Early Cenozoic topography, morphology, and tectonics of the northern Sierra Nevada and western Basin and Range

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ABSTRACT

Debate surrounds the origin, uplift, and evolution of the northern Sierra Nevada and western Basin and Range. The studies presented here integrate different scales of observation, from local paleovalley morphology, estimation of local slopes, and braided stream alluvial architecture, to regional assessments of sediment and volcanic provenance and paleoelevations across the proposed ancestral Sierra Nevada–Nevadaplano region to gain a better understanding of early Cenozoic topography, morphology, and landscape evolution of the region, and to assess the possible tectonic and climatic drivers for this evolution. Results from sedimentologic analysis of Eocene fluvial deposits show diachronous, localized paleovalley incision and braided stream aggradation in a system influenced by Eocene climate, eustasy, and Laramide tectonism, and suggest that previous estimates of the timing and amount of range uplift based on paleochannel gradients may be invalid. Overlying Oligocene ignimbrites deposited in the Sierra Nevada were geochemically and geochronologically correlated to sources in central Nevada, and results from this work show that ignimbrites traveled over 200 km from their source calderas across what is now the crest of the Sierra Nevada, and that in the Oligocene, no drainage divide existed between Nevada source calderas and sample locations 200 km west. Hydrated volcanic glass from these units was used as a proxy for the isotopic composition (δD) of Oligocene meteoric water, which reflects the effect of ancient topography on precipitation. δD decreases from west to east across the Sierra Nevada by ~48‰, which is similar to the isotopic gradient of precipitation over the area today. δD across Nevada decreases at a significantly lower gradient, reflecting a significant reduction in the rate of increase of paleoelevation with distance, and may reflect a gradual increase in mean elevation from west to east or partially closed system hydrology. This multidisciplinary approach provides a detailed reconstruction of the evolution of the ancestral Sierra Nevada drainage system from Eocene to Oligocene time and multiple lines of evidence to support the conclusion that the northern Sierra Nevada likely acted as the steep western flank of a gradually sloping high-elevation plateau (“Nevadaplano”) in the Oligocene. Miocene to Holocene extension lowered elevations across what is now the Basin and Range, possibly associated with gravitational spreading of overthickened, magmatically and radiogenically heated crust.

INTRODUCTION

The Sierra Nevada and Basin and Range regions of western North America reflect a complex tectonic history of late Mesozoic–early Cenozoic convergence, subsequent Laramide flat-slab subduction and batholith unroofing, and Neogene–Quaternary extension and translation (Fig. 1). Although the high topography of the modern Sierra Nevada has been thought to be the product of a substantial increase in mean elevation and relief in the late Cenozoic (e.g., Lindgren, 1911; Axelrod, 1980; Watabayashi and Sawyer, 2001; Jones et al., 2004), some research indicates that the Sierra Nevada has persisted as an elevated topographic feature throughout the Late Cretaceous and Cenozoic, requiring little to no late Cenozoic increase in mean elevation (e.g., Wernicke et al., 1996; House et al., 2001; Poage and Chamberlain, 2001, 2002; Mulch et al., 2006; Mulch et al., 2008; Cassel et al., 2009b; Hren et al., 2010). The Late Cretaceous to early Cenozoic Sierra Nevada is now widely regarded to have formed the western edge of a high elevation plateau that covered much of what is now Nevada and western Utah (cf. Wolfe et al., 1997; Wolfe et al., 1998; Dilek and Moores, 1999; DeCelles, 2004; Best et al., 2009; Cassel et al., 2009a). The topography, extent, and drainage of this plateau remain uncertain (Henry, 2008; Cecil et al., 2010; Mix et al., 2010), and questions persist as to the tectonic history, timing, and magnitude of changes in paleoelevations, and drivers for erosion and deposition in the ancient Sierra Nevada.

Eocene–Oligocene fluvial deposits (“auriferous gravels”) are locally exposed in the northern part of the range, overlain by a sequence of Oligocene rhyolitic ignimbrites that is preserved across the modern crest of the range and into central Nevada (Figs. 1 and 2). These geologic units provide an important record of the topographic and geomorphic evolution of the range during a controversial time period in the history of the region. Paleolandscape reconstructions based on a combination of methods, including sedimentology, stratigraphy, geochemistry, and geochronology, can improve our understanding of the timing and drivers of mountain building, sedimentary basin creation, and landscape evolution associated with Mesozoic–Cenozoic batholith emplacement and compressional deformation, Miocene–Holocene large-scale extension, and possible mantle delamination events at the western North American margin. This paper compares and evaluates results from multiple tools and data sets at local and regional scales, gathered from these Cenozoic sedimentary and volcanic deposits spanning ~25 Ma of deposition in the northern Sierra Nevada and western Basin and Range, to evaluate the accuracy of the individual methods and interpretations. In the Previous Research, Paleoaltimetry, and Discussion sections, we have outlined the
assumptions and weaknesses in the common methods of paleoelevation estimation. One of these methods (i.e., geomorphic gradients, incision estimates, stable isotope paleoaltimetry, paleobotany, etc.) alone cannot answer questions of Sierra Nevada paleotopographic history with enough accuracy or detail. It is only by combining this study, our previous work, and a host of other previous research that we can gain an accurate understanding of the history of surface uplift, denudation, and landscape change in the area that is now the Sierra Nevada and western Basin and Range. This approach provides both a detailed reconstruction of the evolution of the ancestral Sierra Nevada drainage system from Eocene–Oligocene time and multiple lines of evidence to support the conclusion of recent research that the northern Sierra Nevada likely acted as the steep western flank of a gradually sloping high-elevation plateau (“Nevadaplano”) in the Oligocene. Miocene to Holocene extension lowered elevations across what is now the western Basin and Range, possibly associated with gravitational spreading of overthickened, magmatically and radiogenically heated crust.

GEOLOGIC SETTING

Pre-Cenozoic basement on the western slope of the northern Sierra Nevada consists of an amalgam of Paleozoic–Mesozoic continental margin and accretionary prism metasedimentary and metavolcanic rock, arranged in N-S–trending belts defined by major pre-Cenozoic faults (Fig. 2; Girty et al., 1996), intruded by Jurassic to Late Cretaceous tonalite and granodiorite emplaced during continental arc magmatism (Figs. 1 and 2; Stern et al., 1981; Chen and Moore, 1982; Girty et al., 1996; Duca, 2001). The westernmost “Smartville Belt” is composed of primarily weakly foliated and metamorphosed mafic to intermediate volcanic rocks (Fig. 2). Across the Big Bend–Wolf Creek fault zone, the basement is composed of both foliated chert-argillite and mafic to ultramafic ophiolitic rocks, with variable, but generally higher grade, metamorphism (Girty et al., 1996; Day and Bickford, 2004). East of the Dogwood Peak fault, a band of highly deformed, serpentinitized ultramafic rocks—the Feather River Peridotite Belt—divides the Calaveras Complex to the west from the Shoo Fly Complex to the east (Fig. 2). The Shoo Fly Complex is composed of strongly foliated phyllites, metacherts, and quartzites, and is older overall than the Calaveras Complex to the west (Schweickert and Cowan, 1975; Hanson et al., 1988; Saucedo and Wagner, 1992; Girty et al., 1996).

The northern part of the Sierra Nevada batholith is a composite of predominantly Middle Jurassic and mid-Cretaceous plutons, which become younger and more felsic from west to east (cf. Saleeby, 1989; Girty et al., 1996; Day and Bickford, 2004). The younger plutons (ca. 100–85 Ma), intruded during the Sierra Crest magmatic event, occur as a series of large plutonic centers that extend from the southern part of the batholith to northwestern Nevada.
Figure 2. Distribution of geologic units in the northern Sierra Nevada of California (Interstate Highway 80 shown in red for reference); inset shows map location and major tectonic regions of northern California and Nevada. The Big Bend-Wolf Creek fault zone (BB-WC FZ), Pinoli Ridge-Tahoe fault zone (PR-T FZ), Dogwood Peak fault (DPF), and other faults, shown with black dashed lines, divide north-south-trending bedrock belts. Distribution of Eocene fluvial sediments is shown in orange, with important sedimentologic study locations noted. Digital elevation model from U.S. Geological Survey Seamless Data Warehouse; geology based on Yeend (1974), Saucedo and Wagner (1992), and Girty et al. (1996), modified from Cassel and Graham (2011).
Shallow erosion depths (<8 km) predominate in the northern batholith (Ague and Brimhall, 1988; Ague, 1997; Van Buer et al., 2009). The northern part of the range has lower mean and peak elevations than the south, and the extent of Cenozoic cover strata is limited to the north.

After the cessation of arc magmatism, the region that is now the Sierra Nevada and western Basin and Range underwent a period of significant erosion. The volcanic cover produced by the Cretaceous arc was eroded, as well as the Cretaceous plutonic rock—based on granitoid depths of emplacement—creating a 3–8 km of Cretaceous plutonic rock—based on 

The “Chalk Bluffs flora,” from the auriferous gravels at You Bet Diggings (Fig. 2), have been interpreted as proximal equivalent and deposits of the fluvial feeder system to the Ione Formation (Fig. 2; Dickinson et al., 1979), a unit of fine-grained marine, marginal-marine, and estuarine sediments composed of predominantly anauxilite clay and quartz sand, occurring at the western base of the Sierra Nevada (cf. Allen, 1929). The Ione Formation is assigned to the middle Eocene, based on molluscan fauna and correlation to the Domengine Formation of the California Coast Ranges (Morris, 1966; Creely and Force, 2007). Oligocene rhyolitic ignimbrite, tuffaceous paleosol horizons, and fluvial volcanic clays, generally referred to as either the Delleker (northern Sierra) or Valley Springs (central Sierra) Formations (Dalrymple, 1964; Wagner et al., 2000), overlie the prevolcanic gravel or lie in buttressed unconformity on the basement (Figs. 2 and 3). The Oligocene age was confirmed by 

Figure 3. Chronostratigraphic column showing ages and depositional relationships of Cenozoic units in the northern Sierra Nevada. Units are bounded by unconformities (wavy lines; base of section unknown) and the complete sequence is only found in some locations. Stratigraphy modified from Saucedo and Wagner (1992), Girty et al. (1996), Wagner et al. (2000), and Cassel and Graham (2011).

PREVIOUS RESEARCH

Although a number of studies have shown that the area that is now the Basin and Range was a region of eroding high topography in the mid-Cenozoic (Dilek and Moores, 1999; DeCelles, 2004; Best et al., 2009; Cassel et al., 2009a; Colgan and Henry, 2009; Ernst, 2010; Henry and Faulds, 2010; Mix et al., 2010), the timing of development, the maximum paleoelevations, the extent of paleovalleys, and the timing and pattern of structural demise of this high topography are still controversial. Much of the previous research has focused primarily on the western flank of this highland, in what is now the northern and central Sierra Nevada. A diverse array of methods focused on determining paleo-topography, although often applied to the same area, has provided somewhat conflicting results. These methods include comparisons of modern topographic gradients, timing of erosion and incision patterns, fault distributions, and projections of tilted strata in the northern Sierra Nevada that have estimated the timing and magnitude of range uplift using Cenozoic and older units (Lindgren, 1911; Hudson, 1960; Huber, 1981, 1990; Unruh, 1991; Wakabayashi and Sawyer, 2001; Jones et al., 2004), paleobotanical and stable isotopic proxies for paleo-erosion (Axelrod, 1957, 1962; Chamberlain and Poage, 2000; Poage and Chamberlain, 2002; Horton et al., 2003; Retallack et al., 2004a; Prothero, 2008).
2004; Mulch et al., 2006; Crowley et al., 2008; Mulch et al., 2008; Cassel et al., 2009b; Hren et al., 2010), and estimates of long-term exhumation rates through (U-Th)/He thermochronology of bedrock (cf. Wernicke et al., 1996; Cecil et al., 2006). All of these methods require assumptions as to how the regional Earth system, such as the drainage system or regional climate, operated in the mid-Cenozoic.

Stratigraphic studies used the current dip, orientation, location, depth of erosion, and/or total thickness of Cenozoic sedimentary sections from multiple locations to estimate the timing and magnitude of surface uplift (Hudson, 1960; Christensen, 1966; Huber, 1981, 1990; Unruh, 1991; Wakabayashi and Sawyer, 2001; Jones et al., 2004), resulting in estimates of late Cenozoic (5–2.5 Ma) surface uplift of 1500–2500 m. This method is based on the assumptions that (1) the Sierra Nevada landscape had reached equilibrium after Late Cretaceous–early Cenozoic denudation of the volcanic arc, and was then tilted to the southwest as a rigid block in the late Cenozoic, causing uplift at the range crest; (2) individual fluvial sedimentary sections and/or paleovalley contacts, with geographically limited exposures, were at one time part of the same paleochannel; and (3) that all modern valley incision topographically below Miocene–Pliocene volcanic rocks represents late Cenozoic basin incision. Some studies (e.g., Lindgren, 1911; Huber, 1981; Unruh, 1991) have also mistaken rock uplift for surface uplift by projecting gradients and tilted units directly up to the modern crest position to estimate changes in elevation (see England and Molnar, 1990, for detailed explanation of this problem), and assumed no erosion of the range, ignoring Eocene–Oligocene fluvial deposition (as described in Cassel and Graham, 2011) and the tilt produced due to erosional loading (Small and Anderson, 1995).

Paleoaltimetry studies based on isotopic or botanical proxies (Wolfe et al., 1997; Wolfe et al., 1998; Mulch et al., 2006; Mulch et al., 2008; Cassel et al., 2009b; Hren et al., 2010), however, indicate a relatively high Sierra Nevada and Basin and Range in the Eocene, Oligocene, and middle Miocene, requiring little to no late Cenozoic surface uplift. The isotopic composition of ancient hydration waters has been used to estimate paleoelevation in the northern Sierra Nevada in Eocene kaolinite from the foothills (Mulch et al., 2006), and in Oligocene volcanic glass from the entire western flank of the range (Cassel et al., 2009b). Both studies found evidence for significant pre-Eocene surface uplift and mid-Cenozoic surface elevations similar to the modern. Similar studies of authigenic minerals in strata east of the Sierra Nevada (Chamberlain and Poage, 2000; Poage and Chamberlain, 2002; Horton et al., 2004), and of mammalian tooth enamel (Crowley et al., 2008) and volcanic glasses (Mulch et al., 2008) from both sides of the current range crest, indicate the presence of a rain shadow effect produced by high topography since the middle Miocene (12–18 Ma). The isotopic proxy method is based on the assumptions that (1) once the proxy mineral has incorporated hydrogen (or oxygen) into its structure, it does not later exchange those hydrogen ions with hydrogen ions from ambient waters (Friedman et al., 1993b; Mulch et al., 2008), and (2) isotopic distillation processes can be reasonably modeled with a Rayleigh distillation equation (e.g., Chamberlain and Poage, 2000; Poage and Chamberlain, 2001; Rowley et al., 2001; Rowley, 2007).

Crustal structure modeling and thermochronology from the southern Sierra Nevada support early Cenozoic high topography (Wernicke et al., 1996; House et al., 1998, 2001). Using (U-Th)/He ages, House et al. (1997, 1998, 2001) concluded that ancient fluvial relief and mean elevation in the southern Sierra Nevada have either not changed or gradually lowered in the Cenozoic. Clark et al. (2005) confirmed this pattern of cooling ages, but noted the lack of temporal constraints on elevation changes between 32 and 3.5 Ma. Thermochronology along the Yuba River in the northern Sierra Nevada (Cecil et al., 2006) showed that total exhumation in the past 60 My was <3 km, with faster exhumation rates in the Cretaceous, but these data do not provide the resolution necessary to deduce smaller elevation changes within the Cenozoic, such as 1–2 km of late Cenozoic uplift (Farley, 2002). Furthermore, determining paleorelief and exhumation rates from (U-Th)/He ages requires an assumption of geothermal gradient, and surface uplift cannot be uniquely inferred from thermochronologic methods alone (Small and Anderson, 1995; Farley, 2002).

Considering the resolution of each method and the assumptions required, only one of these methods alone is not enough to thoroughly understand the landscape evolution of this paleohighland. By comparing multiple methods, including previously published and new data, we aim to provide further insights into the landscape evolution of this region, as well as to discuss the validity of the assumptions required with both stratal dip and isotopic proxy methods. A summary of work from Cassel and Graham (2011) will address the paleolandscapge geomorphology and drainage system evolution of the Eocene sedimentary sequence, showing the diachronous, localized nature of paleovalley incision and braided stream aggradation within the northern Sierra Nevada, which bears on previous estimates of the timing and amount of range uplift based on paleochannel reconstructions and gradients. Results from Cassel et al. (2009a), correlating overlying Oligocene ignimbrite deposits to Nevada using trace and rare earth element geochemical analyses of volcanic glass and 40Ar/39Ar ages, show that ignimbrites traveled over 200 km from their source calderas across what is now the crest of the Sierra Nevada (Cassell et al., 2009a; Henry and Faulds, 2010). Hydrated volcanic glass separated from these ignimbrites was also used for a hydrogen stable isotope paleoaltimetry study across the Sierra Nevada and western Basin and Range, to determine elevations across the area in the Oligocene, the timing of development of topography, and tectonic drivers for uplift. Previous results from across the northern Sierra Nevada, showing a 48% decrease in the isotopic composition of hydrated glass from ignimbrites located near paleo–sea level to ignimbrites 100 km to the east (Cassell et al., 2009b), are compared to new isotopic data from across the western Basin and Range. Analyses of 31–24 Ma ignimbrites, from Reno to Carlin, Nevada, show a gradual decrease in δD from west to east at a significantly lower gradient than that documented to the west of the crest.

**EOCENE FLUVIAL SEDIMENTS**

Eocene–Oligocene fluvial sediments provide a record of paleolandscapge geomorphology and drainage system evolution (cf. Cassel and Graham, 2011). Local topography and bedrock lithology controlled the location of broad, terraced paleovalley segments, which predominantly formed in the Paleozoic–Mesozoic higher-grade metamorphic bedrock east of the Big Bend–Wolf Creek fault zone, in the Calaveras Complex, and in the Shoo Fly complex (Fig. 2), and subsequently became areas of significant fluvial deposition. High-gradient, high-energy deposits are present near the edges of broad paleovalleys, suggesting that tributaries and outlets resided in steeper, narrower paleovalleys (Cassel and Graham, 2011). Multiple strath terraces are common in broad paleovalleys, indicating that incision events occurred both before deposition and during paleovalley widening, followed by periods of aggradation (Cassell and Graham, 2011). Within the paleovalley system, the location and morphology of the valleys were controlled by the mechanical strength and structure of the bedrock, and multiple strath terraces recorded generations of fluvial incision resulting from tectonically driven changes in base level (cf. Burbank et al., 1996; Strong and Paola, 2008). Macroscale strata sets in upper levels of the sections are continuous.
across terraces, indicating that terrace formation occurred prior to deposition of the upper part of the section, either during the initial incision of the paleovalley or as channel-widening events during deposition. Differences in resistance to weathering in differing lithologies, and the prevailing dip of bedding and cleavage in those lithologies, allowed paleorivers to widen in easily weathered, steeply dipping schists, phyllites, and chert-argillites, and confined paleorivers to steep canyons within more gently dipping bedrock, especially weakly foliated metavolcanics and quartzites (Figs. 2 and 4). Clast composition within the lower part of the Eocene fluvial sequence reflects local metamorphic bedrock lithologies (Dickinson et al., 1979), and some locations preserve “quarried” bedrock blocks at the base of the section (cf. Hancock et al., 1998).

Depositional Architecture

Broad paleovalley deposits contain predominantly braided stream facies deposited in an energetically variable system, and consist of multiple upward-fining cycles bounded by erosional surfaces (Cassel and Graham, 2011). The typical vertical profile observed in Eocene fluvial sediment sections consists of (1) a thick section of coarse-grained channel belt deposits, containing both channel fill and channel bar deposits (Willis, 1993a), with an erosional base that generally fines upwards from gravel- to sand-size lithofacies. Channel belt deposits are composed of predominantly crudely bedded pebble-cobble gravel and cross-bedded sandy gravel, both occurring near the base of the unit, often with rip-up clasts of silt-mud. Channel belt deposits can also have graded low-angle, often with rip-up clasts of silt-mud. Channel belt deposits is dominated by channel bar deposits, is controlled by the grain sizes of channel belt deposits, but instead consistently conformably overlie them across km-scale exposures (Fig. 4E), indicating a decrease in water discharge, the number of active channels within the channel belt, and the overall amount of in-channel deposition (Bridge, 1984; Willis, 1993a, 1993b; Bridge, 2003). In a climate with a high intensity of precipitation and high physical weathering rates, rapid coarse-grained transport and valley aggradation typically accompany an increase in runoff intensity (Tucker and Slingerland, 1997). These changes may coincide with climatic shifts between a system dominated by physical weathering with high runoff intensity, and a system dominated by chemical weathering with high fine-grained sediment supply. These anomalously thick (up to 11 m), laterally continuous (1–3 km) floodplain units, containing kaolinite-rich clay and oxisol intervals, record high fine-grained sediment supply, driven by high rates of chemical weathering in the warm, wet Eocene climate.

Channel Gradients

The mean grain size of channel belt and overbank deposits is controlled by the grain sizes available and the hydraulics of the channel-floodplain system (Bridge, 1993). The largest boulder-size grains likely represent blocks quarried from high-relief, incising bedrock channels upstream (Fig. 5A; cf. Hancock et al., 1998). The existence of paleovalleys filled with extremely coarse, locally sourced, poorly sorted, traction-structured deposits (Fig. 4A) suggests the presence of high-gradient tributary and headwater channels feeding coarse-grained sediment directly into broad paleovalleys, and moderate to high hillslope relief outside of the broad paleovalleys (Hancock et al., 1998; Whipple et al., 1999; Atta and Lavé, 2006). Rounding of large clasts and traction structures preserved within coarse-grained lithofacies in broad paleovalley deposits indicates that, at least during higher flow periods, paleochannel slopes were high enough to transport boulder-size grains (up to 3 m diameter) as bedload (e.g., Paola and Mohrig, 1996; Dade and Friend, 1998). Using the equations and method described by Paola and Mohrig (1996), paleochannel slope estimates of 0.004–0.055 m m−1 were obtained based on average grain sizes of coarse channel fills. These paleoslope estimates are similar to, although slightly lower overall than, the range of slopes of individual terrace treads exposed today in Omega (0.020), Upper Dutch (0.044 to 0.060), and Spanish Flat (0.030 to 0.063) (Fig. 2). The terrace morphology and sedimentology of the lower part of the section are consistent with transport and deposition of cobble-boulder grain sizes in braided streams of moderate gradients, and suggests that paleovalley surfaces have likely not been tilted or significantly changed since the Eocene. These results are also consistent with horizontally bedded, flat-lying floodplain deposits, and with the presence of cobble-boulder–size granitic clasts within the section (Figs. 4A and 5A), which indicate transport on the order of at least 10–20 km from their nearest possible sources (i.e., granitic clasts present at Malakoff, likely sources ~20 km east, as shown in Fig. 2).

OLIGOCENE IGNIMBRITES

Cassel et al. (2009a) studied the sequence of rhyolitic ignimbrites preserved across the modern crest of the Sierra Nevada and into the western foothills of the range to address the implications of their sourcing and travel from calderas located in what is now central Nevada (Fig. 6). Recent studies have shown that these units comprise the distal equivalent of ignimbrites sourced from calderas in central Nevada during the Oligocene “ignimbrite flare-up” (e.g., Brooks et al., 2003; Faulds et al., 2005; Brooks et al., 2008; Cassel et al., 2009a; Henry and Faulds, 2010). Multiple studies have suggested that the location of the regional drainage divide shifted significantly to the west in the late Cenozoic (Yeend, 1974; Dilek and Moores, 1999; Henry, 2008).
Figure 4. Photos showing typical Eocene Sierra Nevada braided stream depositional architecture and sedimentary structures: (A) traction-structured boulder conglomerate at Bear Creek, near You Bet: large clasts outlined, many were kaolinitized in situ in the Eocene (stick is 1.5 m with 10 cm stripes); (B) fine-grained sheet (FS), overlain by channel belt deposits (CB) (note sharp erosive contact between dark- and light-colored deposits) at Malakoff Diggings; (C) oxisol with ferric nodules, color mottling, and burrowing at Omega Diggings; (D) smoothed bedrock surface at base of section at Omega Diggings: cleavage dipping 74°–75° W; Eocene fluvial sediments, including paleosol pictured in (C) in background; (E) panoramic photo from a mine wall at Malakoff Diggings: thick lines delineate contacts between channel belt deposits (CB) and fine-grained sheets (FS). Modified from Cassel and Graham (2011).
Figure 5. Photos showing typical Eocene Sierra Nevada braided stream lithofacies: (A) large, well-rounded boulders in traction-structured conglomerate at Bear Creek, near You Bet (stick is 1.5 m with 10 cm stripes); (B) cross-bedded sandy cobbles and pebbles, overlain by clay-rich fine-grained sheet, at Omega Diggings; (C) cross-bedded sand at Morris Ravine; (D) nature of contacts between three upward-fining packages of channel belt deposits (CB) and fine-grained sheets (FS) at Malakoff Diggings; and (E) close-up of uppermost package set from (D): contact between a clay-rich fine-grained sheet and overlying cross-bedded sandy gravel. Modified from Cassel and Graham (2011).
Figure 6. Digital elevation map of the northern Sierra Nevada and western Basin and Range, showing distribution of 34–17 Ma ignimbrites (pink), previously mapped calderas (blue), and all volcanic glass samples used in stable isotopic and geochemical analyses (circles). Color indicates average isotopic value as shown in legend ($\pm 3\sigma$ error). Sample points are projected onto the red transect line for Figure 7. Extensional domains used for the paleogeographic reconstruction also shown along red line, assuming generally E-W–directed extension, and include percent extension (top) and reconstructed distance (estimates from Seedorf, 1991; Dilles and Gans, 1995; John et al., 2000; Surpless et al., 2002; Colgan et al., 2006, 2008; John et al., 2008; Colgan and Henry, 2009; Van Buer et al., 2009).
Cassel et al. (2009a) correlated Sierra Nevada ignimbrites to ignimbrites described and isotopically dated in the Walker Lane fault zone and central Nevada (Henry et al., 2004; Faulds et al., 2005) using trace and rare earth element geochemical analyses of volcanic glass and $^{40}$Ar/$^{39}$Ar isotopic ages to trace the extent of paleodrainages and characterize the Oligocene landscape morphology of the Sierra Nevada and Nevadaplano.

Geochemical and Radiometric Age Correlations

Sierra Nevada ignimbrite samples were broken into two populations based on geochemical composition (Cassel et al., 2009a). Three different radiometric ages match with these two geochemical populations. Samples from Ignimbrite 1, which is widespread throughout the Sierra Nevada (Fig. 6), were correlative with a sample of the tuff of Rattlesnake Canyon, including the type locality of the Delleker Formation. One sample in group 1 yielded a slightly older $^{40}$Ar/$^{39}$Ar age, but is geochemically similar to all other Ignimbrite 1 samples, suggesting that Ignimbrite 1 consists of two units that came from the same source caldera complex. Samples from Ignimbrite 2, including a sample taken from Tuff F of Brooks et al. (2003, 2008) at Haskell Peak, are correlative with the tuff of the Campbell Creek sample from western Nevada (see also Henry et al., 2012).

Cassel et al. (2009a) determined ten $^{40}$Ar/$^{39}$Ar radiometric ages of sanidine and plagioclase phenocrysts to evaluate the geochemical correlations discussed above. One sample yielded 12 concordant analyses with a weighted mean of 31.24 ± 0.06 Ma, and is within error of previous results for the tuff of Axehandle Canyon (31.2 ± 0.1 Ma; Henry et al., 2004; Faulds et al., 2005). Four samples grouped near 30.9 Ma and likely represent the downstream equivalent of the tuff of Rattlesnake Canyon (31.0 ± 0.1 Ma in Faulds et al., 2005). Crystals from four samples yielded discordant analyses at 28.7 Ma and correlate well with the tuff of Campbell Creek (28.8 ± 0.1 Ma in Faulds et al., 2005; see also Henry et al., 2012). Thus the majority of samples from within the northern Sierra Nevada have compositions and ages that correlate to the tuffs of Axehandle Canyon or Rattlesnake Canyon, both likely sourced from the same caldera complex in the Clan Alpine Mountains or the Stillwater Range, or to the tuff of Campbell Creek, sourced from a caldera in the Desatoya Mountains (Brooks et al., 2008; Cassel et al., 2009a; Henry et al., 2012).

**STABLE ISOTOPE PALEOALTIMETRY**

To determine elevations across the northern Sierra Nevada and western Basin and Range in the Oligocene, the timing of uplift, and the orogenic processes that drove the evolution of topography across both regions, we compare and integrate this previously published data together with an analysis of landscape morphology based on a regional hydrogeology stable isotope paleoaltimetry study (Cassel et al., 2009b), including new data from across what is now the Basin and Range. We estimated paleomeetric water values using hydrated volcanic glass separated from rhyolitic ignimbrites, based on the assumptions previously mentioned: that (1) once the glass has incorporated hydrogen into its structure, it does not exchange those hydrogen ions with hydrogen ions from the atmosphere, and (2) isotopic distillation processes can be reasonably modeled with a Rayleigh distillation equation.

Stable isotope paleoaltimetry is based on the observation that the isotopic composition of precipitation significantly changes with altitude, following a Rayleigh distillation, resulting in progressive depletion of $^{18}$O and $D$ in meteoric water at higher elevations (Craig, 1961; Dansgaard, 1964; Ingraham and Taylor, 1991; Chamberlain and Poage, 2000; Rowley et al., 2001). Increasing elevation drives adiabatic cooling of water vapor, and thus acts as the dominant control on the hydrogen isotopic composition ($\delta D$) of precipitation (Craig, 1961; Dansgaard, 1964). $\delta D$ scales at a predictable rate with change in elevation (Poage and Chamberlain, 2001). Thus, the change in elevation along a transect can be estimated based on the change in $\delta D$ of hydration water in volcanic glass. The simple Rayleigh distillation model of orographic precipitation (as discussed in Molnar, 2010) is dependent on the initial temperature and relative humidity at the site of evaporation (Rowley et al., 2001; Rowley, 2007). As Molnar (2010) points out, this distillation model may only apply to cooler climate regimes, similar to the modern.

Evidence for a significantly cooler Oligocene, in comparison to the Eocene, is found in floral and marine sedimentary records across North America and the world, with 5–8 °C of cooling estimated from late Eocene to Oligocene west coast flora and widespread paleosols (e.g., Sheldon and Retallack, 2004; Miller et al., 2008; Sheldon, 2009). According to Miller et al. (2009), global cooling began in the middle Eocene and culminated at 33.55 Ma (Oi-1). This is generally referred to as the transition to an icehouse Earth, with up to 12 °C of cooling overall, the establishment of the Antarctic ice sheet, and mean annual temperatures (MATs) decreasing to near modern levels (Lear et al., 2000; Lear et al., 2008; Zanazzi et al., 2009).

From oxygen isotopes in tooth enamel, Zanazzi et al. (2009) calculated a MAT drop of 8.2 ± 3.1 °C (from 21.0 ± 10.1 °C to 13.1 ± 9.5 °C) in the North American mid-continent, which is similar to estimates produced by Wolfe (~8 °C; from ~20 °C to 12 °C; 1994) and Sheldon and Retallack (5–6 °C; 2004) using western North American fossil floras. Modern western North American (California-Nevada) MAT, averaged from 1895 to 1995, was 12.7 °C, while the combined land and ocean global annual temperature, averaged from 1901 to 2000, was 13.9 °C (Smith and Reynolds, 2005; NOAA, 2009). Paleotemperature records for the Northern Hemisphere show that the MAT over the past 1000 yr ranged from 0.2 to 0.7 °C below current MAT (cf. Briffa and Osborn, 2002), and paleotemperature records from Antarctica indicate ±2 °C of variation from current MAT throughout the Holocene (Petit et al., 1999). Thus early Oligocene isotopic and fossil flora data record MATs similar to those recorded throughout the Holocene. North American continental ice sheets, like the Laurentide and Cordilleran ice caps of the Quaternary (Clark et al., 1999), did not develop in the Oligocene, of course, but this has been attributed to less amplification of orbital forcing due to differences in greenhouse gas levels (Petit et al., 1999; Retallack et al., 2004b).

The early Oligocene was also a time of voluminous supervolcanic eruptions across North America, with at least a dozen supervolcanic (>1000 km$^3$) eruptions in the Basin and Range alone (Best et al., 1989; Henry, 2008; Best et al., 2009), which have been linked to additional cooling trends (e.g., Cather et al., 2009; Jicha et al., 2009). The icehouse climate of the early Oligocene is considered similar to Holocene climate, especially across western North America (Lear et al., 2000; Katz et al., 2008; Lear et al., 2008; Miller et al., 2008). Thus volcanic glass erupted and deposited in the Oligocene is suitable for comparison of isotopic fractionation trends. As many factors, such as climate, can influence the rate of change in $\delta D$, employing this method requires confirmation of results based on other data sets, as we have previously presented in this paper.

In felsic volcanic glass, magmatic water possesses ~0.2 wt% of volcanic glass upon eruption and is present as hydroxyl groups (Dobson et al., 1989). Posteruption, at near surface temperatures, meteoric water enters glass first through H$^+$ (or D$^+$) exchange for primarily Na$^+$, K$^+$, and Al$^{11+}$ ions, followed by hydration through H$O_2$ absorption (Cerling et al., 1985; Oelkers, 2001; Rébiscoul et al., 2007; Valle et al., 2010). In basaltic glass, dissolution occurs in up to five successive steps, depending on glass composition.
(1–3) univalent (Na, K) and divalent (Ca, Mg) cations are quickly removed from near the surface of the glass; (4) three aqueous H⁺ ions exchange for Al while maintaining the glass structure; and (5) the now partially liberated silica slowly detaches from the glass surface during H₂O absorption, resulting in full dissolution of the glass (solid destruction) at a rate proportional to its geometric surface area (Oelkers, 2001; Oelkers and Gislason, 2001). Wolff-Boenisch et al. (2004) found that the dissolution process is similar for both basaltic and rhyolitic glass.

Silicate glasses have significant chemical durability, and have demonstrated sharp decreases in alteration rates at long time scales (Cailleteau et al., 2008). Multiple geologic studies have found this to be true based on empirical data. Based on their study of North American volcanic ash-fall tuffs, Friedman et al. (1993b) concluded that H⁺ exchange occurs in glass shards smaller than 150 μm within 5 ka of deposition, and those protons do not exchange further once the cations have been removed. Mulch et al. (see Table S2, 2008) also found that volcanic glass samples from different ash-fall tuffs collected from nearby locations, with ages varying from 0.6 to 12.1 Ma, preserved variable isotopic signatures. Recently, many geochemical studies (e.g., Crovisier et al., 2003; Cailleteau et al., 2008; Casey, 2008; Valle et al., 2010) have found that molecular changes in near-surface glass structure are critical to improving long-term glass durability and slowing water corrosion and glass dissolution. Cailleteau et al. (2008) found that the chemical durability of silica-rich glasses drastically increases during the process of aqueous dissolution. After coming into contact with water, the more soluble components of the glass exchange for protons (as outlined above, steps 1–4), resulting in a porous silicate structure on the surface (Cailleteau et al., 2008). These silicate groups then bond together, driving substantially increased density and decreased porosity of the outer layers of the glass (referred to as a gel in Cailleteau et al., 2008), thus preventing further dissolution, proton exchange, and the release of soluble elements (see fig. 2 in Casey, 2008). Valle et al. (2010) found that a gel layer up to 5 μm thick develops within basaltic glasses within three months at 90 °C. The isotopic composition of the glass inside of this gel layer is preserved during interactions with both ²⁸SiO₂ and H,¹⁸O solutions. Thus, after formation, this gel layer serves to lock in the isotopic signature of the water in contact with glass during proton exchange. Similar to Valle et al. (2010), we also do not find evidence of further isotopic exchange with ambient waters used to soak and wash rhyolitic glasses during the separation process on short time scales (7–14 days).

Methods: Volcanic Glass

Bulk unwelded ignimbrite samples were taken from natural surface outcrops throughout the study area, from ignimbrites sourced from the central Nevada caldera complex (see Cassel et al., 2009a). Typically, basal unwelded portions of ignimbrites are preserved beneath strongly welded sections across the Basin and Range, and, as welding decreases with distance from the source, most Sierra outcrop locations are composed completely of unwelded deposits. These ignimbrites were originally deposited in paleovalleys with active fluvial systems, either blanketing the entire paleovalley or filling channels within the section, and ash-rich fluvial deposits overlies some ignimbrites (Henry, 2008; Cassel et al., 2009a; Henry and Fauds, 2010).

In the Sierra Nevada, these ignimbrites are now typically located a few meters (i.e., Mohawk Valley and Great Valley) to tens of meters in the western Sierra foothills (i.e., Nevada City) to 100–300 m in the high Sierra (i.e., Emigrant Gap) above modern streams (Fig. 2). At each outcrop location, bulk samples were selected from 10 to 30 cm under the surface exposure, to avoid any surface clays or other weathering products that may have formed due to increased surface area. Materials formed through full dissolution of volcanic glasses (silica detachment during H₂O absorption; see above) are recognizable in hand sample and with a petrographic microscope, and were not analyzed in these studies.

Volcanic glass fractions of 99% purity were prepared by crushing each sample in a ceramic mortar and then wet-separating into size fractions using plastic sieves and nylon mesh screens with opening diameters of 150, 70, and 38 μm. The 70–150 μm mesh fraction was treated three times in 10% HCl for 60 seconds, then soaked and washed thoroughly in distilled water to remove any carbonate minerals. Samples were treated twice with 8% HF for 10 seconds to remove the exposed surface and near-surface rinds (outside of and the outermost gel layer), washed repeatedly in distilled water, and dried. Repeated passes through a Frantz isodynamic separator removed magnetic minerals. Glass was gravity separated from quartz and feldspar in lithium metatungstate liquid. The purity and degree of alteration of each sample was checked with a petrographic microscope. Only samples with no visible alteration minerals, dissolution rinds, or birefringence were analyzed, to eliminate the potential of analyzing waters acquired during the final stage of glass dissolution. This treatment is thought to have little to no effect on glass chemistry (Cerling et al., 1985; Sarna-Wojcicki and Davis, 1991).

Hydrogen isotope ratios were measured in the Biogeochemistry Laboratory at Stanford University by high-temperature thermal combustion and continuous-flow gas mass spectrometry using a Thermo Electron TC-EA coupled to a Thermo Electron delta XL mass spectrometer. One and one-half to 3.5 mg of pure glass shards were enclosed in silver foil, dried under vacuum at 80 °C for at least 48 hours, and then immediately flushed with dry He gas within a zero-blank auto sampler. Four internationally referenced standard materials and in-house working standards were run with the samples, and three to 14 repeat analyses were run on each glass separate for consistency and error analysis. The raw isotope data were corrected for mass bias, drift over time of the thermal combustion reactor, and offset from the certified reference values. After correction, multiple geologic studies have found this to be true based on empirical data. Based on their study of North American volcanic ash-fall tuffs, Friedman et al. (1993b) con-cluded that H⁺ exchange occurs in glass shards smaller than 150 μm within 5 ka of deposition, and those protons do not exchange further once the cations have been removed. Mulch et al. (see Table S2, 2008) also found that volcanic glass samples from different ash-fall tuffs collected from nearby locations, with ages varying from 0.6 to 12.1 Ma, preserved variable isotopic signatures. Recently, many geochemical studies (e.g., Crovisier et al., 2003; Cailleteau et al., 2008; Casey, 2008; Valle et al., 2010) have found that molecular changes in near-surface glass structure are critical to improving long-term glass durability and slowing water corrosion and glass dissolution. Cailleteau et al. (2008) found that the chemical durability of silica-rich glasses drastically increases during the process of aqueous dissolution. After coming into contact with water, the more soluble components of the glass exchange for protons (as outlined above, steps 1–4), resulting in a porous silicate structure on the surface (Cailleteau et al., 2008). These silicate groups then bond together, driving substantially increased density and decreased porosity of the outer layers of the glass (referred to as a gel in Cailleteau et al., 2008), thus preventing further dissolution, proton exchange, and the release of soluble elements (see fig. 2 in Casey, 2008). Valle et al. (2010) found that a gel layer up to 5 μm thick develops within basaltic glasses within three months at 90 °C. The isotopic composition of the glass inside of this gel layer is preserved during interactions with both ²⁸SiO₂ and H,¹⁸O solutions. Thus, after formation, this gel layer serves to lock in the isotopic signature of the water in contact with glass during proton exchange. Similar to Valle et al. (2010), we also do not find evidence of further isotopic exchange with ambient waters used to soak and wash rhyolitic glasses during the separation process on short time scales (7–14 days).

The Sierra Nevada

Cassel et al. (2009b) used Oligocene volcanic glass samples from northern Sierra Nevada ignimbrites to estimate paleoelevations. The 8D of ancient hydration water in glass shard separates was determined at sample locations across the range, from deposits near paleo–sea level at the edge of the Great Valley to samples in western Nevada (Fig. 6; Table 1). Sample values cannot be interpreted as a direct measure of 8D of precipitation at the sample location, because higher elevation precipitation is incorporated into meteoric waters downstream through ground and surface transport. Glass thus records the 8D of the hypsometric mean of the elevation of the drainage basin above that sample location (cf. Rowley et al., 2001; Rowley, 2007). Lower in the basin, less-enriched precipitation is mixed with higher elevation, more-enriched meteoric waters, resulting in a smaller difference in meteoric water 8D between low and high elevation samples than is present in precipitation 8D, and an overall minimum estimate of topographic gradient (Rowley, 2007). This estimate is nonetheless useful for discerning between tectonic models.

The 8D of the volcanic glass samples from Cassel et al. (2009b) decreases at a steady rate from −95‰ ± 3‰ to −115‰ ± 4‰e at the most westerly locations, to values ranging from −152‰ ± 3‰e to −153‰ ± 3‰e at the locations ~100 km east of the Ione marine deposits (Fig. 7; Table 1: all values relative to SMOW). This
### TABLE 1. δD OF VOLCANIC GLASS FROM ALL ANALYSES, WITH UNCERTAINTIES. SAMPLE LOCATIONS IN THE NORTHERN SIERRA NEVADA, CALIFORNIA, AND DISTANCE FROM THE GREAT VALLEY (GV)

<table>
<thead>
<tr>
<th>Distance to GV (km)</th>
<th>Sample</th>
<th>Latitude (°N)</th>
<th>Longitude (°W)</th>
<th>Sample elevation (m)</th>
<th>Corrected individual analyses (see Methods)</th>
<th>Average δD (‰)</th>
<th>Std. dev.</th>
<th>2σ error</th>
</tr>
</thead>
<tbody>
<tr>
<td>Feather River drainage</td>
<td>SN07-005</td>
<td>39.6952</td>
<td>120.9899</td>
<td>1561</td>
<td>–132.8</td>
<td>–131.6</td>
<td>–131.6</td>
<td>–137.8</td>
</tr>
<tr>
<td>Yuba River drainage</td>
<td>SN07-019</td>
<td>39.6623</td>
<td>119.7768</td>
<td>1608</td>
<td>–153.8</td>
<td>–155.3</td>
<td>–155.0</td>
<td>–154.0</td>
</tr>
<tr>
<td>Mokelumne River drainage</td>
<td>SN07-001</td>
<td>39.6538</td>
<td>119.7760</td>
<td>1627</td>
<td>–145.0</td>
<td>–144.5</td>
<td>–145.5</td>
<td>–137.5</td>
</tr>
<tr>
<td>SN07-039</td>
<td>39.3273</td>
<td>120.7980</td>
<td>1341</td>
<td>8</td>
<td>–119.4</td>
<td>–126.0</td>
<td>–116.8</td>
<td>–120.4</td>
</tr>
</tbody>
</table>

~48‰ decrease in δD of meteoric water is similar to the isotopic gradient of precipitation over the range today (Ingraham and Taylor, 1991).

### The Western Basin and Range

Samples of 31–24 Ma ignimbrites from the California border east to Carlin, Nevada, were recently analyzed to determine the topography along transect across this region in the Oligocene, and to compare to the transect studied to the west of the modern crust (Figs. 6 and 7; Cassel et al., 2009b). The δD of volcanic glass separates sampled from 31 to 24 Ma ignimbrites distributed across what is now western Basin and Range are shown in Figure 7, plotted versus distance of each sample from the paleomarine shoreline as indicated by Ione Formation deposits (Dickinson et al., 1979; Creely and Force, 2007), and listed with location information in Table 2. An accurate estimate of changes in δD along the transect in the Oligocene requires removal of post-24 Ma extension across the Basin and Range. The onset of major faulting and crustal extension in the western Basin and Range occurred in the middle Miocene (e.g., Zoback et al., 1994; John et al., 2000; Colgan et al., 2008). Extension estimates vary widely across the province, from 10% to 25% in western Nevada and 15% to 100% in central Nevada (Fig. 4; Seedorf, 1991; Smith et al., 2002; Colgan et al., 2008; Van Buer et al., 2009). Using averages of these estimates in each extensional domain, we reconstructed pre-extensional distances along the transect and, for comparison, δD of modern meteoric waters (see also Table 2). Glass values are offset from the δD of hydration waters due to an isotopic fractionation of ~30‰ at surface temperatures and pressure as water is incorporated into the glass structure (Friedman et al., 1993a; Friedman et al., 1993b).

δD of the volcanic glass samples gradually decreases across Nevada from values ranging from ~139‰ ± 3‰ to ~146‰ ± 3‰ at locations that are now ~180 km east of the Great Valley, to values ranging from ~153‰ ± 3‰ to ~167‰ ± 3‰ at locations ~470 km to the east (now near Carlin, Nevada, Fig. 6; reconstructed distances shown in Fig. 7). This 7–28 ± 3‰ decrease in δD of meteoric water is slightly larger than the isotopic gradient of precipitation over the area today (Fig. 7; Ingraham and Taylor, 1991; Friedman et al., 2002a), and occurs over less distance (postextension) from west to east than a similar magnitude isotopic change in 5.5–2 Ma old hydrated volcanic glass (Mulch et al., 2008). This decrease, however, does not necessitate orographic fractionation, as discussed in the Paleotopography section below.

### DISCUSSION

**Fluvial System Evolution**

The large-scale evolution of the Eocene Sierra Nevada fluvial system is a coupling of warm, wet climate contributing high sediment supply as fluvial valleys backfilled, with a larger tectonically controlled shift from an incisional to a locally aggradational system (Cassel and Graham, 2011). The onset of aggradation roughly correlates with the end of the Laramide orogeny (Bird, 1998; Sigloch et al., 2008) and with the global shift to a warmer, wetter climate.
Early Cenozoic topography, morphology, and tectonics of the northern Sierra Nevada and western Basin and Range

in the Eocene (Miller, 1992; Zachos et al., 2001). Changes in sea level may have exerted some control on deposition as well. Maximum sea level (70–170 m above present) occurred in the early Eocene, increasing from the Paleocene (30–50 m above present) and subsequently falling through the Oligocene (Van Sickel et al., 2004; Miller et al., 2005), and may have triggered paleovalley backfilling, similar to Eocene units in the Great Valley succession (e.g., Dickinson et al., 1979; Johnson et al., 2007). Although Sierra arc volcanism stopped in the Late Cretaceous (e.g., Ducea, 2001), Laramide compression and rock uplift may have continued in the area of the Sierra Nevada (Moxon and Graham, 1987) and Basin and Range (cf. Humphreys et al., 2003; DeCelles, 2004; Humphreys, 2009), sustaining steep upstream river gradients through the Eocene (Henry, 2008; Cassel et al., 2009a; Cassel and Graham, 2011). Early-middle Eocene topographic uplift, driven by removal of the Farallon slab and the end of the Laramide, has been proposed for the western U.S. (Jones et al., 2011) and NevadaPlano region (Mix et al., 2010). Regional rock uplift, whether sustained throughout the early Cenozoic or a product of end-Laramide deep mantle processes, could have driven relief formation and paleovalley incision, and when uplift slowed or halted, rivers would drive toward equilibrium by propagating knickpoints upstream and aggrading in low elevation areas (Burbank et al., 1996; Whipple et al., 1999; Heller et al., 2001).

An overall upward-fining succession suggests that the locus of fluvial deposition shifted and depositional energy waned over time as the system backfilled successive broad paleovalleys from west to east (Cassel and Graham, 2011). This is consistent with paleoflora age estimates and detrital zircon grain age distributions throughout the basin (Wing and Greenwood, 1993; Wilf et al., 1998; Cassel et al., 2012). Based on sedimentary characteristics, combined with floral and detrital zircon data, braided stream deposition within the basin occurred from late early Eocene through late Eocene, and aggradation primarily occurred in broad, lower gradient reaches behind a major bedrock knickpoint, created in the late Mesozoic–early Cenozoic, possibly due to asymmetric batholith unroofing (Cassel and Graham, 2011). The youngest single-grain detrital zircon ages are similar to the range of ages of Eocene volcanic rocks in central Nevada (Cassel et al., 2012). The presence of these grains suggests that paleoriver headwaters extended into central Nevada by late Eocene. This provides support for the existence
of a high elevation, westward draining plateau at the latitude of Nevada (“Nevadaplan”) from Eocene through Oligocene time.

Paleogeomorphic Reconstructions

With a better understanding of Paleogene fluvial system evolution in the northern Sierra Nevada, we question the assumptions required in the method of comparing reach gradients to estimate the amount of uplift at the crest of the range (e.g., Lindgren, 1911) that is often cited as evidence for uplift of 1–2 km in the past 5 My (e.g., Wakabayashi and Sawyer, 2001; Jones et al., 2004; Molnar, 2010). The locations of the bedrock paleovalleys surfaces—stratigraphic unconformities and, where not exposed, the lowermost exposures of Eocene fluvial sediments—do not all represent points on the same Eocene paleoriver channel profile. Multiple broad paleovalley strath terraces suggest multiple episodes of valley widening and channel adjustment, followed by aggradation. Changes through the vertical sequence, paleovalley morphology, and spatial variability between locations show the migration of the locus of deposition across the depositional area and diachronous deposition at individual locations. Paleovalley locations are controlled by bedrock structure and topography; paleorivers did not have equilibrium river profiles that can be extrapolated outside of the depositional area, and individual Eocene fluvial deposits and reaches are not necessarily time-correlative (Cassel and Graham, 2011). Previous paleovalley reconstructions may be inaccurate because paleovalleys represent an amalgamation of multiple incisional events and not a topographic surface or time line (Strong and Paola, 2008). Thus measurements between individual incised channel reaches would either produce incorrect estimates of the paleochannel gradient (i.e., younger to older channel would overestimate the overall gradient), or may not represent segments of the same river.

The lack of equilibrium river profiles and variation in timing of incision and deposition throughout the Eocene also highlights problems with the interpretation that there was no incision beneath channel thalwegs from Eocene to Pliocene time and that all modern valley incision topographically below Miocene–Pliocene volcanic rocks represents late Cenozoic basement incision (Wakabayashi and Sawyer, 2001). Incision upstream of local depositional areas likely continued throughout the Eocene and Oligocene, as suggested by the provenance of Eocene sediments (Yeend, 1974; Cassel et al., 2012; Cecil et al., 2010; Cassel and Graham, 2011), which are dominated by local metamorphic and granitic bedrock, the evidence of paleovalley widening and significant cut and fill stratigraphy evident in Eocene sedimentary sections (Cassel and Graham, 2011), and the continued deposition and reworking of locally sourced sediment into the Oligocene (Cassel et al., 2009a; Cassel and Graham, 2011). Incision and headward erosion of drainages across the high-elevation region in what is now Nevada may have occurred through the late Eocene–Oligocene (Henry, 2008; Cassel et al., 2012; Henry and Faulds, 2010). Subsequently, Miocene volcanoclastic deposits blanketed much of the Sierra landscape, filling in paleochannels and smoothing relief, thus incision through these units likely resulted at least in part as a response to constructional topography (Wakabayashi and Sawyer, 2001; Busby et al., 2008a). These deposits themselves, with the exception of a few distinct basalt flows, compose immeasurable individual debris flow events with varying patterns of erosion and deposition, along with landslide and pyroclastic units at higher elevations (Wagner et al., 2000; Busby et al., 2008b). It may be difficult in some locations, such as across the South or Middle Forks of the Yuba River, to determine the original sub-Miocene surface, especially where the Miocene deposits have been removed. Thus estimates of the amount of basement incision may be invalid where rivers, returning to their paleovalleys, may have removed the record of lower elevation Miocene deposition.

The fluvial system also provided sediment to the early Cenozoic forearc basin—sea level extended to what is now the eastern edge of the Great Valley (Dickinson et al., 1979; Creely and Force, 2007; Cassel and Graham, 2011). Wakabayashi and Sawyer (2001) used the amount of sediment deposited within the Great Valley to calculate total sediment accumulation rates. Although this calculation shows significant Pliocene increases in sediment deposited within the basin, the method ignores the well-documented deep-water depositional system present throughout the early Cenozoic, in which submarine channels served as conduits for sediment to move from the forearc basin into the trench beyond (Dickinson et al., 1979; Moxon and Graham, 1987; Williams et al., 1998).
sediment deposited within the fluvial system or in the deep ocean is not fully taken into account when calculating early Cenozoic accumulation rates from Great Valley forearc basin sediments; sediments deposited upstream and in the accretionary wedge must also be included in a mass balance calculation.

Deposition of these fluvial deposits does not necessitate a low-elevation or overall low-gradient stream system (as suggested by Christensen, 1966; Huber, 1981). In fact, the coarse sediment supplied, transported, and deposited within the system requires flows with high shear stress and moderately steep channel slopes (cf. Bridge, 1993). High local relief upstream of the basin likely supplied the cobble–boulder–size clasts to the system, and moderate slopes within the basin would allow for movement of large clasts under flood conditions (Cassel and Graham, 2011). The diachronous, localized nature of paleovalley incision and braided stream aggradation within the northern Sierra Nevada invalidates previous estimates of the timing and amount of range uplift based on paleochannel reconstructions and gradients. We do not find direct evidence for significant increases in surface elevations post-Eocene across the range, although a limited amount of rock uplift balanced with exhumation (e.g., Small and Anderson, 1995) may have occurred since the Eocene, as post-Eocene knickpoint propagation and the post-Miocene response to constructional topography may not account for the total incision to Holocene valley bottoms.

Alternatively, the westward movement of the regional drainage divide as Basin and Range extension progressed (Henry, 2008; Cassel et al., 2009a), the western migration of the shoreline (Creely and Force, 2007), the end of Laramide tectonism and uplift (Bird, 1998; Jones et al., 2011), and the deposition and incision generated from Miocene volcanism (Busby et al., 2008a; Busby et al., 2008b) shifted the river profile westward and dropped base level, which may have driven the post-Eocene valley incision. Both internal and external drivers influenced the evolution of the fluvial system through (1) the pattern of basin filling, as the locus of deposition migrated east to west, (2) the warm, wet Eocene climate, which increased chemical weathering and contributed a high sediment supply, and (3) changes in base level due to Laramide tectonism and, possibly, eustasy. End-Laramide surface uplift may also account for the influx of late Eocene Nevada-sourced detrital sediments to Sierra Nevada sediments in eastern outcrop locations (Cassel et al., 2012). Slowing of regional rock uplift would have driven fluvial aggradation in previously incised valleys (Whipple et al., 1999; Heller et al., 2001).

Oligocene Drainage Divide

The presence of correlative ignimbrites on either side of the modern crest of the Sierra Nevada shows that the Oligocene regional drainage divide was located to the east of its present position, at or east of ignimbrite source calderas (Fig. 6). These results also require that source calderas were located in a region of similar or higher elevations than the drainages in which these ignimbrites were deposited, as ignimbrites can overtop at most ~600 m of topography (Fisher and Schmincke, 1984). Using the range of extension estimates for each region (Seedorf, 1991; Smith et al., 1991; Surpless et al., 2002; Colgan et al., 2006; Van Buer et al., 2009), Cassel et al. (2009a) estimated that the tuffs of Axehandle Canyon, Rattlesnake Canyon, and Campbell Creek traveled 200–250 km from their source calderas (Fig. 6). Based on the texture and lack of bedding, reworking, and sorting of ignimbrites throughout the Sierra Nevada, ignimbrites were deposited up to 180 km from source calderas before their travel was significantly affected by fluvial transport (Cassel et al., 2009a; Henry and Faulds, 2010). These ignimbrites likely flowed downhill from a region of high topography that included what is now the Sierra Nevada and western-central Nevada (Nevadaplano). These studies showed that the regional drainage divide, which now resides west of the California-Nevada border, was located in central-eastern Nevada in the Oligocene, at least 200 km east of its present position, when subsequent Basin and Range extension is removed (Fig. 6; Cassel et al., 2009a).

Paleotopography of the Sierra Nevada and Western Basin and Range

Cassel et al. (2009b) reported that the increase in δD across the northern Sierra Nevada reflected an increase in mean elevation from west to east at 31–28 Ma (Fig. 7). Their results showed that the northern Sierra Nevada had a steep western gradient in the Oligocene, comparable to that of the modern range. Estimates of the Oligocene paleoelevation of the northern Sierra Nevada, based on modern empirical and Oligocene modeled estimates of the isotopic lapse rate (the average change in δD with elevation; Fig. 8), are 2800 m ± 900 m and 3200 m ± 1100/~2000 m, respectively (Cassel et al., 2009b). These results are consistent with stable isotope paleoaltimetry studies from the Eocene (Mulch et al., 2006; Hren et al., 2010) and Miocene (e.g., Cham-

![Figure 8](https://pubs.geoscienceworld.org/gsa/geosphere/article-pdf/8/2/229/3341478/229.pdf)

**Figure 8.** Topographic profile comparison of present day and Oligocene elevations along the Yuba River drainage in what is now the northern Sierra Nevada and western Basin and Range, based on stable isotope paleoaltimetry results. Red line—present-day lapse rate (PDLR); orange line—Oligocene modeled (OMLR); dashed lines—±2σ error bars. Thick dashed lines from 100 to 200 km show possible elevation profiles across Nevada, dependent on drainage hydrology. Location of ignimbrite source calderas noted, assuming maximum extension as shown in Figure 7. Present-day locations of geologic features and faults noted in gray below graph. Modified from Cassel et al. (2009b).
The decrease in slope of δD at ~100 km along the transect likely reflects a significant reduction in the rate of increase in paleoelevation (Figs. 7–9). The change in Oligocene δD across Nevada may reflect a gradual increase in mean elevation across that area, or could be attributed to partially closed hydrologic system behavior, where precipitation is partitioned between runoff and evapotranspiration (Craig, 1961; Ingraham and Taylor, 1991; Ingraham and Craig, 1993; Poage and Chamberlain, 2001). The modern Basin and Range is largely internally drained, and δD of precipitation and meteoric waters shows little variation across the area, displaying closed system hydrologic behavior wherein precipitation and evapotranspiration are in equilibrium (Ingraham and Taylor, 1991; Ingraham and Craig, 1993; Friedman et al., 2002b). The decrease in δD of volcanic glass could be the effect of a somewhat different hydrologic regime in which some fraction of precipitation leaves the system as runoff, leading to gradual fractionation of δD with distance. This could have occurred across an area of low relief and high mean elevations, or across a lower-elevation area in the rain shadow of a paleo–Sierra Nevada (Fig. 9), similar to the Miocene to modern Sierra Nevada (Poage and Chamberlain, 2002; Mulch et al., 2008). Evidence of west-flowing late Eocene–Oligocene drainages extending from what is now central Nevada to the Great Valley (e.g., Faulds et al., 2005; Garside et al., 2005; Brooks et al., 2008; Cassel et al., 2009a; Henry and Faulds, 2010) suggests the former hypothesis is correct.

Samples located at ~150 km along the transect have similar or higher δD values than samples at 100 km, with the exception of one outlier (Fig. 7). This may reflect a hydrologic or topographic configuration different from the modern, as δD values of modern meteoric waters steadily decrease from 0 to 140 km along the transect (Fig. 7). The addition of a less fractionated moisture source at the transition from steep to gradually increasing or flat topography, as indicated by the change in slope of δD versus distance (Fig. 7), could account for the increase in δD. Small elevation changes (<800 m) and local landscape features in areas behind an orographic barrier can be obscured by additional moisture sources not controlled by elevation, and this effect is common on high elevation plateaus (cf. Blisniuk and Stern, 2005; Tian et al., 2007; Mix et al., 2010). Additionally, mean surface temperature and the amount of below-cloud evaporation may have differed between locations west of 100 km and those to the east, and increases in either to the east of 100 km may have contributed to the variation in isotopic compositions along the transect (Kendall and Coplen, 2001; Poage and Chamberlain, 2001; Blisniuk and Stern, 2005).

These differences in the change in δD along the sample transect between modern meteoric water δD and Oligocene volcanic glass δD provide support for the method of using volcanic glass as a stable isotope proxy material. Although Cenozoic volcanic glass has been used as a proxy for paleoprecipitation in a number of isotopic studies, few have thoroughly studied the reliability of this approach (e.g., Friedman et al., 1993a; Shane and Ingraham, 2002; Mulch et al., 2008). Meteoric waters in Oligocene volcanic glass display an increase in δD at 100–150 km that is not present in the trend of δD of modern meteoric waters: the amount of offset from modern values is significantly different to the east of the slope break at ~100 km (Fig. 7). These findings provide additional support for the conclusions of Friedman et al. (1993a) and others that volcanic glass preserves the δD of ancient meteoric waters and does not exchange further once the glass is saturated (within 3–5 ka; Shane and Ingraham, 2002).

Differences in the relative influence of air masses across the Basin and Range are likely driving the increase in δD at 100 km, and may help to account for the spread in data along the eastern part of the transect (Fig. 7). Unlike the Sierra Nevada, the northern Basin and Range currently receives precipitation derived from source waters from distinct regions to the north, west, south, and east of the area (Friedman et al., 2002a). The modern isotopic compositions of these precipitation sources vary up to 54‰ cumulative δD, although 92% of the total precipitation over the northern Basin and Range currently comes from air masses originating from the Pacific (Friedman et al., 2002a). Differences in Oligocene temperatures or topography, however, both across the sample area and over the air-mass trajectories, could have affected both air mass isotopic compositions and the relative influence of each on precipitation (cf. Kendall and Coplen, 2001; Friedman et al., 2002a; Horton et al., 2004; Blisniuk and Stern, 2005; Mix et al., 2010). Using flora collected from mid-latitudes of western North America, Wolfe et al. (1998), however, estimated cooler temperatures than Eocene and Oligocene global averages in the region that is now the Basin and Range, which they attribute to high altitudes across the region. Less-depleted air masses from the south or east may have had a stronger influence on isotopic composition of precipitation. This could have been driven by differences in the topographic configuration of the region in the Oligocene and may account for the significant difference between the Oligocene and the modern isotopic trends.

CONCLUSIONS

As the participants of the Sierra Nevada Penrose Conference presented and discussed, significant questions and controversies remain regarding the topographic and geomorphic history of the Sierra Nevada. A number of recent publications (e.g., Bennett et al., 2009, Earth and Planetary Science Letters; Cassel et al., 2009b, Geology; Hren et al., 2010, Geology; Molnar, 2010, Geological Society of America Bulletin; and the International Geology Review Special Volume: Rise and Fall of the Nevadaplano, 2009) and conference sessions (Cordilleran Section GSA 2010: Sierra Nevada Microplate: Basement and Basins; Penrose Conference 2010: Origin and Uplift of the Sierra Nevada, California, USA) highlight the continued interest in the tectonic, climatic, and topographic history of the Sierra Nevada and its connection to the Basin and Range province of Nevada and western Utah. Many recent studies have cited previous geomorphic evidence from the northern part of the range as the primary argument for late Miocene to Pliocene tilting and uplift of 1–2 km at the Sierran crest (see Jones et al., 2004 or Molnar, 2010). The existing geomorphic and sedimentologic studies (e.g., Whitney, 1880; Lindgren, 1911; Christensen, 1966; Yeend, 1974; Huber, 1990; Wakabayashi and Sawyer, 2001), however, have not provided sufficient modern sedimentologic and topographic analyses to interpret the complex record of Eocene–Oligocene fluvial and volcanic deposits. Through our detailed research into the tectonic processes and geomorphic evolution of the Eocene–Oligocene in the area that is now the Sierra Nevada and western Basin and Range, we have gained a better understanding of the history of surface uplift, denudation, and landscape change.

A comparison of data and results from these multiple studies provides us with a reconstruction of Eocene–Oligocene landscape morphology in the northern Sierra Nevada and western Basin and Range, and can help to address questions of the deeper mantle processes that drove the formation of that region through the Cenozoic. Based on the large time span of ages from flora, correlative strata, and youngest detrital zircon grains found within Eocene sediments, fluvial deposition within the braided stream system continued from late early Eocene through the late Eocene, and the system remained active into the Oligocene when overlying volcanic and volcaniclastic sediments were deposited.
Figure 9. Proposed landscape reconstructions of the area that is now the northern Sierra Nevada and western Basin and Range, based on the results of this study and previous research. (A) Middle to late Eocene landscape shows rivers incising in high-elevation areas in steep, high-gradient paleovalleys, feeding sediment into braided streams aggrading in broad, lower-gradient paleovalleys in the west, and headward erosion extending drainages across a low-relief, high-elevation plateau to the east. (B) Oligocene landscape: ignimbrites sourced from calderas ~200–250 km to the east (now central Nevada) travel and deposit in paleovalleys in northern Sierra Nevada. Approximate location of southern end of the Quaternary Mohawk Valley fault, based on Eocene outcrop locations, shown in gray for reference. Elevation estimates and landscape reconstructions based on this study, Henry (2008), Cassel et al. (2009a, 2009b), Henry and Faulds (2010), and Cassel and Graham (2011).
in paleovalleys (Fig. 9). This long period of aggradation was a consequence of a high sediment supply, driven by both the warm, very wet Eocene climate and high channel relief within the upper part of the drainage basin. The diachronous, localized nature of paleovalley incision and braided stream aggradation within the northern Sierra Nevada invalidates estimates of the timing and amount of range uplift based on paleochannel reconstructions and gradients. Aggradation in the system was likely triggered by a slowdown or halt in rock uplift, although the age of basal sediments is unknown.

Trace and rare earth element compositions in volcanic glass and 40Ar/39Ar radiometric ages of the age of basal sediments is unknown. Problems with paleovalley reconstructions and gradients.

Rhyolitic, localized nature of paleovalley incision and braided stream aggradation within the northern Sierra Nevada invalidates estimates of the timing and amount of range uplift based on paleochannel reconstructions and gradients. Aggradation in the system was likely triggered by a slowdown or halt in rock uplift, although the age of basal sediments is unknown.

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