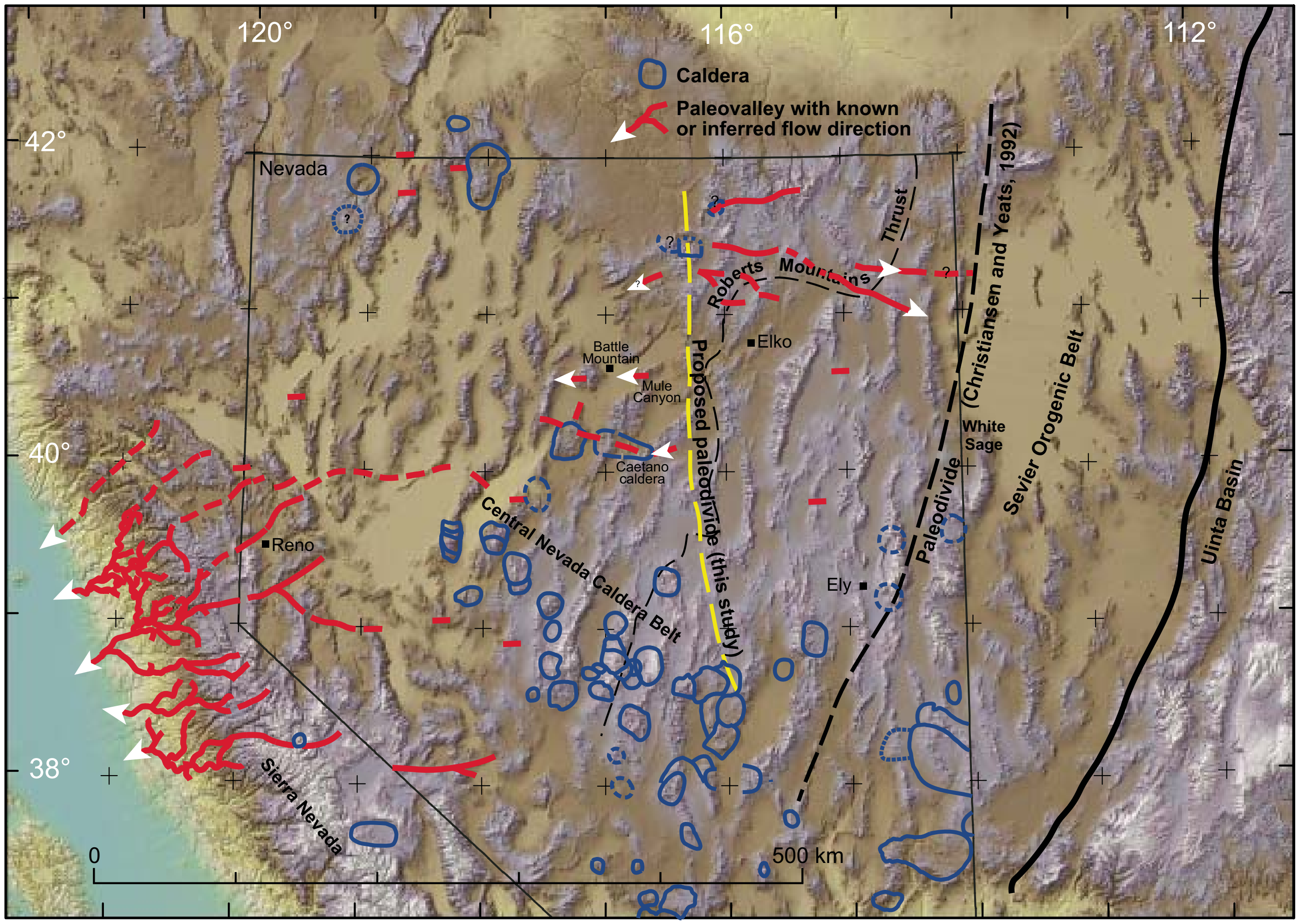

Figure 1. Digital elevation map of the Great Basin showing the distribution of known paleovalleys and a few segments (from Lindgren, 1911; Faulds et al., 2005; Garside et al., 2005; and this study), the Eocene paleodivide proposed from this study, and the paleodivide of Christiansen and Yeats (1992). The east-draining paleovalleys flowed to the Uinta basin; the west-draining paleovalleys flowed to the Pacific Ocean, which was in the Great Valley of California at the time. Arrows show flow direction where known or reasonably inferred.

in the region probably occurred in the middle Miocene or later.

GEOLOGIC SETTING

Northeastern Nevada underwent multiple episodes of contraction from the Late Devonian-Early Mississippian Antler orogeny through the Late Cretaceous Sevier orogeny (Roberts et al., 1958; Armstrong, 1968; Miller et al., 1992; Poole et al., 1992; Camilleri et al., 1997; Taylor et al., 2000; DeCelles, 2004; Dickinson, 2006). During the Antler orogeny, deep-water siliciclastic rocks (western assemblage) were thrust eastward over coeval shelf- and slope-facies marine rocks (eastern assemblage) along the Roberts Mountains thrust (Fig. 1). A thick, clastic wedge was deposited in front of the thrust belt (Burchfiel et al., 1992). Subsequent thrust belts developed progressively eastward, and eastern Nevada was in the hinterland of the Sevier orogenic belt, the youngest contractional deformation in the region (Armstrong, 1968; Miller and Gans, 1989; Vandervoort and Schmitt, 1990; Wright and Snoke, 1993; Howard, 2003; DeCelles, 2004).

Jurassic and Cretaceous plutons are scattered through northeastern Nevada (Coats, 1987; Wright and Snoke, 1993; Barton, 1996; Mortensen et al., 2000). They are relatively localized near the surface but presumably are more abundant at depth, as indicated by the abundance of granitic rocks in the Ruby Mountains metamorphic core complex (Snoke et al., 1997; McGrew et al., 2000; Howard, 2003; Lee et al., 2003). Late Cretaceous peraluminous granites formed by crustal anatexis in the thick crust are common (Miller et al., 1990; Lee et al., 2003).

Cenozoic magmatism began ca. 45 Ma in northeastern Nevada and was part of a southward-migrating belt of magmatism that swept from Washington and Idaho, through northeastern Nevada and northwestern Utah, and into central Nevada in the Oligocene (Stewart, 1980; Christiansen and Yeats, 1992; Brooks et al., 1995a, 1995b; Humphreys, 1995; Henry and Ressel, 2000a; this study). Eocene magmatism in northeastern Nevada was dominated by andesitic to dacitic lavas and compositionally similar intrusions in numerous centers. Several rhyolitic ash-flow tuffs also erupted, notably from one or more calderas near Tuscarora (Figs. 1 and 2; Henry et al., 1999), but the total number of tuffs and, with few exceptions, the location of their source calderas are uncertain. The early phase of Cenozoic magmatism in northeastern Nevada ended by ca. 35 Ma, and, with the exception of 29 Ma sills in the Ruby Mountains core complex (Wright and Snoke, 1993; MacCready et al., 1997; Howard, 2000), magmatism did not resume until the middle Miocene.

Sedimentary basins began to form at about the same time as magmatism began and are mostly interpreted to have resulted from extension (Solomon et al., 1979; Clark et al., 1985; Solomon, 1992; Satarugsa and Johnson, 2000; Haynes et al., 2002; Haynes, 2003). Eocene sedimentary rocks crop out widely in the region (Fig. 2) and are variably interpreted as contiguous parts of a single, large basin, generally referred to as the Elko basin (Smith and Ketner, 1976; Solomon, 1992; Christiansen and Yeats, 1992), or several isolated basins (Solomon et al., 1979; Nutt and Good, 1998). Haynes (2003), Cline et al. (2005), and Hickey et al. (2005) show a large, composite Elko basin that encompasses most Eocene sedimentary deposits but is partly separated by local topographic highs. As discussed below, I interpret many of the Eocene sedimentary deposits to have accumulated in relatively small basins along paleovalleys.

Eocene sedimentary rocks in the Elko Hills and Piñon Range have been called the Elko Formation (Smith and Ketner, 1978; Solomon et al., 1979; Haynes, 2003) and are interpreted to have accumulated in the hanging wall of the Ruby Mountains detachment fault (Fig. 3; Solomon et al., 1979; Satarugsa and Johnson, 2000; Haynes et al., 2002; Haynes, 2003). The Elko Formation in the Elko Hills consists of ~200 m of basal conglomerate overlain by ~600 m of lacustrine shale, oil shale, claystone, siltstone, and minor, water-laid tuff (Fig. 3; Solomon et al., 1979; Ketner and Alpha, 1992; Haynes, 2003). Zircon U-Pb ages on water-laid tuffs in the basal conglomerate and near the top of the lacustrine sequence are 46.1 ± 0.2 Ma and 38.9 ± 0.3 Ma, respectively (Haynes et al., 2002; Haynes, 2003). Outcrop, seismic-reflection, and borehole data indicate that Eocene sedimentary rocks considered equivalent to the Elko Formation are discontinuous in the subsurface along the Ruby Mountains front (Smith and Ketner, 1976; Satarugsa and Johnson, 2000; Haynes, 2003).

The little studied Bull Run basin is the other large area of Eocene sedimentary rocks (Fig. 3). The Bull Run basin is interpreted to have formed through extension by ca. 44 Ma and existed until 36 Ma (Axelrod, 1966a, 1966b; Clark et al., 1985). Basin fill consisting of basal conglomerate overlain by lacustrine shale, oil shale, siltstone, and marl is more than 950 and possibly 1500 m thick (Decker, 1962).

Interpreted episodes of extension began with initial exhumation of the Ruby Mountains between the Late Cretaceous and Eocene (McGrew and Snee, 1994; Camilleri and Chamberlain, 1997; McGrew et al., 2000) and development of the Elko basin of the Elko Hills at ca. 46 Ma (Haynes et al., 2002; Cline et al., 2005; Hickey et al., 2005). An episode of extension that tilted the Elko Formation near Elko ~10° to 15° to the southeast before ca. 38 Ma is well established (Brooks et al. 1995a; Henry and Faulds, 1999; Henry et al., 2001; Haynes, 2003; Hickey et al., 2005). Most cooling and presumed uplift from extension of the Ruby Mountains is interpreted to have occurred between 36 and 20 Ma based on K-Ar and $^{40}Ar/^{39}Ar$ ages (Kistler et al., 1981; Dokka et al., 1986; McGrew and Snee, 1994) and at ca. 15 Ma based on apatite fission-track and (U-Th)/He ages (Colgan and Metcalf, 2006). The apparent 36–20 Ma uplift generated no basins or sedimentary deposits, however, and any sediment was transported out of the region (Wallace et al., 2008). Major uplift and erosion of the Ruby Mountains in the middle Miocene (ca. 15 Ma) is also documented by the first appearance of clasts of Paleozoic metasedimentary rocks and garnet-muscovite–bearing granites in the Humboldt Formation west of the Ruby Mountains (Smith and Ketner, 1976). Extension of variable magnitude was widespread in northern Nevada at ca. 15 Ma and includes development of the northern Nevada rift beginning ca. 16 Ma (Zoback et al., 1994; John et al., 2000), extension of the nearby Shoshone and Toiyabe Ranges (Colgan et al., 2008), and major exhumation of the Snake Range metamorphic core complex (Miller et al., 1999).

Volcanism in northeastern Nevada renewed at 16.5 Ma with small eruptions of Steens-type basalt near Copper Basin (Coats, 1964; Rahl et al., 2002). The Jarbidge Rhyolite, a widespread and voluminous group of rhyolite lavas (Fig. 2), began to erupt by 16.2 Ma (my unpublished $^{40}Ar/^{39}Ar$ age in Bull Run basin).

The giant Carlin-type gold deposits formed contemporaneously with Eocene magmatism (for example, 40–37 Ma in the Carlin trend, Fig. 3; Ressel and Henry, 2006) and possibly are coeval with extension (Seedorff, 1991; Cline et al., 2005; Hickey et al., 2005). The origin of Carlin-type deposits and their relation to magmatism and extension are strongly debated (Muntean et al., 2004; Cline et al., 2005; Hickey et al., 2005; Ressel and Henry, 2006).

ASH-FLOW TUFF STRATIGRAPHY

Geochronologic, Geochemical, and Petrographic Data

I use the distribution and character of regional ash-flow tuffs and interbedded sedimentary rocks to determine the Eocene paleogeography and extensional history of the region. Ash-flow tuff correlation is based on my own and published

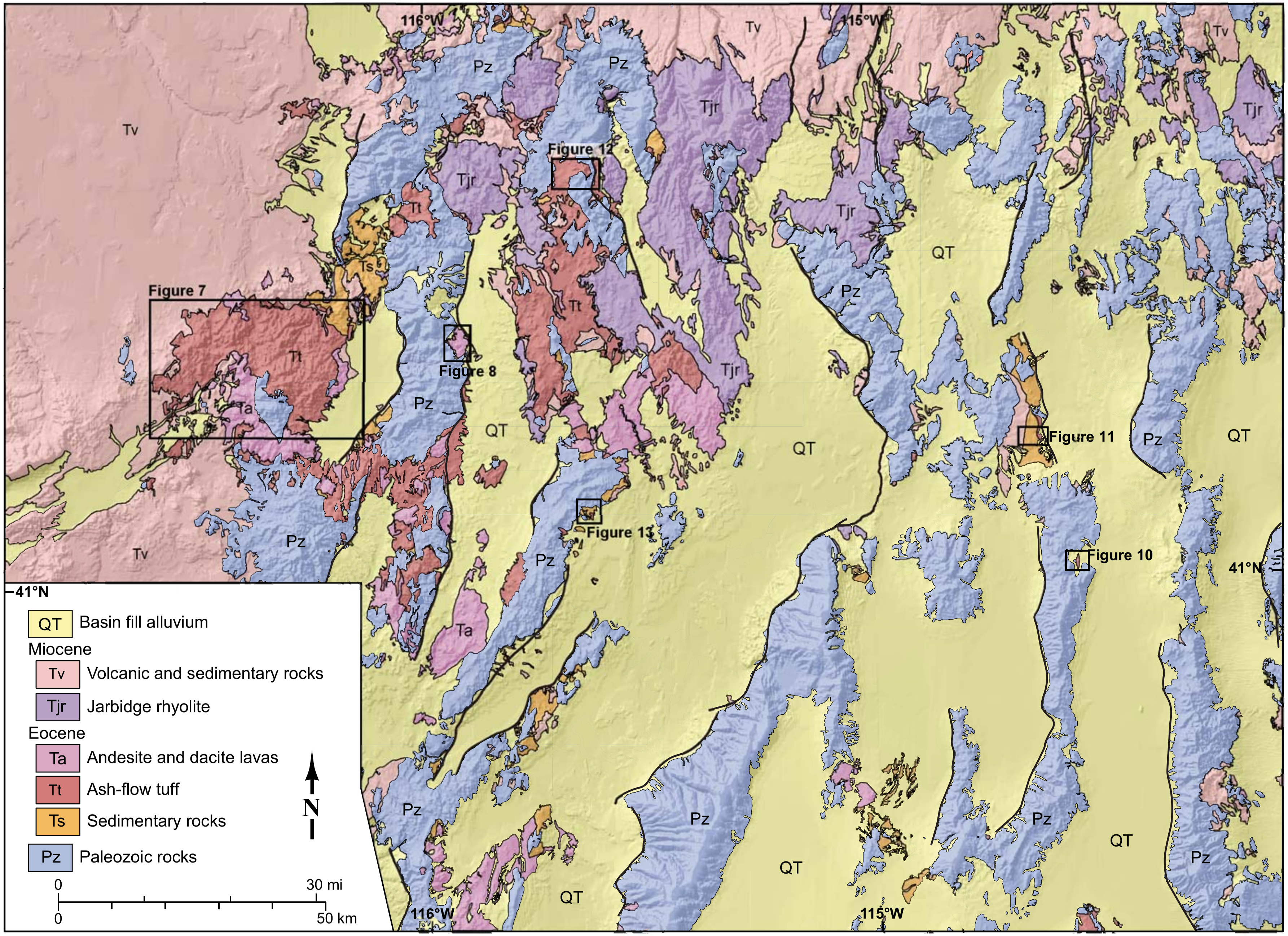

Figure 2. Geologic map of northeastern Nevada (simplified from Coats, 1987), showing locations of more detailed maps.

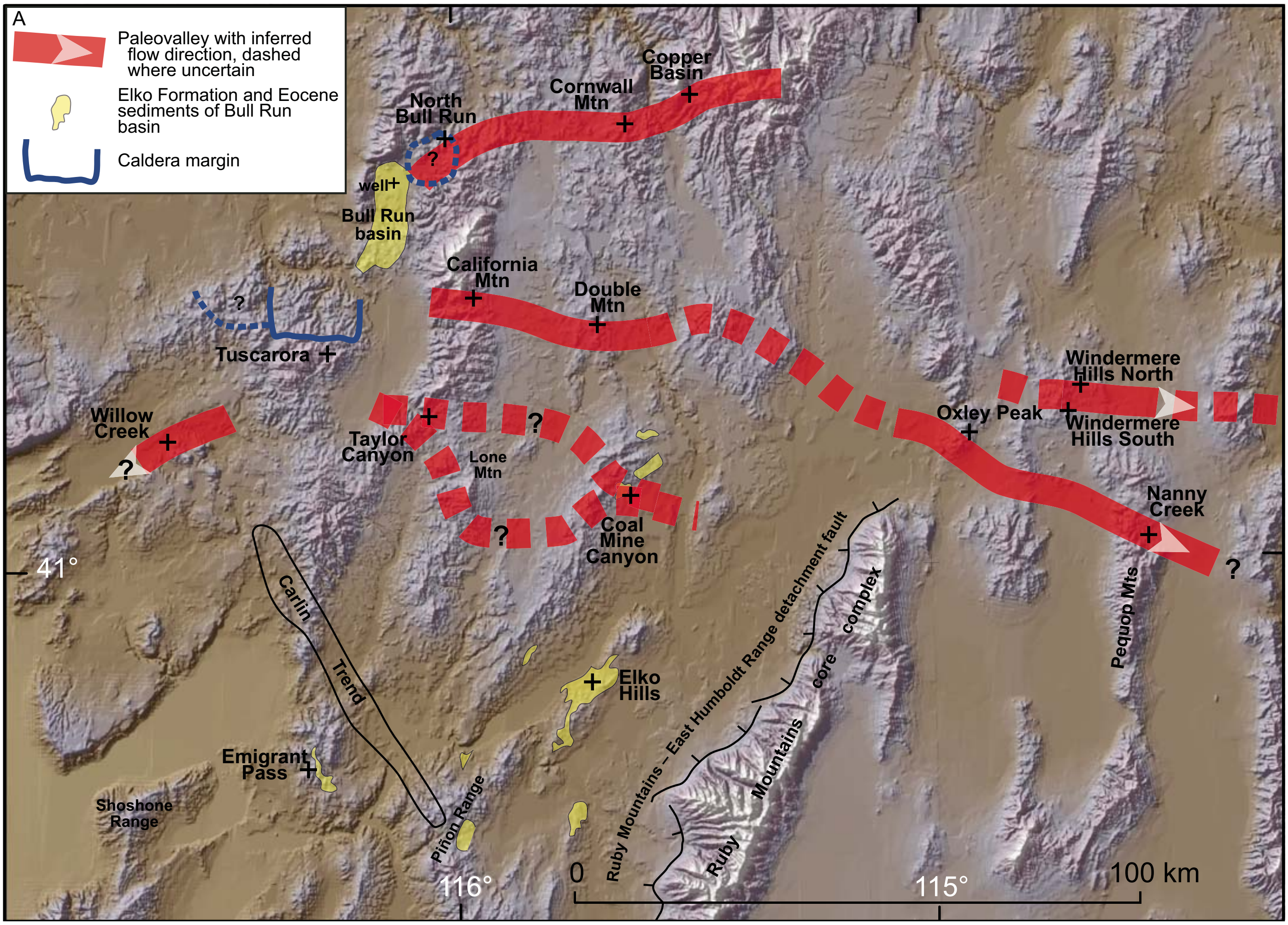

Figure 3 (*on this and following page*). **(A) Map of interpreted paleovalleys showing locations of simplified stratigraphic columns for paleovalley segments and correlations of tuffs along the paleovalleys (B).**

geologic mapping, stratigraphy, petrography, geochemistry, and especially $^{40}Ar/^{39}Ar$ ages. This section focuses on $^{40}Ar/^{39}Ar$ ages, petrography, and geochemistry. Stratigraphy and other field relations are discussed in more detail in the descriptions of individual paleovalley segments.

Table 1 lists more than 30 new and published $^{40}Ar/^{39}Ar$ ages on ca. 40–45 Ma ash-flow tuffs in northeastern Nevada (analytical data are available in Table S1[1]). All ages have been normalized to an age of 28.02 Ma for the common Fish Canyon Tuff sanidine monitor (Renne et al., 1998). Some published ages that were disturbed or could not be assigned unequivocally to a specific tuff because of location uncertainty are not listed. Ages obtained in this study are primarily on replicate single grains of sanidine, because they provide highly precise, reproducible ages that can distinguish volcanic events separated by as little as 100,000 yr at the ca. 40 Ma age of these rocks. The K/Ca ratio of sanidine also provides an additional correlation tool, and analyses of single grains permit the identification of xenocrysts (Fig. 4) (McIntosh et al., 1990; John et al., 2008). All published ages cited in Table 1 are by step-heating of bulk samples of sanidine or biotite (Hofstra, 1994; Brooks et al., 1995a, 1995b; Mueller et al., 1999). Most of these ages agree with the results obtained in this study, although bulk samples can contain xenocrysts. Fortunately, Mesozoic granitic rocks, the only significant potential source of biotite or feldspar xenocrysts other than other tuffs, are not major constituents of bedrock in the region. Biotite dates can be significantly disturbed by minor alteration, especially chloritization (DiVincenzo et al., 2003). Data presented in this paper illustrate that sanidine ages provide the most certain and straightforward correlations, whereas comparison between sanidine and biotite dates can be uncertain. Most published K-Ar ages are insufficiently precise to help with correlation and are not considered.

Thirty-five whole-rock, ash-flow tuff samples were analyzed for major and trace elements by XRF (X-ray fluorescence) at Washington State University (Table 2) and were combined with published analyses in Mueller (1992), Brooks et al. (1995a, 1995b), and Wallace (2003a). Comparison of the new data with published analyses of the same units and, in a few cases, the same outcrops, indicates some interlaboratory bias, which complicates correlation. Alteration, which has affected some samples, is another complication. For

[1]If you are viewing the PDF of this paper or reading it offline, please visit http://dx.doi.org/10.1130/GES00122.S1 or the full-text article on www.gsajournals.org to view Table S1.

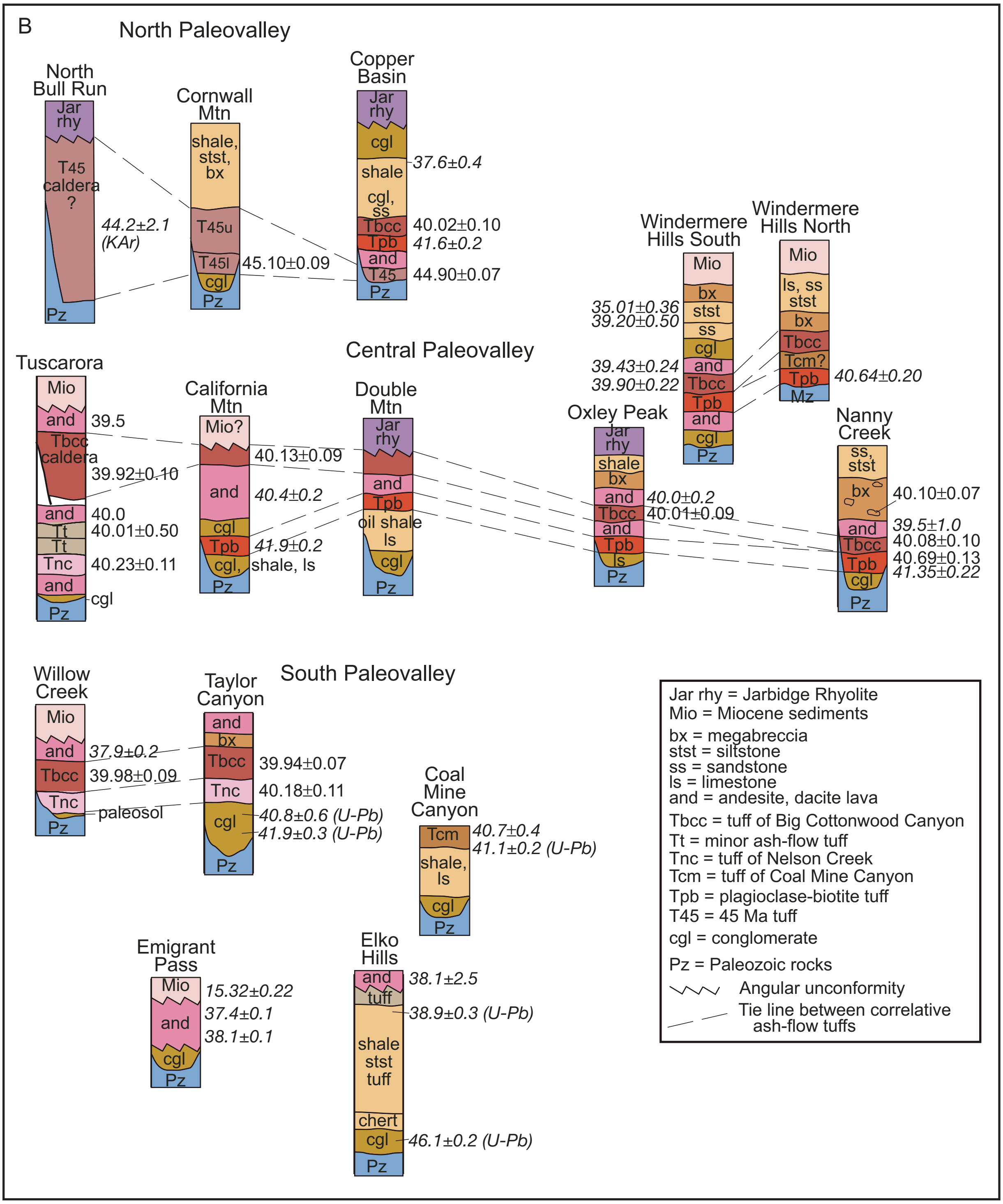

Figure 3 (*continued*). (B) Dates in regular type are from this study; dates in italics are from Axelrod (1966a), Brooks et al. (1995a, 1995b), Henry and Faulds (1999), Mueller et al. (1999), Haynes (2003), and Wallace (2003a).

example, a plot of SiO_2 versus $Na_2O + K_2O$, a standard classification scheme for volcanic rocks, shows some clustering but considerable overlap between different tuffs and that some samples of the tuff of Big Cottonwood Canyon are silicified (Fig. 5). Selected immobile elements from these analyses are still useful for correlation—for example, Zr versus TiO_2. Obviously, analyses of altered samples should not be used for petrogenetic studies.

Regionally Widespread Ash-Flow Tuffs

The tuff of Big Cottonwood Canyon (Tbcc in all figures) is the youngest, most distinctive, and most widely distributed, and therefore, best

TABLE 1. $^{40}Ar/^{39}Ar$ AGES, ASH-FLOW TUFFS OF NORTHEASTERN NEVADA

Paleovalley	Location	Sample #	Lab	Mineral	Age (Ma)	±2σ	K/Ca	±2σ	n	%^{39}Ar	Latitude	Longitude	Reference
Tuff of Big Cottonwood Canyon													
Intracaldera Tuff													
	Tuscarora volcanic field	H96-42	New Mexico Tech	sanidine	39.92	0.12	64.1	15.6	15		41.38000	–116.26000	This study
	Single pumice	H96-80	New Mexico Tech	sanidine	39.92	0.10	64.5	11.1	11		41.40833	–116.34333	This study
Outflow Tuff													
C	Millsite volcanics	H05-90	New Mexico Tech	sanidine	40.13	0.09	67.8	9.7	10		41.40210	–115.91145	This study
C	Oxley Peak	H03-111	New Mexico Tech	sanidine	40.01	0.09	67.4	19.8	15		41.20005	–114.91047	This study
C	Oxley Peak	88T41	USGS Denver	biotite	40.02	0.26				63	41.20111	–114.91000	Brooks et al. (1995)
C	Nanny Creek	H03-102	New Mexico Tech	sanidine	39.93	0.08	71.7	24.4	11		41.02788	–114.54582	This study
C	Nanny Creek megabreccia	H05-38	New Mexico Tech	sanidine	40.10	0.07			14		41.03117	–114.52611	This study
C	Nanny Creek	88T55	USGS Denver	biotite	40.15	0.24				94	41.02472	–114.53444	Brooks et al. (1995)
C	Nanny Creek megabreccia	90B9B	USGS Denver	biotite	39.87	0.26				60	41.02611	–114.53417	Brooks et al. (1995)
C	Windermere Hills	HOL-4	USGS Denver	sanidine	39.90	0.22				91	41.24561	–114.63131	Mueller et al. (1999)
C	Wood Hills	92BWH1	USGS Denver	biotite	40.0	0.2				d	41.08139	–114.88194	Brooks et al. (1995)
N	Copper Basin	H06-121	New Mexico Tech	sanidine	40.02	0.10	61.1	15.0	12		41.74840	–115.49297	This study
S	Willow Creek Reservoir	96 WC28	New Mexico Tech	sanidine	39.98	0.09	69.1	7.8	15		41.22656	–116.53853	Wallace (2003a); this study
S	South of Lone Mountain	S48	New Mexico Tech	sanidine	39.94	0.07	66.3	8.4	19		41.04948	–116.00676	This study*
Minor tuff													
Toe Jam Mountain		H97-51	New Mexico Tech	plagioclase	40.01	0.49			39		41.28000	-116.46000	Henry and Boden (1999)
Tuff of Nelson Creek													
C	Toe Jam Mountain	H97-71	New Mexico Tech	sanidine	40.14	0.11	71.2	14.6	12		41.25167	–116.42000	Henry and Boden (1999)
C	Toe Jam Mountain	H97-110	New Mexico Tech	sanidine	40.23	0.11	67.6	12.9	9		41.26333	–116.50167	Henry and Boden (1999)
C	Sugarloaf Butte	H96-32	New Mexico Tech	sanidine	40.10	0.10	70.8	13.6	15		41.22333	–116.25167	Henry et al. (1999)
C	Taylor Canyon	H97-30	New Mexico Tech	sanidine	40.18	0.11	71.7	18.1	15		41.27333	–116.13333	Henry et al. (1999)
C	Singletree Creek	1057	New Mexico Tech	sanidine	40.13	0.08	74	14.8	15		41.17706	–116.05273	This study*
C	Singletree Creek	1054	New Mexico Tech	sanidine	40.07	0.08	70	16.7	14		41.19663	–116.01547	This study*
Tuff of Coal Mine Canyon													
S	Coal Mine Canyon	12624	USGS Denver	hornblende	40.66	0.40				93	41.11444	–115.62944	Brooks et al. (1995)
Pyroclastic-fall tuff, Copper Basin													
N	Copper Basin	970728-3	Ohio State	sanidine	37.6	0.4				95	41.76450	–115.46798	Rahl et al. (2002)
Plagioclase-biotite tuff(s)													
C	Windermere Hills	WC-1	USGS Denver	biotite	40.64	0.20				64	41.31444	–114.64167	Mueller et al. (1999)
C	Nanny Creek	H03-101	New Mexico Tech	plagioclase	40.69	0.13			18		41.02745	–114.54837	This study
C	Nanny Creek	88T56	USGS Denver	biotite	41.34	0.22				86	41.02500	–114.53444	Brooks et al. (1995)
C	Millsite volcanics	MSV91-4	USGS Denver	biotite	41.9	0.4				?	41.39215	–115.93672	Hofstra, 1994
N	Copper Basin	960702-1A	Ohio State	biotite	41.6	0.2				70	41.73496	–115.48191	Rahl et al. (2002)
45 Ma tuff													
N	Copper Basin	H06-123	New Mexico Tech	sanidine	44.90	0.07	66.2	5.2	9		41.74391	–115.48560	This study
N	Mount Velma	H05-71	New Mexico Tech	sanidine	45.10	0.09	67.4	5.4	9		41.70302	–115.62411	This study

Note: C—Central; N—North; S—South paleovalleys; n—number of individual grains used to calculate weighted mean age. %^{39}Ar = percentage of ^{39}Ar used in plateau; d = disturbed. Decay constants and isotopic abundances after Steiger and Jäger (1977); λ_β = 4.963 × 10-10 yr-1; $\lambda_{\varepsilon+\varepsilon'}$ = 0.581 × 10^{-10} yr^{-1}; 40K/K = 1.167 × 10-4. Minerals were separated from crushed, sieved samples by standard magnetic and density techniques; sanidine was leached with dilute HF to remove matrix and handpicked. Analyses at the New Mexico Geochronological Research Laboratory used the methods of McIntosh et al. (2003). 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—28.02 Ma (Renne et al., 1998). Single sanidine grains were fused with a CO_2 laser operating at 10 W. Extracted gases were purified with SAES GP-50 getters. Argon was analyzed with a Mass Analyzer Products (MAP) model 215-50 mass spectrometer operated in static mode. Weighted mean $^{40}Ar/^{39}Ar$ ages calculated by the method of Samson and Alexander (1987).

*Sample provided by Ken Hickey.

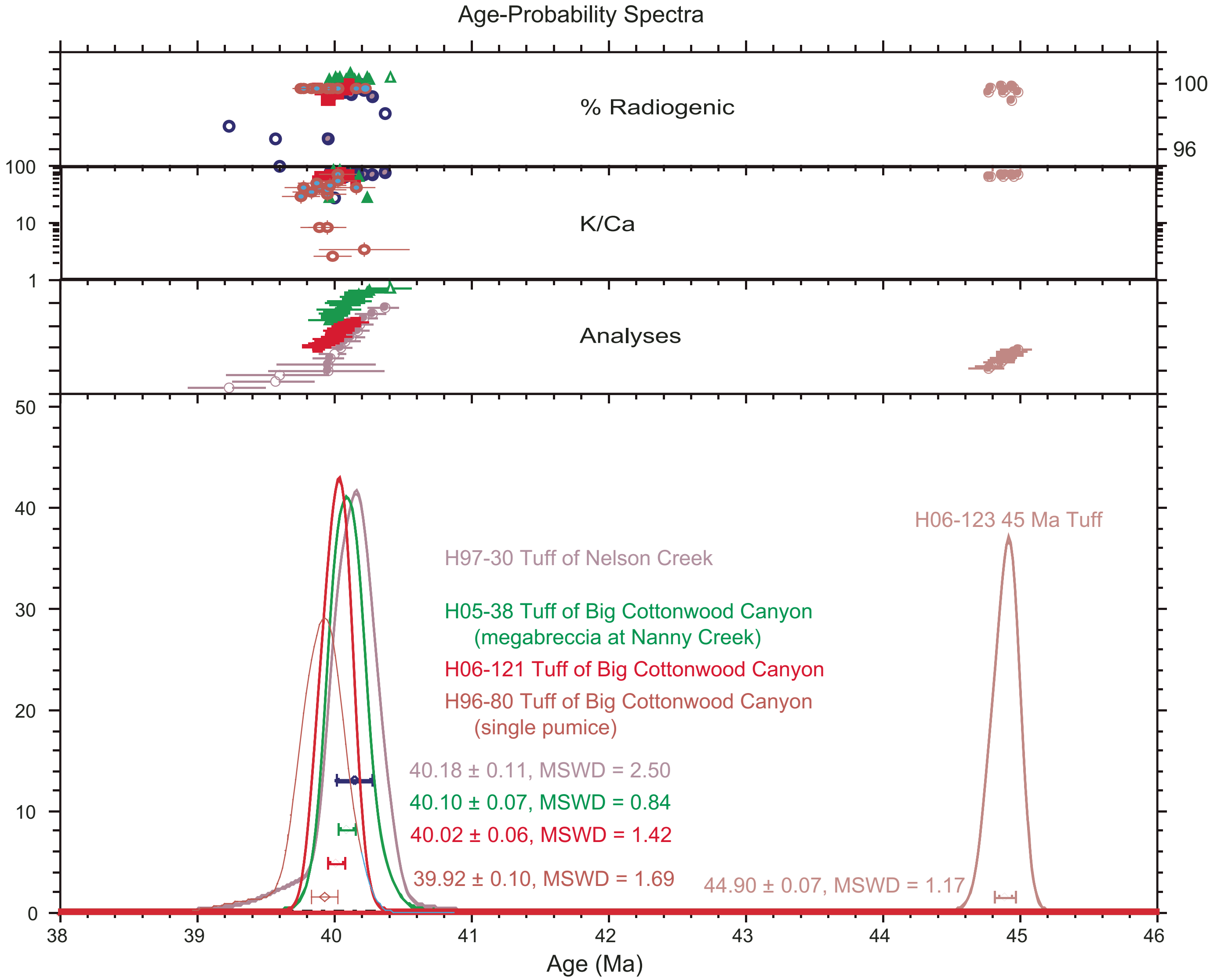

Figure 4. Probability plots of representative sanidine $^{40}Ar/^{39}Ar$ ages of ash-flow tuffs dated in this study showing different ages of the 45 Ma tuff, tuff of Nelson Creek (40.14 ± 0.06 Ma; n = 6), and tuff of Big Cottonwood Canyon (39.99 ± 0.08; n = 9). Unfilled data points were not used in age calculation (for example, analyses that are outside 2σ uncertainty or are of plagioclase).

marker of the major ash-flow tuffs of the region (Tables 1 and 2; Figs. 3, 5, and 6). The sparsely to moderately porphyritic tuff is high-silica rhyolite with distinctively low, whole-rock TiO_2, P_2O_5, Ba, and Zr contents, although samples with more than 78% SiO_2 are silicified (Fig. 5; Table 2). The compositionally most unusual sample is a pumice fragment from intracaldera tuff; the fragment has the lowest SiO_2 content and is notably enriched in Ba and Zr. Nine sanidine $^{40}Ar/^{39}Ar$ ages of this study range from 39.92 ± 0.10 to 40.13 ± 0.10 Ma and give a mean and standard deviation of 39.99 ± 0.08 Ma (Table 1). Published biotite and sanidine step-heating ages are similar. Paleomagnetic data at Nanny Creek, Windermere Hills, and Oxley Peak support correlation (Palmer and MacDonald, 2002). The tuff erupted from a caldera in the northern part of the Tuscarora volcanic field, the only well-established caldera for the tuffs described in this paper (Figs. 2 and 7; Henry and Boden, 1998; Castor et al., 2003).

The tuff of Nelson Creek (Tnc) underlies the tuff of Big Cottonwood Canyon near the Tuscarora volcanic field (Figs. 3 and 7). It contains less sanidine and quartz and more biotite and hornblende than the tuff of Big Cottonwood Canyon (Fig. 6). It is a low-silica rhyolite that contains distinctly more TiO_2, P_2O_5, Ba, Zr, and Nb than the tuff of Big Cottonwood Canyon (Fig. 5). Six sanidine $^{40}Ar/^{39}Ar$ ages range from 40.07 ± 0.08 to 40.23 ± 0.11 Ma and give a mean and standard deviation of 40.14 ± 0.06 Ma. The tuff of Nelson Creek has only been found within about a 50 km radius southeast and southwest of the Tuscarora volcanic field. A volcanic center at Lone Mountain could be the source (Henry and Ressel, 2000b). Henry et al. (1999) suggested that a large area of thick, undivided ash-flow tuff west of the Big Cottonwood Canyon caldera (Fig. 2) could be a caldera and source for the tuff of Nelson Creek. However, absence of the tuff in the central paleovalley (Fig. 3) argues against that area being the source.

The tuff of Coal Mine Canyon (Tcm) is a high-silica dacite characterized by abundant phenocrysts of plagioclase, biotite, and hornblende (Figs. 5 and 6; Table 2). Petrographically

TABLE 2. CHEMICAL ANALYSES, ASH-FLOW TUFFS OF NORTHEASTERN NEVADA

	Tuff of Big Cottonwood Canyon								
Sample:	H96-80	H05-90	H06-47	H06-48	H03-111	H06-114	H06-46	H06-110	H06-111
Map symbol:	Tbcc	Tbcc	Tbcc	Tbcc	Tbcc	Tbcc	Tbcc	Tbcc	Tbcc
Location:	Tuscarora	California Mountain	Double Mountain	Double Mountain	Oxley Peak	Wood Hills	Windermere Hills	Windermere Hills	Windermere Hills
Latitude:	41.40833	41.40210	41.35316	41.35187	41.20005	41.08266	41.24649	41.27440	41.28992
Longitude:	–116.34333	–115.91145	–115.72384	–115.72468	–114.91047	–114.88235	–114.62424	–114.66672	–114.63294
Note:	single pumice		silicified				silicified	breccia	silicified
SiO_2	73.08	76.09	77.97	77.37	76.67	76.67	78.80	75.45	78.81
TiO_2	0.21	0.117	0.095	0.114	0.125	0.120	0.107	0.139	0.127
Al_2O_3	14.48	12.99	11.93	12.32	12.48	12.76	11.81	13.12	11.78
FeO*	1.80	1.26	1.16	1.18	1.20	1.09	0.95	1.29	0.68
MnO	0.02	0.013	0.044	0.016	0.008	0.026	0.006	0.009	0.020
MgO	0.63	0.12	0.12	0.15	0.47	0.14	0.32	0.57	0.39
CaO	1.38	0.89	0.82	0.92	0.39	0.91	0.75	1.32	0.86
Na_2O	2.97	3.23	3.10	3.20	1.14	3.33	3.02	0.94	1.13
K_2O	5.38	5.25	4.74	4.72	7.48	4.93	4.21	7.15	6.17
P_2O_5	0.06	0.032	0.016	0.023	0.034	0.027	0.023	0.020	0.027
Sum	98.62	98.06	97.44	97.56	94.53	97.10	97.01	92.89	94.91
Sc	4	2	2	1	2	2	2	3	2
V	19	11	8	8	10	9	9	11	10
Cr	0	1	1	2	2	2	4	3	4
Ni	9	6	2	2	7	4	2	4	4
Cu	6	8	1	1	1	2	5	3	3
Zn	40	24	29	33	29	35	26	34	20
Ga	19	15	13	15	15	14	12	14	13
Rb	139	159	145	151	204	166	94	137	126
Sr	292	140	115	128	84	122	272	500	130
Y	15	17	11	13	14	13	13	10	10
Zr	225	117	101	101	109	106	92	112	101
Nb	13.4	12.2	10.0	11.2	13.5	11.9	10.2	12.2	11.2
Ba	2819	805	781	654	768	695	830	746	718
La	73	50	41	42	42	49	38	37	39
Ce	117	81	71	77	68	77	68	69	76
Nd	39.5	30	22	25	26	25	25	24	25
Pb	23	24	23	26	22	28	17	27	16
Th	21.5	24	24	27	23	31	28	25	27

(*continued*)

TABLE 2. CHEMICAL ANALYSES, ASH-FLOW TUFFS OF NORTHEASTERN NEVADA (*continued*)

	Tuff of Big Cottonwood Canyon			Minor tuffs		Tuff of Nelson Creek		
Sample:	H03-102	H05-38	H06-121	H97-53	H97-51	H97-30	H97-71	H97-110
Map symbol:	Tbcc	Tbcc	Tbcc	Tt	Tt	Tnc	Tnc	Tnc
Location:	Nanny Creek	Nanny Creek	Copper Basin	Tuscarora	Tuscarora	Tuscarora	Tuscarora	Tuscarora
Latitude:	41.02788	41.03117	41.74840	41.302	41.280	41.273	41.252	41.263
Longitude:	–114.54582	–114.52611	–115.49297	–116.487	–116.460	–116.133	–116.420	–116.502
Note:	silicified	breccia						
SiO_2	78.06	79.86	75.32	72.19	69.46	73.64	71.94	73.98
TiO_2	0.115	0.119	0.162	0.42	0.41	0.51	0.57	0.40
Al_2O_3	11.96	10.84	13.78	14.23	16.08	14.69	14.28	13.38
FeO*	0.83	0.39	0.98	2.10	2.46	1.21	2.66	2.28
MnO	0.014	0.007	0.027	0.02	0.07	0.01	0.02	0.02
MgO	0.13	0.06	0.21	0.32	0.48	0.38	0.69	0.50
CaO	0.93	0.47	1.16	1.36	2.20	2.54	3.11	2.45
Na_2O	2.81	1.90	2.06	3.53	3.36	3.08	2.99	2.93
K_2O	5.09	6.31	6.27	5.68	5.36	3.81	3.55	3.91
P_2O_5	0.059	0.035	0.039	0.13	0.11	0.13	0.19	0.15
Sum	98.13	98.38	96.32	99.34	97.92	99.13	98.69	98.73
Sc	2	3	3	6	9	8	9	3
V	10	12	12	22	20	75	73	34
Cr	2	4	4	2	0	13	14	6
Ni	5	6	4	8	8	13	10	8
Cu	3	5	4	2	3	2	3	5
Zn	27	21	42	52	74	21	50	31
Ga	14	14	15	19	20	16	19	18
Rb	169	159	183	150	136	111	98	120
Sr	111	83	262	296	478	508	541	416
Y	15	13	16	23	23	40	18	13
Zr	100	106	154	244	284	191	201	168
Nb	12.0	10.6	11.7	16.1	17.7	12.9	11.9	10.8
Ba	535	797	1789	1830	1830	1890	1940	1380
La	36	42	52	48.2	49.1	45.8	45.2	39.7
Ce	64	72	85	87.5	88.5	78.9	77.2	68.4
Nd	22	23	29	34.9	33.4	35.6	34.1	28.2
Pb	34	17	27	20.2	23.6	16.7	13.8	16.2
Th	26	21	27	15.4	15.1	13.6	12.2	14

(*continued*)

TABLE 2. CHEMICAL ANALYSES, ASH-FLOW TUFFS OF NORTHEASTERN NEVADA (*continued*)

	Tuff of Coal Mine Canyon			Plagioclase-biotite tuff			
Sample:	H06-41	H06-42	H06-113	H03-101	H03-112	H06-106	H06-112
Map symbol:	Tcm	Tcm	Tcm	Tpb	Tpb	Tpb	Tpb
Location:	Coal Mine Canyon	Coal Mine Canyon	Windermere Hills	Nanny Creek	Oxley Peak	Windermere Hills	Windermere Hills
Latitude:	41.12634	41.11827	41.31401	41.02745	41.20300	41.25542	41.31387
Longitude:	–115.63617	–115.63367	–114.64253	–114.54837	–114.91130	–114.65904	–114.64150
Note:						breccia	
SiO_2	68.41	70.67	65.51	69.52	69.45	66.67	71.61
TiO_2	0.686	0.665	0.705	0.467	0.433	0.458	0.364
Al_2O_3	15.84	15.41	15.81	16.26	16.49	16.61	14.46
FeO*	2.86	1.73	4.51	1.95	1.73	3.54	1.51
MnO	0.028	0.035	0.044	0.032	0.011	0.026	0.019
MgO	0.99	0.70	1.22	0.72	0.56	0.73	0.53
CaO	4.45	3.77	4.47	2.75	2.17	2.20	3.70
Na_2O	3.11	3.02	3.18	3.82	3.31	3.19	3.57
K_2O	3.36	3.75	4.30	4.35	5.70	6.43	4.11
P_2O_5	0.259	0.248	0.238	0.135	0.135	0.136	0.138
Sum	96.82	96.15	95.74	98.19	96.85	96.84	96.25
Sc	12	11	12	7	5	5	4
V	107	68	112	42	42	52	22
Cr	21	21	22	4	4	4	3
Ni	8	5	11	8	7	5	4
Cu	7	7	13	4	4	3	3
Zn	64	51	66	69	39	69	46
Ga	19	16	17	19	16	19	16
Rb	60	91	108	120	150	162	119
Sr	602	539	508	493	426	441	447
Y	16	14	14	19	15	19	17
Zr	162	165	171	255	246	258	207
Nb	8.9	8.8	10.2	12.8	11.8	11.2	9.6
Ba	1157	1303	1380	1642	2217	1591	1449
La	36	34	36	46	45	36	37
Ce	59	64	71	76	71	69	69
Nd	26	25	27	32	31	28	27
Pb	17	17	19	24	23	22	21
Th	15	14	14	18	18	17	20

(*continued*)

TABLE 2. CHEMICAL ANALYSES, ASH-FLOW TUFFS OF NORTHEASTERN NEVADA (*continued*)

	Plagioclase-biotite tuff (*continued*)						
Sample:	H05-40	H05-79	H05-86	H05-89	H06-49	H06-50	H06-117
Map symbol:	Tpb	Tpb	Tpb	Tpb	Tpb	Tpb	Tpb
Location:	California Mountain	California Mountain	California Mountain	California Mountain	Double Mountain	Double Mountain	Copper Basin
Latitude:	41.42218	41.39703	41.39966	41.40289	41.39005	41.39005	41.74492
Longitude:	–115.94983	–115.93274	–115.91345	–115.91287	–115.71927	–115.71927	–115.48117
Note:	vitrophyre						
SiO_2	69.16	71.92	70.16	71.05	69.12	70.55	70.00
TiO_2	0.441	0.410	0.443	0.430	0.454	0.423	0.434
Al_2O_3	16.19	14.52	16.01	15.93	16.16	15.66	15.28
FeO*	2.94	2.84	1.79	1.34	2.69	2.40	3.27
MnO	0.065	0.033	0.018	0.016	0.033	0.027	0.041
MgO	0.66	0.41	0.94	0.38	0.73	0.54	0.83
CaO	2.71	2.43	2.05	2.52	2.71	2.35	2.90
Na_2O	3.17	3.38	3.15	4.10	3.96	3.71	3.27
K_2O	4.52	3.91	5.20	4.10	3.99	4.19	3.85
P_2O_5	0.146	0.147	0.247	0.140	0.142	0.150	0.132
Sum	94.90	96.48	93.94	97.62	97.63	96.98	96.94
Sc	5	5	5	5	6	6	6
V	37	33	34	28	45	37	48
Cr	3	3	5	3	4	4	5
Ni	5	6	6	8	3	3	3
Cu	4	3	3	4	3	1	4
Zn	71	59	61	72	65	50	70
Ga	19	19	18	19	19	16	17
Rb	120	88	116	114	125	129	105
Sr	508	455	329	501	480	430	543
Y	21	17	16	19	18	19	15
Zr	249	217	236	250	259	241	230
Nb	13.4	11.6	12.4	13.2	11.6	11.1	9.2
Ba	1872	1433	1452	1578	1607	1588	1534
La	47	38	38	47	41	41	35
Ce	79	67	68	70	78	75	63
Nd	34	29	27	32	30	31	23
Pb	21	16	17	18	23	24	20
Th	16	14	16	18	18	22	12

(*continued*)

TABLE 2. CHEMICAL ANALYSES, ASH-FLOW TUFFS OF NORTHEASTERN NEVADA (*continued*)

	45 Ma tuff				
Sample:	H06-127	H06-128	H05-71	H05-69	H06-122
Map symbol:	T45	T45	T45	T45	T45
Location:	Bull Run	Bull Run	Cornwall Mountain	Cornwall Mountain	Copper Basin
Latitude:	41.67658	41.67425	41.70302	41.68312	41.74413
Longitude:	–116.03831	–116.02642	–115.62411	–115.66932	–115.48768
Note:					
SiO_2	72.17	73.21	71.93	73.64	72.32
TiO_2	0.316	0.347	0.138	0.292	0.326
Al_2O_3	15.32	15.11	14.92	14.57	15.48
FeO*	1.89	1.18	2.24	1.49	2.04
MnO	0.015	0.041	0.011	0.041	0.011
MgO	0.41	0.30	0.21	0.60	0.38
CaO	1.87	1.88	0.93	1.92	1.70
Na_2O	3.32	3.48	3.33	3.36	3.32
K_2O	4.61	4.37	6.23	4.01	4.35
P_2O_5	0.081	0.083	0.054	0.072	0.076
Sum	95.26	96.48	95.96	94.69	96.68
Sc	4	4	2	3	4
V	23	27	12	23	32
Cr	6	5	4	4	5
Ni	4	4	7	6	4
Cu	5	5	3	4	5
Zn	48	36	42	37	42
Ga	16	16	19	16	17
Rb	135	120	152	144	121
Sr	348	385	126	364	373
Y	12	12	18	10	10
Zr	152	175	151	167	155
Nb	9.7	8.7	12.1	10.9	8.6
Ba	1275	1576	1082	1363	1308
La	34	42	51	42	39
Ce	63	74	82	72	64
Nd	20	28	32	22	22
Pb	24	21	24	24	23
Th	23	20	19	21	20

Note: All analyses by X-ray fluorescence at the Geoanalytical Laboratory, Washington State University. All analyses normalized to 100% anhydrous. FeO*—total Fe reported asFeO. Sum—total before normalization to 100% anhydrous.

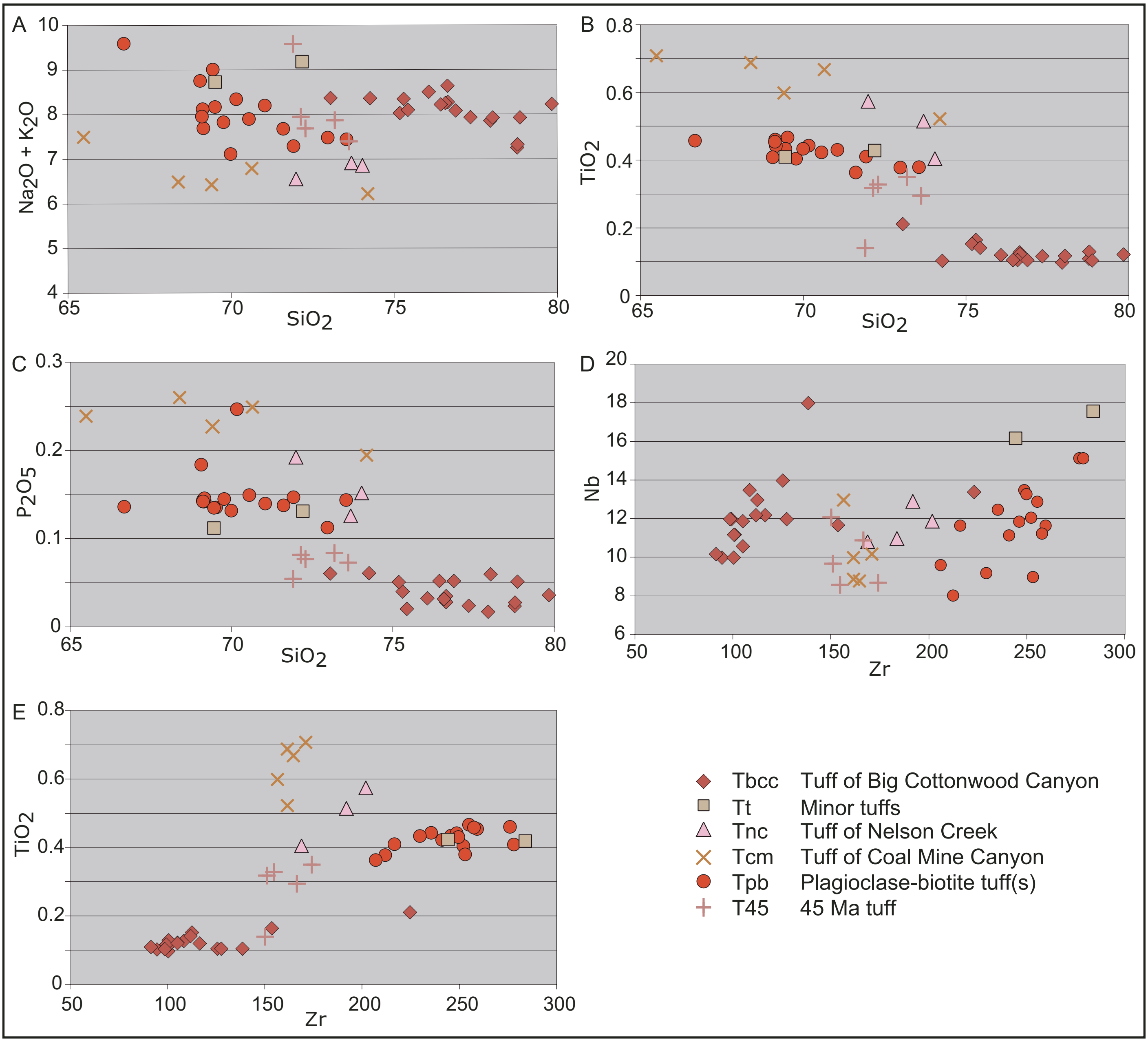

Figure 5. Compositional variation diagrams for whole-rock samples of Eocene ash-flow tuffs of northeastern Nevada. Data are from this study (Table 2) and Mueller (1992), Brooks et al. (1995a, 1995b), and Wallace (2003a). Major elements normalized to 100% volatile-free. (A) SiO_2/Na_2O+K_2O. (B) SiO_2/TiO_2. (C) SiO_2/P_2O_5. (D) Zr/Nb. (E) Zr/TiO_2.

and compositionally similar tuff crops out in the Windermere Hills. Although compositionally variable, these tuffs form a group distinct from other tuffs, for example, in Zr and TiO_2 abundances (Fig. 5). Reported ages from Coal Mine Canyon are 40.7 ± 0.4 Ma (hornblende, $^{40}Ar/^{39}Ar$; Brooks et al., 1995a, 1995b) on densely welded tuff and 41.1 ± 0.2 (zircon, U-Pb; Haynes et al., 2002) on an immediately underlying poorly welded tuff that we interpret to be a basal part of the densely welded tuff. The probably correlative tuff in the Windermere Hills underlies the tuff of Big Cottonwood Canyon and overlies a tuff dated at 40.64 ± 0.20 Ma.

Ash-flow tuff dominated by plagioclase and biotite phenocrysts with lesser and variable amounts of hornblende, pyroxene, quartz, and sanidine is widespread through the central and eastern part of Figure 2. Despite minor phenocryst variations, these tuffs are compositionally similar and mostly occupy the same stratigraphic position (Table 2; Figs. 3, 5 and 6). Five $^{40}Ar/^{39}Ar$ ages fall into two groups (Table 1). Two ages, one on biotite and one on plagioclase, are ca. 40.7 Ma (40.64 ± 0.20, Windermere Hills; 40.69 ± 0.13 Ma, Nanny Creek), whereas three others, all on biotite, are 41.34 ± 0.22 (Nanny Creek), 41.9 ± 0.4 (California Mountain), and 41.6 ± 0.2 Ma (Copper Basin)**.** Despite the two age groups, two separate plagioclase-biotite tuffs are not present at any location east of the Tuscarora volcanic field. Moreover, the 40.69 Ma plagioclase date and

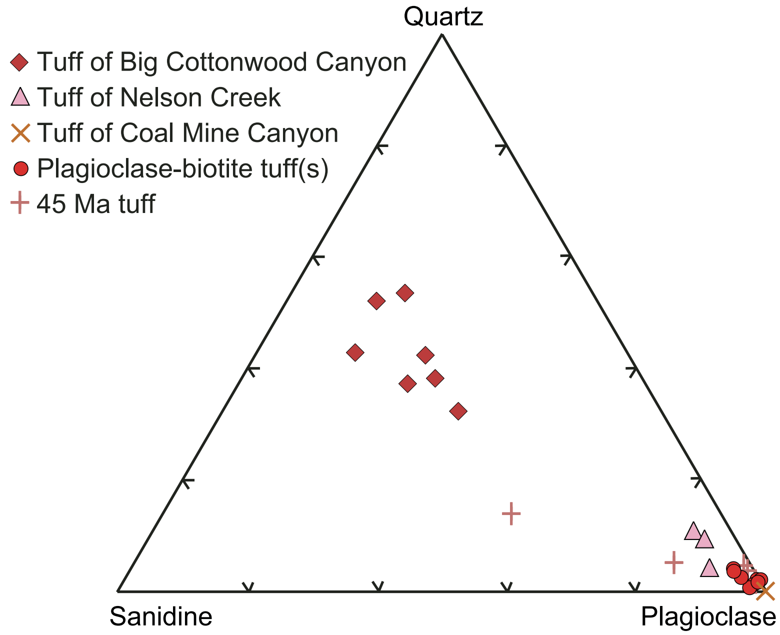

Figure 6. Ternary plot showing relative modal abundances of quartz, plagioclase, and sanidine phenocrysts in Eocene ash-flow tuffs of northeastern Nevada.

the 41.34 Ma biotite date are from the same tuff at Nanny Creek. The 40.64 Ma-tuff crops out in the Windermere Hills, where it underlies the plagioclase-biotite-hornblende tuff that is probably the tuff of Coal Mine Canyon. Paleomagnetic data support correlation of the Nanny Creek and Windermere Hills plagioclase-biotite tuffs (Palmer and MacDonald, 2002). Haynes (2003) obtained a zircon U-Pb date of 40.8 ± 0.6 Ma on a thin, altered plagioclase-biotite tuff in Taylor Canyon. All these data support an age of ca. 40.7 Ma for the plagioclase-biotite tuff at Nanny Creek and in the Windermere Hills. Whether the two other tuffs with ca. 42 Ma ages are an older but petrographically and chemically similar tuff is uncertain. In the following discussions, all plagioclase-biotite tuffs are treated as one and informally termed the plagioclase-biotite tuff (Tpb). But the possibility that they are two separate tuffs must be kept in mind.

The oldest recognized, Cenozoic ash-flow tuff in Nevada, informally termed the 45 Ma tuff (T45), crops out in a belt from Copper Basin westward past Cornwall Mountain to a possible caldera source north of Bull Run basin (Fig. 3). It overlaps in SiO_2 content with several other tuffs but has lower TiO_2 and P_2O_5 contents (Table 2; Figs. 5 and 6). Moreover, two sanidine $^{40}Ar/^{39}Ar$ ages are 44.90 ± 0.07 and 45.10 ± 0.09 Ma (Table 1). Although the two ages do not overlap at 2 sigma, their similarity to each other, their difference from all other tuff ages, and the stratigraphic, petrographic, and compositional evidence for correlation seem incontrovertible. The tuff may have erupted from a caldera at the north end of Bull Run basin, where petrographically and compositionally similar tuff is at least 1000 m thick and contains probable andesite megabreccia. However, descriptions by Decker (1962) suggest the tuff north of Bull Run basin may occupy a northeast-trending paleovalley (discussed in the next section). Axelrod (1966a) reported a biotite K-Ar age of 44.2 ± 2.1 Ma from this tuff in the northern Bull Run basin. Although not diagnostic, this K-Ar age is one of the oldest in the region and supports correlation.

Two minor, plagioclase-biotite tuffs (Tt) in the western part of the Tuscarora volcanic field do not correlate with any of the more extensive tuffs to the east. Both of these tuffs overlie the tuff of Nelson Creek; therefore, they are younger than the plagioclase-biotite tuff(s), and one has a plagioclase $^{40}Ar/^{39}Ar$ age of 40.01 ± 0.49 Ma (Table 1). They are compositionally similar to the plagioclase-biotite tuff but are somewhat distinguished by high Nb concentrations (Table 2; Fig. 5).

PALEOVALLEY GEOLOGY

Ash-flow tuffs are widely distributed and thickest around Tuscarora, Taylor Canyon, and north of Bull Run basin, areas of known or suspected calderas, and generally thin away from the source areas (Figs. 2 and 3). At distances of only 20 km from these areas, the tuffs and sedimentary rocks are confined to narrow channels cut into Paleozoic bedrock. This section describes three paleovalley systems, starting with the best understood, central paleovalley. All paleovalleys are discontinuously exposed across a series of ranges that formed during middle Miocene or later Basin and Range extension. Figure 3 shows the generalized stratigraphy and $^{40}Ar/^{39}Ar$ and U-Pb ages of tuffs and lavas in the paleovalleys, and Table 3 summarizes the characteristics of paleovalleys, conglomerates, and lacustrine deposits at each location. The thickness of tuffs, lavas, and sedimentary rocks in the paleovalleys is used as the best approximation of the depth of the paleovalley. Obviously, paleovalleys could be either underfilled or overfilled, although overfill other than by intermediate lavas is probably minor.

These descriptions are based on my detailed mapping of the Tuscarora volcanic field, California Mountain, and Nanny Creek, and on detailed published maps in Willow Creek, Windermere Hills, Oxley Peak, and Coal Mine Canyon, and on more generalized or regional maps north of Bull Run basin, at Cornwall Mountain, Copper Basin, Double Mountain, and Taylor Canyon. In all areas, I examined stratigraphy and rock types and, where appropriate, collected samples for chemical analysis or $^{40}Ar/^{39}Ar$ dating. In some cases noted below, my interpretations are significantly different than those of the original workers.

Central Paleovalley

The central paleovalley can be traced ~150 km from near the Tuscarora volcanic field eastward over the Independence Mountains at California Mountain as far as Nanny Creek in the Pequop Mountains (Fig. 3). Although gaps exist across several valleys, continuity is based on proximity and trend of individual segments and similar stratigraphy, especially the presence of the tuff of Big Cottonwood Canyon.

Tuscarora Volcanic Field

The Tuscarora volcanic field, the largest Eocene volcanic center in the region, erupted rhyolitic to dacitic tuffs and andesitic to dacitic lavas during several episodes between ca. 40.2 and 39.5 Ma (Fig. 7; Henry et al., 1999; Castor et al., 2003). Basal, pre-volcanic conglomerate crops out in two locations near the south edge of the field and may be covered elsewhere. However, the presence of a buried paleovalley in this area is speculative. Otherwise, the oldest volcanic rocks, the tuff of Nelson Creek and andesite lavas, rest directly on a low-relief surface developed on Paleozoic rocks west of the field. An extensive dacite-andesite complex, a group of

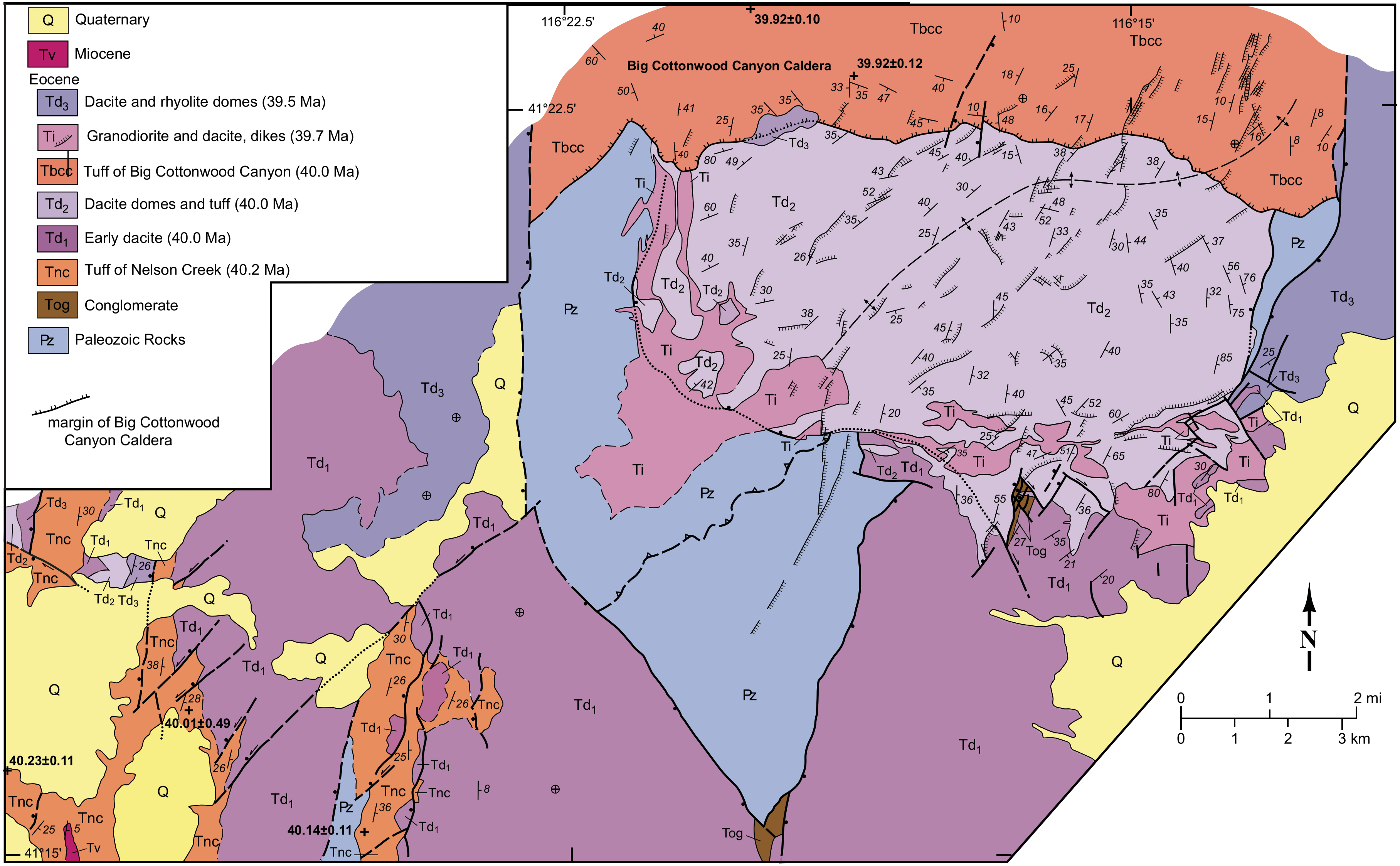

Figure 7. Simplified geologic map of the Tuscarora volcanic field with $^{40}Ar/^{39}Ar$ dates of ash-flow tuffs (Henry et al., 1999). Magmatism, including eruption of the tuff of Big Cottonwood Canyon at 40.0 Ma, occurred over a brief, intense period between 40.2 and 39.5 Ma. Any paleovalley is largely buried, but pre-volcanic conglomerate is exposed along the southern and southeastern flank of the volcanic field.

TABLE 3. CHARACTERISTICS OF PALEOVALLEYS IN NORTHEASTERN NEVADA

	Central Paleovalley			
Location:	Tuscarora	California Mountain	Double Mountain	Oxley Peak
Width:	Mostly covered	5 km	7.5 km	Two small channels separated by 4 km: northern 1.7 km; southern 1 km
Depth:		≥800 m	≥600 m	Northern, ≥500 m; southern, ?
Tilt:	Main part of volcanic field is not tilted	SE, E, ~20°–30°	W, ~25°	E to SE, ~40°
Total thickness:		~800 m	~600 m	~500 m
Lacustrine deposits:	None known	finely laminated shale and micrite	Calcareous to silicified shale, oil shale, sandstone, limestone, minor pyroclastic-fall tuff	Northern channel has basal 1-m-thick, oolitic limestone and as much as 100 m of tuffaceous siltstone and shale that overlie 40.0 Ma tuff of Big Cottonwood Canyon and dacite lavas
Thickness:		≤5-m-thick lenses interbedded with basal conglomerate; ~25 m overlying basal conglomerate	≥250 m	Northern channel, 100 m
Age:		≥41.9 Ma	≥41 Ma	≤40.0 Ma
Fossils, leaves:		Metasequoia, gastropods, ostracodes		
Conglomerate:	Basal, prevolcanic	Basal, prevolcanic and overlying 41.9 Ma ash-flow tuff	Basal, prevolcanic	Southern channel has basal pebble conglomerate
Thickness:	≥30 m	Basal: 15 m; intravolcanic: 15 m	≤100 m	~10 m
Clast types:	Chert, quartzite, argillite, and minor limestone	Basal: quartzite, chert, chert-quartzite pebble conglomerate; intervolcanic: quartzite, chert, chert-quartzite pebble conglomerate		quartzite, chert, chert-pebble conglomerate, tuff of Nanny Creek
Clast sizes:	Quartzite to 1.3 m, but mostly ≤20 cm	Basal: to 70 cm; intervolcanic: to 1.5 m		Pebble
Rounding:	subrounded to angular	Basal: well-rounded to subrounded; intravolcanic: subrounded to subangular		Well-rounded
Relation to Miocene:	Angular unconformity, ~15°, west of volcanic field	Angular unconformity with possible Miocene sediments, ~15°	Angular unconformity with Jarbidge Rhyolite, ~15°	Conformable with Jarbidge Rhyolite dated at 15 ± 1.2 Ma (zircon fission-track; Thorman et al., 2003)
Notes:	39.8 to 39.6 Ma, northeast-striking dike swarm indicates northwest least principal stress	Episode of northwest extension tilted rocks about 20° to southeast after 40.4 Ma, possibly after 40.0 Ma	Common soft-sediment deformation in lacustrine deposits	Overlain conformably by Jarbidge Rhyolite dated at 15 ± 1.2 Ma (zircon fission-track; Thorman et al., 2003)
Sources (this study for all):	Henry and Boden, 1998, 1999; Henry et al., 1999	Hofstra, 1994	Coats, 1987 and unpublished	Thorman et al. (2003); C.H. Thorman (2007, personal commun.)

(*continued*)

TABLE 3. CHARACTERISTICS OF PALEOVALLEYS IN NORTHEASTERN NEVADA (*continued*)

	Central Paleovalley		Northern Paleovalley	
Location:	Windermere Hills	Nanny Creek	North Bull Run	Cornwall Mountain
Width:	≥10 km	>6 km	4 to 7 km	>7 km
Depth:	1.5 km?	1.6 km?	≥1 km	As much as 1.5 km, but some fault repetition
Tilt:	W, 30°–40°	E, 30°–60°	W, ~25°	W, 40°–50°
Total thickness:	>1 km(?) but large uncertainty in amount of fault repetition		45 Ma tuff is >1 km	~1.5 km, but some repetition by faulting
Lacustrine deposits:	Thick sequence consisting of tuffaceous sandstone passing upward to tuffaceous siltstone and shale	Undated tuffaceous silstone and shale at top of section		Thick sequence of tuffaceous siltstone, shale, sandstone, and minor pebble conglomerate with numerous interbedded breccia lenses composed of subrounded to subangular clasts of probable 45 Ma tuff to 4 m diameter
Thickness:	600 m(?)	?		~1 km
Age:	≤39.4 Ma	≤39.5 Ma, if present		≤45.0 Ma
Fossils, leaves:	Abundant paleoflora, fish fossils, gastropods, and pelycypods		Abundant paleoflora in Bull Run basin between 44 and 36 Ma based on KAr dates (Axelrod, 1966a, 1966b)	
Conglomerate:	Several thin intratuff and major post–39.4 conglomerates	basal, prevolcanic	Base not observed	Basal, prevolcanic
Thickness:	Post–39.4: ~100 m	30 m		20–50 m
Clast types:	Post–39.4 Ma: mostly tuffs and other volcanic rocks and some quartzite and Triassic? Intratuff: mostly volcanic and some granitoid	Chert ± quartzite-pebble conglomerate, quartzite, chert, coarse-grained granodiorite, *crinoidal limestone*; sparse porphyritic andesite, silicified lake deposits with ostracodes		Dark and light quartzite, chert-pebble conglomerate, limestone, and granodiorite
Clast sizes:	Post–39.4 Ma: pebbles to boulders	Chert ± quartzite-pebble conglomerate to 6 m, chert to 1.4 m, granitoid to 1.7 m		Quartzite to 4 × 6 m; granodiorite to 2 × 3 m
Rounding:	Rounded	Rounded		Rounded to well-rounded
Relation to Miocene:	Conformable with oldest middle Miocene sediments	No Miocene	Probable angular unconformity with Jarbidge Rhyolite	No Miocene
Notes:	Megabreccia composed of tuff of Big Cottonwood Canyon and tuff of Nanny Creek; stratigraphic position uncertain. Mueller (1992) noted abrupt along and across strike variation in rock types. Imbricated clasts in post–39.4 Ma conglomerate indicate flow toward east. Overlying middle Miocene rocks tilted the same.	Thick megabreccia consisting of clasts mostly of tuff of Big Cottonwood Canyon, lesser tuff of Nanny Creek, and minor andesite.		Coash (1967) notes that this area was a topographic basin because volcanic rocks are much thicker here than immediately to north and south.
Sources (this study for all):	Mueller (1992, 1993); Mueller and Snoke (1993); Mueller et al. (1999)	Brooks et al. (1995a, 1995b); C.H. Thorman (2007, personal commun.)	Decker (1962)	Coash (1967)

(*continued*)

TABLE 3. CHARACTERISTICS OF PALEOVALLEYS IN NORTHEASTERN NEVADA (*continued*)

	Northern Paleovalley	Southern Paleovalley		
Location:	Copper Basin	Willow Creek	Taylor Canyon	Coal Mine Canyon
Width:	>5 km	≥10 km	≥10 km	>3 km
Depth:	≥1.6 km	≥450 m	800 m(?)	>450 m
Tilt:	NW, ~20°–40°	E, ~30°	SE, ~30°	SE, ~20°–40°
Total thickness:	~1.6 km but much of it is covered	>450 m	~400 m	420 m
Lacustrine deposits:	Thick lacustrine sequence above tuff of Big Cottonwood Canyon consists of tuffaceous shale and sandstone	Not present	Minor limestone in scattered locations	Claystone, siltstone, shale, oil shale, limestone, sandstone, and minor pebble conglomerate
Thickness:			10 m	350 m
Age:	Between 40.0 and 37.6 Ma		≥41 Ma	≥41.0 Ma
Fossils, leaves:	Abundant paleoflora interpreted initially to indicate an elevation of 1.1 km (Axelrod, 1966a, 1966b), then reinterpreted as 2.0 ± 0.2 km (Wolfe et al., 1998) or 2.8 ± 1.8 km (Chase et al., 1998)			Ostracods and abundant paleoflora
Conglomerate:	Base not exposed; intravolcanic conglomerate overlies tuff of Big Cottonwood Canyon and underlies thick lacustrine section. Additional conglomerate post ~30 Ma	Paleosol with angular quartzite	Basal, prevolcanic and early intravolcanic	Basal, prevolcanic
Thickness:	Intravolcanic, ≥20 m; post–30 Ma, ~400 m	1–3 m	140 m	60–80 m
Clast types:	Intravolcanic: quartzite, chert, limestone, tuff of Big Cottonwood Canyon. Post–30 Ma: granite and quartzite, with lesser marble, phyllite, and schist	Quartzite	Chert and quartzite with volcanic rocks in stratigraphically higher parts	Quartzite, chert, chert-quartzite-pebble conglomerate
Clast sizes:	Intravolcanic: to 20 cm; post–30 Ma, to 1.5 m		30 cm	2 m
Rounding:	Intravolcanic: Paleozoic are well-rounded, tuff of Big Cottonwood Canyon subrounded to subangular. Post–30 Ma; granite is rounded, others angular to subangular		Well-rounded	Well-rounded
Relation to Miocene:	Angular unconformity with Jarbidge Rhyolite, ~25°	Angular unconformity with 16 Ma tuffs and sediments, ~20°	No Miocene?	No Miocene
Notes:	Eocene sediments continue at least 10 km east-northeast but are mostly covered by Jarbidge Rhyolite. Coats (1964) interpreted a narrow trough with steep walls because of the abrupt changes in thickness and rock type.		Conglomerate occurs discontinuously over wide area	
Sources (this study for all):	Coats (1964); Axelrod (1966a, 1966b); Rahl et al. (2002); McGrew and Foland (2004)	Wallace (2003a, 2003b)	Lovejoy (1959); Ketner (1998); Henry et al. (1999); Haynes (2003)	Moore et al. (1982); Ketner and Ross (1990); Brooks et al. (1995a, 1995b); Haynes (2003)

dacite lava domes and small-volume ash-flow tuffs, and the tuff of Big Cottonwood Canyon erupted sequentially but all indistinguishably at 40.0 Ma. Numerous small to moderate-sized intrusions, including a northeast-striking dike swarm, were emplaced between ca. 39.8 and 39.6 Ma, and the final igneous activity consisted of more dacite lavas and domes erupted at 39.5 Ma.

California Mountain

Eocene rocks occupy an ~5-km–wide, ≥800-m–deep, southeast- and east-tilted paleovalley at California Mountain in the Independence Mountains (Fig. 8; Muntean and Henry, 2006). A basal gravel is marked by a lag of well-rounded to subangular boulders of quartzite, chert, and chert-quartzite-pebble conglomerate up to 70 cm in diameter. Thin (≤5 m-thick) lenses of finely laminated shale (Fig. 9A) and micrite are interbedded with the gravel, and a thicker (~20 m) section of shale overlies gravel. The whole sequence is ~40 m thick. The plagioclase-biotite tuff dated at 41.9 ± 0.2 Ma (Hofstra, 1994) is the oldest volcanic rock. More gravel containing rounded boulders of quartzite and conglomerate to 1.5 m overlies this tuff in the center of the paleovalley. A thick sequence of locally derived, dacitic lavas and tuffs is dated at 40.4 ± 0.2 Ma (Hofstra, 1994). The tuff of Big Cottonwood Canyon is the youngest Eocene rock. Maximum thickness of the Eocene rocks is ~800 m, and units thin and wedge out against paleovalley walls. A 6-km–long, 500-m–deep swale in the crest of the Independence Mountains is just west of the east-tilted paleovalley and may indicate its westward continuation (Fig. 9B).

The Eocene rocks at California Mountain have undergone two episodes of faulting and tilting. All units with the possible exception of the tuff of Big Cottonwood Canyon were tilted ~20° to the southeast along a series of east-northeast–striking, mostly northwest-dipping faults (Fig. 8). The faults are spaced every 1–2 km and have displacements of 300–500 m. Faulting is younger than 40.4–40.0 Ma. Later (probably post-middle Miocene), the rocks were tilted eastward ~20° during formation of the present Independence Mountains. Scarps in Quaternary alluvial fans demonstrate that a similar style of faulting has continued to the present day (Muntean and Henry, 2006).

Double Mountain

The central paleovalley continues eastward to an unnamed, west-tilted range at Double Mountain, ~20 km east of California Mountain (Fig. 3). The Eocene section here is identical to that at California Mountain, with a basal sedimentary sequence overlain by the plagioclase-biotite tuff, andesite lava, and the tuff of Big Cottonwood Canyon. However, the basal sedimentary section is considerably thicker. Unpublished details of mapping by Coats (1987; Nevada Bureau of Mines and Geology information office files) show a basal conglomerate overlain by calcareous to silicified shale, oil shale, sandstone, limestone, and minor pyroclastic-fall tuffs. Finely laminated deposits showing soft-sediment deformation are common. The sedimentary rocks are so poorly exposed that even measuring an attitude was not possible, but based on an estimated dip of 25° and outcrop width, the section is ~300 m thick. The ash-flow tuffs are overlain unconformably by nearly flat-lying Miocene Jarbidge Rhyolite.

Oxley Peak

Two parts of what is probably a single paleovalley are separated by ~4 km near Oxley Peak in the southern Snake Mountains. The northern channel is ~1.7 km wide and has a 1-m–thick, oolitic limestone at the base. This is overlain by the plagioclase-biotite tuff and the tuff of Big Cottonwood Canyon, conglomerate or breccia containing clasts of the tuffs, and dacite lavas dated at 40.0 ± 0.2 Ma (Brooks et al., 1995a, 1995b; Thorman et al., 2003). Eocene(?) tuffaceous siltstone and shale overlie the dacite, apparently conformably but on an eroded surface; thickness varies greatly from ~1 to 100 m (Thorman et al., 2003). Jarbidge Rhyolite overlies the sedimentary rocks without angular discordance (Thorman et al., 2003).

The southern channel is only ~1 km wide. It has a basal pebble conglomerate containing well-rounded Paleozoic and volcanic clasts, including the plagioclase-biotite tuff, but the tuff is not exposed in the area examined by me. The overlying tuff of Big Cottonwood Canyon is discontinuously exposed and locally has been eroded and reworked into coarse, fluvial conglomerate containing angular blocks of the tuff up to 1 m in diameter. Some blocks are only partly broken away from outcrop, and none have undergone significant transport. Field relations between the tuff of Big Cottonwood Canyon and overlying conglomerate are well preserved and exposed because both were silicified.

Although Oxley Peak is ~70 km east of Double Mountain, the correlation of the tuff of Big Cottonwood Canyon is certain based on composition and age (Tables 1 and 2), and the sequence of the plagioclase-biotite tuff overlain by tuff of Big Cottonwood Canyon is identical to that at Double Mountain and California Mountain.

Nanny Creek

Nanny Creek is a key locality in understanding paleovalleys and their significance in northeastern Nevada because it has the most distal known outcrops of the tuff of Big Cottonwood Canyon and it best illustrates megabreccia consisting of reworked ash-flow tuff. Brooks et al. (1995a, 1995b) produced a sketch geologic map of the Nanny Creek paleovalley and recognized several of the key relationships. They identified a coarse, basal conglomerate containing some clasts that could not be locally derived, overlain by two ash-flow tuffs and a sequence of andesite flows and flow breccias, all these overlain by a heterogeneous unit containing blocks of ash-flow tuff. C.H. Thorman (2003, personal commun.) recognized that the rocks occupied a paleovalley that could be as much as 1.6 km deep based on the dip of the volcanic rocks and the distribution of surrounding Paleozoic rocks. This study (Fig. 10) expands upon their interpretations.

The exposed width of the Nanny Creek paleovalley is ~6 km, but it is faulted against Paleozoic rocks on the south side, and the original width is thus unknown. The 20- to 30-m–thick basal conglomerate consists of a lag of rounded boulders commonly up to 1.5 m in diameter with one 6 m across (Figs. 9C and 9D). Most clasts, including the largest ones, are chert ± quartzite-pebble conglomerate similar to adjacent Paleozoic bedrock. However, numerous clasts are present of rocks that do not crop out locally, including coarse granite up to 1.7 m in diameter, andesite, and silicified, finely laminated, lake-bed sediments (Brooks et al., 1995a).

Conglomerate is overlain by the plagioclase-biotite tuff and tuff of Big Cottonwood Canyon (Tables 1 and 2; Fig. 10; Brooks et al., 1995a, 1995b; Palmer and MacDonald, 2002). The thickness of the tuffs is difficult to estimate. Foliation in the tuffs dips 30°–60° eastward, but dips are probably partly primary, resulting from compaction of the tuffs against paleovalley walls (Henry et al., 2004). Using the 30°–60° range, the plagioclase-biotite tuff and tuff of Big Cottonwood Canyon could be 60–140 and 90–190 m thick, respectively. Both values are significantly greater than the thicknesses of 25 and 55 m, respectively, reported by Brooks et al. (1995a). The tuff of Big Cottonwood Canyon here is 150 km from its source in the Tuscarora volcanic field, yet it is at least 55 m thick and restricted to a narrow paleovalley within which it wedges out against the sides (north of the area of Fig. 10; Brooks et al., 1995a). Andesite and andesite flow breccia dated at 39.5 ± 1.0 Ma (Brooks et al., 1995a, 1995b) overlie the tuff of Big Cottonwood Canyon.

The uppermost unit in Nanny Creek (Fig. 10) is a breccia composed of blocks, mostly of silicified tuff of Big Cottonwood Canyon with lesser plagioclase-biotite tuff and andesite (Fig. 9E). Paleozoic clasts are rare. Clasts of tuff of

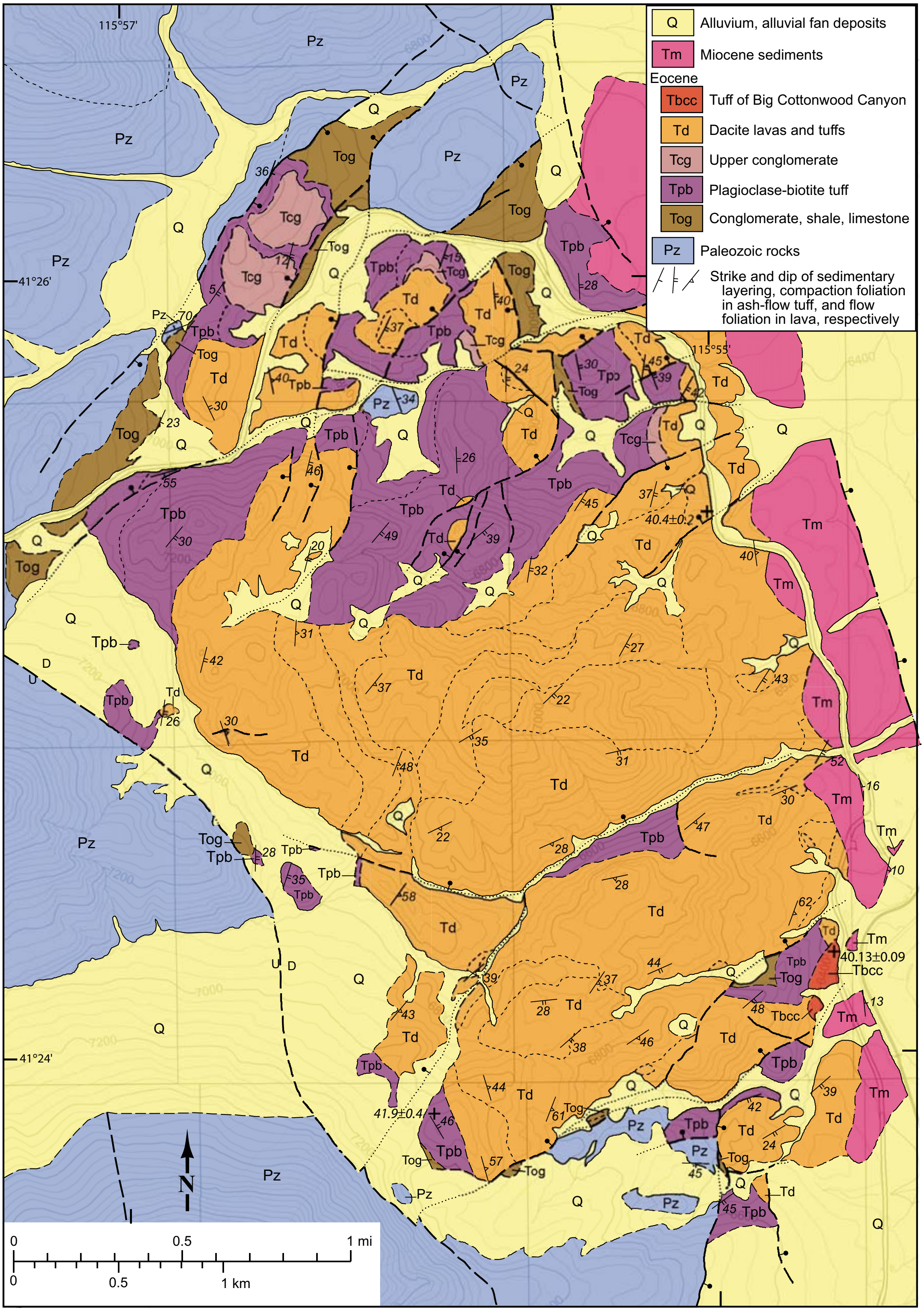

Figure 8. Simplified geologic map of the Eocene paleovalley near California Mountain, east side of the Independence Mountains (Muntean and Henry, 2006) with $^{40}Ar/^{39}Ar$ dates from this study (regular type) and Hofstra (1994; italics). The paleovalley, which was 5 km wide and at least 800 m deep, was filled by conglomerate containing rounded boulders to 1.5 m diameter, shale and limestone, the plagioclase-biotite tuff and tuff of Big Cottonwood Canyon, and a thick dacite sequence between the tuffs. The rocks were tilted gently southeastward after 40.0 Ma by several small-displacement, northeast-striking, mostly down-to-the-northwest normal faults. Dashed lines within unit Td are contacts between individual lavas and tuffs.

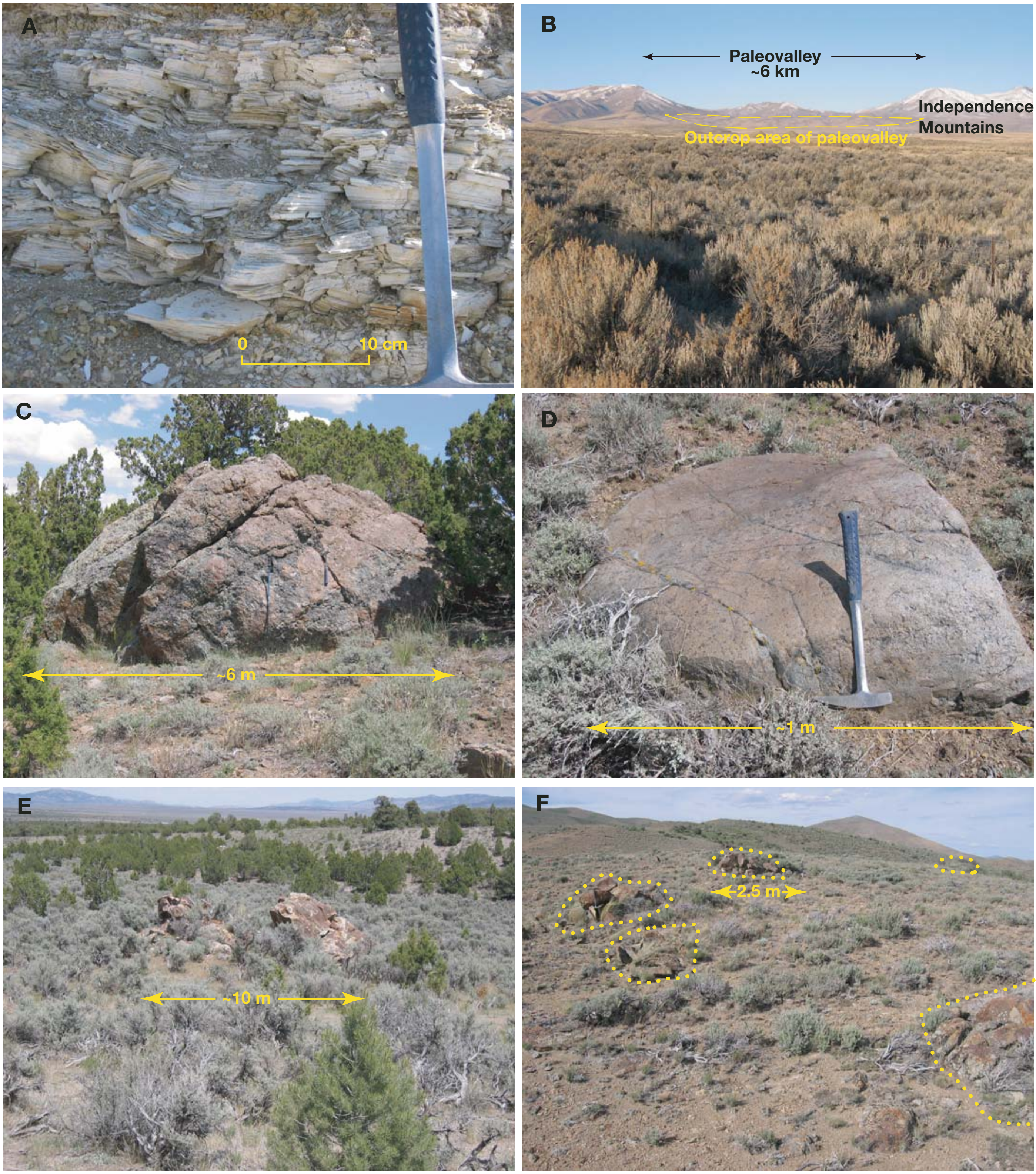

Figure 9. (A) Finely laminated, "paper" shale in the California Mountain paleovalley (Fig. 8). (B) Photo looking west at an ~6-km–wide, 500-m–deep swale in the Independence Mountains west of the exposed paleovalley sequence near California Mountain (Fig. 8). The swale is probably the western continuation of the paleovalley. (C) Rounded, 6-m–wide boulder of chert-pebble conglomerate in basal Tertiary conglomerate of the Nanny Creek paleovalley (Fig. 10). (D) Well-rounded, 1-m–wide boulder of quartzite in the Nanny Creek paleovalley (Fig. 10). (E) Isolated block of tuff of Big Cottonwood Canyon in megabreccia in Nanny Creek paleovalley (Fig. 10). (F) Scattered blocks of tuff of Big Cottonwood Canyon in megabrecccia in Taylor Canyon paleovalley (Fig. 3). Compaction foliation is oriented differently in each block, showing that each is a separate block.

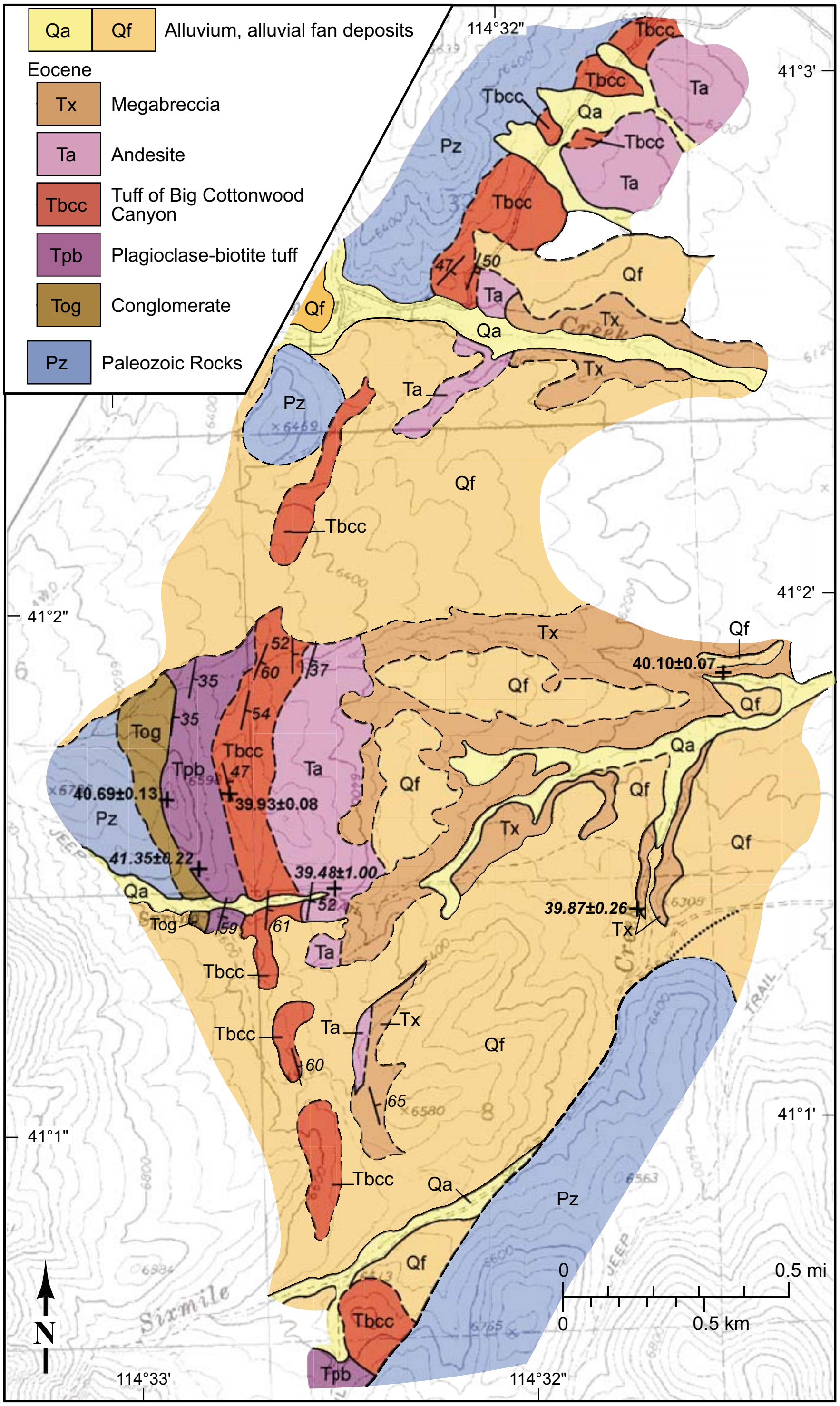

Figure 10. Geologic map of the paleovalley at Nanny Creek, east side of the Pequop Mountains (Brooks et al., 1995b; this study) with $^{40}Ar/^{39}Ar$ dates from this study (regular type) and Brooks et al. (1995a, 1995b; italics). The paleovalley, which was at least 6 km wide and possibly 1.6 km deep, has basal conglomerate containing rounded boulders up to 6 m in diameter, overlain by the plagioclase-biotite tuff and tuff of Big Cottonwood Canyon, andesite lavas, and a thick megabreccia consisting of angular blocks mostly of the tuff of Big Cottonwood Canyon. That megabreccia consists of reworked blocks is confirmed by petrographic, chemical, $^{40}Ar/^{39}Ar$, and paleomagnetic data (this study; M.R. Hudson, in Brooks et al., 1995a; Palmer and MacDonald, 2002).

Big Cottonwood Canyon are up to 12 m across, plagioclase-biotite tuff is up to 8 m long, and andesite is as much as 2 m across. Blocks range from angular to rounded, and many are internally broken but not disaggregated. Rarely exposed matrix consists of coarse sand and variably rounded pebbles and cobbles. Brooks et al. (1995a) recognized the hummocky character of this unit and the possibility that the hummocks might be landslide blocks. Chemical analysis and dating of one clast confirm that it is tuff of Big Cottonwood Canyon (Sample H05-38, Tables 1 and 2; Figs. 4 and 10). Furthermore, highly discordant compaction foliations and scattered magnetization directions from several breccia clasts (M.R. Hudson, in Brooks et al., 1995a; Palmer and MacDonald, 2002, their sites W02, W15, and W18) confirm that these blocks are not in place.

Windermere Hills

The Windermere Hills are also a key area to illustrate paleovalley relationships. Mueller (1992, 1993) mapped the area, and Mueller et al. (1999) used the mapping and $^{40}Ar/^{39}Ar$ geochronology to conclude that a thick sequence of sedimentary rocks accumulated in a half graben resulting from major extension between ca. 39 and 35 Ma. These sediments overlie a pre-extensional volcanic sequence that includes the regional ash-flow tuffs of this study. I reinterpret the area to be in an ~10-km–wide part of the central paleovalley (Figs. 3 and 11) that was dammed by minor extension (see Discussion).

The volcanic and volcaniclastic strata are similar across the width of the paleovalley but, as noted by Mueller (1992), show abrupt changes in rock type and thickness both along and across strike. A basal conglomerate and andesite are exposed only in the south and are overlain by the plagioclase-biotite tuff and tuff of Big Cottonwood Canyon. According to Mueller et al. (1999), the oldest synextensional deposits are pebble-cobble conglomerate and sandstone that overlie the tuffs (Tdc in Fig. 11). Clasts in the conglomerate are well rounded and mostly consist of the underlying volcanic rocks with some Paleozoic quartzite. Mueller (1992) found algal limestone and laminated siltstone in the conglomerate unit and imbricated clasts that indicate deposition in an east-flowing stream. A thick sequence of tuffaceous sandstone, siltstone, and shale overlies the conglomerate. Mueller et al. (1999) obtained dates of 35.01 ± 0.36 and 39.20 ± 0.50 Ma on sanidine and biotite, respectively, from a pyroclastic-fall tuff in the upper part of this sequence and favored the less disturbed sanidine age. They interpreted the sedimentary rocks to have been deposited in lacustrine fan-delta and lacustrine environments in a half graben.

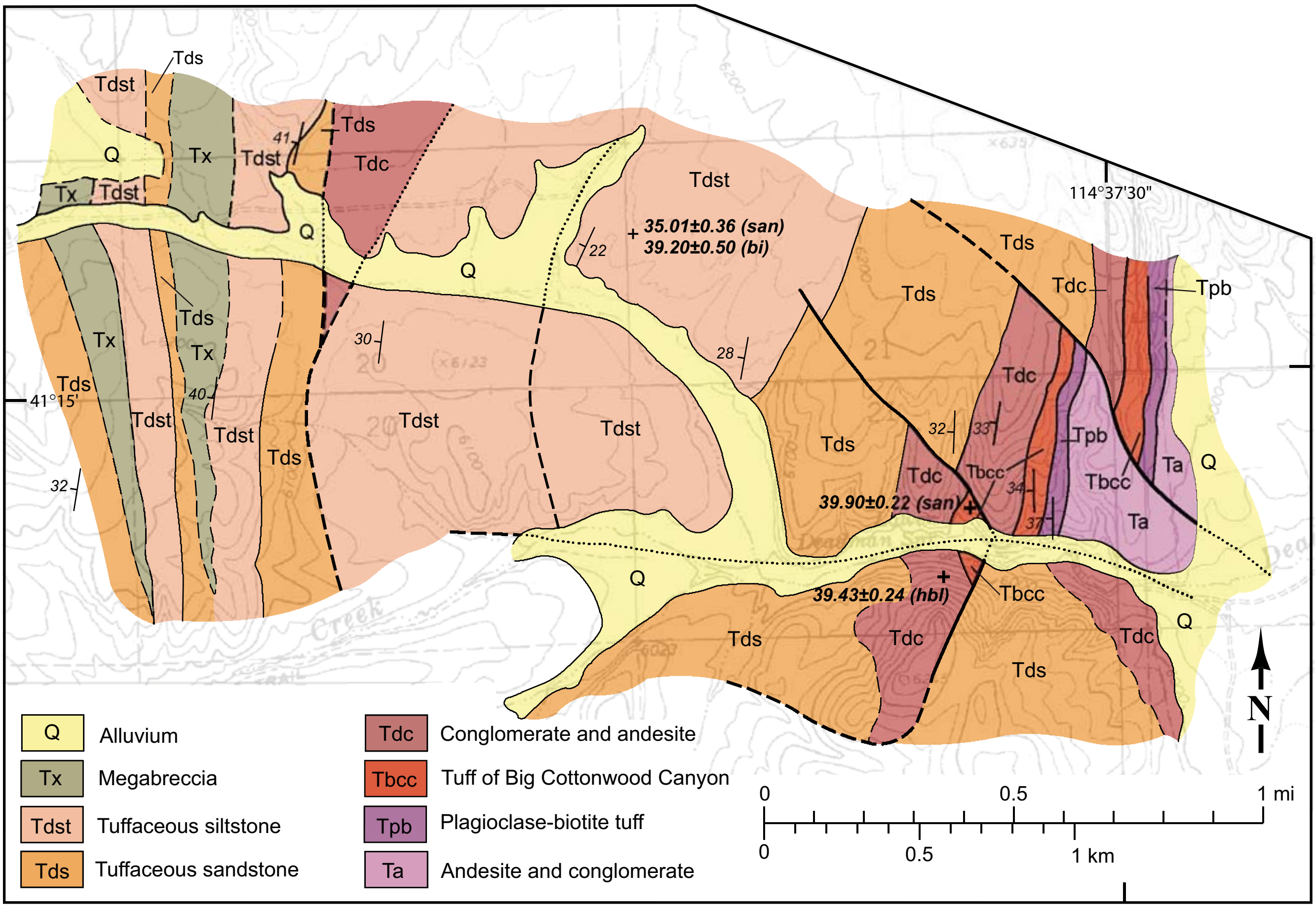

Figure 11. Geologic map of part of the Eocene paleovalley in the Windermere Hills (modified from Mueller, 1993, with $^{40}Ar/^{39}Ar$ dates from Mueller et al., 1999). Imbricated clasts in cobble conglomerate (Tdc) indicate eastward flow (Mueller, 1992). Mueller (1993) and Mueller et al. (1999) interpreted major normal faults repeating the ash-flow tuffs along the west side of this map. I reinterpret these as megabreccia (Tx) composed of blocks of the plagioclase-biotite tuff and tuff of Big Cottonwood Canyon. That they are blocks is confirmed by petrography, chemical analysis, discordant compaction foliations, and by scattered magnetization directions (Palmer and MacDonald, 2002). Therefore, their stratigraphic position and the amount of repetition by faulting are uncertain.

Middle Miocene, tuffaceous sedimentary rocks overlie the Eocene rocks in the paleovalley but rest directly on Paleozoic rocks just to the south in the northwestern Pequop Mountains (Mueller, 1993). Both Eocene and Miocene rocks are similarly tilted 30° to 40° west (Fig. 11; Mueller, 1993; Mueller et al., 1999).

Mueller et al. (1999) interpreted exposures in the western part of Figure 11 as intact ash-flow tuffs (plagioclase-biotite tuff and tuff of Big Cottonwood Canyon of this study) that were repeated by several closely spaced, east-dipping normal faults. However, where examined by the author, both in the area of Figure 11 and farther north in the map area of Mueller (1993), these outcrops are actually lenses of megabreccia (unit Tx) consisting of variably disaggregated blocks of the tuffs. The blocks form hummocks in a matrix of variably coarse clastic deposits, similar to those in the upper unit at Nanny Creek. As at Nanny Creek, the blocks have highly variable and discordant compaction foliations and scattered magnetization directions (Palmer and MacDonald, 2002, their sites W06 and W07) that confirm that they are not in place. Conglomerates composed of well-rounded pebbles of quartzite and chert are interbedded with the megabreccias, but megabreccias contain no clasts of Paleozoic rocks. The age and stratigraphic position of these breccias and the amount of repetition by faulting are therefore uncertain, but no fault repetition is required. The angular conformity of all Eocene rocks and at least the lower part of the middle Miocene rocks indicates negligible tilting during their deposition.

The Windermere Hills lie north of, and parallel to, the paleovalley at Nanny Creek (Fig. 3). It seems unlikely that both locations are part of a single, east-trending paleovalley. They could, however, be tributaries of the same paleovalley that joined downstream, which I conclude was to the east (see Discussion).

North Paleovalley

The north paleovalley has been investigated only in reconnaissance, and published maps commonly combined Tertiary volcanic rocks into one or two units interpreted to be Miocene (Decker, 1962; Coash, 1967). Therefore, interpretations about the geology are more tentative than for the central paleovalley. Nevertheless, the presence of a distinctive ash-flow tuff and sedimentary sequence filling narrow valleys cut into Paleozoic rocks over a 60 km length from Bull Run to the Copper Basin support the paleovalley interpretation. Miocene or younger rocks cover older rocks west of the Bull Run area.

North Bull Run Area

An area of ~30 km^2 at the northeast edge of the Bull Run basin is underlain by ash-flow tuff (Decker, 1962). A section across the northern part of this area appears to be a single ash-flow tuff that is petrographically similar to the 45 Ma tuff at Cornwall Mountain and Copper Basin (Fig. 6). The tuff dips uniformly 20°–30° westward and is repeated by one, north-striking, east-dipping fault. Based on the dip and outcrop width, tuff in each fault block is 600–800 m thick, and the base is not exposed. Total displacement on the fault is uncertain but does not appear to be more than 200 m. Therefore, the tuff may total more than 1 km thick. The easternmost part of the tuff contains several lenses of andesite boulders. Although Decker (1962) shows this tuff in depositional contact with Paleozoic rocks to the west, this contact must be a fault because the tuff dips uniformly into the contact.

In describing the overall area, Decker (1962) stated that tuff fills a 4 × 8 km, northeast-trending "ancient topographic sag" and also noted steep irregular attitudes in the tuff, which he thought could be primary. Decker also noted that an exploration well ~3 km west of a down-to-the-west fault bounding the southwestern edge of ash-flow tuff encountered only Eocene sedimentary rocks of the Bull Run basin above Paleozoic rocks (Fig. 3A).

I interpret that most of this tuff is the 45 Ma ash-flow tuff based on its similar phenocryst assemblage and composition (Fig. 5), as well as its location only 30 km west of dated 45 Ma tuff at Cornwall Mountain (see discussion of Cornwall Mountain and Copper Basin). Also, Axelrod (1966a) reported a biotite K-Ar age of 44.2 ± 2.1 Ma from tuff near the southwestern edge of outcrops. Although he did not describe the tuff and the age is not diagnostic, it is one of the oldest K-Ar ages in the region.

The tuff fills either a paleovalley or possibly a caldera. The paleovalley interpretation is consistent with the northeast-elongate "topographic sag" and with possible primary dips. The caldera interpretation is consistent with the apparent great thickness, which seems too large for a paleovalley, and with the andesite breccia lenses, which could be mesobreccia. Abrupt termination of the tuff to the southwest could be consistent with either interpretation.

Cornwall Mountain

The Tertiary section at Cornwall Mountain occupies a paleovalley that has been tilted moderately westward and folded into a west-plunging syncline (Fig. 12; Coash, 1967). Based on an average 40°–50° west dip and width of the outcrop belt, the overall section is ~1.5 km thick. However, the section is repeated by at least one major, northwest-striking fault.

A discontinuous basal conglomerate contains well-rounded boulders up to 4 × 6 m, mostly of quartzite, chert-pebble conglomerate, limestone, and granite. Overlying the conglomerate is a thick, ash-flow tuff that is zoned from an ~30-m–thick lower part with relatively abundant sanidine and quartz phenocrysts to an upper part ≥500 m thick with relatively more abundant plagioclase and mafic phenocrysts (Figs. 5 and 6). A sample from the lower part has a sanidine $^{40}Ar/^{39}Ar$ age of 45.10 ± 0.09 Ma (Table 1).

The ash-flow tuff is overlain by a thick sequence of sedimentary rocks consisting of two distinct rock types (Fig. 12). Fine-grained, commonly tuffaceous siltstone, shale, sandstone, and minor pebble conglomerate make up most of the section. Interbedded with these fine deposits are numerous breccia lenses composed of ash-flow tuff indistinguishable from the upper part of the underlying tuff. Clasts are up to 4 m in diameter and are subrounded to subangular. Paleozoic clasts are absent; therefore, the footwall of a fault that exposed basement rock was not the source.

Copper Basin

Copper Basin is an ~12 km^2 area of Eocene ash-flow tuff and tuffaceous sedimentary rocks in the hanging wall of the Copper Basin fault, a major, east-dipping normal fault (Coats, 1964; Axelrod, 1966a, 1966b; Rahl et al., 2002). Eocene strata are divided into the Dead Horse Formation and the overlying Meadow Fork Formation (Coats, 1964; Rahl et al., 2002). The Dead Horse Formation is as much as 1600 m thick and consists of ash-flow tuff overlain by tuffaceous, lacustrine sedimentary rocks. The Meadow Fork Formation is as much as 400 m thick and consists of coarse conglomerate containing clasts of quartzite, marble, phyllite, and granitic rocks up to 1 m in diameter. The entire section is conformable but overlain with angular unconformity by Steens-type, coarsely porphyritic basalt, dated at 16.6 ± 0.4 Ma, and Jarbidge Rhyolite (Coats, 1964; Rahl et al., 2002). Shale in the Dead Horse Formation not far above the ash-flow tuffs contains the fossil plant assemblage that is variably interpreted to have lived at 1.1–2.8 km (Axelrod, 1966b; Chase et al., 1998; Wolfe et al., 1998).

Three ash-flow tuffs are present in the lower part of the Dead Horse Formation. They consist of the 45 Ma tuff, a plagioclase-biotite-hornblende tuff, and the tuff of Big Cottonwood Canyon (Table 1). Rahl et al. (2002) obtained a biotite $^{40}Ar/^{39}Ar$ age of 41.6 ± 0.2 Ma on what may be the middle, plagioclase-biotite tuff, which I tentatively correlate with the regional plagioclase-biotite tuff. A channel cut into the tuff of Big Cottonwood Canyon contains pebbles up to 4 cm of Paleozoic quartzite and chert and Tertiary volcanic rocks, mostly the tuff of Big Cottonwood Canyon. Paleozoic clasts are mostly well rounded, whereas the Tertiary clasts are subrounded to subangular. A pyroclastic-fall tuff from the uppermost part of the Dead Horse Formation gave an age of 37.6 ± 0.4 Ma (Rahl et al., 2002). The base of the Tertiary section is not exposed.

Coats (1964) interpreted the Meadow Fork Formation to have accumulated in a narrow, northeast-trending, fault-bounded basin because of its coarse clasts and the fact that both it and the Dead Horse Formation wedge out abruptly. Rahl et al. (2002) interpreted that extension and displacement on the Copper Basin fault began during deposition of the upper part of the Dead Horse Formation. McGrew and Foland (2004) revised this interpretation to one in which most extension occurred in the Oligocene during deposition of the Meadow Fork Formation, for which they obtained a 30 Ma date. I suggest that the northeast-trending, fault-bounded basin was initially a paleovalley that was not bounded by faults but connected with the Cornwall Pass and Bull Run area to the west. Based on the 1600-m thickness of the Dead Horse Formation (Coats, 1964; Rahl et al., 2002), the paleovalley may have been at least 1.6 km deep. Extensional faults in the Eocene and Miocene partly to eventually totally disrupted the paleovalley. This interpretation is developed more thoroughly in the Discussion.

South Paleovalley

Three segments of a southern paleovalley were investigated. Detailed maps are available for Willow Creek and Coal Mine Canyon (Wallace, 2003a; Moore et al., 1982; Ketner and Ross, 1990). Ash-flow tuffs are widely distributed around Taylor Canyon in the center of this area, and Lone Mountain was a major volcanic center (Lovejoy, 1959; Ketner, 1998; Henry and Ressel, 2000b). How the segments connect is unclear (Fig. 3A). The paleovalley at Willow Creek may have flowed westward. Taylor Canyon may have flowed eastward to Coal Mine Canyon, which probably flowed into the Elko basin, or southward directly into the Elko basin.

Willow Creek Reservoir

The only paleovalley segment exposed west of the Tuscarora volcanic field is near Willow Creek Reservoir, where the tuff of Nelson Creek, the tuff of Big Cottonwood Canyon, and 37 Ma

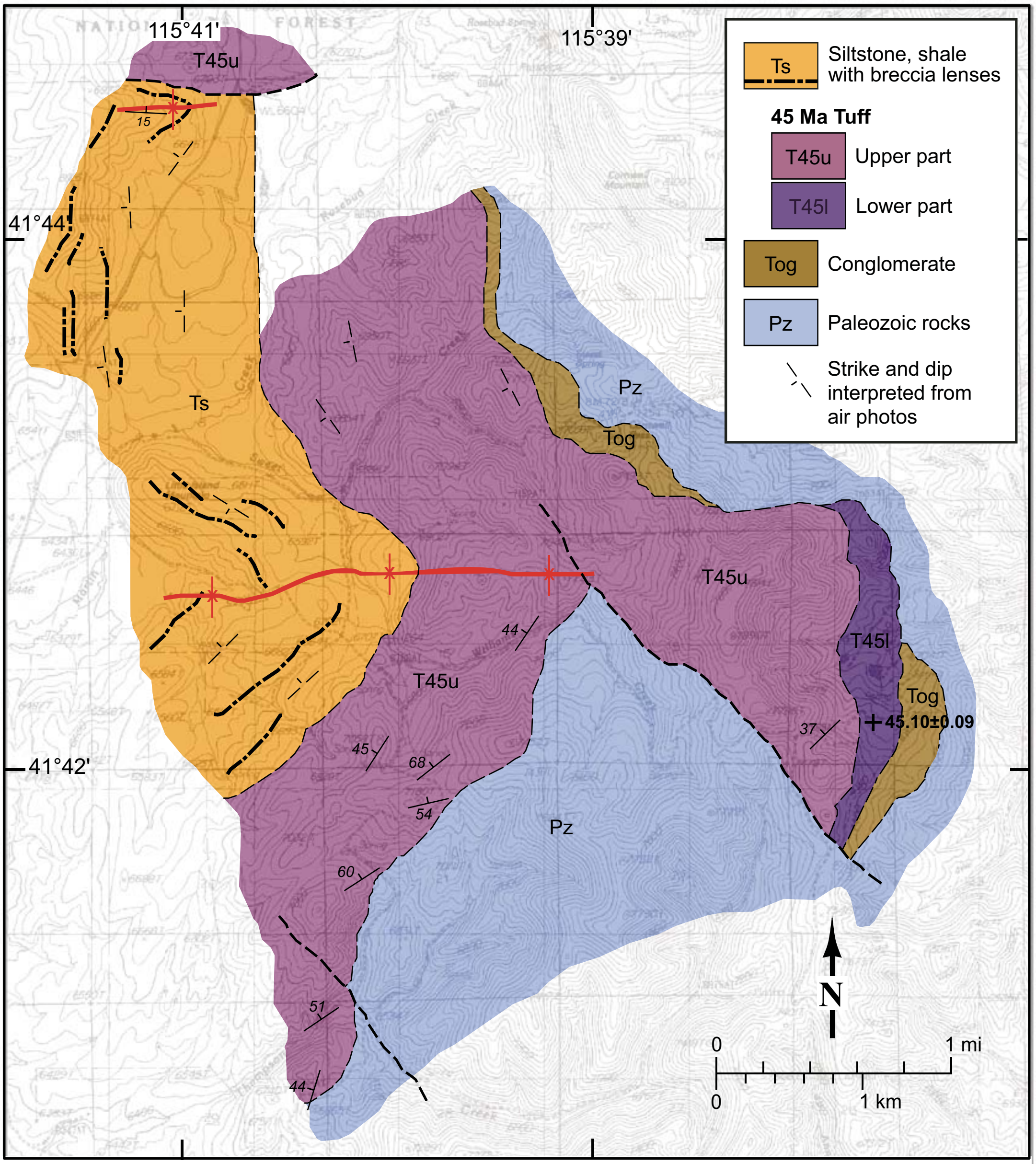

Figure 12. Simplified geologic map of the Eocene paleovalley near Cornwall Mountain (modified from Coash, 1967). Basal conglomerate with rounded clasts up to 4 × 6 m is overlain by the zoned, 45 Ma tuff ($^{40}Ar/^{39}Ar$ date from this study). An upper, probably lacustrine sequence of tuffaceous siltstone and shale has numerous interbedded breccia lenses containing clasts up to 4 m diameter of the upper, less silicic part of the 45 Ma tuff.

andesite lavas, with an aggregate thickness of ~450 m, are preserved in a west-trending belt (Fig. 3; Wallace, 2003a, 2003b). A red paleosol containing angular fragments of quartzite is developed on top of the Paleozoic basement. Less than 5 km to the south, middle Miocene rocks rest directly on Paleozoic rocks (Wallace, 2003a, 2003b). Eocene rocks are absent for ~50 km to the southeast, where 38–36 Ma lavas overlie Paleozoic rocks in the Emigrant Pass volcanic field (Henry and Faulds, 1999), and 60 km to the south, where the 34.2 Ma tuff of Cove Mine overlies Paleozoic rocks in the northern Shoshone Range (Fig. 3) (John et al., 2008). Pre-middle Miocene rocks are covered for a considerable distance to the west (Fig. 2).

Taylor Canyon

Detailed mapping is available for only the northwestern edge of the Taylor Canyon area (Henry et al., 1999). Other information comes from published data and from examination of rocks at scattered locations. A locally thick but discontinuous, cobble-to-boulder conglomerate rests on Paleozoic rocks and is overlain by the tuff of Nelson Creek, the tuff of Big Cottonwood Canyon, and andesite lavas. These rocks occupy a broad low in the present Independence Mountains and eastern flank of the Tuscarora Mountains. Elevations drop from 2761 m 5 km north of Taylor Canyon to 1950 m in the canyon, then rise to 2676 m at Lone Mountain 10 km south of the canyon. Lone Mountain is part of a ca. 39 Ma igneous dome and volcanic center (Lovejoy, 1959; Henry and Ressel, 2000b); therefore, presumably it was not a topographic high during earliest Eocene sedimentation.

Conglomerate is discontinuously exposed along the Paleozoic-Tertiary contact over a wide area, locally reaching 140 m thick (Lovejoy, 1959; Coats, 1987; Ketner, 1998; Haynes, 2003). It contains well-rounded clasts, mostly of chert and quartzite with volcanic rocks in stratigraphically higher parts. Fresh-water limestone is present southwest of Lone Mountain (Ketner, 1998). Pyroclastic-fall tuffs near the base of the conglomerate in two locations have U-Pb zircon ages of 41.9 ± 0.3 and 40.8 ± 0.6 Ma (Haynes, 2003), which suggest correlation with the plagioclase-biotite tuff (Table 1). Conglomerate is overlain by the tuffs of Nelson Creek and Big Cottonwood Canyon and by andesite lavas dated at 39.63 ± 1.58 Ma (Haynes, 2003). Megabreccia composed of variably oriented blocks of tuff of Big Cottonwood Canyon is present in the northern part of the paleovalley (Fig. 9F).

The wide distribution of conglomerate and ash-flow tuff in this area suggests a broad paleovalley or possibly two separate valleys. Notably, neither major ash-flow tuff is found at Coal Mine Canyon or in the Elko basin just 30 km to the southeast (Moore et al., 1982; Haynes, 2003; this study).

Coal Mine Canyon

A paleovalley cut into Mississippian Chainman Shale at Coal Mine Canyon along the southeast flank of the Adobe Range contains some of the most thoroughly studied Eocene sedimentary rocks of the region (Fig. 13; Moore et al., 1982; Ketner and Ross, 1990; Haynes, 2003). A basal conglomerate and sandstone unit contains well-rounded boulders to cobbles of quartzite, chert, and chert-quartzite-pebble conglomerate up to 1.1 m in diameter. Moore et al. (1982) estimated the conglomerate to be 60–80 m thick and interpreted it to have been deposited in a fluvial or alluvial-fan environment. Overlying deposits consist of claystone, siltstone, shale (including oil shale), limestone, sandstone, and minor pebble conglomerate. These deposits are up to ~350 m thick and were deposited in an open lacustrine to fluvial environment (Moore et al., 1982). The ca. 41 Ma tuff of Coal Mine Canyon, a poorly to densely welded, dacite ash-flow tuff, overlies the sedimentary rocks (Fig. 13; Table 2). Although the ash-flow tuff–Elko Formation contact has been interpreted as an angular unconformity (Wingate, 1983; Brooks et al., 1995a), the tuff dips the same as the underlying Elko Formation. Total relief on the paleovalley is at least 400 m, because all the sedimentary rocks wedge out against Paleozoic basement (Fig. 13).

DISCUSSION

Distribution of Ash-flow Tuffs and Sedimentary Deposits in Paleovalleys

Ash-flow tuffs are widely distributed around Tuscarora, Taylor Canyon, and north of Bull Run basin, areas of known or suspected calderas (Fig. 2). However, just 20 km away from these areas, the tuffs and overlying and underlying sedimentary rocks are restricted to narrow belts within Paleozoic bedrock. For the following reasons, I interpret these belts to be segments of three separate, approximately east-west paleovalleys (Fig. 3).

(1) Most locations preserve a paleovalley geometry where the tuffs and sedimentary rocks crop out within valleys incised into Paleozoic rocks. Both tuffs and sedimentary rocks wedge out along strike against the Paleozoic rocks. This geometry is well shown at California Mountain, Double Mountain, and Nanny Creek in the central paleovalley, at Cornwall Mountain and more poorly at Copper Basin in the north paleovalley, and at Coal Mine Canyon in the south paleovalley. Preserved thicknesses of ash-flow tuff and sedimentary deposits, which would pond in low areas, indicate that some paleovalleys were as much as 1.6 km deep (Table 3). The paleovalley is not as well exposed at Willow Creek, but surface and subsurface data demonstrate its presence (Wallace, 2003a, 2003b). The paleovalley at Taylor Canyon is also less clear, because of considerable faulting and because tuffs are so widely distributed in this near source location.

(2) The distinctive ash-flow tuff stratigraphy within each paleovalley and the restriction of distal tuffs to paleovalleys are further evidence for continuous paleovalleys. Topography generally controls flow and distribution of ash-flow tuffs (Wilson and Walker, 1985; Baer et al., 1997; Ort et al., 2003). With the exception of the tuff of Big Cottonwood Canyon, all tuffs are restricted to one or two adjacent paleovalleys. The 45 Ma tuff is restricted to the north paleovalley. Whether or not its source was near Bull Run basin, the tuff did not reach the central paleovalley. The tuff of Coal Mine Canyon is only present at Coal Mine Canyon and in the Windermere Hills. The tuff of Nelson Creek is only exposed at Taylor Canyon and Willow Creek Reservoir. The tuff of Big Cottonwood Canyon is more widely distributed but is still mostly restricted to the central paleovalley. Its source caldera was well located to send ash flows into all three paleovalleys. Whether a single or multiple unit, the plagioclase-biotite tuff is present only in the northern and central paleovalleys.

(3) The thickness and dense welding of ash-flow tuffs and their confinement to narrow outcrop belts far from their sources preclude the region having overall low relief—for example, a low-relief plateau without paleovalleys or a large contiguous basin, or having been broken into a series of approximately north-striking basins and ranges similar to the present topography. The tuff of Big Cottonwood Canyon at Nanny Creek is 150 km from its source caldera at Tuscarora and is restricted to a narrow wedge within the paleovalley. The tuff's thickness and dense welding suggest that it traveled significantly farther. Even if Miocene and younger extension totaled an unrealistically large 50%, the Eocene distance was at least 100 km. Tuffs flowing in paleovalleys can travel farther than can unconfined tuffs, because they are channelized and therefore disperse less and because they interact less with air and do not cool as rapidly (Bursik and Woods, 1996).

Ash-flow tuffs erupted onto a low-relief surface would have spread radially and would not show the linear distribution that they do in northeastern Nevada. Radial spreading means tuffs disperse rapidly and therefore cannot travel as far as channelized tuffs (Cas and Wright, 1987). If northeastern Nevada had low relief in the Eocene, individual tuffs would have been distributed much more radially and therefore would have overlapped each other more commonly.

If erupted into a lake, these tuffs would have formed water-laid tuffs similar to the tuffs of the Elko basin (Solomon et al., 1979; Haynes, 2003) instead of thick, densely welded deposits. Some of the Elko basin tuffs could be pyroclastic-fall deposits equivalent to the regional ash-flow tuffs of this study.

Tuffs erupted into basin-and-range topography similar to present-day topography also would have been channelized, but into wide, north-trending paleovalleys, and would not have been able to reach distant locations to the east. Although very low density pyroclastic flows can surmount major ridges far from their source, such flows are so dilute that they do not generate thick, densely welded deposits (Woods et al., 1998).

Characteristics and Origin of Sedimentary Deposits

Three general types of sedimentary deposits are preserved in the paleovalley segments: coarse basal, mostly pre-volcanic conglomerate; fine-grained lacustrine shale, siltstone, and limestone; and coarse megabreccia deposits composed mostly of reworked ash-flow tuff.

Basal Conglomerate

Coarse, basal conglomerates are present in almost all paleovalley segments. Clasts are

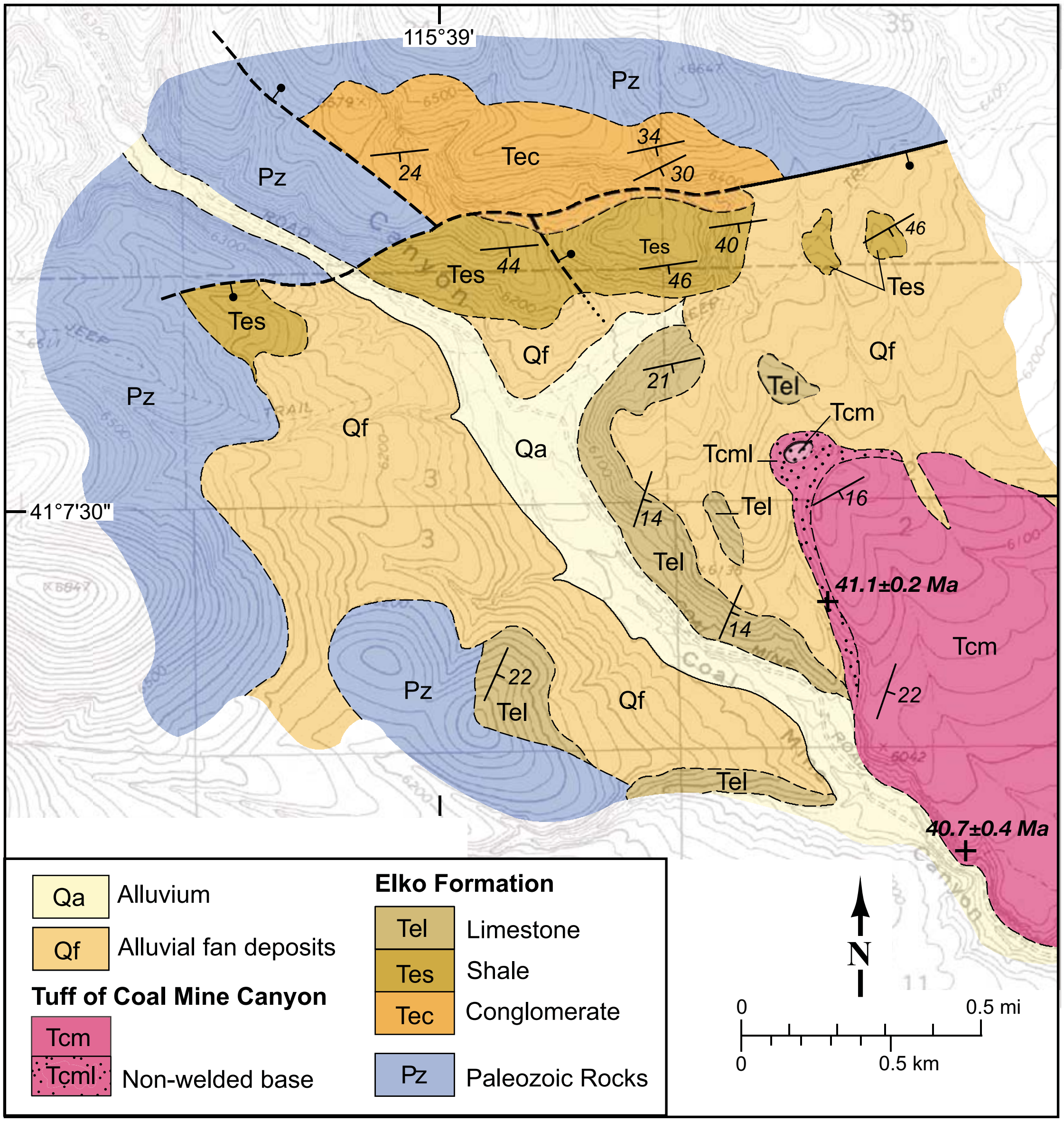

Figure 13. Geologic map of the paleovalley at Coal Mine Canyon, southeast flank of the Adobe Range (modified from Moore et al., 1982). The sedimentary sequence consists of fluvial, boulder-to-cobble conglomerate overlain by lacustrine shale, claystone, and limestone. The tuff of Coal Mine Canyon is dated at 40.7 ± 0.4 Ma ($^{40}Ar/^{39}Ar$, hornblende; Brooks et al., 1995a, 1995b) and 41.1 ± 0.2 Ma (U-Pb, zircon; Haynes, 2003).

almost entirely of Paleozoic rocks, particularly resistant quartzite and chert-quartzite pebble conglomerate. They are up to 6 m in diameter but are notably rounded (Figs. 9C and 9D; Table 3). The rounding of such resistant rocks indicates the clasts underwent significant transport and were not derived from nearby fault scarps, further supporting significant transport in a through-flowing river system. Transport of such large clasts indicates that the fluvial systems were high energy, through high gradient and/or high discharge; both are possible. The Eocene headwaters may have been at elevations of several kilometers (Coney and Harms, 1984; Dilek and Moores, 1999; DeCelles, 2004); therefore, high gradient is possible. The Eocene was notably warm and humid, with enhanced runoff, so high discharge is also possible (Zachos et al., 2001; Kelly et al., 2005). It is possible that coarse clasts were being continually transported through paleovalleys but were only preserved as deposits after being blanketed by welded ash-flow tuffs, which protected them from subsequent erosion. Similar coarse conglomerates, commonly of resistant Proterozoic quartzite up to 3 m in diameter, occupy two middle Eocene paleovalleys that drained eastward from the Sevier fold-and-thrust belt in Idaho and Montana (Janecke et al., 2000).

Lacustrine Deposits

Lacustrine shale and oil shale, other fine-grained clastic rocks, and limestone are common but not ubiquitous constituents of the sedimentary sequence (Figs. 3 and 9A; Table 3). Their stratigraphic position relative to dated rocks indicates two episodes of lacustrine deposition. The oldest lacustrine deposits may be in the Elko basin, where a 46 Ma, water-laid tuff is interbedded with basal conglomerate; however, the lower part of lacustrine deposits there is undated and only known to be older than ca. 39 Ma (Solomon et al., 1979; Haynes et al., 2002; Haynes, 2003). Lacustrine deposits are older than 41 or 42 Ma at California Mountain, Double Mountain, Coal Mine Canyon, and probably at Taylor Canyon. They are between 40 and 37.6 Ma in Copper Basin and younger than 39.4 Ma, possibly as young as ca. 35 Ma, in the Windermere Hills. The deposits at Cornwall Mountain are younger than 45 Ma and possibly the same age as the Copper Basin deposits. The bimodal age and the absence of lacustrine deposits at Nanny and Willow Creeks indicate that a single basin could not have covered the entire region. Reworking of clasts of silicified lakebeds into the basal conglomerate at Nanny Creek is further evidence that the lacustrine basins were along paleovalleys.

Either faulting or nonextensional processes could have created lacustrine basins along the paleovalleys, but Eocene faulting was likely responsible for formation of some basins. Eocene faults demonstrably contemporaneous with lacustrine deposits have not been identified, but the post–40 Ma, east-northeast–striking, down-to-the-northwest faults at California Mountain are probably representative (Fig. 8). Other evidence for active Eocene faults with this orientation consists of the southeast tilt of Elko Formation at Emigrant Pass and near Elko (Henry et al., 2001; Haynes, 2003; Cline et al., 2005; Hickey et al., 2005), west-northwest–trending, stretching lineations of the Ruby Mountains (although the lineations also affect 29 Ma rocks and therefore may have developed over an extended period; Snoke et al., 1997; Howard, 2003), and the northeast strike of 39.8–39.6 Ma dikes in the Tuscarora volcanic field (Fig. 7). Faults like those at California Mountain would have generated a series of bedrock dams that would have backed up lakes along the paleovalleys. Conglomerate would have accumulated initially in the lakes but then would have been overlain by thick sections of fine sediments, similar to the deposits observed at many locations (Fig. 3).

Landslides, lavas, and ash-flow tuffs in nontectonically active areas commonly have blocked paleovalleys and modern rivers to generate short- to long-lived lakes (Costa and Schuster, 1988; Waythomas, 2001; Wallace et al., 2008). The most famous Tertiary examples are the Florissant and other lakebeds in Colorado, where paleovalleys similar to those in Nevada were blocked by andesite mudflows or silicic ash-flow tuffs (McLeroy and Anderson, 1966; Wobus and Epis, 1978; Gregory and McIntosh, 1996). The Colorado lakebeds consist of conglomerate overlain by finely laminated shale and mudstone containing abundant plant fossils, similar to the Nevada deposits. Blockage by resistant lavas seems more likely to generate long-lived lakes than blockage by easily eroded tuffs. Wallace et al. (2008) interprets several middle Miocene basins in northern Nevada to have formed where rhyolite lavas blocked exterior drainages. Oxbow lakes along the paleovalleys are another possible lacustrine environment. An ox-bow setting might account for the close interbedding of thin layers of shale and limestone with coarse conglomerate in the California Mountain paleovalley segment.

Megabreccia

Megabreccias at Nanny Creek, Windermere Hills, Cornwall Mountain, Oxley Peak, and Taylor Canyon are distinctly different than the basal conglomerates because they are composed almost entirely of coarse, angular clasts of ash-flow tuff (Figs. 9E and 9F)—thus post-dating ash-flow tuffs and other volcanic rocks—and comprise minor constituents of fine-grained sedimentary sequences in the Windermere Hills and at Cornwall Mountain. I interpret megabreccia to have formed during catastrophic, dam-burst type floods resulting from short-term blockage of paleovalleys by volcanic rocks or landslides (for example, Costa and Schuster, 1988). Dam-burst floods are common, even in historic times, in volcanically or tectonically active regions worldwide, have far greater magnitudes than normal floods, and commonly generate extremely coarse fluvial deposits (Wilson and Walker, 1985; Costa and Schuster, 1988; Manville et al., 1999; Waythomas, 2001).

Dams created by ash-flow tuffs blocking paleovalleys probably would be short lived relative to fault-created dams and would be more likely to fail catastrophically. Catastrophic failure is supported by observation of modern examples and by consideration of the type of dam they create. For example, an ash-flow tuff deposited in one of two tributaries of a paleovalley would block the other tributary to create a nearly instantaneous dam as much as 100 m high. However, the upper part of such a tuff would be poorly or non-welded and easily eroded. The lake behind the dam would overtop the tuff and erode rapidly through it, generating a dam-burst flood. Clasts reworked into megabreccia would be dominantly of the tuff itself. Basement clasts would be sparse or absent, because dams created by these processes would not expose basement. Coarse, angular blocks of the tuff of Big Cottonwood Canyon only slightly detached from underlying tuff outcrop at Oxley Peak may be a preserved example of intense erosion that cut through a tuff dam.

Fault dams are unlikely to break catastrophically. West-facing faults would have east-tilted basement blocks making the dam. A lake overtopping such a dam in an east-draining river would erode slowly through Paleozoic bedrock in the footwall and would not, therefore, create a dam-burst flood. Also, because faulting is incremental, deep lakes would not form quickly. Even east-facing faults might make rollover basins and lakes, but these would also not break catastrophically.

Regional Drainage Patterns and Location of the Paleodivide

Sparse evidence for paleoflow direction suggests that most of the paleovalleys drained to the east. Sedimentary paleocurrent data are available only in the Windermere Hills, where Mueller (1992) found imbricated cobbles that indicate eastward flow during deposition of the post–39.4 Ma conglomerate (Tdc, Fig. 11). Elsewhere, sedimentary matrices are too poorly indurated to provide outcrop.

Clast types also support eastward flow. Most clasts are quartzite or chert-quartzite-pebble conglomerate, the most resistant rock types in the region. These rocks crop out widely across the region, with the eastern limit of their distribution near the longitude of Nanny Creek (Fig. 1; Coats, 1987). Notably lacking are cratonic or eastern assemblage rocks, which crop out extensively near the Nevada-Utah border. Their absence suggests paleovalleys did not drain westward.

Flow and deposition of the tuff of Big Cottonwood Canyon ~100 km to the east also suggests eastward-draining paleovalleys. Tuffs can flow uphill variable distances, because they have a component of kinetic energy generated during column collapse (Cas and Wright, 1987). However, uphill flow for ≥100 km in narrow, potentially steep channels is very unlikely. If erupted near a paleodivide, the tuff could have initially crested the divide, then flowed downstream; therefore, a divide east of but near the caldera is possible. Eastward flow of the tuff of Big Cottonwood Canyon contradicts the conclusions of Palmer and MacDonald (2002) from anisotropy of magnetic susceptibility (AMS) data that it flowed westward in the Windermere Hills and at Nanny Creek. However, AMS data depend on proper correction to paleohorizontal orientation, and Palmer and MacDonald recognized that the tuffs might have had primary dips. Small over or under correction for tilt can change inferred flow directions by 180°, which may have happened in this case. The sources of other tuffs are unknown or uncertain but are more likely in the western part of the area shown in Figure 2 than in the eastern part where Eocene tuffs are sparse.

I infer that these paleovalleys drained eastward to the remnants of the Uinta and Green River basins. Isotopic ages of probably water-lain tuffs in the Uinta basin range from ca. 47 to 30 Ma (Bryant et al., 1989). These ages and stratigraphic data (Baars et al., 1988; Hintze, 1993) indicate the Uinta basin likely was accumulating sediments throughout the time of emplacement of the 45–40 Ma ash-flow tuffs and during the later paleovalley fill of this study. Similar east-draining rivers fed the Eocene Claron Formation in southwestern Utah at the same time (Goldstrand, 1994).

In contrast, the paleovalley at Willow Creek Reservoir may have drained westward. The tuffs of Big Cottonwood Canyon and Nelson Creek are thick and densely welded there (Wallace, 2003a, 2003b). This location is ~30 km southwest of the Big Cottonwood Canyon caldera, and tuff could have flowed upstream that distance. However, tuffs that had flowed upstream even 30 km more likely would be thin and poorly welded.

A west-draining system of paleovalleys, also filled with ash-flow tuffs and sedimentary rocks, extended to the Pacific Ocean (Fig. 1; Lindgren, 1911; Henry et al., 2004; Faulds et al., 2005; Garside et al., 2005). If the interpretation that the northeastern Nevada paleovalleys drained eastward is correct, a paleodivide must have separated the two systems, and its location can be placed within an ~100-km–wide belt through north-central Nevada (Fig. 1).

Clast-type distribution and size demonstrate that paleovalleys at Battle Mountain (Doebrich, 1995), Mule Canyon (John and Wrucke, 2003), and adjacent to the Caetano caldera (Gilluly and Masursky, 1965; John et al., 2008) drained to the west (Fig. 1). Farther south, ash-flow tuffs of the central Nevada caldera belt were able to flow far to the west, into the Sierra Nevada (Deino, 1989; Garside et al., 2000; Brooks et al., 2003; Henry et al., 2004; Faulds et al., 2005), which indicates most calderas were probably west of the divide. The Nine Hill Tuff (unit D of the Bates Mountain Tuff) spread from near Ely on the east to the foothills of the Sierra Nevada on the west (Deino, 1989; Best et al., 1989). Assuming Ely was east of the paleodivide, this distribution suggests the source caldera of the Nine Hill Tuff was probably near the divide. The caldera's location, unfortunately, is unknown (Deino, 1989; Best et al., 1989).

This proposed divide is as much as 200 km west of that shown by Christiansen and Yeats (1992) (Fig. 1). Given that east-drainage is based only on one paleocurrent analysis (Mueller, 1992), the reasonable but no doubt challengeable inference that an ash-flow tuff is unlikely to have flowed upstream ~100 km, and the absence of carbonate eastern assemblage clasts, additional paleocurrent and clast-type data are necessary to evaluate the paleodivide location.

Eocene paleovalleys that drained Sevier or Laramide uplifts and have similar sedimentary deposits ± ash-flow tuff are widely recognized elsewhere in the western USA. Paleovalleys containing coarse debris and partly filled with 49 Ma ash-flow tuff drained >100 km eastward from the Sevier fold-and-thrust belt in Idaho and Montana into Wyoming (Janecke et al., 2000). Paleovalleys drained northeastward off what is now Basin and Range Province of central Arizona onto the Colorado Plateau (McKee, 1951; Cooley and Davidson, 1963; Elston and Young, 1991). The Eocene erosion surface in Colorado had low relief and was incised by eastward-draining paleovalleys 50–300 m deep (Epis and Chapin, 1975; Chapin and Lowell, 1979). Paleoflora analysis of the Florissant lake beds, which accumulated in a Colorado paleovalley that was blocked by andesite mudflows, indicates the surface was at 2.4–2.7 km (Gregory and Chase, 1992). A 36 Ma ash-flow tuff flowed at least 140 km eastward in one Colorado paleovalley (Chapin and Lowell, 1979). The Florissant lake beds and their paleovalley setting are analogous to the lacustrine beds in paleovalleys in northeastern Nevada.

Paleotopography and Elevation

The distribution of ash-flow tuffs and sedimentary rocks demonstrates that a series of deep (up to 1.6 km), wide, east-trending paleovalleys existed by at least 45 Ma, the oldest ash-flow tuff found in this study, but probably formed much earlier. The broad paleovalleys and low-relief interfluves seem characteristic of relatively deeply eroded topography. However, widespread preservation of uppermost crustal rocks, i.e., western assemblage Paleozoic rocks or the clastic wedge that formed after the Antler orogeny, as paleovalley walls indicates little total stripping of the surface. Deep weathering and erosion could have been enhanced by the humid climate and high runoff characteristic of the Eocene (Kelly et al., 2005). Nevertheless, considerable time also seems required. Therefore, formation of the paleovalleys following uplift during the Sevier orogeny at the end of the Cretaceous is plausible. This interpretation is consistent with inferences of maximum crustal thickening and uplift following Sevier contraction (Coney and Harms, 1984; Dilek and Moores, 1999; DeCelles, 2004) and with common Cretaceous apatite fission-track ages in Paleozoic rocks (Hickey et al., 2003). Janecke et al. (2000) also concluded a Late Cretaceous origin for the two middle Eocene paleovalleys that drained eastward from the Sevier fold-and-thrust belt in Idaho and Montana.

The region probably was a high plateau, dissected by numerous broad valleys, similar to the Eocene erosional surface of the Rocky Mountains (Epis and Chapin, 1975) but with generally deeper valleys. Where exposed outside of paleovalleys, the Eocene erosional surface is relatively flat. Eocene lavas rest on low-relief surfaces cut into pre-Cenozoic rocks in the Piñon Range (Smith and Ketner, 1978), western part of the Tuscarora volcanic field (Henry and Boden, 1999), and Emigrant Pass volcanic field (Henry and Faulds, 1999), and the late Eocene erosional surface has low relief near the Caetano caldera in the Shoshone Range south of Figure 3 (John et al., 2008). Eocene rocks are absent in many areas, for example, the northwestern flank of the Carlin trend (Evans, 1974) and south of Willow Creek (Wallace, 2003b), which were probably topographic highs in the Eocene (Wallace et al., 2008). In these areas, Miocene rocks rest on a similar low-relief surface on pre-Cenozoic rocks. However, the ~500-m to 1600-m depths of paleovalleys probably require some relief on interfluves (Table 3).

Approximate absolute elevations of the high plateau can be inferred from the depths of the paleovalleys and paleoflora-based estimates of the elevations of lacustrine deposits. If the paleoelevations of 2.0 ± 0.2 km (Wolfe et al., 1998) or 2.8 ± 1.8 km (Chase et al., 1998) estimated from flora in the paleovalley at Copper Basin are correct, and the paleovalley there was ~1.6 km deep, then the adjacent interfluve was between 3.6 and 4.4 km high. A paleoelevation for Lake Uinta, to which the rivers drained, would place a minimum elevation for northeastern Nevada but is not available. Paleoelevation of the Green River basin in Colorado, which probably connected with the Uinta basin during much of their history, has recently been reinterpreted from 0.5 to 1 km (Greenwood and Wing, 1995) to 2–3 km (Chase et al., 1998; Wolfe et al., 1998). The 2–3 km estimates also require that the northeastern Nevada paleovalleys had to be at high elevation. These calculations represent somewhat circular reasoning because Chase et al. (1998) and Wolfe et al. (1998) reinterpreted higher elevations from the initial low-elevation interpretation of Axelrod (1966b) and Greenwood and Wing (1995). Using the older estimates of low elevations for the Green River basin would shift everything down 1–2 km. Nevertheless, both the absolute elevation and interpretation of a relatively low-relief surface cut by broad paleovalleys are consistent with interpretations about the character of the Sevier hinterland and with northeastern Nevada being an older, eroded analog to modern Tibet and the Andean plateau (Coney and Harms, 1984; Dilek and Moores, 1999; DeCelles, 2004).

Two studies propose that the region underwent uplift in the Eocene contemporaneous with magmatism, and this study does not preclude renewed uplift of a slightly eroded Sevier highland. If initial uplift of the region occurred during the Sevier orogeny, erosion would have had 20–40 Ma to reduce average elevations. However, stripping of more than 1–2 km of material is precluded by the widespread preservation of upper crustal rocks. Based on the southward sweep of magmatism through the northwestern USA between ca. 60 and 30 Ma, Humphreys (1995) proposed that the shallow Laramide slab detached at what are now the USA-Canadian and USA-Mexican borders and sank in the middle. As the northern edge pulled southward, asthenosphere upwelled in its wake, driving both magmatism and uplift. Horton et al. (2004) used stable-isotope data to interpret ~2 km of uplift in northeastern Nevada between ca. 55 and 25 Ma and cited Eocene magmatism as the origin of uplift. The ages of sediments analyzed by Horton et al. (2004) were based mostly on assumed sediment accumulation rates, constrained by only a few isotopic ages, most done ~30 yr ago. More modern dating indicates that the analyzed rocks probably range only from 46 to 37.5 Ma (Haynes et al., 2002; Haynes, 2003; Ressel and Henry, 2006) and require that the proposed uplift occur over a span of 1–3 Ma. Uplift of 2 km would require either that the region reached elevations of ~5–6 km or that it had declined to elevations of ≤2 km by early Eocene before the proposed uplift. A test for both asthenospheric upwelling and a relation between uplift and magmatism or upwelling would be to look for a southward sweep of uplift in parallel with the southward sweep of magmatism.

Extension and Orogenic Collapse

The data presented in this paper are consistent with two episodes of minor extension, one before ca. 41 Ma and possibly as old as 46 Ma that formed the Elko basin and small lacustrine basins at Double Mountain and Coal Mine Canyon, and a second between ca. 40 and 38 Ma that formed lacustrine basins at Cornwall Mountain, Copper Basin, and the Windermere Hills. Although lacustrine basins could have formed by blockage of the paleovalleys by lavas or tuffs, the younger basins formed contemporaneously with a well-documented, 40–38 Ma episode of extension. An extensional origin thus seems plausible for the older basins.

The timing of older lacustrine deposition and presumed extension is only constrained to be older than ca. 41 Ma, the age of tuffs at Double Mountain and Coal Mine Canyon that overlie lacustrine deposits. Haynes et al. (2002), Haynes (2003), Cline et al. (2005), and Hickey et al. (2005) interpret extension of the Elko basin to have started before 46 Ma, based on the age of the basal conglomerate. However, this conglomerate is similar to the basal paleovalley conglomerates described earlier in this paper, and by itself is not indicative of extension. Lacustrine deposits in the Elko basin are only known to be as old as 39 Ma and no older than 46 Ma (Solomon et al., 1979; Haynes, 2003; Cline et al., 2005; Hickey et al., 2005). The absence of pre–40 Ma lacustrine deposits in northern and eastern paleovalley segments suggests either that the older episode of extension did not affect those areas or any extension did not generate lacustrine basins.

Lacustrine deposits and presumed extension are more precisely dated at Copper Basin, where they lie between the 40.0 Ma tuff of Big Cottonwood Canyon and a 37.6 ± 0.4 Ma tuff (Rahl et al., 2002). Lacustrine deposits are younger than 39.4 Ma in the Windermere Hills and may be as young as ca. 35 Ma (Mueller et al., 1999).

Several factors indicate that extension was minor in both episodes. (1) Lacustrine and other sedimentary deposits are everywhere thin (<1 km, mostly ≤600 m) and laterally discontinuous. Although most obvious for the paleovalley basins, lacustrine deposits and the entire Elko Formation in the Elko basin are only ~500 and ~850 m thick, respectively (Solomon et al., 1979; Haynes, 2003). Outcrop, seismic-reflection, and borehole data indicate that Elko basin deposits are discontinuous along the Ruby Mountains front (Smith and Ketner, 1976; Satarugsa and Johnson, 2000). I conclude that a large, contiguous Elko basin depicted by some (Solomon, 1992; Christiansen and Yeats, 1992; Cline et al., 2005) never existed. The Elko basin may have been a series of discrete, extensional basins that formed in the hanging wall of an early Ruby Mountains fault, although the possibility that they were also erosional low spots like the paleovalleys should be considered. In either case, the similarity of deposits across the region probably reflects similar processes acting in a series of small, separate basins.

(2) Angular unconformities are generally absent or minor. The only demonstrable angular unconformity, of ~15°, is between Elko Formation and 38 Ma volcanics in the adjacent Piñon Range, Elko Hills, and Emigrant Pass (Smith and Ketner, 1976; Solomon et al., 1979; Henry and Faulds, 1999; Haynes, 2003; Cline et al., 2005). Otherwise, Eocene rocks are conformable. Minor, post–40 Ma tilting at California Mountain would have generated an angular unconformity, but younger Eocene rocks are absent there. Eocene and Miocene rocks are conformable at Oxley Peak and in the Windermere Hills.

(3) Paleovalleys acted as conduits for ash-flow tuffs and fluvial sediments at least through emplacement of the 40 Ma tuff of Big Cottonwood Canyon. Continuity after 40 Ma is less certain, but deposition of fluvial sediments and megabreccia in the Windermere Hills and at Nanny Creek indicate the paleovalleys continued to be active, although for how long is uncertain. Examination of sedimentary strata in the western Uinta basin to determine if significant material was derived from the west could help evaluate this question.

My interpretation from surficial evidence for minor extension in the Eocene or earlier or even pre-middle Miocene is starkly different than interpretations from thermochronology and paleobarometry of at least 7 km of pre-Late Eocene tectonic exhumation of the Ruby Mountains and possibly 30 km of total exhumation there (Hodges et al., 1992; McGrew and Snee, 1994; Snoke et al., 1997; McGrew et al., 2000: Howard, 2003). Rocks formerly at depths of ~30 km are now exposed in the Ruby Mountains core complex (McGrew and Snee, 1994; McGrew et al. 2000). However, this considerable uplift seems to have had variable but mostly little expression at the surface, except possibly immediately above the Ruby Mountains. The Elko basin adjacent to the Ruby Mountains is shallow and discontinuous. The prolonged period of cooling and inferred uplift and extension between 36 and 20 Ma (Kistler et al., 1981; Dokka et al., 1986; McGrew and Snee, 1994; McGrew et al., 2000) apparently generated no contemporaneous basins, and any eroded material must have been transported out of the region (Wallace et al., 2008). The proposed, low-angle Pequop fault, which is interpreted to have accommodated ~10 km of crustal thinning between ca. 84 and 41 Ma (Camilleri and Chamberlain, 1997), is one of the few cited surface-breaking, early Cenozoic faults in northeastern Nevada. However, numerous well-rounded quartzite boulders are present along part of the fault, which suggests that it may be a Paleozoic unconformity. Several episodes of extension have been inferred in the Windermere Hills (Mueller and Snoke, 1993; Mueller et al., 1999), but the angular conformity of Eocene and middle Miocene deposits (discussed above) would seem to preclude significant deformation over this interval. Possibly pre-middle Miocene extension was mostly limited to the lower and middle crust (MacCready et al., 1997; Wernicke and Getty, 1997; McGrew et al., 2000), or early uplift and cooling of the Ruby Mountains may have been by diapirism (Howard, 1980). Clearly, further work is needed to resolve the disparity between interpretations.

The driving mechanisms for extension are debated and probably varied, with gravitational spreading of thickened crust being one of the prominent possibilities (Wernicke, 1992; Sonder and Jones, 1999; Dickinson, 2002). The record of Cenozoic extension in northeastern Nevada, which most likely had thick (up to 60 km) crust resulting from Mesozoic contraction, should help test the possibilities. Based on this study, extension with surface expression is limited to possibly two episodes of minor faulting in the Eocene. Most tilting, uplift, and extension in a broad region around northeastern Nevada appear to have occurred in the middle Miocene (Thorman et al., 1991; Zoback et al., 1994; Brooks et al., 1995a, 1995b; Snoke et al., 1997; John et al., 2000; Colgan and Metcalf, 2006; Colgan et al., 2008). Overthickening alone seems to have been insufficient to drive extension, in agreement with calculations of Sonder and Jones (1999), unless extension was restricted to the lower and middle crust. Even the addition of intense Eocene magmatism to weaken overthickened lithosphere seems to have been unable to drive major, surface-breaking extension, consistent with the observation of little extension during middle Cenozoic magmatism in the Great Basin (Best and Christiansen, 1991). Rahl et al. (2002) also concluded that gravitational collapse did not drive extension and called upon the asthenospheric upwelling model of Humphreys (1995) to weaken the mantle lithosphere.

Implications for Carlin-Type Gold Deposits

Contentious issues about Carlin-type deposits have been their age, which is now thoroughly documented as Eocene (Hofstra et al., 1999; Ressel and Henry, 2006), their relation to magmatism or extension (Muntean et al., 2004; Cline et al., 2005), and their depth of formation, which was variably interpreted to be as shallow as 1 km to as deep as 5–8 km (Cline et al., 2005). Carlin-type gold deposits of the Carlin trend formed between ca. 40 and 37 or 36 Ma (Cline et al., 2005; Ressel and Henry, 2006), contemporaneous with the intense, 40–36 Ma episode of magmatism in the trend and with the younger, ca. 40–38 Ma episode of what I interpret to be small-magnitude extension in the region. Magmatism remains a compelling heat source for Carlin-type deposits of the Carlin trend, because the deposits are intimately associated in space and time with intrusions (Ressel and Henry, 2006). To what extent the pre–41 Ma and the 40–38 Ma episodes of extension affected the area of the Carlin trend remains to be determined. An origin calling on extension alone would seem to require that only the 40–38 Ma episode of extension affected the trend or was somehow different than the pre–41 Ma episode, for which there are no known contemporaneous deposits. An additional question is whether small-magnitude extension is sufficient to drive the hydrothermal systems. Possibly both magmatism, as a heat source, and extension-driven fracturing, to provide conduits for hydrothermal circulation, were necessary.

Preservation of uppermost crustal rocks throughout northeastern Nevada, which demonstrates only minor removal of material above the areas of Carlin-type deposits, extrapolation of the Eocene paleosurface over the Carlin trend (Henry and Ressel, 2000a; Ressel and Henry, 2006), and fission-track data that confirm that the present-day surface in the trend is ~500–1500 m below the Eocene paleosurface (Hickey et al., 2003; Cline et al., 2005) require that the deposits formed at depths no greater than 2 km and probably commonly less than 1 km. Finally, the Carlin trend was probably within 50 km of the paleodivide and was not covered by a lake (Figs. 1 and 3). This supports the conclusion of Haynes et al. (2003) and Hickey et al. (2005) that the Elko basin was probably not a source for hydrothermal fluids for the deposits of the Carlin trend.

ACKNOWLEDGMENTS

I greatly appreciate discussions with David John, Joe Colgan, Keith Howard, Chuck Thorman, Allen McGrew, Alan Wallace, Chuck Chapin, Dave Boden, Jim Faulds, Patrick Goldstrand, John Muntean, Fred Zoerner, Ken Hickey, Dick Tosdal, and Mike Ressel, very helpful reviews of an initial draft by David John and Joe Colgan, and insightful journal reviews by Allen McGrew, Keith Howard, and Alan Wallace. Geologic mapping was supported by the STATEMAP program of the U.S. Geological Survey. $^{40}Ar/^{39}Ar$ dating was done at the New Mexico Geochronology Research Laboratory (New Mexico Institute of Mining and Technology) under the patient guidance of Bill McIntosh, Matt Heizler, Rich Esser, and Lisa Peters.

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Manuscript Received 13 April 2007
Revised Manuscript Received 22 August 2007
Manuscript Accepted 23 August 2007