Late Cenozoic paleogeographic evolution of northeastern Nevada: Evidence from the sedimentary basins

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ABSTRACT

Field and geochronologic studies of Neogene sedimentary basins in northeastern Nevada document the paleogeographic and geologic evolution of this region and the effects on major mineral deposits. The broad area that includes the four middle Miocene basins studied—Chimney, Ivanhoe, Carlin, and Elko, from west to east—was an upland that underwent prolonged middle Tertiary exposure and moderate erosion. All four basins began to retain sediments at ca. 16 Ma. Eruption of volcanic flows in the Chimney and Ivanhoe basins produced short-lived (ca. 2 Ma), lacustrine-dominated basins before the dams failed and the streams drained to the southwest. In contrast, early, high-angle, normal faulting induced fluvial to lacustrine sedimentation in the Carlin and Elko basins, and volcanic flows further blocked drainage in the Carlin basin until the basin drained at ca. 14.5 Ma. The Elko basin, with continued synsedimentary faulting, retained sediments until ca. 9.8 Ma and then drained west into the Carlin basin. Sediment buildup in all basins progressively buried existing highlands and created a subdued landscape.

Relatively minor post-sedimentation extension produced early north-northwest–striking normal faults with variable amounts of offset, and later east-northeast–striking normal faults with up to several kilometers of vertical and left-lateral offset. The earlier faults are more pronounced east of the Tuscarora Mountains, possibly reflecting a hanging-wall influence related to uplift of the Ruby Mountains-East Humboldt core complex on the east side of the Elko basin. The later faults are concentrated along the north-northwest–trending northern Nevada rift west of the Tuscarora Mountains. The area west of the rift contains major tilted horsts and alluvium-filled grabens, and differential extension between this more highly extended region and the less extended area to the east produced the intervening east-northeast–striking faults.

The Humboldt River drainage system formed as the four basins became integrated after ca. 9.8 Ma. Flow was into northwestern Nevada, the site of active normal faulting and graben formation. This faulting lowered the base level of the river and induced substantial erosion in upstream regions. Erosion preferentially removed the poorly consolidated Miocene sediments, progressively reexposed the pre-middle Miocene highlands, and transported the sediments to downstream basins. Thus, some ranges in the upstream region are exhumed older highlands rather than newly formed horsts. In addition, the drainage system evolution indicates that northern Nevada has become progressively lower than central Nevada since the middle Miocene.

Mineral belts with large Eocene gold deposits are exposed in uplands and concealed beneath Neogene basin units in the study area. Also, numerous epithermal hot-spring deposits formed at and near the paleosurface in the Chimney, Ivanhoe, and Carlin basins as those basins were forming. The Neogene geologic and landscape evolution had variable effects on all of these deposits, including uplift, weathering, supergene enrichment, erosion, and burial, depending on the events at any particular deposit. As such, this study documents the importance of evaluating post-mineralization processes at both regional and local scales when exploring for or evaluating the diverse mineral deposits in this area and other parts of the Basin and Range region.

Keywords: sedimentary basins, tectonics, geomorphology, Nevada, Miocene, Pliocene, gold, Humboldt River.

INTRODUCTION

Any physiographic map of northern Nevada clearly shows numerous elongate mountain ranges separated by sedimentary basins, perfect examples of the Basin and Range physiographic province. Quaternary alluvium derived from the ranges blankets many of the basins, and Pliocene and Miocene sediments comprise part of the underlying Tertiary sedimentary sequence (Stewart and Carlson, 1978; Effimoff and Pinezich, 1981; Gordon and Heller, 1993; Hess, 2004; Wallace, 2005). Historically, the formation of these basin-range pairs has been attributed to crustal extension and faulting that began in the middle Miocene and has continued to the

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Late Cenozoic paleogeography, northeastern Nevada

Until recently, however, little work has focused on the actual ages of the horsts and grabens. Recent field and thermochronologic studies have demonstrated that the ages of basin-range pairs in this area are diverse. The Ruby Mountains (Fig. 1) experienced major uplift at ca. 15–14 Ma (Colgan and Metcalf, 2006), other ranges did not begin to form until ca. 10 Ma (Wallace, 1991, 2005; Colgan et al., 2004, 2006), and some ranges, such as the Adobe Range, may be relics of early Tertiary uplands (Haynes, 2003; Wallace, 2005). In addition, some faulting events did not produce major uplift (Gordon and Heller, 1993; Wallace, 1993, 2005; Colgan et al., 2008).

These relations indicate that the history of “basin-range” tectonics is not straightforward. To begin to understand the late Cenozoic history of the area, this study has examined the Neogene geology, with a focus on the sedimentary basins, along a 200-km-long transect that extends across north-central Nevada from the Santa Rosa Range east to the western base of the Ruby Mountains (Fig. 1). The primary goal of the study was to use the facies relations and ages of the sedimentary units to define the evolution of the late Cenozoic paleogeography and the relative effects of faulting, uplift, sedimentation, and erosion during long-term landscape evolution. While this paper does not attempt to explain the broader Basin and Range region, it does provide process-oriented information that may be applicable to the region, as a whole, and extensional terrains, in general.

The study area includes world-class, Paleogene and older mineral deposits in the basement rocks and Miocene epithermal epiments in or adjacent to the Miocene basins (Fig. 1). Consequently, northern Nevada is the third-largest producer of gold in the world. An important goal of the study was to determine the effects of late Cenozoic landscape evolution on these mineral deposits, with implications for the formation and modification of the deposits and deposit- to regional-scale mineral exploration and assessment.

The study area includes four middle and late Miocene sedimentary basins. From west to east, they include what are referred to in this paper as the Chimney, Ivanhoe, Carlin, and Elko basins, which were actively receiving sediments between ca. 16.5 and 9.8 Ma (Figs. 1 and 2). All of the Miocene strata in each basin were examined on at least a reconnaissance basis, and appropriate samples of the sedimentary and coeval volcanic rocks were collected and dated to provide a time-stratigraphic framework. In addition, the Neogene sedimentary units in Pine and Independence Valleys (Fig. 1) were studied and dated on a reconnaissance basis.

This paper describes the geology and histories of the basins from the early Miocene to the present, including their paleogeographic setting, sedimentation, faulting, erosion, and any coeval volcanism. Later sections of the paper discuss the similarities and differences between the basins, leading to a synthesis of the late Cenozoic evolution of the study area as a whole, its effect on mineral deposits, and implications for the more regional paleogeographic evolution. Each major section begins with a short summary of the details presented in the ensuing descriptions.

In all of the basins, the two principal components of the sediments are air-fall ash and pumice that were derived from distal to, less commonly, local eruptions, and materials that were eroded from pre-Miocene bedrock exposures in uplands that surrounded the basins. The bedrock-derived materials are generically referred to as epiclastic sediments to highlight their more

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Figure 1. Locations of the Miocene Elko, Carlin, Ivanhoe, Chimney, and adjacent sedimentary basins, shown with the light blue color, and major geographic features in northeastern Nevada. Areas with the “V” pattern are underlain by middle Miocene rhyolitic and basaltic volcanic rocks; JR—Jarbidge Rhyolite; NNR—northern Nevada rift; SR-C—Santa Rosa–Calico volcanic field. The dotted yellow lines indicate the major late Eocene gold-deposit trends; BM-E—Battle Mountain–Eureka; C—Carlin; G—Getchell; J—Jerritt Canyon. The solid red circles are the locations of middle Miocene epithermal systems that were active at the same time as the sedimentary basins; other epithermal systems beyond the limits of this study area are not shown. Abbreviated geographic locations: B—Beowawe; EH—Elko Hills; LT—Lone Tree mine; MC—Mule Canyon; MM—Marys Mountain; P—Pebble mine; PH—Peko Hills; PV—Paradise Valley; R—Rain mine; TC—Twin Creeks mine; R-EH—Ruby Mountains–East Humboldt Range detachment and high-angle faults. Inset map shows the location of the study area (gray); W—Winnemucca; PF—Pine Forest Range.
local derivation in comparison to the air-fall materials. Air-fall deposits that landed on the uplands were redistributed during erosion and stream transport and were intermixed with epiclastic materials at the depositional sites. Both sediment components were deposited in fluvial and lacustrine environments, and the distinctions between air-fall and bedrock-derived sediments in the strata are described in the text.

REGIONAL GEOLOGIC SETTING

A lithologically and structurally complex assemblage of pre-Tertiary accreted terranes, sedimentary sequences, and plutons, a discontinuous cap of Paleogene sedimentary and volcanic rocks, and Tertiary plutons underlie the Neogene and Quaternary units along the transect. Most of the pre-Neogene history does not pertain to the Neogene story, but some Paleogene events and features did lead into and (or) pertain to the Neogene story, but some Paleogene events and features did lead into and (or) pertain to the Neogene story.

In the Eocene, moderate extension produced modest-relief uplands and lowlands. Broad, shallow lakes covered some lowlands, and basin-filling sediments flanked the uplands, including the southern Tuscaraora Mountains, Adobe Range, East Humboldt Range, and Ruby Mountains (Haynes, 2003). Late Eocene flows and ash-flow tuffs were erupted onto and across this still-subdued landscape, in places filling east- and west-draining paleovalleys (Henry and Ressell, 2000; Henry, 2008). Virtually all of this igneous activity had ceased by ca. 38 Ma in northeastern Nevada (Haynes, 2003). Major gold deposits, some of which formed near coeval late Eocene igneous centers, are hosted by Paleozoic rocks and comprise the northwest-trending Carlin and Battle Mountain-Eureka mineral-deposit trends, the north-trending Jerritt Canyon trend, and the north-north-east-trending Getchell trend (Fig. 1; Cline et al., 2005). The tops of the central Carlin trend deposits were ~500–1500 m below the late Eocene paleosurface (Haynes, 2003; Ressell and Henry, 2006), and the younger porphyry-related systems near Battle Mountain were a few hundred meters below the paleosurface at that location (Fig. 1; Theodore and Blake, 1975).

Middle Tertiary extension affected much of northern Nevada, although the amounts varied from minimal (Colgan et al., 2008) to at least 50 percent (Muntean et al., 2001). In most places, the age of extension can only be constrained to between the late Eocene and middle Miocene (Smith and Ketner, 1976; Smith and Howard, 1977; Wallace, 1993, 2003c). At Mule Canyon (Fig. 1), tilting took place between ca. 34 and 16 Ma (John et al., 2003), and, at Marys Mountain near Carlin (Fig. 1), tilting occurred both before and after the emplacement of a 25 Ma welded tuff and before the deposition of 15.3 Ma rhyolite flows (Henry and Faulds, 1999). In the southern Tuscaraora Mountains, more deeply formed plutonic rocks were exposed by 16.5 Ma and shed clasts into the western part of the Carlin basin. Supergene alunite dates from gold deposits along the Carlin and Getchell trends indicate that there was enough uplift and erosion to erode and weather the deposits between 30 and 18 Ma (Hofstra et al., 1999; Cline et al., 2005).

As extension relaxed the crust, over-thickened parts of the crust became more buoyant. This allowed deep-seated metamorphic rocks to ascend to shallower levels in the Ruby Mountains and East Humboldt Range and produced a major, west-northwest-dipping detachment fault along the west flanks of those ranges. Displacement along this fault may have been more than 50 km (Howard, 2003). The thermal, metamorphic, and igneous history of this complex dates back to before the Eocene, and major uplift related to high-angle faulting peaked between 14 and 15 Ma (Dokka et al., 1986; Snoucke et al., 1997; Howard, 2003; Colgan and Metcalf, 2006). Uplift has continued into the Holocene (Wesnousky and Willoughby, 2003).

Much more subdued extension stretched the crust by ~10% between the middle and late Miocene (Muntean et al., 2001), although some areas just to the south were extended more than 100% (Colgan et al., 2008). This extension generally was to the west, and pre-Miocene basement structures undoubtedly affected the apparent local extension direction. For instance, north-northwest-striking dikes along the northern Nevada rift (Fig. 1; see ensuing section) suggest a west-southwest extension direction (Zoback et al., 1994), whereas reconstructed offsets for a large number of Miocene fault blocks in north-eastern Nevada indicate a generally northwest but locally highly variable extension direction (Muntean et al., 2001).

Extension and the ascent of the Yellowstone mantle plume into the crust along the Oregon-Nevada border induced widespread bimodal volcanism starting at ca. 16.5 Ma (Christiansen et al.,

![Regional Geologic Setting Diagram](image-url)
In northern Nevada, some of the early mafic magmas ascended and erupted along generally north-northwest–striking crustal breaks, such as the northern Nevada rift and related mafic dike complexes (John et al., 2000; Pierce et al., 2002; Ponce and Glen, 2002). Partial melting of the crust, coupled with magma mixing, and fractional crystallization, created more widespread rhyolitic volcanism (Coats, 1987; John et al., 2000; Brueseke and Hart, 2007). The heat from this magmatism, coupled with conduits provided by the faulting and water from the wet climate and lakes, generated a number of small to very large, epithermal gold + silver ± mercury deposits. Most of these deposits formed near the interface between volcanic centers and lacustrine basins (Fig. 1; John, 2001; Wallace et al., 2004a, 2004b).

Extension continued after 10 Ma and led to much of the fault-bounded horst-and-graben physiography of northeastern Nevada (Stewart, 1998). Uplift rates may have been greater in the late Neogene than in the Quaternary (Colgan et al., 2004; Personius and Mahan, 2005). Total extension across northern Nevada was not great, and the formation of the horst-graben pairs was more vertical than in parts of southern and western Nevada, where the horsts are highly tilted fault blocks that formed during significant extension (Anderson, 1971; Proffett, 1977).

Miocene Climate

As noted by Smith (1994), climate can have equal or greater influences on basin sedimentation than do other geologic processes and events. Precipitation causes erosion, creates channels and lakes, and floods, followed by felsic calderas, domes, and flows. The greatest amount of volcanic activity was focused along the north-northwest–trending northern Nevada rift and the related Santa Rosa-Calico volcanic field to the northwest (Fig. 1). Domes and flows of the Jarbidge rhyolite were erupted across the northeastern tier of the area as the mantle plume migrated to the east-northeast, in some areas producing extensive and thick masses of coarsely porphyritic rhyolite with a minor amount of underlying basalt. Volcanic activity occurred at the same time that the sedimentary basins described in this paper were forming and expanding, and, in many areas, volcanism and sedimentation overlapped in both time and space.

Northern Nevada Rift

The north-northwest–trending northern Nevada rift extends from east-central Nevada to near the Oregon border (Fig. 1), a distance of ~500 km. The rift originally was identified by a pronounced positive aeromagnetic anomaly related to a narrow (~7 km wide), deep-seated middle Miocene mafic dike complex that was intruded along a preexisting crustal structure (Zoback and Thompson, 1978; Hildenbrand et al., 2000). Intense bimodal volcanism occurred along the rift and adjacent areas from ca. 16.0 to 14.9 Ma (John et al., 2000; Leavitt et al., 2004). In general, thick mafic flow sequences formed early, followed in many, but not all, areas by the eruption of more viscous dacite to rhyolite flows and domes. The very similar and likely related Santa Rosa–Calico volcanic field is just west of the magnetic anomaly near the Oregon-Nevada border. This field was active from ca. 16.7 to 14 Ma (Brueseke and Hart, 2007), and volcanic units related to both systems overlap in the area of the Chimney basin. Volcanism along the northern Nevada rift was more extensive, especially to the east, than the aeromagnetic anomaly, indicating that the deeper crustal processes related to magma genesis were not confined to just the mafic dike complex.

Normal faulting along the rift occurred during the entire period of volcanism. Some of this faulting was related to the “rift,” and some likely was related to regional extension. In the northern Shoshone Range and at the latitude of Midas, offset along north-northwest–striking faults along and near the locus of volcanism produced a central downdropped zone. This zone is 14–19 km wide in the northern Shoshone Range and ~30 km wide near Midas. Displacement along the faults generally was about a kilometer, with little to no fault-related tilting of the Miocene volcanic units. Overall, the total amount of extension across the structural zone was no more than a few kilometers (John et al., 2000). Rift-related volcanism produced more than 1 km of volcanic rocks in 300,000 yr in the southwestern Sheep Creek Range (John et al., 2000) and at least that much over 500,000 yr in the central Snowstorm Mountains (Wallace, 1993; Leavitt et al., 2004; this study). With as much as 1 km of synvolcanic downdropping, the relief of the volcanic assemblage produced along the rift may have been slight. Specific details on the relation between volcanism, faulting, and sedimentation in the Snowstorm Mountains and the adjacent Chimney and Ivanhoe basins are presented in later sections of this paper.

Jarbidge Rhyolite

The Jarbidge rhyolite loosely includes coarsely porphyritic flows and domes, with some related tuffaceous units (Coats, 1987). The type area is in the Jarbidge area of northeastern Nevada (Fig. 1), and similar rhyolites are scattered throughout northeasternmost Nevada and as far east as the Utah border. Because of the highly viscous nature of the magmas and mode of emplacement, flow across the landscape probably was not great, and the distribution of the rhyolite likely reflects the distribution of the eruptive centers. Miocene basin sediments locally overlie, underlie, or encase rhyolite flows. North of the Independence Mountains, the rhyolite overlies 16.6 Ma basaltic andesite flows (Rahl et al., 2002), and some nearby rhyolite domes were erupted at ca. 15.8–16 Ma (Coats, 1987; C. Henry, 2004, written commun.). Jarbidge-like flows and dikes just southwest of Wells in the East Humboldt Range were emplaced between 14.8 and 13.4 Ma (Snake et al., 1997), and some data suggest that rhyolitic volcanism at Jarbidge, the type area, occurred at ca. 14 Ma (Bernt, 1998).

Rhyolite flows were erupted at ca. 15.3 Ma west and south of Carlin (Fig. 1) to form the Pali- sade Canyon rhyolite. These volcanic units have an uncertain relation to both the northern Nevada rift and Jarbidge systems, although their mineralogies and chemistries suggest an origin similar to rhyolites in those systems. This unit is described in more detail in the Carlin basin section.
Post-Basin Volcanic Units

Extensive sheets of Miocene rhyolite and basalt flows underlie much of the Owyhee Plateau in northernmost Nevada and are related to the Yellowstone mantle plume along the Snake River Plain to the north (Fig. 1; Wood and Clemens, 2002). These volcanic units were erupted between ca. 15.4 and 8 Ma, and they conceal virtually all underlying Miocene and older units, including the northern end of the northern Nevada rift and any basin sediments that may be present. These volcanic units are largely unaltered and retain their original low to horizontal dips.

GEOCHRONOLOGY

The geochronology used for this study included both new and previously published data obtained by several methods. The new data were obtained from $^{40}Ar/^{39}Ar$ dates on mineral separates from ash beds and volcanic rocks or, in the case of aphanitic mafic volcanic rocks, on whole rocks or groundmass separates, and from tephra correlations of volcanic ash beds within the sedimentary units. Previously published data included tephra correlations and $^{40}Ar/^{39}Ar$ and conventional K-Ar dates, as well as four fission-track dates on detrital zircons collected in the Snake Mountains near Wells (Thorman et al., 2003). Table 1 provides all of the dates and correlation ages cited in this paper, and Appendix I gives brief descriptions of unpublished samples collected and dated for this study. Several figures in the text show the approximate locations where samples were collected, along with the date or date range. In addition, Table 2, introduced later in the paper, summarizes published and unpublished K-Ar and $^{40}Ar/^{39}Ar$ dates on supergene alunite from several gold deposits in the study area. All $^{40}Ar/^{39}Ar$ dates in the tables, including the dates used for the tephra correlations, were calculated or recalculated using 28.02 Ma for the Fish Canyon Tuff sanidine (Table 1). The units reach a thickness of ~100 m and are continuous along strike for more than 15 km. They extend south into the Midas district, where they have been dated at 15.7 Ma (Table 1; Leavitt et al., 2004), and to the north to within 4 km of 16.1–15.5 Ma strata related to the Chimney basin.

At Snowstorm Mountain, the sedimentary units overlie and are interbedded with 16.1 Ma basaltic andesite flow units and underlie 15.5–15.7 Ma rhyolite flow units. The sedimentary units are as much as 100 m thick. Small areas of possibly related pyroclastic materials are interbedded with basaltic andesite flows southwest of Snowstorm Mountain. Sediments of this age are absent only in the west-southwestern part of the volcanic field, and the overall field relations suggest that the sedimentary rocks may be continuous beneath younger volcanic rocks throughout much of the field (Wallace, 1993).

Offset along predominantly north-northwest-striking normal faults occurred during volcanism in the Snowstorm Mountains. As exposed in both the Snowstorm Mountain and Castle Ridge areas, faulting offset and tilted the 15.7 Ma and older volcanic and sedimentary units prior to the eruption of 15.5 Ma and younger rhyolite flows, some of which filled a broad, fault-controlled basin between Castle Ridge and Snowstorm Mountain. The younger units are relatively unaltered and have very modest dips, and a few ca. 15.1 Ma flows near Snowstorm Mountain contain directional flow-related folds that indicate flow westward away from the central part of the volcanic center (Wallace, 1993).

The older sedimentary rocks on the northwestern side of the Snowstorm Mountains overlie the 15.5 Ma Little Humboldt rhyolite, and they underlie and interfinger with 9.8 Ma basalt flows of the Big Island Formation (Wallace, 1993). A broader study of the mixed sedimentary and basaltic Big Island Formation in northernmost Nevada showed that sedimentation closely preceded and coincided with the eruption of the basalt flows (Coats, 1985). On the northeastern side of the Snowstorm Mountains, the fine- to medium-grained sedimentary units are weakly to moderately consolidated. Most of the materials are air-fall tuff and reworked tuffaceous material, although some pebble-rich layers contain rhyolitic clasts. Clast imbrications and compositions, as well as cross-bedding directions, indicate that these clasts were derived from the Snowstorm Mountains area to the south. Additional post-Little Humboldt rhyolite sedimentary units are widespread throughout the Snowstorm Mountains area, but their correlation to those in the Big Island Formation can only be inferred from their general stratigraphic position.

These younger sedimentary and basaltic units in the Snowstorm Mountains area are not faulted, and they overlie faulted and tilted volcanic and
### TABLE 1. ISOTOPIC, TEPHRA CORRELATION, AND FISSION-TRACK GEOCHRONOLOGIC DATA FOR MIDDLE MIocene SEDIMENTARY AND RELATED UNITS, NORTHEASTERN NEVADA

<table>
<thead>
<tr>
<th>Unit or rock type</th>
<th>General location</th>
<th>Latitude</th>
<th>Longitude</th>
<th>$^{39}$Ar/$^{39}$Ar age (Ma)</th>
<th>Tephra corr. age (Ma)</th>
<th>Relation to basin sediments</th>
<th>Reference</th>
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</thead>
<tbody>
<tr>
<td><strong>Snowstorm Mountains (Fig. 3)</strong></td>
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</tr>
<tr>
<td>Big Island basalt</td>
<td>Northeast</td>
<td>46°26.51</td>
<td>116°57.47</td>
<td>9.8 ± 2.5*</td>
<td>–</td>
<td>Above all units</td>
<td>Wallace et al. (1990)</td>
</tr>
<tr>
<td>Little Humboldt rhyolite</td>
<td>Northwest</td>
<td>41°27.45</td>
<td>116°59.13</td>
<td>15.45 ± 0.04</td>
<td>–</td>
<td>Above, within</td>
<td>This study (ChB-1)</td>
</tr>
<tr>
<td>Red rhyolite flows</td>
<td>Midas</td>
<td>41°15.98</td>
<td>116°45.55</td>
<td>15.69 ± 0.12</td>
<td>–</td>
<td>Above</td>
<td>Leavitt et al. (2004)</td>
</tr>
<tr>
<td>Tuffaceous strata</td>
<td>Midas</td>
<td>41°14.08</td>
<td>116°46.38</td>
<td>15.80 ± 0.09</td>
<td>–</td>
<td>Within</td>
<td>Leavitt et al. (2004)</td>
</tr>
<tr>
<td>Basaltic andesite</td>
<td>Midas</td>
<td>41°14.58</td>
<td>116°45.41</td>
<td>16.06 ± 0.2</td>
<td>–</td>
<td>Below</td>
<td>Leavitt et al. (2004)</td>
</tr>
<tr>
<td><strong>Chimney Basin (Fig. 3)</strong></td>
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<tr>
<td>Air-fall tuff</td>
<td>Twin Creeks</td>
<td>41°14.20</td>
<td>117°10.80</td>
<td>14.17 ± 0.08</td>
<td>–</td>
<td>Fluvial at top</td>
<td>Breit et al. (2005)</td>
</tr>
<tr>
<td>Tephra</td>
<td>East</td>
<td>41°23.15</td>
<td>117°06.24</td>
<td>14.7 ± 0.1</td>
<td>Lacustrine-fluvial contact</td>
<td>This study (ChB-1)</td>
<td></td>
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<tr>
<td>Kelly Creek rhyolite</td>
<td>East</td>
<td>41°22.36</td>
<td>117°03.42</td>
<td>15.1 ± 0.04</td>
<td>–</td>
<td>Above; age estimated</td>
<td>Wallace (1993)</td>
</tr>
<tr>
<td>Little Humboldt rhyolite</td>
<td>Northeast</td>
<td>41°27.45</td>
<td>116°59.13</td>
<td>15.45 ± 0.04</td>
<td>–</td>
<td>Interfingered and above</td>
<td>This study (ChB-2)</td>
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<tr>
<td>Andesite flow</td>
<td>West, north</td>
<td>41°26.60</td>
<td>117°20.44</td>
<td>15.85 ± 0.22</td>
<td>–</td>
<td>Interfingered</td>
<td>This study (ChB-3)</td>
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<td>Basaltic andesite</td>
<td>West</td>
<td>41°26.86</td>
<td>117°04.11</td>
<td>16.07 ± 0.2</td>
<td>–</td>
<td>Below; age estimated</td>
<td>Leavitt et al. (2004)</td>
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<td>Welded tuff</td>
<td>West</td>
<td>41°27.55</td>
<td>117°19.88</td>
<td>16.11 ± 0.04</td>
<td>–</td>
<td>Below</td>
<td>This study (ChB-4)</td>
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<tr>
<td>Rhyolite flow</td>
<td>Center</td>
<td>41°23.44</td>
<td>117°10.6</td>
<td>16.33 ± 0.08</td>
<td>–</td>
<td>Below</td>
<td>This study (ChB-5)</td>
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<td>Andesite flow</td>
<td>South</td>
<td>41°19.52</td>
<td>117°11.09</td>
<td>22.1 ± 0.09*</td>
<td>–</td>
<td>Below</td>
<td>Wallace and McKee (1994)</td>
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<td><strong>Ivanhoe basin (Fig. 5)</strong></td>
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<tr>
<td>Rhyolite dome</td>
<td>Ivanhoe</td>
<td>41°11.44</td>
<td>116°36.38</td>
<td>15.18 ± 0.05</td>
<td>–</td>
<td>Above</td>
<td>Wallace (2003)</td>
</tr>
<tr>
<td>Craig rhyolite (east)</td>
<td>Santa Renia Fields</td>
<td>41°04.87</td>
<td>116°29.59</td>
<td>15.24 ± 0.34</td>
<td>–</td>
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Late Cenozoic paleogeography, northeastern Nevada

Although these Big Island-related units are somewhat younger than those in other basins described in this paper, they do indicate that 10 Ma streams drained northeastward from a modest highland that had been faulted by that time.

CHIMNEY BASIN

Chimney basin is generally west of the Snowstorm Mountains, east of Paradise Valley, and north of the Osgood Mountains and Hot Springs Range in eastern Humboldt County and westernmost Elko County (Fig. 3). Widespread Miocene sedimentary units are centered on Chimney Reservoir. The most extensive exposures are north of and along the Little Humboldt River. Pliocene and younger alluvial sediments largely conceal the Miocene units south of the river. On the basis of gravity, depth-to-basement data (Ponce, 2004), the aggregate thickness of the Miocene units and any underlying Tertiary volcanic rocks is much less than 1 km throughout the entire basin. Geochronologic data for the basin are provided in Table 1.

In general, the Chimney basin was active from ca. 16.3 Ma until 14.2 Ma (Fig. 2). The basin was a broad lowland, and coeval volcanic activity blocked westward streamflow to form a shallow, ephemeral lake. The basin in its early stages extended to the southeast across the northern Nevada rift and connected with the Ivanhoe basin near Midas. Continued volcanism along the fringes of the basin limited expansion of basin sedimentation, and late-stage uplift in the Snowstorm Mountains produced fluvial sedimentation in the eastern part of the basin. Most fault activity and uplift, however, took place after sedimentation, and uplift of two major horsts just to the south did not affect the basin. Despite a temperate, moist climate, no sediments were deposited after 14.2 Ma, and the basin presumably began to drain externally at that time.

Stratigraphy

The majority of the exposed sedimentary rocks in the Chimney basin are composed of fine-grained, evenly bedded, tuffaceous sediments and somewhat coarser unwelded, pumiceous, air-fall deposits. The finer grained sedimentary layers generally are thinly bedded, whereas the coarser, more pumiceous beds are thicker and structureless (Fig. 4A); bedding is usually planar. The sediments are almost entirely air fall in origin; epiclastic materials are rare, except adjacent to the Snowstorm Mountains to the east, where clasts of Snowstorm-derived volcanic rocks are scattered through the otherwise ash-rich sediments. Evidence of
fluvial reworking, such as graded bedding, cross bedding, and channel features, is also uncommon, except in the area between the Snowstorm Mountains and Chimney Reservoir. Mud cracks and siliceous sinter deposits with reed and leaf fragments are common in sedimentary rocks in the northern and western parts of the basin and indicate periodic subaerial exposure. Diagenetic zeolites and Magadi-type chert replaced many of the fine-grained, ash-rich sediments (Shippeard and Gude, 1983).

At Chimney Reservoir, ash-rich sediments were deposited on a 16.3 Ma rhyolite flow sequence. This relation provides a maximum, but not necessarily the initial, age for sedimentation. In the western and northwestern parts of the basin, the ash-rich sedimentary units interfinger with 15.9 Ma andesite flow units, which, in turn, overlie 16.1 Ma rhyolite tuffs. These volcanic units are part of the Santa Rosa–Calico volcanic field, which is extensive to the north and northwest (Brueseke and Hart, 2007). Near Martin Creek (Fig. 3), sediments are absent at the contact between 5 Ma basalt flows (Hart et al., 1984) and the 16.1 Ma tuffs; therefore, the margin of the sedimentary basin was south of that point. To the north and northeast, the regionally extensive 15.5 Ma Little Humboldt rhyolite sequence directly overlies the 15.9 Ma andesite units with no intervening sedimentary units (Wallace, 1993; W. Walker, 2001, unpublished mapping), but the rhyolite extends south from that area across the sedimentary units. Therefore, the ash-rich sediments in the western and northern parts of the basin apparently were deposited between ca. 16.1 and 15.5 Ma and permissively as early as 16.3 Ma in the center of the basin at Chimney Reservoir.

In the eastern and northeastern parts of the basin, the ash-rich sedimentary rocks both underlie and interfinger with flow units of the Little Humboldt rhyolite sequence. Farther east, toward the Little Humboldt ranch and at Rodear Flat (Fig. 3), ash-rich strata overlie basaltic andesite flow units that, in the Midas area to the southeast, were dated at 16.1 Ma (Leavitt et al., 2004), and the strata interfinger with and underlie the 15.5 Ma Little Humboldt rhyolite. Near the Little Humboldt ranch, the earliest rhyolite flow contains a thick phreatic breccia at its basal contact with the underlying strata, and several tens of meters of additional sedimentary units overlie the brecciated flow unit; these sediments, in turn, underlie a second, unbrecciated rhyolite eruptive unit (Fig. 4B). The sedimentary units may extend for an unknown distance to the northeast beneath the upper eruptive unit, and the Rodear Flat strata likely correlate with the fine-grained sediments exposed just to the south at the north end of Castle Ridge (see the previous section on the Snowstorm Mountains). In all of these northeastern exposures, the sediments are ash rich, fine grained, and altered to zeolites and chert, similar to those in the main part of the basin.

In the east-central part of the basin, just west of the Snowstorm Mountains and south of the Little Humboldt River, a thick sequence of epiclastic sediments conformably overlies the ash-rich lacustrine sediments (Figs. 4C and 4D). The contact between the two facies is sharp, and the base of the sequence above fine-grained ash beds contains abundant volcanic cobbles. A tephra layer, 2 m beneath the contact, produced a tephra correlation age of 14.7 Ma (Table 1). Near the Snowstorm Mountains, the epiclastic strata include alternating cobble-rich and sand-pebble beds. The coarser beds contain cross bedding, graded bedding, and channel cut-and-fill structures. Clasts in all parts of the fluvial sequence are composed of volcanic lithologies exposed in the Snowstorm Mountains directly to the east.

To the west, this sequence becomes finer grained and is composed of planar-bedded, structureless sandstones and thin, pebble-rich beds with weak cross bedding and graded bedding. This finer grained sequence projects west across the incised Little Humboldt River valley to the top of the sedimentary section exposed on the north shore of Chimney Reservoir. That section grades upward from thinly bedded, ash-rich bedded sediments into the epiclastic strata derived from the Snowstorm Mountains. This section, albeit finer grained and more gradational, mimics the transition from ash-rich to coarser epiclastic facies near the Snowstorm Mountains.

### Table 2: Supergene Aluminate Dates from Late Eocene Gold Deposits, Northeastern Nevada

<table>
<thead>
<tr>
<th>Gold trend</th>
<th>General location</th>
<th>Age (Ma)</th>
<th>Method</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Battle Mountain</td>
<td>Lone Tree mine</td>
<td>8.86 ± 0.5</td>
<td>*Ar/Ar</td>
<td>A. Hofstra, L. Sne (unpubl. data, 2005)</td>
</tr>
<tr>
<td>Carlin</td>
<td>Gold Quarry area</td>
<td>6.7 ± 0.2</td>
<td>K-Ar</td>
<td>P. Vikre (unpubl. data, 2005)</td>
</tr>
<tr>
<td>Carlin</td>
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<td>K-Ar</td>
<td>P. Vikre (unpubl. data, 2005)</td>
</tr>
<tr>
<td>Carlin</td>
<td>Post mine</td>
<td>8.6 ± 0.2</td>
<td>K-Ar</td>
<td>Arehart et al. (1992)</td>
</tr>
<tr>
<td>Carlin</td>
<td>Post mine</td>
<td>9.5 ± 0.2</td>
<td>K-Ar</td>
<td>Arehart et al. (1992)</td>
</tr>
<tr>
<td>Carlin</td>
<td>Genesis mine</td>
<td>11.0</td>
<td>K-Ar</td>
<td>Heitt (1992)</td>
</tr>
<tr>
<td>Carlin</td>
<td>Rain mine</td>
<td>12.6 ± 0.5</td>
<td>K-Ar</td>
<td>Williams (1992)</td>
</tr>
<tr>
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<td>18.5 ± 0.8</td>
<td>K-Ar</td>
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</tr>
<tr>
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</tr>
<tr>
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<tr>
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<td>Mike deposit</td>
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<td>Arehart et al. (1992)</td>
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<td>Carlin</td>
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<td>Heitt (1992)</td>
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<tr>
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<td>K-Ar</td>
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<td>K-Ar</td>
<td>Arehart and O'Neil (1993)</td>
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<tr>
<td>Getchell</td>
<td>Preble mine</td>
<td>23</td>
<td>K-Ar</td>
<td>Arehart and O'Neil (1993)</td>
</tr>
</tbody>
</table>

*Note: Gold trends are shown on Figure 1. *Ar/Ar dates recalculated to 28.02 Ma Fish Canyon tuff standard. *Error not published.
Late Cenozoic paleogeography, northeastern Nevada

Figure 3. Map of the Chimney basin and Snowstorm Mountains areas, showing the original known and inferred extents of the sedimentary basin, middle Miocene volcanic fields, dates of various sedimentary and volcanic units, and post-sedimentation normal faults and synforms. Areas that do not have an overlay color are underlain by pre-Miocene basement rocks (especially in the Santa Rosa and Hot Springs Ranges and Osgood Mountains), post-middle Miocene sedimentary cover (Paradise, Eden, and Kelly Creek Valleys), and post-sedimentation volcanic cover (Owyhee Plateau region). Some volcanic units along the northern Nevada rift and in the Santa Rosa–Calico volcanic fields were erupted after sedimentation ended, but they are included within the volcanic fields regardless of relative age. See Table 1 for geochronologic information. The modern digital elevation map is used as the base.
At the spillway for Chimney Reservoir (Fig. 3), sand- and pebble-rich clastic units overlie the 16.3 Ma rhyolite flow unit, and angular clasts derived from the rhyolite are abundant at the basal contact. Additional fluvial sediments are exposed in low hills along the Little Humboldt River downstream from the reservoir. All of these deposits are isolated from other fluvial deposits; they could be related to the Miocene epiclastic units described above or are much younger deposits related to the late Miocene to Pliocene development of the Little Humboldt River.

Miocene units are poorly to not exposed south of the Little Humboldt River. This area includes the Hot Springs Range, the Osgood Mountains, and the intervening Eden and Kelly Creek Valleys. Strata exposed between the Dry Hills (Fig. 3) and the Little Humboldt River are composed of fine-grained, poorly exposed sediments. Both epiclastic and ash-rich sedimentary units are exposed near the Snowstorm Mountains in the southeastern part of the basin, but the section thins and laps onto Paleozoic basement to the south. At the Twin Creeks mine (Fig. 3), Miocene units above the Paleozoic basement include basal colluvial and regolith deposits and overlying epiclastic sand and gravel deposits derived from the nearby Dry Hills; a 14.2 Ma, air-fall ash is interbedded with these units (Breit et al., 2005). Ash-rich basin strata extend a few kilometers south into Eden Valley between the Osgood Mountains and Hot Springs Range, but most of that area has an extensive cover of Pliocene and younger alluvial sediments that masks the southern extent of the Miocene units.

Figure 4. Photographs of middle Miocene sedimentary and volcanic units in the Chimney basin. A: Fine-grained, thin-bedded, ash-rich lacustrine sedimentary strata overlying a thick ash bed at the base of the exposure. Photo was taken in the western part of the basin. B: Phreatic, vitric megabreccia in base of basal flow of the 15.5 Ma Little Humboldt rhyolite above lacustrine sedimentary units (blocky hill), possibly indicating eruption onto wet sediments or water. The overlying, younger flow unit (left background) overlies both the breccia and the sedimentary units but does not have a basal breccia. Photo was taken at the Little Humboldt ranch along the Little Humboldt River (Fig. 3). C: Conformable contact (dashed white line) between the lacustrine and fluvial facies; a tephra sample from 2 m below the contact produced a 14.7 Ma correlation age. Photo taken between Chimney Reservoir and the Snowstorm Mountains (right far background). D: Closer view of contact between lacustrine (below, light colored) and conformably overlying fluvial sediments (tan). The cobbles concentrated along the contact zone were derived from volcanic units exposed in the Snowstorm Mountains to the east (right).
Faulting

Numerous small, normal faults offset the Miocene strata throughout the northern half of the Chimney basin. Alluvial cover and very poor exposures conceal any possible faults in the southern part of the basin. Dips on all Miocene sedimentary units generally are less than 15°, and many units are nearly horizontal. The basal sediments dip as gently as strata higher in the section, indicating little or no synsedimentary, fault-related tilting.

Faults strike predominantly to the north to north-northeast, although faults of all orientations are present in the basin. Offsets range from a few tens of meters to ~100 m. Several down-to-the-east normal faults strike northerly through the middle of the basin, including the Chimney Reservoir area. The most obvious of these faults forms the east side of the Dry Hills (Fig. 3), extends north to near the dam at Chimney Reservoir, and continues to the north. Juxtaposed pre-sedimentation volcanic units and the basal parts of the Miocene section indicate offset of only ~100 m. Along the southern projection of this fault at the Twin Creeks mine, a scarp along the fault appears to have controlled the deposition of Pliocene alluvial deposits (Breit et al., 2005). A series of both down-to-the-east and down-to-the-west normal faults parallel the west side of the Snowstorm Mountains, but displacement along each fault was less than 100 m.

Along the west side of the basin, east of Paradise Valley (Fig. 3), down-to-the-west normal faults with individual offsets of less than 100 m offset the basin strata and underlie Miocene volcanic rocks and tilted them modestly to the east. At Martin Creek, a series of west-dipping normal faults tilted 16.1 Ma volcanic units ~20°, but produced only minor offset of overlying, nearly horizontal 5 Ma basaltic flows. To the south, these faults tilted the Hot Springs Range to the east. To the west, the bulk of the uplift of the Santa Rosa Range and formation of the east-dipping Paradise Valley half graben took place between ca. 10 Ma and 5 Ma (Colgan et al., 2004). The Martin Creek faults continue northwest and bisect the Santa Rosa Range, and the southern part of the range is now structurally lower than the northern half.

The Osgood Mountains are composed of two, west-dipping blocks. The largest is the main north-northeast–trending, west-tilted range, with a major range-front fault along its east side (Fig. 3). The north-striking Getchell normal fault truncates the northeast end of this block (Hotz and Willden, 1964). This fault is not evident in the main part of the Chimney basin to the north. Early Miocene (ca. 22 Ma) andesite flow units on the west side of the range likely were deposited at a very low angle, now dip ~20° to the west, and project east to just above the crest of the range (Hotz and Willden, 1964). As such, the range crest was just below the early Miocene paleosurface, and the Osgood Mountains were not a highland at 22 Ma. The second block is the Dry Hills, which is bounded by the Getchell fault to the west and a relatively minor, north-striking fault to the east. This fault, as noted above, does cut the Miocene strata to the north. West-dipping, 22 Ma andesite flows (Wallace and McKee, 1994), identical to those along the west side of the Osgood Mountains, are present in the Dry Hills and may be the downfaulted remnants of a broader volcanic field.

Eden Valley separates the northeast-trending Osgood Mountains and the north-trending Hot Springs Range, which converge to the south. East-dipping, early Miocene andesite flows on the east side of the Hot Springs Range mirror those along the west side of the Osgood Mountains (Jones, 1997). Gravity data in Eden Valley indicate that the thickness of Pliocene and Miocene units above Paleozoic bedrock is generally less than 1 km, consistent with the valley being a shallow synform between the two opposite-tilted ranges. This synform is narrow at the south end of the ranges, and it broadens to the north as the ranges diverge. North of the Little Humboldt River (north of the limits of the ranges), all Miocene strata dip very gently or are flat lying, and abundant small normal faults with widely varying strikes obscure the synform, if present. Farther north, west of Whiskey Springs (Fig. 3), any semblance of a synform in 10 Ma and older volcanic units is absent. As with the Osgood Mountains, the Eden Valley synform appears to die out into and is not a noticeable structural element in the Chimney basin.

Paleogeography and Sedimentation

Deposition of ash-rich, waterlain sediments in the Chimney basin began sometime after 16.3 Ma and was actively taking place from 16.1 Ma to ca. 14.7 Ma. Deposition of sandstones and coarser materials in the southeastern part of the basin occurred from 14.7 to 14.2 Ma. At limits of the exposed basin sediments, the sedimentary units interfere with and pinch out against coeval volcanic flow units along the western, northwestern, and northern margins of the basin, and the volcanic activity likely limited the extent of the basin in those areas. The early sedimentary environment may have extended to the northeast, but eruption of the 15.5 Ma Little Humboldt rhyolite created a new, more proximal basin margin in that direction. In the eastern part of the basin and in all Snowstorm Mountain exposures, sediments were deposited on 16.1 Ma basalts andesite flow units until ca. 15.7 Ma.

With the exception of the late epiclastic sediments in the southeastern part of the basin, the generally planar, laterally continuous bedding in the ash-rich units indicates deposition in a low-energy, lacustrine environment, and mud cracks and sinter deposits point to episodic subaerial exposure. The general absence of epiclastic material or evidence of fluvial reworking until late in the basin history suggests that the highlands that constrained the margins of this lacustrine environment had low relief. The shallow lacustrine connection of the Chimney and Ivanhoe basins across the northern Nevada rift in the Snowstorm Mountains demonstrates that rift-related volcanic and fault activity did not create any substantial positive or negative relief.

The lacustrine environment was shallow, became ephemeral, and eventually transitioned into a fluvial environment. The sinter deposits and mud cracks indicate periodic subaerial exposure during the lacustrine stage, and early, but not later, flows of the Little Humboldt rhyolite were erupted onto wet or submerged sediments. The upsection increase in diagenetic minerals indicative of more saline and alkaline conditions may reflect a progressively more evaporative lacustrine environment (Sheppard and Gude, 1983; R.A. Sheppard, 1992, personal commun.), and the transition from lacustrine to fluvial environments at and east of Chimney Reservoir point to possible external drainage starting at ca. 14.7 Ma.

The evolution of the western part of the Snowstorm Mountains volcanic field significantly affected the eastern margin of the basin. At Snowstorm Mountain (Fig. 3), the eruption of 15.7 Ma rhyolite flows ended sedimentation in that area, and post–15.5 Ma pre–15.1 Ma faulting tilted the older units to the west (Wallace, 1993). However, this faulting did not generate epiclastic sediments, which did not appear in the basin until ca. 14.7 Ma; the sudden influx of coarse epiclastic sediments at that time indicates some uplift in the Snowstorm Mountain area, although ca. 15.1 Ma flow units at Snowstorm Mountain dip only ~5° to the west. Faults along the west side of the range may have been active at this time, although there is no direct evidence of such activity. Farther south, at the Twin Creeks mine, sediments were shed from the Dry Hills at ca. 14.2 Ma (Breit et al., 2005), but the relation of fluvial sedimentation there and that west of Snowstorm Mountain is unknown due to the lack of intervening exposures.

On the basis of the southward thinning of the basin strata onto Paleozoic rocks, the southern margin of the basin at the time of sedimentation was most likely a low Paleozoic-cored upland or
bedrock sill. Overall, the available stratigraphic and structural data indicate that the Osgood Mountains and Hot Springs Range formed after Miocene sedimentation but did not significantly affect the Chimney basin, despite its proximity and the projection of major structures toward the basin. Uplift possibly started near the end of sedimentation, producing 14.4 Ma supergene alunite at the Twin Creeks mine (Arehart and O’Neil, 1993) and the nearby 14.2 Ma clastic sediments. Alternatively, the major uplift took place between ca. 10 and 5 Ma, based upon evidence of that uplift age in the Santa Rosa Range and the tilting constraints near Martin Creek. Sedimentary and fault-scarp data (Breit et al., 2005; Wesnousky et al., 2005) indicate that uplift continued into the Pliocene and Quaternary.

Why and when sedimentation ended in the Chimney basin are unclear. The climate was moist and temperate, the depositional environment was becoming increasingly subaerial and fluvial in the eastern part of the basin, and major rhyolite eruptions were significantly disrupting the northern and eastern parts of the basin. The absence of younger sediments argues against inflowing streams, which would have carried sediments, simply evaporating when they reached the basin. This absence of younger sediments in this topographically low area, despite the presence of uplands to the east, strongly suggests that the confining volcanic or topographic dams along the southwestern part of the basin were breached and the basin began to drain externally. Unfortunately, uplift of the Osgood Mountains and Hot Springs Range and Pliocene and younger alluvial sedimentation in Eden and Kelly Creek Valleys have largely obliterated the middle Miocene record in those areas.

IVANHOE BASIN

The Ivanhoe basin is located in southwestern Elko County and northernmost Lander and Eureka Counties (Fig. 5), stretching east to west from the Tuscarora Mountains to the Sheep Creek Range, and south to north from Boulder Valley into the Snowstorm Mountains. Basin-related sedimentary units are well exposed in the Ivanhoe mining district and areas to the east, south, and west, as well as in the Midas district in the southeastern Snowstorm Mountains. Post-sedimentation faulting and Pliocene and younger sediments conceal basin-related units along the Midas trough and along Rock Creek southwest of Ivanhoe.

In general, lacustrine and late, low-energy, fluvial sedimentation in the Ivanhoe basin took place from before 16.1 Ma to after 14.7 Ma (Fig. 2) as volcanism related to the northern Nevada rift blocked streams flowing westward.

Figure 5. Map of the Ivanhoe basin area, showing the original known and inferred extents of the sedimentary basin, middle Miocene volcanic units, dates of various sedimentary and volcanic units, major post-sedimentation normal faults, and the late Eocene Carlin gold trend. Uncolored areas include areas of pre-Miocene basement rocks in the Tuscarora Mountains and post-middle Miocene sedimentary cover in Boulder and Kelly Creek Valleys; the extent of volcanic units beneath sediments in Boulder Valley is based upon exposures of identical volcanic units in the cliffs on either side of the valley. See Table 1 for geochronologic information. The modern digital elevation map is used as the base.
from the Tuscarora Mountains. Relief was low, the lake level rose, and sediments progressively blanketed the entire area. Most of the sediments were air-fall ash and pumice, and fine-grained epiclastic materials were carried into the southern part of the basin from the Tuscarora Mountains only late in the basin history. Sedimentation likely ended when volcanic dams were breached and the basin drained externally. Minor faulting occurred only during late sedimentation in the northern part of the basin. Major east-northeast–striking faults formed much later, perhaps after 10 Ma.

**Stratigraphy**

The oldest sediments in the basin were deposited shortly before 16.1 Ma in the Willow Creek Reservoir area (Fig. 5; Table 1; Perkins et al., 1998; Wallace, 2003c). Sedimentation in that area continued until after 15.4 Ma, depositing more than 200 m of sediments above late Eocene and Paleozoic units. The basal Miocene sediments are composed of relatively minor epiclastic sand and silt and moderate amounts of reworked ash. The majority of the overlying units are composed of thinly planar-bedded, air-fall ash and pumice with local to pronounced soft-sediment deformation textures. One of these ash beds was dated at 16.1 Ma (Table 1; Wallace, 2003a). Thin conglomerate beds with clasts derived from the Tuscarora Mountains to the northeast locally are interbedded with the ash-rich units (Henry and Boden, 1999; Wallace, 2003a), and some beds are a mixture of reworked small pumice and epiclastic sand, indicating periodic influxes of non-ash material. The ash-rich section grades up into thin-bedded mudstones, thick-bedded sandstone, and a sub-aerial, 15.4 Ma vitric tuff that is a widespread marker unit throughout the northern part of the basin. Very thin-bedded, air-fall ash beds with locally pronounced, soft-sediment deformation, as well as some fine-grained epiclastic sediments, were deposited above the vitric tuff. Faulting truncated the top of the section at Willow Creek Reservoir, but a poorly exposed, thick section of fine-grained, ash-rich sediments overlies the vitric tuff southeast of the reservoir (Wallace, 2003a).

The Willow Creek Reservoir section extends discontinuously ~15 km north and northeast, where it overlies both Eocene volcanic rocks and Paleozoic rocks. Closer to the Tuscarora Mountains (Fig. 5), the lower part of the section thins to ~20 m above Eocene volcanic rocks and underlies the 15.4 Ma vitric tuff, which is the youngest preserved unit in that area (Henry and Boden, 1999).

Coeval volcanic activity produced rhyolite flow units and domes west of Willow Creek Reservoir. These include the Rock Creek rhyolite in the central part of the Ivanhoe basin and the June Bell rhyolite in the Midas area (Fig. 5). Neither rhyolite has been dated, but surface exposures and drilling data indicate that they are interbedded with or underlie the oldest sedimentary units in both areas and thus were erupted just before or during early sedimentation (Wallace, 1993; Goldstrand and Schmidt, 2000; Wallace, 2003c). The Rock Creek rhyolite is exposed over a large area and, based on exposed thicknesses, likely had relief on the order of several hundred meters, greater than the thickness of the sediments to the east.

The lower part of the Willow Creek Reservoir section thins to the south into the Ivanhoe mining district (Figs. 5, 6A, and 6B), where only ~50 m of thin-bedded, ash-rich strata lie between the widespread 15.4 Ma vitric tuff and the Paleozoic basement rocks. Units above the vitric tuff in this area include thin- to thick-bedded, air-fall ash and epiclastic sand beds; locally abundant pebble conglomerate and debris-flow units contain clasts derived from Paleozoic rocks that cropped out a few kilometers to the east (Wallace, 2003b).

Rhyolite and andesite flow units and domes were emplaced between ca. 15.5 and 15.2 Ma throughout the Ivanhoe district, and the volcanic units are interbedded with the sedimentary units. Several large, rhyolite porphyry domes were erupted at 15.2 Ma between Ivanhoe and Willow Creek Reservoir, and they overlie most of the sedimentary units in that part of the district (Fig. 6A). Therefore, the sedimentary units in the Ivanhoe district include only the upper part of the section exposed at the reservoir, and they were deposited between ca. 15.5 and 15.2 Ma (Wallace, 2003c).

The volcanic flow units in the district were emplaced subaerially, although the distal ends of a few anesite flow units have hyaloclastic breccias. Sinter deposits are common throughout most of the pre- and post–15.4 Ma sedimentary units (Fig. 6C; Wallace, 2003c), and they also indicate periodic, subaerial exposure during sedimentation. The locus of sinter activity shifted to the east with time, reflecting changes in the ground-water table and paleotopography produced by progressive volcanic activity, sedimentation, and minor late synsedimentary faulting (Wallace, 2003c).

In the Santa Renia Fields area east and southeast of Ivanhoe (Fig. 5), the oldest sediments were deposited shortly before 15.4 Ma, and sedimentation continued to after 14.7 Ma (Fleck et al., 1998). The sedimentary units are composed of a mixture of thin-bedded, air-fall ash and pumice deposits and fine-grained, epiclastic sediments derived from the Tuscarora Mountains to the east, including materials eroded from exposed gold deposits along the Carlin trend (Theodore et al., 1998, 2006). Air-fall materials are common in the lower half of the section, decrease in abundance upslope, and are absent in the upper quarter of the section, which was deposited after 14.7 Ma and is composed of epiclastic sand and silt. The total sedimentary thickness exceeds 500 m.

The large Goldstrike mine along the northern Carlin gold trend is in the southeastern part of the basin adjacent to the Tuscarora Mountains (Fig. 5). There, Miocene sedimentary units include a relatively thin basal conglomerate and overlaying, thin-bedded, ash-rich strata (Betts and Lauha, 1996). These sediments largely overlie 15.2 Ma rhyolite flow units (Table 1; Theodore et al., 2007). The rhyolite flow units extend northwest across the northern end of Boulder Valley and north into the Santa Renia Fields area, where they underlie the middle Miocene sedimentary units (Theodore et al., 1998, 2006).

The Craig rhyolite, a large, thick assemblage of rhyolite flows and domes, was erupted between the Santa Renia Fields area and the Ivanhoe district between 15.4 and 15.2 Ma and likely is related to the Boulder Valley rhyolites. In the Ivanhoe district, the flow units overlie 15.4 Ma and older sedimentary units, and younger strata overlie the flows. In the Santa Renia Fields area, 15.2 Ma sediments were deposited on the rhyolite (Theodore et al., 1998), and clasts derived from the rhyolite are abundant in overlying, otherwise fine-grained epiclastic sediments along Antelope Creek south of the rhyolite exposures (Fig. 6D). Strata in the Ivanhoe area are continuous to the south, along the west side of the Craig rhyolite exposures, and they connect in the Antelope Creek area with the sedimentary units that extend west from the Santa Renia Fields area.

The strata along Antelope Creek extend west beyond Rock Creek and southwest to nearly the topographic rim above Boulder Valley (Fig. 5). Many of the sedimentary units just east of Rock Creek contain thin, extensive surface travertine deposits and pervasive carbonate cement in underlying strata. In the southwestern Sheep Creek Range, 15 Ma basalt flows conformably overlie a thin sequence of fine-grained, epiclastic sedimentary units (Fig. 6E), which contain an ash bed dated at 15.2 Ma (Table 1; John and Wrucke, 2002, 2003). However, with the exception of two, very small exposures, sedimentary units are absent at the extensive contact between 15.6 Ma dacite flows and overlying 15.1 Ma rhyolite flows in the westernmost Sheep Creek Range (John and Wrucke, 2002).

Sedimentary units related to the basin are exposed in the Midas district (Figs. 5 and 6F),
Figure 6. Photographs of middle Miocene sedimentary and volcanic units in the Ivanhoe basin. A: Area between Willow Creek Reservoir and the Ivanhoe district, looking east. RR—Rimrock mercury mine; a—andesite flow unit; vt—15.4 Ma vitric tuff unit, both of which were erupted subaerially. Light-colored, ash-rich strata below the andesite were deposited subaqueously, as were strata between the andesite and the vitric tuff. Dark capping units in distance are 15 Ma rhyolite porphyry flow units and domes, which dip ~10° less than the underlying units, indicating tilting between ca. 15.4 and 15 Ma. The Rimrock mercury deposit formed in silica replacement bodies in lacustrine sedimentary units that overlie the vitric tuff unit. B: Thin-bedded, ash-rich lacustrine sedimentary units exposed in the north wall of the open pit of the Hollister gold mine, Ivanhoe district. These beds are stratigraphically between the andesite and the vitric tuff unit (see Fig. 6A). Draping near the base of the exposure is above the irregular top of an andesite flow unit that is exposed just below the photo. Note the progressive upward change to planar bedding. The variegated colors are due to hydrothermal and supergene alteration. C: Sinter and silicified lacustrine sedimentary units (light, massive unit in lower part of photo), overlain by unsilicified lacustrine sediments (middle of photo) and a capping 15.4 Ma rhyolite flow unit. The silicified horizon is widespread throughout the Ivanhoe district and formed prior to the deposition of the 15.4 Ma vitric tuff. Photo was taken 2 km east of the Hollister gold mine. D: Tan, pebble-rich sandstone bodies along Antelope Creek south of the Ivanhoe district. The pebbles are subangular to angular and were derived from the ca 15 Ma rhyolite exposures in the left background, although the main transport direction for the sand component was from right to left. Arrows point to soil horizons between sand bodies. E: Tan—epiclastic sedimentary units overlain by a 15 Ma basalt flow in the southwestern part of the Ivanhoe basin just southwest of the confluence of Rock and Antelope Creeks (Fig. 5). The basalt flow was emplaced subaerially. F: Midas district, looking to the north-northwest. Unwelded tuffs and lacustrine sedimentary units (light tan in foreground) are overlain by 15.7 Ma red rhyolite flows at the skyline. A wide, north-northwest–striking dike fed the flows and intruded the tuffs and sedimentary units in that area. The mine workings are along veins that filled north-northwest–striking faults in the tuffs and sediments; the veins formed at ca. 15.4 Ma (Leavitt et al., 2004).
where they have been called the Esmeralda formation (Rott, 1931; Goldstrand and Schmidt, 2000). The units include fine-grained, thin-bedded, ash-rich strata and sand- and pebble-rich conglomerate units. Clast compositions and imbrications in the latter units indicate transport from Paleozoic exposures to the northeast (Wallace, 1993; Goldstrand and Schmidt, 2000). The sedimentary units interfinger with pyroclastic and 16.1 Ma mafic flow units and underlie extensive 15.7 Ma and younger rhyolite flows (Wallace, 1993; Goldstrand and Schmidt, 2000; Leavitt et al., 2004). A thin tuff at the base of the sedimentary section was deposited at 15.7 Ma (Leavitt et al., 2004). About 6 km northeast of Midas, beds of air-fall ash and epiclastic sediments are exposed between 15.7 and 15.3 Ma rhyolite flow units (Leavitt et al., 2004). Although late Cenozoic alluvial sediments cover most of the area between Midas and Willow Creek Reservoir, isolated exposures of Miocene strata in the intervening area, as well as the similar ages of the sediments, indicate that the strata in the two areas are related. Also, as described earlier, the sedimentary units in the Midas district are continuous with 16.1 to 15.5 Ma strata exposed in the Snowstorm Mountains immediately to the northwest.

Faulting

A minor amount of faulting took place in the Ivanhoe basin near the end of middle Miocene sedimentation, but most of the faulting occurred much later. In addition, late Eocene volcanic rocks near Willow Creek Reservoir dip as much as 30° more steeply than overlying Miocene sediments, indicating a post–38 Ma, pre–16 Ma period of fault-related tilting (Wallace, 2003c). Throughout most of the Ivanhoe basin, evidence of synsedimentary faulting, such as decreasing stratigraphic units and overlap of faults by younger strata, was not observed. In the Ivanhoe district, lateral migration of hydrothermal fluids produced widespread, stratiform, silica-replacement zones in strata that directly underlie the 15.4 Ma vitric tuff unit, and silicification took place shortly before the deposition of the latter unit (Wallace, 2003c). Post–15.4 Ma faults modestly cut and tilted the silicified horizon, but the lateral wide extent of the that horizon indicates that the ground-water table was not hydraulically segmented by faults until after 15.4 Ma. South of Willow Creek Reservoir, the dips of the 15.4 Ma vitric tuff and enclosing tuff units are, on average, 10° greater than those of overlying 15.2 Ma rhyolite flows (Fig. 6A). As such, some faulting in the Ivanhoe area took place during the later stages of sedimentation and volcanism in the area, but it did not produce significant offset of the units.

At Midas, north-northwest–striking faults began to develop by ca. 15.8 Ma, during early sedimentation, and thus may have affected sedimentation (Goldstrand and Schmidt, 2000; Leavitt et al., 2004). Faults in the Midas area and the Snowstorm Mountains area to the north and northwest served as conduits for 15.8 Ma rhyolite dikes (Fig. 6F) and cut volcanic and sedimentary units as young as 15.5 Ma. Just west of Midas, rhyolite flows erupted shortly after 15.5 Ma are nearly flat lying and unconformably overlie tilted older flow units (Wallace, 1993). A few kilometers northeast of Midas, modest east-dip dikes on 15.3 Ma flow units indicate continued faulting in that area.

Along the northern Carlin trend, near the Goldstrike mine (Fig. 5), generally north-striking, east-dipping normal faults have as much as 150 m of displacement. These faults are in part related to a west-dipping half graben on the west side of the Tuscarora Mountains. Early sediments were deposited during the formation of the graben, but later sediments were deposited on top of the faults (Leonardson and Rahn, 1996). These sediments overlie 15.2 Ma rhyolite flow units; therefore, faulting likely occurred shortly after that time.

The largest faults in the Ivanhoe basin are along the Midas trough and the boundary between the southern Sheep Creek Range and Boulder Valley (Fig. 5). Movement along these east-northwest–striking faults produced as much as 1 km of vertical offset and several kilometers of left-lateral displacement (Wallace, 1991, 1993; John et al., 2000). These faults formed after sedimentation. Zoback and Thompson (1978) estimated that faulting took place between ca. 10 and 6 Ma, although faults along the Midas trough cut and modestly tilted 4.4 Ma basalt flows (Hart et al., 1984; Wallace, 1993). Fault scarps indicate continued offset into the Quaternary (Wallace, 1993; Wensnousky et al., 2005).

Several of these faults along the Midas trough extend from the western side of the Sheep Creek Range and Snowstorm Mountains east to the Willow Creek Reservoir area. Volcanic and sedimentary units on the north and south sides of the fault-controlled trough dip north and south, respectively, with dips decreasing away from the trough. To the east-northeast, offset along individual faults diminishes, and displacement is distributed across numerous faults spread over a broad area (Wallace, 2003a, 2003b). In and around the Ivanhoe district, displacement along individual faults usually is less than 100 m and more commonly is less than 50 m, and the faults die out near Willow Creek Reservoir. Fault-related offset and southerly tilts are greatest in the northern part of the district and decrease toward Antelope Creek, where the sedimentary units are horizontal and largely unaffected (Theodore et al., 1998; Wallace, 2003b).

Similarly, several closely spaced, east-northeast–striking faults form the steep escarpment between the southern Sheep Creek Range and Boulder Valley (Fig. 5). Volcanic and sedimentary units near the escarpment dip modestly to the north, decreasing to horizontal along Antelope Creek. As such, Antelope Creek generally follows the axis of an east-northeast–trending synform, and Rock Creek is an antecedent stream that incised into the northern and southern limbs of the synform. These faults die out to the east-northeast and are not present along the west side of the Tuscarora Mountains.

Sedimentation and Paleogeography

Sediments in the Ivanhoe basin were deposited in shallow lacustrine, low-relief subaerial, and low-energy fluvial environments. In the areas where the basement-sediment contact is visible, post-Eocene, pre-middle Miocene sedimentary units are not present. In the Ivanhoe area, the top of the Paleozoic basement is composed of a variably thick regolith (Bartlett et al., 1991) that formed during exposure and weathering prior to ca. 16 Ma sedimentation. In the eastern part of the Ivanhoe district, a late Eocene pluton (the Hatter stock) had been exposed by the ca. 15.2 Ma inception of sedimentation in that area (Wallace, 2003b). Sediments older than ca. 15.5 Ma occur only in the northern half of the basin at Willow Creek Reservoir and in the Midas area; the sediments in the southern half of the basin were deposited after ca. 15.3 Ma and are laterally continuous with the upper part of the northern section.

Basin sedimentation began shortly before 16.1 Ma in the Willow Creek Reservoir area, which, as far back as the late Eocene, was a west-draining Eocene paleovalley (Wallace, 2003a; Henry, 2008). On the basis of the differences in stratal thickness, the initial depocenter in the area of the reservoir was ~200 m lower than areas to the south and possibly the north. Clast lithologies and imbrication show that inflow into the early basin was from the Tuscarora Mountains to the northeast and east. The epiclastic materials are generally fine grained and uncommon, in contrast to the abundant, coarse sediments that were being carried into the Carlin basin on the east side of the range at the same time (see description of the Carlin basin). Air-fall materials comprise the bulk of the sediments deposited up to 15.3 Ma. By ca. 15.3 Ma, lacustrine and low-energy fluvial environments had expanded to cover most of the adjacent, lower lying areas, and sedimentation continued primarily in the southern half of the

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basin. It was only after ca. 15.3 Ma that significant amounts of fine-grained, epiclastic sediments were carried into the basin, primarily in west-flowing, low-energy streams in the southern part of the basin.

Coeval volcanic flows and domes erupted along the entire margin of and within the Ivanhoe basin strongly affected the timing and distribution of sedimentation. The eruption of the Rock Creek and June Bell rhyolites west of the Willow Creek Reservoir area (Fig. 5) just prior to sedimentation may have blocked westward streamflow and, combined with the modest upland area to the south, produced the early lacustrine environment centered on the reservoir area. That early depocenter filled, and the lacustrine environment expanded, reaching the highest point in the Ivanhoe district to the south by ca. 15.6 Ma and the Midas area to the west by ca. 15.7 Ma.

Rhyolite to andesite flows were erupted in the Ivanhoe area between ca. 15.6 and 15.3 Ma. These eruptions were largely subaerial despite nearby coeval lacustrine sedimentation (Wallace, 2003c), but deposition of lacustrine sediments on top of the flow units indicates that the flows had a low relief and the lake was able to expand over them after eruption. This cycle between volcanism and lacustrine sedimentation continued until the eruption of large rhyolite domes at 15.1 Ma between Willow Creek Reservoir and Ivanhoe. The eruption of the extensive Craig and Boulder Valley rhyolites at ca. 15.2 Ma may have created a volcanic upland that focused streamflow into the southwestern part of the basin.

A thick sequence of basalt to dacite flows was erupted in the western and southwestern Sheep Creek Range between ca. 16.1 and 15.6 Ma (John et al., 2000; John and Wrucke, 2002; Leavitt et al., 2004). A thin sequence of 15.2 Ma sediments between those flows and overlying 15 Ma basalt flows pinches out to the west, and the older basalt-dacite flow sequence likely defined the southwestern margin of the basin. An extensive field of 15 Ma rhyolite domes forms the southwestern and western margins of the basin. These domes were erupted onto 15.8 Ma andesite basalt flows to the north (Wallace, 1993; Leavitt et al., 2004). Sedimentary units are absent at volcanic contacts in both areas, suggesting that eruption of these volcanic rocks produced the western boundary of the basin.

In the Midas area and adjacent parts of the Snowstorm Mountains, the sediments were deposited on 16.1 Ma basaltic andesite flow units along the axis of the northern Nevada rift. These basaltic andesite units are absent east of the rift, such as at Scraper Springs (Fig. 5), which is east of the rift, and the earliest basin sediments were deposited directly on late Eocene volcanic units. At both Midas and Scraper Springs, the eruption of extensive rhyolite flow units starting at 15.7 Ma ended further sedimentation. Farther west, in the southern Snowstorm Mountains and northern Sheep Creek Range, the eruption of large rhyolite domes and extensive flows continued to create a basin margin until ca. 15.2 Ma.

The locus of sedimentation apparently shifted to the south with time, but the reasons why post–15.2 Ma sediments were not deposited in the northern half of the basin are unknown. Given the moist, temperate climate, water undoubtedly continued to flow into the northern area from the Tuscarora Mountains. At the same time, epiclastic sediments were being carried westward from this highland into the low-relief, southern part of the basin. It is possible that all post–15.2 Ma sediments in the northern part of the basin were removed by later erosion, but the total efficiency of such erosion over a broad area seems improbable. The only possible outlet to the northern part of the basin would have been to the west, and the cessation of volcanism in the western part of the basin at 15.2 Ma may have allowed streams to breach those dams and exit the basin. Unfortunately, that area is along the Midas trough, which was significantly modified by later faulting and covered by Quaternary sediments.

The floor of the southern half of the basin was topographically higher, although modestly so, than the northern part, and the absence of sediments older than ca. 15.4 Ma indicates that rainfall, runoff, and any air-fall or epiclastic sediments left the area and were not retained. As noted above, the 15.2 Ma eruption of rhyolites in and along the western side of Boulder Valley may have blocked more southward streamflow and diverted it to the west to merge with the basin that was expanding from the north.

The post–15.4 Ma, westward-transported, epiclastic sediments did not reach the southwestern margin of the basin. Related streamflow may have either ponded and evaporated before reaching that area (although there is no evidence of evaporative conditions), or it may have drained south out of the basin, perhaps in the vicinity of the modern Rock Creek canyon.

CARLIN BASIN

The Miocene Carlin sedimentary basin was centered on an area in southwestern Elko County that includes the town of Carlin. The basin extended over a broad area bounded by what now are, in clockwise order from the southwest, Marys Mountain, the southern Tuscarora Mountains, Swales Mountain, the Adobe Range, and the Piñon Range (Fig. 7).

In general, sedimentation in the Carlin basin began at ca. 16.5 Ma and continued until after 14.6 Ma (Fig. 2). Faulting within a remnant Eocene upland produced the early basin, and coarse fluvial sediments were carried into the new lowland and southward toward Pine Valley. Fault-related subsidence and damming and the later 15.3 Ma eruption of the Palisade Canyon rhyolite at the south end of the basin produced a lake that expanded over much of the basin and across low divides in surrounding uplands, connecting with adjacent basins. The lake largely drained at ca. 15.2 Ma, and fluvial sediments then blanketed much of the basin, although this transition did not occur in the western part of the basin until ca. 14.6 Ma. Except for the initial basin-related faulting, all of the faulting within the basin and adjacent highlands took place after sedimentation. Late Miocene and younger erosion has preferentially removed large quantities of Miocene sediments and exhumed the pre-sedimentation paleotopography.

Stratigraphy

Air-fall ash and more locally derived epiclastic sediments comprise the sedimentary units in the Carlin Basin, which can be divided into four broad units. These include three interfingering, but increasingly younger, lower units—a basal, coarse epiclastic unit, a mixed epiclastic and ash-rich unit, and a fine-grained, ash-rich unit—and an upper, sandy to conglomeratic epiclastic unit (Wallace, 2005). Each unit, as described below, is extremely variable.

The basement rocks include discontinuous, late Eocene volcanic rocks that overlie various Paleozoic sedimentary rocks. Eocene volcanic rocks are widespread in the Swales Mountain area to the north and at Marys Mountain to the west (Evans and Ketner, 1971; Henry and Faulds, 1999). Only the Swales Mountain volcanic rocks extend beneath Miocene strata in the basin, and a few Eocene rhyolite flow units are present beneath the eastern part of the basin. Where Miocene and Eocene units are in depositional contact, the older rocks dip 15° to 20° more steeply than the younger strata.

Rhyolite to andesite flows were erupted during Miocene basin sedimentation. At the southwestern and southeastern margins of the basin, eruption of the 15.3 Ma Palisade Canyon rhyolite produced an immense pile of flows that is now exposed in an arcuate belt from Marys Mountain southeast to the northwestern end of the Piñon Range (Fig. 7). On the basis of thickness, number of flows, and flow features, the main eruptive centers were just north of Palisade (Fig. 7), and
Figure 7. Map of the Carlin basin area, showing the original known and inferred extents of the sedimentary basin, the middle Miocene andesite and Palisade Canyon rhyolite flow units, and dates of various sedimentary and volcanic units. Uncolored areas include areas of pre-Miocene basement rocks in the surrounding ranges. See Table 1 for geochronologic information. The modern digital elevation map is used as the base.
the flow package generally thins in all directions away from that area. Near Emigrant Pass, the rhyolite underlies 14.8 Ma and younger strata and overlies a very thin sequence of Miocene sedimentary units. Between Emigrant Pass and Palisade (Fig. 7), as well as on the west side of the Piñon Range, the rhyolite directly overlies Eocene and older rocks. At the north end of Pine Valley, the rhyolite overlies fluvioluvial Miocene units and underlies fine-grained strata that were deposited at ca. 15.3 Ma. In a small area north of Carlin, the rhyolite is exposed beneath some basin strata, and rocks in these exposures likely connect beneath sedimentary cover to the main rhyolite bodies to the southwest (Fig. 7).

Thin andesite flows were erupted along the eastern margin of and in the southwestern part of the basin (Figs. 7 and 8A). The eastern flow units now extend over the low crest of the southern Adobe Range into the southwestern margin of the Elko basin (Fig. 7). Flow units in both areas are at the contact between the ash-rich unit and the upper epiclastic unit; the top of the ash-rich unit in the eastern part of the basin was dated at 15.2 Ma. Petrographically identical andesite flow units in the Peko Hills, east of the Adobe Range, were dated at 15.1 Ma (Table 1; see the Elko basin descriptions). In the southwest- western exposures, north of Carlin, andesite overlies the rhyolite with a few meters of intervening ash-rich strata, and magnetic data indicate that the flow units extend a few kilometers to the southwest beneath the sedimentary cover (Plume, 1995).

The basal epiclastic unit is exposed in the northwestern, western, and northeastern parts of the basin (Wallace, 2005); it is concealed, if present, in the central and southern parts of the basin. In general, the unit includes thick sand beds, debris flows, and braided stream deposits typical of the medial to distal parts of an alluvial fan (Fig. 8B; Nilsen, 1982; Einsele, 1992). Thin-bedded, ash-rich sediments that might indicate lacustrine conditions are absent except in small, isolated areas. The basal epiclastic unit is older than 16.3 Ma tephas at the base of the overlying mixed epiclastic and ash-rich unit, but the age of initial sedimentation is unknown.

In the northwestern part of the basin, clast compositions and imbrications indicate source areas in the Tuscarora Mountains and southwestern Swales Mountain and southeastward streamflow toward the north-central part of the basin (Fig. 7). The unit thickens between Cottonwood Creek and the north end of Schroeder Mountain (Fig. 7) and may have been deposited in a southeast-draining paleovalley.

In the northeastern part of the basin, epiclastic sedimentary units include siltstone, pebble-rich sandstone, and conglomerate. Clast imbrications indicate southerly streamflow, and clast lithologies are identical to those in Swales Mountain and, to a lesser extent, the Adobe Range, similar to clasts in modern Susie Creek. A distinct, kilometer-wide horizon of conglomerate extends south-southwest toward the north-central part of the basin from the southeastern base of Swales Mountain before thickening into a 200- m-thick section of alternating conglomerate and finer grained beds, possibly representing a wide channel that fed a depocenter. Basal Miocene sediments in the northern part of Pine Valley to the south (Fig. 7) are epiclastic and underlie the Palisade Canyon rhyolite (see the later section on Pine Valley).

Most epiclastic beds contain at least some reworked air-fall material, indicating deposition of the pyroclastic materials on the source areas and simultaneous transport and deposition of both types of sediments. In the northwestern part of the basin, as well as in the Camp Creek area to the northeast (Fig. 7), some thick sand bodies are composed partially to largely of reworked, small, mafic to intermediate-composition pumice clasts with locally abundant small chips of Paleozoic rocks. In places, dominantly epiclastic sand grades upward over several meters into pumice-rich material, possibly the result of rafting of the more buoyant pumice. The ages and compositions of these coarser air-fall deposits are similar to those of volcanic rocks that were erupted along the northern Nevada rift to the west (John et al., 2000).

The ash-rich unit consists of thin- to thick-bedded, air-fall ash, locally abundant diatomite at the top of the unit, and thin, uncommon beds of limestone and chert. Epiclastic materials are extremely rare. Beds or packages of beds can be traced laterally for several kilometers. Most textures indicate deposition in a lacustrine environment, and wind ripple marks, eolian cross bedding, mud cracks, hot-spring sinter deposits, and algel mats indicate periodic subaerial exposure. Soft-sediment deformation is rare but can be pronounced locally, such as along the west side of the Adobe Range (Fig. 8C). The ash was derived from distant eruptions (Perkins and Nash, 2002), and coarser air-fall materials, such as those found in the basal fluvial unit, are absent. Almost all diatoms are planktonic (predominantly *Aulacoseira granulata*), with minor benthic diatoms near the bases of the diatomite sections, indicating fresh, deeper water. However, diagenetic formation of zeolites and Magadi-type chert in the basal ash-rich sediments in the middle of the basin indicate early alkaline, perhaps evaporative conditions (Sheppard and Gude, 1983; Wallace, 2005).

The ash-rich unit was deposited throughout much of southern two-thirds of the basin. This is the basal sedimentary unit in the eastern and southwestern parts of the basin (Fig. 7), as well as on topographic highs in the northern part of the area that were not initially covered by earlier sediments. The top of the ash-rich unit was deposited at ca. 15.5 Ma in the northern part of the basin, at ca. 15.2 Ma in the southeastern part, and after 14.8 Ma in the southwestern and westernmost parts (Table 1).

The mixed epiclastic and ash-rich unit is exposed throughout much of the northern part of the basin and is composed of alternating beds typical of both the ash-rich and the basal epiclastic units (Fig. 8D). This unit was deposited between ca. 16.3 and 15.7 Ma (Table 1). Given the southerly streamflow indicated by clast imbrications and source lithologies in this and the basal epiclastic unit, the strata likely are present beneath younger units in the central and southern parts of the basin. The proportions of both sediment types vary considerably both laterally and vertically through the section, reflecting the dynamic interplay between the fluvial and lacustrine environments. The first appearance of laterally persistent, thinly bedded, ash-rich beds defines the base of the mixed unit; the top of the unit is gradational into the ash-rich unit and is defined somewhat arbitrarily where epiclastic sediments are only a minor component of the whole. Southeast of Cottonwood Creek (Fig. 7), debris flows in the lower epiclastic unit interfinger southward into the mixed unit and indicate a southward change from purely fluvial to mixed depositional environments. Combined epiclastic and pumiceous beds, similar to those in the lower fluvial unit, are common, but the beds are much thinner, and some beds are composed of primary air-fall pumice.

The upper epiclastic unit was deposited beginning at ca. 15.5 to 15.1 Ma, depending on location; the upper age is unknown. Clast lithologies are similar to those in the lower epiclastic unit and indicate similar source areas and transport directions. Clasts derived from the Adobe Range, which contributed only minor amounts of fine-grained, epiclastic sediments to the lower epiclastic unit, are very common in the upper unit in the eastern part of the basin. All sediments in this unit become finer grained toward the center of the basin.

The upper unit is composed of alternating sand and gravel beds. The sand generally lacks sedimentary structures and probably represents overbank deposits that, on the basis of abundant rhyzoliths and some burrows, were bioturbated. Weak soil horizons, identified by their lighter color, increased carbonate cement, and upward termination of burrows and rhyzoliths, are common in some exposures. The unit typically is reddish and commonly calcareous; early hematite
Figure 8. Photographs of middle Miocene sedimentary units in the Carlin basin. A: Thin andesite flow units underlain by ca. 15.1 Ma, ash-rich strata (white); sediments of the upper epiclastic unit (not shown) overlie the andesite. Faulting and tilting occurred after basin sedimentation. Photo taken looking north along the west side of the Adobe Range. B: Sand, pebbles, and cobbles in braided stream and channel deposits of the basal epiclastic member; hammer for scale. Photo taken east of the Cottonwood Gulch area in the northwestern part of the basin. C: Soft-sediment deformation in the ash-rich unit along the west side of the Adobe Range. The deformed bed is ~1 m thick. D: Alternating fluvial, epiclastic sand beds and lacustrine ash-rich beds of the mixed epiclastic and ash-rich unit. The thick ash bed being sampled was deposited at 16.3 Ma. The underlying sand bed fills a 2-m-deep channel just to the right of the photo. Photo taken along upper Susie Creek in the northeastern part of the basin. E: Contact between the ash-rich unit (white) and the upper epiclastic unit (tan) along Interstate 80 west of Carlin. A tephra in the ash-rich unit near where the photo was taken produced a 14.7 Ma correlation age. The contact is conformable and, where exposed, sharp. F: The upper epiclastic unit (f) and a local zone of the underlying ash-rich unit (a, arrow), both of which overlie weathered and oxidized Paleozoic sedimentary rocks and the Gold Quarry gold deposit in the left side of the photo. The dashed white line delineates the sediment-bedrock contact. Photo taken looking north in the Gold Quarry mine on the west side of the Carlin basin.
coats sand grains, and later calcite cement fills the remaining pore spaces.

The contact between the upper fluvial and ash-rich units is conformable throughout most of the basin (Fig. 8E). In the eastern half of the basin, the contact is consistently just above a laterally persistent, ca. 15.2 Ma diatomite zone in the upper part of the ash-rich unit. In the center of the basin, the contact zone is only a few centimeters thick and represents a sudden influx of epiclastic sediments. Soil horizons in sandstone beds above the contact indicate a change from lacustrine to terrestrial environments.

Elsewhere, the contact zone varies in time and character. East of Cottonwood Creek (Fig. 7), the top of the ash-rich unit is ca. 15.5 Ma, older than at other locations in the basin. As such, part of the ash-rich sequence may have been eroded prior to deposition of the upper epiclastic unit, or the shift in depositional environment began earlier there than in other places. In the southeastern part of the basin, the contact zone immediately above the 15.2 Ma diatomite horizon spans several tens of meters, with alternating ash-rich beds and coarse conglomerates and debris flows derived from the southermost Adobe Range. This high-energy environment was restricted to this area and thus must have resulted from events, possibly a series of localized storms that were centered on and affected only the southernmost Adobe Range.

Fluvial Miocene sediments are exposed in the upper Maggie Creek area north-northeast of Cottonwood Creek. These nearly horizontal sediments, although poorly exposed, resemble the upper epiclastic unit. Gravity data indicate that the depth to basement in the Maggie Creek area is as much as 1700 m (Ponce and Morin, 2000). Given the thickness of and sedimentological relations in the Miocene sedimentary sections near Cottonwood Creek just to the east, much of that 1700 m must be pre-Miocene units. The juxtaposition of upper fluvial unit in the valley against the base of the lower epiclastic unit just to the east indicates a west-dipping normal fault along the east side of the Maggie Creek valley.

In the far western part of the basin, between Welches Canyon and the Gold Quarry mine (Fig. 7), all four units are present, although the ash-rich unit is much thinner. The overall form of this sedimentary package and the along-strike contact relations with basement rocks to the south and north suggest that this area was a moderately deep, east-northeast–trending paleovalley. The basal epiclastic unit is much thinner and finer grained than it is elsewhere, although conglomerates with clasts derived from nearby gold deposits comprise some of the basal beds (Norby and Orobona, 2002). The 14.8 to 15.2 Ma mixed unit overlies the basal unit and is composed primarily of fine-grained epiclastic sediments, some pebble conglomerates, and locally thick, air-fall ash and pumice deposits.

On the east side of Schroeder Mountain, northeast of the Gold Quarry mine (Fig. 7), a distinct lacustrine unit is absent, and massive epiclastic sand beds in its place are extremely ash rich. These beds resemble the upper epiclastic unit but contain the abundant ash typical of the ash-rich unit, suggesting a mixed but homogenized environment. These units overlie 14.8 Ma tephras. On the basis of dates and field relations in this and the Welches Canyon areas, both ash-rich and epiclastic sediments were being deposited in the western part of the basin at the same time that the upper epiclastic unit was being deposited in the eastern half of the basin.

Broad areas along the west side of the Adobe Range and the east side of Marys Mountain remained exposed until covered by the ash-rich unit at ca. 15.3 Ma. The resulting sedimentary cover in those areas is much thinner than in the rest of the basin. As such, these areas were topographically higher during early sedimentation than the central part of the basin. The basal epiclastic unit was deposited along the western base of the Adobe Range bench, and lacustrine sediments overlie both the bench and the older epiclastic strata. The western edge of the bench in part coincides with a major post-sedimentation fault (Wallace et al., 2007a; this fault may have been active before or during sedimentation, or, alternatively, the bench may have been a pediment formed during pre-middle Miocene erosion.

On the east side of Marys Mountain, 14.8 Ma and younger ash-rich and upper epiclastic sediments were deposited on the Palisade Canyon rhyolite and, to the north, Paleozoic rocks. At the Gold Quarry mine, some ash-rich strata overlie the Paleozoic basement and gold deposit, but the upper epiclastic unit directly overlies many parts of basement rocks (Fig. 8F). Surface exposures east of Emigrant Pass and drilling data from south of Gold Quarry indicate that the base of the Miocene section thins to the south-west onto the area underlain by the rhyolite. The upper epiclastic unit gradationally overlies the ash-rich unit, extends out into the middle of the basin northwest of Carlin, and conceals any underlying relations.

Connections with Other Basins

Strata in the Carlin basin extend past modern ranges and connect with strata in adjacent basins. In the Camp Creek area to the north-northeast on the east side of Swales Mountain (Fig. 7), 15.5 to 15.3 Ma sediments are similar to and continuous with the basal epiclastic and mixed units in the Carlin basin and contain both epiclastic sediments and air-fall ash and dark pumice similar to that in Carlin basin (Lovejoy, 1959). These strata extend north into the main part of Independence Valley, as described in a later section.

At the south end of the Adobe Range, andesite flow units and the ash-rich and upper epiclastic units extend across the low crest of the range and connect with similar units on the east side of the range, which is part of the Elko basin. Two tephra beds stratigraphically beneath the andesite on the east side were deposited at 15.3 and 15.1 Ma, similar to the age of the ash-rich unit in the southeastern part of the Carlin basin. Thus, a low-energy, presumably horizontal sedimentary environment extended across what is now the modern crest of the Adobe Range, and the andesite lavas were able to flow over the entire area. Epiclastic sediments, including coarse debris flows, on the east side of the range, were derived from rocks exposed nearby, indicating a highland just north of the interbasin connection. These epiclastic deposits overlie the ash-rich unit and thus formed at approximately the same time that coarse conglomerates were deposited at the base of the upper epiclastic unit in the nearby Carlin basin.

Ash-rich and epiclastic sediments are exposed at and west of Emigrant Pass, tentatively connecting with similar sediments in the southwestern part of the Carlin basin. Some of these mixed sediments underlie the 15.3 Ma Palisade Canyon rhyolite. The base of the section locally is a conglomerate, which underlies ash-rich units. Ash-rich sediments are exposed as far west as Bobs Flat (Fig. 7; Peters, 2003), and the depositional environment may have extended beyond the present Marys Mountain highland and connected with a Miocene depocenter on the west side of the Cortez Range (Stewart and Carlson, 1978).

Sedimentation, Paleogeography, and Faulting

A late Eocene topographic high extended north-northwest from the northern Piñon Range, through the future site of the Carlin basin, and along the Tuscaraora Mountains (Fig. 9; Haynes, 2003). The Swales Mountain area was on the northeast flank of the paleohigh, and the Marys Mountain area was on the southwest flank. The Adobe Range also was a modest topographic high that extended northeast from the other topographic high, creating a “V”-shaped highland.

Late Eocene volcanic rocks overlie the paleosurface in at least the Marys Mountain and Swales Mountain areas. Their absence in other areas could be due to nondeposition or erosion.
The late Eocene gold deposits in the Tuscarora Mountains, at Gold Quarry, and in the northern Piñon Range (Rain deposit), all of which formed within the paleohighs, were deposited at least 800 m beneath the paleosurface (Hickey et al., 2003; Ressel and Henry, 2006).

A series of generally north-striking faults cut and tilted the late Eocene volcanic and sedimentary units in the Marys Mountain and Swales Mountain areas, as well as a 26 Ma tuff in the Marys Mountain area (Henry and Faulds, 1999; Wallace, 2005). The angular unconformities between the volcanic units and overlying Miocene basin strata, as well as the apparent lack of fanning of stratal dips in the latter units, indicate that this faulting ceased before the onset of basin sedimentation (Henry and Faulds, 1999; Wallace, 2005). Supergene alunite ages of 30.0–18.5 Ma from the gold deposits (Table 2) indicate that protracted middle Tertiary erosion induced supergene weathering of the deposits.

The tops of the gold deposits were exposed by the middle Miocene, and clasts from the deposits were shed into the basal Carlin basin sediments. Therefore, at least 800 m of erosion took place in the highlands prior to middle Miocene sedimentation.

Figure 9. Paleogeographic map of the Carlin basin area during its early stages of formation. Shown are late Eocene highlands (estimated from page-size figure in Haynes, 2003), possible early-basin normal faults, and the distribution of the basal fluvial unit of the Humboldt Formation, which was the first Miocene unit to be deposited in the basin. Shown also is the approximately outline of the Carlin gold trend (yellow lines), from which numerous Oligocene to early Miocene supergene alunite dates were obtained.

The deposition of the coarse, fluvial sediments of the basal epiclastic unit in at least the western and northern parts of the basin indicates the creation of relief and a depocenter. The thickest parts of the basal epiclastic unit are just east of the Tuscarora Mountains, which might suggest that uplift of the Tuscarora Mountains initiated early fluvial sedimentation. However, low-energy lacustrine sedimentation was taking place in the Ivanhoe basin, on the western side of the Tuscarora Mountains area, at the same time that this much higher energy fluvial sedimentation was taking place in the western part of the Carlin basin. Uplift of the highland should have shed coarse epiclastic sediments in both directions, not just one. Therefore, fault-related downdropping of the east side of the Tuscarora Mountains highland shortly before 16.3 Ma, possibly along the Tuscarora fault (Evans, 1980; Ressel and Henry, 2006), most likely created the early basin into which the sediments were deposited.

A comparison of the areas where early fluvial sediments were deposited and areas where sediments were not deposited until later provides some idea of where possible controlling faults, as well as topographically higher areas, may have been located (Fig. 9). As noted above, the Tuscarora fault likely provided a structural control to the northwestern part of the basin. Thickness distributions of the basal epiclastic unit suggest the presence of an elongate upland at and north of Schroder Mountain during early sedimentation. To the north, the early epiclastic unit laps onto the southwest, south, and east sides of Swales Mountain. On the basis of angular unconformities between the Eocene volcanic rocks and overlying Miocene units, that block was tilted to the west prior to sedimentation, possibly by a fault on its east side (Evans and Ketner, 1971). However, only ash-rich sandstones are present along the eastern flank of Swales Mountain, indicating little relief.

Early sedimentation apparently was concentrated in the central part of the basin, flowed to the south toward Pine Valley, and did not begin to cover flanking areas to the east and west until ca. 15.3 Ma. As noted earlier, facies and age relations suggest that the west side of the Adobe Range bench may have been fault controlled just prior to or during early sedimentation (Fig. 9). To the southwest, a major north-striking fault offset the 15.3 Ma Palisade Canyon rhyolite along the Humboldt River north of Palisade (Fig. 9) and extends north into the Carlin basin east of Marys Mountain. Although not exposed, the facies relations imply that it may have been
able overlying upper epiclastic unit in that area. However, based on source lithologies for concentration of lacustrine sedimentation in that amount of westward tilting of the basin and These western relations could indicate a mod-

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rhyolite flows did not even begin until 14.8 Ma. of the basin, sedimentation above some of the post-Eocene bedrock are exposed in the Elko and Peko Hills and Cedar Ridge, and a paleoknoll of Jarbidge rhyolite is exposed at Black Butte. According to the volcanic pile was thick, raised the base level of the basin. At ca. 15.2 Ma, the lake that covered most of the basin drained, and fluvial sediments derived primarily from the east, north, and northwest were deposited across much of the basin floor to form the upper epiclastic unit. The general sharpness of the contact between the lacustrine and fluvial units indicates that the change in depositional environment happened quickly. Given that the ensuing fluvial sedimentation in the basin was directed to the south, failure of the volcanic dam at the south end of the basin was the probable cause for lake drainage.

The coarseness of the sediments in the upper fluvial unit could indicate an increase in relief in the source areas relative to the Carlin basin. However, no faults that can be attributed to uplift at that time have been identified along the flanks of or within the highlands. The failure of the dam may have lowered the base level and perhaps increased the stream gradients in the basin, allowing the transgression of coarser sediments across the basin.

On the western side of the basin, a mixed fluvial and lacustrine environment persisted until after 14.8 Ma, and, in the southwestern part of the basin, sedimentation above some of the rhyolite flows did not even begin until 14.8 Ma. These western relations could indicate a modest amount of westward tilting of the basin and concentration of lacustrine sedimentation in that area. However, based on source lithologies for the clasts, streams that deposited the conformably overlying upper epiclastic unit in that area flowed to the east. More likely, the western part of the basin may have been somewhat iso-

basin-sediment contacts and subsequent preferential erosion of the Miocene sediments relative to the resistant basement since the late Miocene (see later section; Wallace, 2005).

Some sediment-basement contacts, such as those on the east side of the Tuscaraora Mountains and west side of Schroeder Mountain, are faulted. Schroeder Mountain (Figs. 7 and 9) was a low upland that diverted early stream flow and, at its north end, remained exposed throughout basin sedimentation. West-dipping, post-sedimentation normal faults along the west side of Schroeder Mountain extend to the north, in part creating the major gravity low along Maggie Creek (Ponce and Morin, 2000). The sediment-basement contact along the east side of Schroeder Mountain is depositional, but the sediments in that immediate area dip west into and ramp onto the basement block. This most likely indicates an east-dipping normal fault within the block.

ELKO BASIN

The Elko basin is an extensive, elongate basin located between the Ruby Mountains and East Humboldt Range to the east and the Adobe Range and Piñon Range to the west. Its southern end is south of Jiggs along Huntington Creek, and it extends far to the north along the Marys River Valley (Fig. 10). Small islands of pre-

Miocene bedrock are exposed in the Elko and Peko Hills and Cedar Ridge, and a paleoknoll of Jarbidge rhyolite is exposed at Black Butte. In general, the area of the Elko basin was in the hanging wall of the Tertiary, west-dipping Ruby–East Humboldt detachment and high-angle fault systems. Any sediments produced during detachment-related uplift are largely absent in or near the Elko basin, and the area either drained externally or, like the Carlin basin, was not a lowland. The oldest post-Eocene sediments were deposited just before 15.4 Ma, approximately the same time that high-angle faulting along the range front began; the rate of uplift peaked between ca. 15 and 14 Ma and then continued at a slower rate. During early sedimentation, alluvial fans drained eastward from the structurally passive Adobe and Piñon Ranges into the developing half-graben basin, and a lacustrine environment began to form in the southern half of the basin at ca. 15.4 Ma. The alluvial sediments partially buried paleohills of the Elko and Peko Hills and Cedar Ridge, and uplift of the Ruby Mountains and East Humboldt Range progressively shed coarse materials into the eastern part of the basin. Sedimentation continued until ca. 9.8 Ma, when the basin began to drain externally, likely between the Adobe and Piñon Ranges, and connected with the Carlin basin. This drainage was blocked temporarily in
the middle Pliocene, forming a shallow lake that covered much of the Elko basin before draining at ca. 2 Ma. Late Miocene and younger erosion and removal of enormous amounts of Miocene sediments from the Elko basin have reexposed much of the pre-middle Miocene paleotopography along the western side of the basin.

**Stratigraphy**

On the west side of the Elko basin, Miocene strata are composed predominantly of conglomerate and clast-rich sandstone that resemble alluvial-fan deposits. At the southeastern end of the Adobe Range, local debris-flow deposits in the ca. 15.2 Ma basal part of the section contain blocks of Paleozoic rocks up to several meters in dimension. Erosional windows through the Miocene units reveal that the sediments were deposited on a gently east-dipping paleosurface on the east side of the range. Underlying units include various Paleozoic, Mesozoic, and Eocene units (Figs. 11A and 11B), and the Eocene units dip more steeply than the overlying Miocene strata. Clast lithologies in the sediments match Paleozoic and Eocene source rocks directly to the west in the range, and the grain size decreases away from the range. The very gentle eastward dip of these beds may be primary, indicating relatively little, if any, post-sedimentation tilting along the east side of the Adobe Range.

The Peko Hills are surrounded and overlapped by similar but somewhat finer grained, epiclastic sediments (Ketner and Evans, 1988), and the sedimentary package on the east side of the Adobe Range projects directly into these deposits (Figs. 10 and 11B). This laterally continuous sequence is preserved across the Devil’s Gate area just north of the Peko Hills, where the epiclastic sediments extend over steeper dipping Eocene volcanic units and continue eastward toward the Twin Buttes area (see later description). At the south end of the Peko Hills, an andesite flow was erupted during sedimentation at 15.1 Ma (Table 1). Similarly, the Adobe Range sedimentary package projects east to the Elko Hills (Fig. 10), which are partially surrounded by and locally overlain by finer grained, epiclastic sediments (Smith and Ketner, 1978; Coats, 1987; Ketner, 1990). As such, the Peko and Elko Hills appear to have been middle Miocene, basement-cored hills that progressively were buried by the east-flowing alluvial fan systems.

East-dipping Miocene basin sediments are exposed extensively in the southern part of the basin, especially along Huntington Creek and other areas between the Piñon Range and the Ruby Mountains (Fig. 10; Smith and Ketner, 1976; Smith and Howard, 1977). In the Huntington Creek area, sedimentation began shortly before 15.4 Ma and continued to after 9.9 Ma (Table 1; Perkins et al., 1998). The strata, with a thickness of more than 560 m, were deposited in both fluvial and lacustrine environments. Conglomerates comprise the basal sediments; due north of Cedar Ridge, coarse conglomerates derived from the northern Piñon Range fill an east-trending paleovalley (Fig. 11D). With the exception of these basal conglomerates, ash-rich units generally are more common in the lower half of the section, and beds are thin bedded and laterally continuous. The sediments have increasingly more epiclastic materials upslope (Smith and Ketner, 1976), and they include sandstone, pebble-rich sandstone, and conglomerate. Clasts were derived from
Figure 11. Photographs of middle Miocene sedimentary units in the Elko basin. A: Basal Miocene, epiclastic sand and conglomerate beds on the west side of the Adobe Range. Photo taken looking north along the western outskirts of Elko. The beds overlie more steeply dipping volcanic units of the late Eocene Indian Wells Formation (light-colored, below). The gentle eastward dip of the Miocene units may be nearly primary. B: Peko Hills (behind the green floodplain of the North Fork Humboldt River) and fluvial Miocene sediments (tan, foreground) on the eastern dip slope of the Adobe Range. The fluvial sediments continue across the river valley, underlie the low bench to the left (north) of the Peko Hills, and depositionally overlie Paleozoic rocks in the Peko Hills, indicating that the hills were paleohighs during sedimentation. The snowcapped peaks in the background are the Ruby Mountains. Late Cenozoic erosion of the Miocene strata during downcutting of the river has progressively exposed the Peko Hills. C: Conglomerate in an east-trending paleovalley north of Cedar Ridge (Fig. 10). The clasts were derived from Paleozoic units exposed in the Piñon Range to the west. When the channel filled, finer grained epiclastic sand and pebbles formed a continuous cover over the channel fill and adjacent bedrock areas. D: Pebble-rich fluvial sediments in the Huntington Creek area in the southern part of the Elko basin. All of the clasts are crystalline metamorphic and igneous rocks derived from the Ruby Mountains to the east. These fluvial beds overlie ash-rich lacustrine units along Huntington Creek. Arrow points to 6-cm pocket knife for scale. E: Ash-rich, thin-bedded lacustrine sediments in a railroad cut near Wells. Green beds to the right have been altered to chert and zeolites. An unaltered tephra from this sequence was dated at 10.5 Ma. Hammer in right-center of photo (arrow) provides scale. F: Horizontal Pliocene (ca. 2.1 Ma) lacustrine units (light, P) cover lowlands cut into gently west-dipping fluvial strata of the middle Miocene Humboldt Formation (M, tan in middle distance). Photo taken looking east-northeast at the Secret Pass area between the East Humboldt Range (left horizon) and the Ruby Mountains (to right of photo).
both the Piñon Range to the west and the Ruby Mountains to the east, and the influx of Ruby-derived gneissic clasts beginning at 14–14.5 Ma records the progressive uplift and incision of the Ruby Mountains (Fig. 11E; Sharp, 1939; Smith and Ketner, 1976). Cedar Ridge, a low, Paleozoic-cored knoll within the basin east of the Piñon Range, contributed clasts to the adjacent basal conglomerates and then was partially covered by ash-rich sediments (Smith and Ketner, 1976, 1978), a burial history similar to that in the Pekol and Peko Hills.

Sharp (1939) studied the Miocene sediments along the Ruby Mountains and the East Humboldt Range between Huntington Creek and the Wells area. Overall, he found that the section grades upward from primarily lacustrine to entirely fluvial, and that the majority of the strata are composed of fluvial sandstone to mudstone. West of Secret Pass (Fig. 10), some of the fluvial units contain turbidites, indicating local subaqueous deposition, but most of the epiclastic units were deposited in streams and alluvial fans derived from the Ruby Mountains and East Humboldt Range to the east. The units dip gently to the east except immediately adjacent to the range front, where they dip steeply east and are in fault contact with moderately west-dipping normal faults related to uplift of the ranges (Sharp, 1939; Snoke et al., 1997).

The Miocene sediments in the northeastern East Humboldt Range and southern Snake Range near Wells (Fig. 10) record a protracted period of sedimentation in the hanging wall of the gently west-dipping Marys River fault system, which is exposed on the west sides of the two ranges and extends beneath the Elko basin (Effimoff and Pinezich, 1981; Robison, 1983; Mueller and Snoke, 1993). Most of the epiclastic sediments in the lower two-thirds of the section were deposited in low-energy, fluvial and lacustrine environments, but thin to thick zones of coarse conglomerate indicate periodic higher energy sedimentation. Abundant air-fall ash in the upper third of the sequence was deposited in a lacustrine to mudflat environment (Fig. 11E). The exposed Miocene section in the southern Snake Mountains is more than 1000 m thick (Thorman et al., 2003), and seismic data in the basin to the west suggest that the Miocene sediments there may be ~3000 m thick (Robison, 1983). Fission-track dates on detrital zircon in the middle of the Snake Mountains sequence indicate a depositional age of less than 11 Ma (Thorman et al., 2003), and the upper, ash-rich section was deposited at ca. 10.3 Ma (Table 1). Fluvial to lacustrine sediments underlie 13.4 to 15.1 Ma rhyolite flow units in the East Humboldt Range and Snake Mountains, but these may be Eocene to Oligocene and unrelated to the Miocene sedimentation (Good et al., 1995; Snoke et al., 1997; Thorman et al., 2003).

Basin sediments in the Marys River area north of the Humboldt River are predominantly fine-grained, epiclastic sands and silts with variable amounts of conglomerate and primary and reworked air-fall ash. At Black Butte (Fig. 10), these sediments largely underlie but locally overlie the Jarbidge Rhyolite, which forms the resistant butte. Clasts were derived from the Snake Mountains to the east, and epiclastic sediments grade northwest into more thin-bedded, ash-rich sediments (Smith et al., 1990). Just to the southwest of Twin Buttes (Fig. 10), andesite flow units identical to the 15.1 Ma flows in the Pekol Hills overlie beds of reworked air-fall ash and small-pebble conglomerate derived from the west. This area is 10 km due east of Devils Gate and “downstream” from the alluvial fan on the east side of the northern Adobe Range, as described earlier; as such, these epiclastic sediments may be a distal part of that fan.

Much of the central part of the Elko basin is directly underlain by ash-rich to epiclastic sediments that, on the basis of limited tephra data, are middle Miocene in age (Fig. 11F; Reheis et al., 2003; Wesnousky and Willoughby, 2003; M.E. Perkins, 2007, unpubl. data). These generally horizontal sediments contain fluvial strata near the Ruby Mountains and fine-grained lacustrine units in the middle of the basin, and they were deposited during a hiatus in episodic late Miocene and younger erosion in the upper Humboldt River basin (Wallace, 2005). One drill hole east of the Elko Hills indicates a thickness of ~275 m. Similar Pliocene sediments are exposed south and southwest of Elko (Smith and Ketner, 1976). The Pliocene sediments are very similar to the Miocene sediments, including the distribution of the sedimentary facies, suggesting deposition in a similar topographic setting.

Despite this blanket of Pliocene sediments, oil drilling, seismic, and gravity data provide some clues about the basal morphology of the Tertiary basin at depth. One complication with using the gravity data is that Miocene sediments are not the only basin fill in most areas, as shown in oil drilling records (Hess, 2004), and the depth-to-basement data reflect the thickness of all Cenozoic units, not just the Miocene sediments. The Eocene Elko Formation alone is consistently at 2 km of that depth as “upper Tertiary” sediments (Effimoff and Pinezich, 1981), but that would include overlying Pliocene and Quaternary units, whose thicknesses are unknown, as well as the Miocene strata. In contrast, the depth to basement in the narrow zone between the Adobe Range and the Elko and Peko Hills to the east is generally just a few hundred meters. In those areas, full exposures of the Miocene sediments are 100 m or less thick and the underlying Elko Formation and Eocene volcanic rocks, where present, comprise the remainder of the basin fill above the Paleozoic basement.

**Structure and Faulting**

The western side of the Elko basin is largely unfaulted and is composed of Miocene strata that overlie a gently west-dipping paleosurface cut into older rocks. These Miocene units extend east to and across outboard pre-Miocene paleohighs. The steeper dips on underlying Eocene volcanic and sedimentary units in all of these areas indicate a poorly constrained, pre-Miocene period of tectonism that may have produced the outboard horsts prior to Miocene sedimentation.

Faults along the west side of the Elko basin are relatively minor and formed after sedimentation. A west-dipping fault along the northern end of Cedar Ridge had ~700 m of post-sedimentation offset, based on offset tephras, but the sediment-basement contacts along the main part of Cedar Ridge are depositional (Smith and Ketner, 1978; M.E. Perkins, 2007, unpubl. data). An east-dipping fault within the type section of the Humboldt Formation along Huntington Creek had ~200 m of displacement. West-dipping normal faults along the southeast flank of the Adobe Range and the southern Elko Hills juxtaposed Paleozoic basement and the basal Miocene sediments, and offsets likely were 200 m or less.

In contrast, the west- to northwest-dipping Ruby–East Humboldt detachment fault and west-dipping normal faults are present along the east side of the Elko basin from Wells to south of Jiggs (Fig. 10). Various isotopic, thermal, and barometric data show significant cooling in the Oligocene and early Miocene, which some workers attributed to detachment-related uplift of those ranges (see summary in Howard, 2003). Fission-track uplift data indicate that uplift and eastward tilting of the southern Ruby Mountains, related to high-angle normal faulting along the west side of the range, peaked at ca. 15–14 Ma; this event may have affected the northern part of the Ruby Mountains as well (Colgan and Metcalf, 2006). Stratigraphic and fission-track data indicate that more than 4 km of Paleozoic sediments were eroded from the
Ruby Mountains during this stage of uplift (Smith and Ketner, 1976; Colgan and Metcalf, 2006). The high-angle faults along the west side of the Ruby Mountains continue south to about Bald Mountain (Fig. 10), at which point the polarity of the intermontane basin reverses. The Newark basin, south of the Elko basin, dips to the west into range-bounding faults along the east side of the Diamond Mountains, and the much shallower, intervening part of the basin at the polarity reversal likely contains a structural transfer zone (Wallace et al., 2007b).

In the vicinity of Wells, the west-dipping Marys River fault controlled the western sides of the southern Snake Mountains and northern East Humboldt Range. The complex tectonic and sedimentary features in this area indicate the formation of an eastward-deepening middle Miocene and younger basin in the hanging wall of the fault system (Effimoff and Piniaczik, 1981; Robison, 1983; Mueller and Snoke, 1993; Thorman et al., 2003). Post-sedimentation (in this area, after ca. 10.3 Ma) faulting has significantly modified the basin margin, which may have extended a few kilometers east of Wells (Mueller and Snoke, 1993), and Pleistocene and younger faults have offset terrace deposits along the entire range front (Dohrenwend et al., 1996).

Farther north, the sediment-basement contact along the west side of the Snake Range is both depositional and fault controlled (Coats, 1987; Smith et al., 1990; Thorman et al., 2003). The range front east of Black Butte forms an arcuate topographic embayment (Fig. 10), and Miocene sediments exposed throughout the embayment dip moderately east into the range front. Late Cenozoic faults are closer to but outboard from the range front west of Wells, and they extend far out into the basin west of the Black Butte embayment (Dohrenwend et al., 1996). The basin deepens west of these faults (Ponce 2004), with a shallowly buried pediment between the faults and the Snake Mountains.

The Wells fault is a west-striking transfer zone that has been projected west from the southern Snake Mountains toward the northern end of the Adobe Range (Thorman and Ketner, 1979). It has been cited as the northern limit of detachment-related faulting in the Ruby Mountains and East Humboldt Range (Mueller et al., 1999; Howard, 2003). If detachment-related faulting was active during Miocene sedimentation, then this fault should have been active as well. However, depth-to-basement data indicate that the mapped fault projects directly into the center of the deepest part of the basin, rather than forming its northern flank as might be expected, if it was a significant depocenter control. Pliocene sediments cover much of the projected trace of the fault through the Elko basin; the Miocene lithologies, where exposed in this area, are monotonous, and detecting offset in them would be extremely difficult.

In contrast to the Carlin basin, significant intra-basin faults have not been identified in the Elko basin, although small-offset faults are evident. In large part, this may be due to the poor exposures of Miocene sediments and to the extensive cover of flat-lying Pliocene sediments. However, the consistent flatness and lack of offset of the younger sediments over a very wide area indicates that any intra-basin faults that may be present were active before Pliocene sedimentation. Locally chaotic to opposing dips in Miocene units east of Huntington Creek (Smith and Howard, 1977) may indicate some concealed faults near the range front, but offset apparently was minimal given the overall lateral continuity of that section.

Paleogeography and Basin Evolution

Similar to the other basins studied, late Eocene to middle Miocene units are very rare in or near the Elko basin. Although thermochronologic data have been interpreted to indicate middle Tertiary, detachment-related uplift of the Ruby Mountains and East Humboldt Range, no sediments of that age that might record that uplift are present in the basin except in a small area just south of Wells (Snoke et al., 1997). Middle Tertiary sediments may be present beneath younger cover in the deep, middle parts of the basin, but these are not evident on seismic lines or in drill-hole records. Sedimentary basins in the hanging walls of detachment systems are common (cf. Wagner and Johnson, 2006); therefore, the apparent absence of such a depocenter in this area is notable. Although the present study did not focus on this issue, the character of the pre-middle Miocene paleogeography is relevant to later sedimentation.

Two scenarios are possible. First, the area that now includes the Elko basin and the Ruby Mountains and East Humboldt Ranges formed a broad upland surface that connected to the west with the remnant uplands of the Adobe and Piñon Ranges and the Carlin basin that were described earlier. Supergene alunite dates of 22–18 Ma from the northern Piñon Range (Table 2; Williams, 1992), indicate that area was exposed and weathering prior to middle Miocene basin formation. As such, the cooling recorded in the thermochronologic data may reflect some middle Tertiary uplift, but without the formation of a hanging-wall basin, and sediments generated during erosion of the upland were transported far beyond the area, as was postulated earlier for the Carlin basin. In this scenario, the middle Miocene formation of the high-angle faults along the west sides of the modern Ruby Mountains and East Humboldt Range (Colgan and Metcalf, 2006) cut the upland, created an east-dipping half graben, and induced early alluvial-fan sedimentation in the new basin.

Alternatively, the cooling data record uplift and the formation of an east-dipping, hanging-wall basin above the detachment fault, but the basin drained externally. In this case, all sediments generated were very efficiently carried out of the basin and produced the same non-sedimentary result. As with the first scenario, a relict older highland remained in the passive Adobe and Piñon Ranges, with supergene weathering in the latter area. In this scenario, the middle Miocene, high-angle faulting accentuated a half-graben setting and closed the once-open basin (or reduced its gradient), possibly along the accommodation zone at the south end. Neither scenario contradicts the presence of an Eocene uplift east of an Eocene Elko basin, as described by Haynes (2003). In both cases, however, erosion did not appreciably remove late Eocene sedimentary and volcanic units that were deposited and still remain in or near many parts of the Elko basin.

Regardless of earlier scenarios, the combined stratigraphic, structural, and depth-to-basement data indicate that the middle Miocene Elko basin was an east-dipping half graben. The west side was structurally passive during sedimentation, and ca. 15.3 Ma, east-dipping alluvial fans extended out across the basin, partially burying outboard, pre-Miocene paleohorsts. At ca. 15.4 Ma, outflow was blocked, possibly at the south end, and an early lake formed along at least the southern axis of the basin. Clast lithologies record the 15 to 14 Ma rapid uplift and erosion of the Ruby Mountains to the east (Smith and Ketner, 1976; Colgan and Metcalf, 2006) as fluvial sediments inundated the lacustrine setting.

The eastern part of the basin probably continued to deepen structurally during sedimentation, as reflected in the protracted sedimentation in that area between 15.4 and 9.9 Ma (Table 1) and the thicker section along the east side of the basin shown by gravity, seismic, and drilling data (see analogous descriptions in Rosendahl, 1987; Chapin and Cather, 1994; Faulds and Varga, 1998). However, the near-original dips of the alluvial fans along the Adobe and Piñon Ranges (Fig. 11A), as well as consistent dips in the full stratigraphic section near Huntington Creek (Smith and Ketner, 1976), do not support a progressive eastward tilting and deepening of the basin. In addition, the ability of the western, non-tectonic side of the basin to connect with the Carlin basin and Independence Valley at ca. 15.2 Ma indicates that the rate of early sedimentation in...
the basin as a whole outpaced the rate of structural subsidence, allowing the basin to fill up.

Miocene sedimentary units are exposed discontinuously between the northwestern part of the basin and the Independence Valley. Immediately north of this area, the middle Miocene Jarbridge Rhyolite is widely exposed, and its present southern extent generally coincides with the northern limit of middle Miocene sedimentary units (Fig. 10). The poorly dated rhyolite overlies most of the sediments at Black Butte and underlies them in the northeastern Independence Valley (see the ensuing section). These extensive, thick, and roughly coeval rhyolite fields may have blocked the expansion of the basin or external drainage to the north.

The youngest dated sediments are 9.9 Ma at Huntington Creek and ca. 10.5 Ma near Wells, but younger sediments were not deposited until the formation of the relatively short-lived middle Miocene lake. However, the Ruby Mountains and East Humboldt Range were notable adjacent highlands that must have shed detritus into the basin. The absence of sediments deposited after ca. 9.9 Ma indicates that the basin began to drain externally at that time, but the outlet or outlets for the basin are unknown. Significant amounts of Miocene sediments were eroded and removed from the basin before the Pliocene lake event (see later section; Wallace, 2005); therefore, the original outlet actually may have been topographically higher than the present base level. Early formed strath terraces are paired around the Humboldt River, which flows southwest out of the basin at Carlin Canyon, suggesting that the west-draining river began to form at that time.

**PINE AND INDEPENDENCE VALLEYS**

Pine and Independence Valleys were examined only in a reconnaissance fashion. In the middle Miocene, these basins connected with adjacent basins and thus merit brief descriptions.

Pine Valley is due south of the Carlin basin and lies between the Piñon Range to the east and the Cortez Range to the west (Figs. 1 and 7). Except at the very north end of the valley, Miocene sediments in Pine Valley are concealed beneath the Pliocene and Pleistocene Hay Ranch Formation but are known to be present based on seismic and drilling data (Gordon and Keller, 1993; Hess, 2004). The Miocene sediments include the Carlin Formation and the upper part of the Raine Ranch Formation as described by Regnier (1960) and reclassified as the middle Miocene Humboldt Formation by Smith and Ketner (1976).

On the basis of field, seismic, and drill-hole data, the Miocene sediments in Pine Valley thicken modestly to the east and likely were deposited in an existing topographic basin between the two ranges (Regnier, 1960; Smith and Ketner, 1976; Gordon and Keller, 1993; Muntean et al., 2001; Hess, 2004). In the northern end of the valley, the lower part of the section is a mixture of fluvial sand, mud, and conglomerate. Identifiable clast lithologies on the eastern side of the basin include Paleozoic sedimentary and Eocene igneous lithologies exposed in the Piñon Range and, permissively, southern Adobe Range; clast imbrications indicate predominantly westward flow from the Piñon Range, but also some southward flow from the area of the Carlin basin. On the western side of the basin, clast lithologies include Jurassic volcanic rocks exposed in the Cortez Range to the west (Smith and Ketner, 1976). In general, the sediments are finer grained and more ash rich upsection. High in the exposed section, a meter of silicified breccia composed of the Palisade Canyon rhyolite resembles a very distal, autobrecciated lava flow or flow-end rubble bed (also noted in Raine Ranch Formation section 3 of Regnier, 1960). One tephra sample collected ~20 m above the rhyolite breccia unit produced a 15.3 Ma correlation age (Table 1; Fig. 7), which also is the age of the rhyolite (Table 1). Therefore, the majority of the fluvial sediments in northern Pine Valley were deposited prior to 15.3 Ma, and their lithologies indicate the presence of highlands to both the west and east.

Previous workers (Regnier, 1960; Smith and Ketner, 1976; Gordon and Keller, 1993) concluded that the basin was not tectonically active during middle Miocene sedimentation. As with the Carlin basin to the north, sedimentary units older than perhaps 16 Ma (based on the 15.3 Ma date near the top of the section) are absent. Eocene and older units on the east side of the Cortez Range dip moderately to the east, beneath the basin and against the Piñon Range. The angular unconformity between the older and the middle Miocene units indicates pre-middle Miocene tilting, presumably related to a west-dipping normal fault along the west side of the Piñon Range. Dips of the Miocene sediments do not consistently decrease upsection; therefore, the tilting event either occurred prior to or during the very early stages of basin sedimentation. If it was before, then the absence of older sediments suggests that the area drained externally until ca. 16 Ma.

The Pliocene and Pleistocene Hay Ranch Formation overlies the Miocene sediments along a very low-angle unconformity (Regnier, 1960; Gordon and Keller, 1993). The distribution and thicknesses of the two units are very similar, although the Hay Ranch Formation is much finer grained and ash rich overall. Coarse epiclastic units in the eastern part of the valley are concentrated near the Piñon Range front, whereas those in the western part extend far out into the basin; lacustrine deposits are closer to the Piñon Range. Gordon and Keller (1993) interpreted these relations to indicate that faulting along the west side of the Piñon Range occurred during sedimentation, although the rate of sedimentation generally kept pace with that of subsidence. The Miocene strata dip ~10° more steeply than the overlying Hay Ranch Formation, which indicates that faulting occurred or began before Pliocene sedimentation.

Seismic data show a west-dipping, range-bounding fault that controls the east side of the valley (Gordon and Keller, 1993), and post-Hay Ranch offset along northwest-dipping normal faults created several intra-basin half grabens. However, the Hay Ranch in places is nearly horizontal, so that the faulting has only minimally tilted the basin sediments since the early Pleistocene. These northwest-dipping faults may be more gravitational than tectonic (Muntean et al., 2001). As such, the Piñon Range appears to have attained much of its altitude before the middle Miocene than later. This supports Smith and Ketner’s (1978) conclusion that the range was present before Miocene sedimentation, as well as supergene alunite dates at the Rain mine that indicate late Oligocene to early Miocene exposure and weathering (Table 2).

The northern half of Independence Valley lies between the Independence Range and Double Mountain, and the southern half is between the Swales–Lone Mountain highland (the southern Independence Mountains) and the Adobe Range (Fig. 10). The North Fork Humboldt River begins at the north end of the valley and drains eastward through a broad gap between the Adobe Range and Double Mountain. Depth to basement typically is less than 1 km, with some small areas deeper than 3 km (Ponce, 2004); how much of this basin fill is composed of Eocene volcanic and sedimentary units is unknown.

Miocene sediments are poorly exposed throughout much of the valley. Highway roadcuts in the middle of the basin southeast of the Independence Mountains expose epiclastic to ash-rich, fine-grained sandstone and siltstone with no coarser material; these sediments were deposited in low-energy, fluvial to lacustrine environments. Dips in this area are nearly horizontal. On the western side of the valley, the fine-grained sediments were deposited unconformably on more steeply dipping, Eocene volcanic rocks. Tephra correlations in this area indicate ages of 14.2–14.8 Ma (Fig. 10; Table 1). Closer to the center of the basin, a small, Paleozioc- and Eocene-cored, intra-basin fault block dips moderately to the east, and one tephra near the
base of the Miocene section correlated with a ca. 15.9 Ma tephra.

The sedimentary units exposed in the middle of the basin are continuous southward to similar 15.8 to 15.3 Ma units in the Camp Creek and Blue Basin areas east of Swales and Lone Mountains (Figs. 7 and 10), which, in turn, extend south into the Carlin basin (see the Carlin basin descriptions). Strata along the eastern flank of the Swales–Lone Mountain highland dip modestly to the east and are largely depositional on Eocene and older rocks that comprise the highland (Lovejoy, 1959; Evans and Ketner, 1971; Ketner, 1998). Strata along the western side of the Adobe Range dip to the west, and sediment-basement contacts are entirely depositional (Ketner and Ross, 1990). The dip reversal takes place several kilometers outbound of the Swales–Lone Mountain range front. Juxtaposition of stratigraphically lower strata on the east side and higher strata on the west side indicates that a steeply west-dipping, intrabasin, normal fault created the dip reversal. This fault may connect with a similar fault in the Carlin basin; its projection to the north is uncertain.

The northern part of the basin, east of the Independence Mountains, has limited exposures. In the center of the basin, the units include nearly horizontal, epiclastic sandstone and siltstone beds, and beds in the northwestern part of the valley, adjacent to the Independence Mountains, dip modestly to the east (Henry, 2008). The Miocene sediments were deposited on and around Jarbridge Rhylolite flow units along the western base of Double Mountain. East of the valley, Miocene sediments are nearly continuously exposed along the broad North Fork Humboldt River valley between Independence Valley and the Devils Gate area in the Elko basin (Fig. 10). As described earlier, those strata overlie tilted, late Eocene volcanic rocks.

Overall, Miocene sediments in the Independence Valley were deposited in low-energy fluvial and epiclastic-rich lacustrine environments. As with other basins, the formation of late Cenozoic strath terraces removed some Miocene sediments from the valley. The limited amounts of coarse fluvial sediments, even close to the source areas, suggests only low relief at the time of sedimentation. Swales Mountain did shed coarse material into the Carlin basin; therefore, it was a highland at the time of sedimentation. Some post-sedimentation faulting occurred within the basin, but not enough to create major offsets or steep dips, and the sediment-basement contacts along the basin margins remain depositional and not fault controlled. As with the Marys Mountain area on the west side of the Carlin basin, the adjacent, high-relief Independence, Lone, and Swales Mountains areas may have been relatively lower relief highlands during sedimentation that experienced intra-highland faulting after sedimentation, followed by preferential late Miocene and younger erosion of the Miocene strata.

**REGIONAL PALEOGEOGRAPHIC EVOLUTION**

The features and histories of the Chimney, Ivanhoe, Carlin, and Elko sedimentary basins provide evidence of the settings and events before, during, and after middle Miocene basin formation. Although each basin evolved in response to events in and around that particular basin, the combined histories create a picture of geologic and landscape evolution that has broader geologic and mineral-deposit implications for this and adjacent areas of the Great Basin. Figure 12 diagrammatically shows four stages of the geologic and paleogeographic evolution of the study area, starting with the setting just prior to middle Miocene sedimentation, and followed by three stages from the middle Miocene to the present.

In summary, the area east of the Tuscaraora Mountains was a broad upland prior to the middle Miocene that experienced modest middle Tertiary faulting but generally was exposed and weathering during that time. Streams drained out of the area. At the same time, the broad area west of the Tuscaraora Mountains was a somewhat lower area with streams that carried sediments westward and out of the area. In the middle Miocene, high-angle faulting segmented the broad upland east of the Tuscaraora Mountains, created the early Carlin and Elko basins, and trapped sediments in those areas. Volcanic activity also dammed some outflowing streams to accentuate sediment retention. Extensive volcanic activity related to the northern Nevada rift blocked westward streamflow out of the Ivanhoe and Chimney, creating those sedimentary basins. Except for the Elko basin, dams failed within a couple of million years and streams resumed their external flow, initiating a proto-Humboldt River drainage system. The Elko basin continued to retain sediments until ca. 9.8 Ma, when it began to drain to the west and integrate with the other streams. The integrated stream system has progressively removed large volumes of Miocene sediments from these basins and carried them to downstream reaches of the Humboldt River in northwestern Nevada. This erosion has exhumed pre-middle Miocene basement that was buried during sedimentation. High-angle faulting affected all areas to some degree after sedimentation, with early minor faulting and more significant offsets later in the Miocene. Overall Neogene crustal extension along the transect was minor except for the somewhat more extended region west of the northern Nevada rift and south of Chimney basin.

**Pre-Sedimentation History**

One striking aspect of the entire 200-km-long transect is the general absence of middle Tertiary sedimentary deposits and only rare volcanic units. Small areas of Oligocene tuffs are exposed in the western Carlin basin (Henry and Faulds, 1999), and sedimentary units that may be as young as Oligocene are preserved just south of Wells beneath middle Miocene volcanic rocks (Snoke et al., 1997). Middle Tertiary units are similarly absent north of the transect in northernmost Nevada (Cosats, 1987; Rahl et al., 2002; Brueseke and Hart, 2007) and southwestern Idaho (Ekren et al., 1981). In contrast, widespread middle Tertiary volcanic rocks and, locally, sedimentary units are present south of the transect, as well as to the west in the Pine Forest Range and adjacent areas (Stewart and Carlson, 1978; Colgan et al., 2006, 2008; Fig. 1). The middle Tertiary climate was generally temperate and moist (Axelrod, 1956; Zachos et al., 2001), and erosion and runoff undoubtedly took place along the transect. Although sediments produced during this period could have been deposited and then entirely removed by erosion just prior to 16 Ma, the more likely scenario is that the area was an upland and streams drained out of the area, carrying the sediments with them.

Middle Tertiary extension tilted all pre-middle Miocene units along the transect. However, with the exception of steeply dipping 33.5 Ma tuffs at Mule Canyon (Fig. 1, John and Wrucke, 2003), the difference between Eocene and Miocene stratal dips generally is 20° or less, indicating relatively small amounts of extension. In most places, the age of tilting is not known. Some deformation took place between ca. 36 and 31 Ma on the east side of the Piñon Range (Palmer et al., 1991), and pre–16.5 Ma tilting took place both before and after the eruption of 25 Ma tuffs at Marys Mountain (Henry and Faulds, 1999), similar to the range of supergene aluminate dates along the Carlin trend that may reflect uplift events (see below). South of the transect, middle Tertiary extension varied from minimal to locally 50% (Muntean et al., 2001; Colgan et al., 2008). Middle Tertiary weathering and erosion affected much of the study area. Pre-middle Miocene regoliths formed in some areas, and materials derived from the regoliths were shed into Miocene basins during early sedimentation. Erosion exposed late Eocene plutons in the Tuscaraora Mountains, eastern Ivanhoe district, and
Figure 12. Diagrammatic geologic and paleogeography history of northeastern Nevada, including uplands (medium brown), areas of basin sedimentation (blue), Miocene volcanic fields (“v” pattern), major normal faults (bar and ball), and stream flow directions (blue arrows); lighter brown areas are uplands that may or may not have been present or were very subdued. Shown also are late Eocene gold trends (gold dotted lines) and middle Miocene epithermal deposits (red dots) from Figure 1. The area shown is the same as in Figure 1, where other geographic locations are presented; county boundaries (dashed lines) can be used as references. Time intervals include: (A) pre-middle Miocene, prior to the inception of middle Miocene basin sedimentation; (B) ca. 16 Ma, when the four basins described in this paper (Chimney (ChB), Ivanhoe (IB), Carlin (CB), and Elko (EB) basins and Independence Valley (IV)) began to form and fill with sediments; (C) ca. 14–9 Ma, the period when the western basins drained externally to form the early Humboldt River system, with continued sedimentation in the Elko basin (EB); and (D) ca. 9 Ma to the present, by which time the Elko basin had integrated into the Humboldt River system, with temporary Pliocene sedimentation in the Elko basin and late Pliocene and Pleistocene sedimentation in Pine Valley (PV) and downstream parts of the Humboldt River (latter two from Reheis, 1999a). The pre-middle Miocene uplands are simplified from Haynes (2003) and include only those related to this study. Uplands had subdued topography, on the basis of only minor amounts of coarse fluvial material except near fault-controlled range fronts. Note the contraction in the exposed extents of some uplands after partial burial by middle Miocene sediments (B to C) and their somewhat increased extent during late Cenozoic erosion of Miocene strata (C to D).
southwestern Adobe Range, and middle Miocene sediments were deposited on and contain clasts derived from those exposed plutons.

The Carlin trend gold deposits formed at least 800 m beneath a paleosurface in the late Eocene (Haynes, 2003; Hickey et al., 2003). Supergene alunite dates of 30–18.4 Ma from some of the Carlin trend gold deposits (Table 2; Fig. 13) record progressive erosion and weathering of that surface; they also may reflect the age(s) of regolith formation. Some of the gold deposits were exposed by the middle Miocene, contributed sediments to the Carlin and Ivanhoe basins, and were in part covered by Miocene sediments. A single, supergene alunite date of 23 Ma from the Preble gold deposit along the southern part of the Getchell trend (Table 2; Fig. 13) reflects weathering in that area, but the paleogeographic relation of that site to the Chimney basin to the north is unknown.

Despite the erosion and weathering, late Eocene volcanic rocks were locally exposed in some of the highlands and all of the Miocene basins. Although these units were not uniformly deposited across the region (Ressel and Henry, 2006; Henry, 2008), their presence demonstrates that some of the late Eocene paleosurface was not significantly eroded in the middle Tertiary. At Willow Creek Reservoir in the Ivanhoe basin, both Eocene and middle Miocene units were deposited in a west-draining paleovalleys, despite an intervening period of tilting (Wallace, 2003a), and the tilting event may have only minimally affected the landscape.

In the Chimney and Ivanhoe basins, the pre-sedimentation topography was subdued and drained generally to the west. These two, lacustrine-dominated basins connected across the northern Nevada rift in the Snowstorm Mountains, which, shortly before sedimentation, was the site of mafic volcanism and concurrent, fault-related subsidence. The lack of sedimentary thickening in the rift zone and the ability of the lacustrine environment to extend across that area shortly after volcanism ceased demonstrate that the eruption rate was similar to the subsidence rate, leaving a fairly level landscape just prior to sedimentation. This contrasts with a proposal that middle Miocene thermal bulging created a kilometer-high upland in the area of the rift (Pierce et al., 2002).

To the north, the absence of northern Nevada-derived zircons in middle Miocene basins along the Oregon-Idaho border led Beranek et al. (2006) to propose a drainage divide, and, thus, a paleo-topographic high, somewhere near the Oregon-Idaho-Nevada border (Fig. 1). Paleotopography around the Santa Rosa–Calico volcanic field, northwest of the Chimney basin, apparently was more pronounced, and uplands and valleys significantly controlled the distribution of volcanic flow units (Brueseke and Hart, 2007).

All of these relations indicate that much of the transect area was a moderately eroding, nonaggradational, topographic upland in the middle Tertiary (Fig. 12A). This feature may have originated in the late Eocene, when a broad paleo-high extended roughly south from the Tuscarora area toward Marys Mountain (Figs. 1 and 9), with streams draining to the east and west (Haynes, 2003; Henry, 2008). As described earlier, the absence of sediments in the hanging-wall region of the Ruby Mountains–East Humboldt Range detachment zone may indicate either an externally draining hanging-wall basin or no basin at all. There, most of the surface uplift may have been related to high-angle faulting that took place during sedimentation

![Figure 13. Supergene alunite dates and geomorphic events in northeastern Nevada. The vertical blue bars indicate the sedimentary lifespans of the basins, simplified from Figure 2. Basin abbreviations: CaB—Carlin basin; ChB—Chimney basin; EB—Elko basin; IB—Ivanhoe basin; IV—Independence Valley. Supergene alunite dates for the Carlin and Getchell gold trends are shown as squares (open, Rain deposit in the northern Piñon Range; solid, all other deposits); data are provided in Table 2. The major geomorphic periods and processes are shown along the right margin.](image-url)
between 14 and 15 Ma (Howard, 2003; Colgan and Metcalf, 2006). Thus, despite the evidence of middle Tertiary faulting along the transect, that faulting apparently did not have a significant paleotopographic effect, such as the formation of uplifts or basins.

**Formation of the Basins**

All of the basins began to retain sediments at nearly the same time: Chimney basin, 16.3–16.1 Ma; Ivanhoe basin, slightly before 16.1 Ma; Carlin basin, ca. 16.5 Ma; Independence Valley, shortly before 15.9 Ma; and Elko basin, slightly before 15.4 Ma (Table 1; Figs. 2 and 12). However, the origins of the basins differ. The floors of the Chimney and Ivanhoe basins drained to the west just prior to sedimentation, and volcanic eruptions in the downstream parts of the basins dammed the streams to create the lakes (Fig. 12B). Ongoing eruptions along the margins of both basins continued to confine inflowing streams and caused the lakes to grow and expand across the landscape. Synsedimentary faulting was relatively minor during the short lifespans of these two basins. However, continued faulting along the northern Nevada rift in the Snowstorm Mountains produced offsets of several hundred meters and, along with coeval rhyolitic volcanism, ended sedimentation in that area. Both basins began to drain externally when the volcanic dams were breached.

In contrast, high-angle faulting along the west side of the Ruby Mountains and East Humboldt Range controlled the formation of the middle Miocene Elko basin (Fig. 12B). The basin evolved as an east-dipping half graben, with the sedimentation rate slightly outpacing the subsidence rate. As tectonism diminished, sediments continued to be shed from the horsts into the eastern part of the basin, and streams cut through the western margin of the basin to connect with those in the Carlin basin. Both faulting and volcanic dams produced the Carlin basin and perhaps part of Pine Valley (Fig. 12B). Early faulting lowered the core of the existing upland to form the south-trending basin. This induced early fluvial and debris-flow sedimentation in the western and northern parts of the basin, and additional faulting in the east-central part of the basin directed streamflow to the south toward Pine Valley. Thus, although the Tuscaraora Mountains were a divide between the Carlin and Ivanhoe basins, faulting was restricted to the east side of the highland, creating very different early sedimentary processes on either side. Eruption of the Palisade Canyon rhyolite further blocked southerly flow out of the basin and created a deepening lake that gradually expanded to the north and connected to adjacent basins. Failure of the dam caused the lake to drain and fluvial sedimentation to blanket most of the basin floor, although a shallow lacustrine environment persisted along the western side of the basin until shortly after 14.6 Ma.

The origin and history of the Independence Valley in part mimics that of the Carlin basin, and the basin eventually filled and connected with both the Carlin and Elko basins through low gaps in the surrounding highlands. However, the lack of coarse episcopal materials and apparent depositional contacts with bedrock along the basin margins suggest a subdued surrounding topography at the time of sedimentation. The Independence Mountains on the west side of the basin now rise more than a kilometer above the valley floor, but there is no sedimentological evidence for such a large upland in the middle Miocene.

**Post-Sedimentation History**

Both faulting and erosion played variably significant post-sedimentation roles in the evolution of the transect. Sedimentation ended in the Chimney, Ivanhoe, and Carlin basins and Independence Valley between ca. 15 Ma and 14.6 Ma, and the youngest sediments in the Elko basin were deposited shortly after 9.9 Ma (Table 1; Fig. 2). In all four basins, almost all of the youngest sediments were deposited in fluvial or mixed fluvial and lacustrine environments (Fig. 2). Highlands partially to completely surrounded each basin; the climate, although gradually drying, remained somewhat temperate and moist; and the precipitation caused erosion and runoff. Therefore, the absence of younger sediments indicates that the streams were able to flow out of the basins, carrying the newly formed sediments with them and likely initiating erosion of older sediments.

**Faulting**

The amount and style of post-sedimentation faulting varied along the transect, and the age in most places is unknown. Along much of the transect, high-angle normal faults cut the highlands and basins equally, and the faults form two general populations—an earlier, north- to north-northwest–striking set and a younger, east-northeast–striking set—although small faults of all orientations are present along the transect. Overall, the early faults are much more common east of the Tuscaraora Mountains than to the west. Conversely, the younger faults are almost exclusively present west of the Tuscaraora Mountains (Figs. 12C and 12D).

West of the Tuscaraora Mountains, the early, north- to north-northwest–striking faults in the Chimney and Ivanhoe basins have less than 100 m of displacement. The low stratal tilts (often nearly horizontal) and minor fault displacements demonstrate relatively little extension in this region. Muntean et al. (2001) estimated that the total amount of Miocene and younger extension in north-central Nevada did not exceed 10%, which is consistent with these field observations.

East of the Tuscaraora Mountains, faults are abundant in the Carlin basin and adjacent highlands, and some have 1 km or more of displacement. However, most of the Miocene strata dip less than 20°, indicating relatively little extension. Similarly, despite post-middle Miocene and also Pliocene faulting along the east side of Pine Valley, all Neogene strata in that basin dip gently or are nearly horizontal. Just to the east, the Elko basin is in the immediate hanging wall of the Ruby Mountains–East Humboldt Range horst, and some faults with a few hundred meters of offset cut basin sediments along the southeastern Adobe Range and the Huntington Creek area. Overall, the dips of exposed Miocene strata are gentle, in places primary. Therefore, the area east of the Tuscaraora Mountains is much more faulted than areas to the west, but the amount of extension was not significant.

The younger faulting event appears to be related to greater amounts of extension west of the northern Nevada rift area and much less to the east. Extension west of the rift created tilted horsts with broad, intervening valleys that formed sometime after ca. 10 Ma (Zoback and Thompson, 1978; Wallace, 1991; Colgan et al., 2004, 2006). Along the transect, these include the Osgood Mountains and the Hot Springs Range and intervening Eden and Kelly Creek Valleys on the southwestern margin of the Chimney basin (Stewart, 1998). These structural features do not extend north into the main part of the basin, indicating a change from significant to minimal extension northward across that area. No transform or accommodation zone is apparent between the more faulted domain to the southwest and the nearly intact domain just to the northeast.

As described above, faulting east of the northern Nevada rift was locally abundant but did not produce much extension. A transition zone between the more extended region to the west and the minimally extended area to the east occurs in the area of the rift (Wallace, 1991). This zone contains a series of east-northeast–striking, oblique-slip faults, such as those along the Midas trough and the north and south sides of Boulder Valley, all of which affected the Ivanhoe basin and cut the earlier north- to northwest-striking faults (Fig. 12D). These faults have up to 1 km of normal displacement and as much as several kilometers of left-lateral movement,
reflecting northwest-directed extension (Zoback and Thompson, 1978; Wallace, 1991). They terminate abruptly to the west at the margin of the more extended terrain described above, and the boundary likely is structurally controlled by west-dipping, normal to oblique-slip faults along the west side of the Sheep Creek Range and part of the Snowstorm Mountains (cf. Wallace, 1993).

The east-northeast–striking faults diminish in magnitude to the east, splay into numerous smaller offset faults, and die out before reaching the Tuscarora Mountains (Peters, 2003; Wallace, 2003a, 2003b). Given the northwest-oriented internal extension within the transition zone, late Cenozoic, north-northeast–striking, strike-slip faults documented along the west side of the Tuscarora Mountains may record the differential offset between the transition zone and the less extended terrain to the east (Wallace, 1991; Leonardson and Rahn, 1996). All of these faults may have begun to form at ca. 10 Ma (Zoback and Thompson, 1978), similar to the uplift age of the Santa Rosa Range to the west (Fig. 1; Colgan et al., 2004, 2006). Movement continued into the Pliocene and Quaternary. Along the Midaus trough and in Paradise Valley, faulting modestly tilted 4 to 5 Ma mafic flow units, and Quaternary fault scarps are present along the Midaus trough (Wallace, 1993) and the eastern end of the Boulder Valley (Wesnousky et al., 2005).

Except for the region west of the northern Nevada rift, extension in the transect area, both during and after sedimentation, was not significant. This supports Muntean et al.’s (2001) conclusion that Miocene and younger extension was ~10%. Dips on Miocene sedimentary units typically are 20° or less, and the faults likely were more vertical than listric. The area west of the rift is more extended, although by how much is unknown. This area is due north of areas south of Battle Mountain that were extended by more than 100% after the middle Miocene (Colgan et al., 2008). As such, and as pointed out by those authors, the rift zone (be it the pre-Miocene crustal structure or the deep mafic dike complex) appears to have exerted a significant control on the amount and location of extension in north-central Nevada.

Erosion

Late Cenozoic erosion affected much of the transect, and it was most pronounced in the areas of the Elko and Carlin basins. Downstepping, erosional strath terraces are abundant in those basins and, to a lesser degree, in the Ivanhoe and Chimney basins. In the Carlin and Elko basins, the progressive formation of as many as a dozen strath terraces paired across both major and minor drainages has lowered the Humboldt River drainage system by as much as 400 m (Wallace, 2005). Terraces are not present along the Humboldt River downstream of about Beowawe (Fig. 1). Erosion has preferentially removed the poorly consolidated Miocene and Pliocene sedimentary units relative to the harder bedrock units. In doing so, it has focused the more modern drainage systems (and erosion) into the areas that originally were the sites of middle Miocene drainages and sedimentation.

In some places, such as along Susie Creek in the Carlin basin, the distribution of modern and Miocene stream deposits shows that the modern stream follows the same route as the middle Miocene stream.

Field relations and limited age control indicate that erosion began after the early faulting that tilted the Miocene sedimentary rocks but prior to the middle Pliocene. Rock Creek, which flows west and then cuts south through the western Ivanhoe basin (Figs. 5 and 12), is an antecedent stream that incised canyons during the formation of the Midaus trough and Boulder Valley fault escarpments and is thus older than the younger faulting event. Similarly, 12.6 to 6.7 Ma supergene alunite dates from the Carlin trend (Table 2; Fig. 13) indicate that supergene processes had begun by the middle to late Miocene. In the Elko basin, as much as 150 m of downcutting took place before the middle Pliocene. At that time, the bedrock sill at Carlin Canyon (Fig. 7) slowed incision of the Humboldt River, and lacustrine and fluvial sediments were deposited on top of incised topography in upstream areas (Figs. 11F and 12D). Strath terrace formation continued when the Pliocene lake drained at ca. 2 Ma, dissecting both the Pliocene and Miocene strata. Along the western base of the East Humboldt Range and Ruby Mountains, Quaternary faulting has offset some of the older strath terraces (Sharp, 1939; Wesnousky and Willoughby, 2003).

The strath terraces in the Elko and Carlin basins have a regionally consistent pattern centered ultimately along the Humboldt River, and they thus formed in response to a drop in the base level of the river downstream of Beowawe (Fig. 1; Wallace, 2005). Because the Chimney and Ivanhoe basins also drain into the Humboldt River, the far-field, base-level drop likely produced the erosion and strath terraces in those areas as well. Therefore, material eroded from the upstream parts of the Humboldt River drainage system has been transported to and deposited in intermontane valleys farther downstream (Wallace, 2005). As a result, Quaternary sediments, which are ubiquitous in most intermontane basins in the region (Stewart and Carlson, 1978), are largely absent in the upper half of the Humboldt River drainage system, where Miocene sedimentary units are exposed over broad areas.

As described earlier, Miocene sediments nearly buried the Adobe Range and the Peko and Elko Hills in the Elko basin. As the erosion progressively removed the Miocene and Pliocene sediments, this previously buried basement was gradually exposed (Fig. 11B). Similar reexposure of basement rocks is evident in all of the other basins as well, such as in the western and southwestern parts of the Carlin basin. This combination of the original middle Miocene paleogeography, relatively minor, post-sedimentation faulting, and significant, later preferential erosion of the Neogene sediments has produced what appear to be structural horsts and grabens but which instead are slightly modified but largely intact middle Miocene or older ranges and basins. The major exception is the Independence Range, west of Independence Valley, which appears to have formed after middle Miocene sedimentation.

Drainage Evolution and Paleoelevations

Previous studies have provided data regarding Cenozoic paleoelevations in the Great Basin, including parts of this study area (Sonder and Jones, 1999; Wolfe et al., 1997; Horton et al., 2004). Wolfe et al. (1997) suggested that central Nevada was 1–1.5 km higher in the middle Miocene than it is present day, and Pierce et al. (2002) proposed a regional, kilometer-high bulge related to the Yellowstone mantle plume. While nothing in the present study can quantitatively determine paleoelevations, water flows downhill, and the evolution of the drainage systems provides evidence of what areas were relatively higher or lower and when.

In the middle Tertiary, the study area was topographically higher than surrounding areas. Based on evidence presented in the basin descriptions, streams in the west flowed west; streams in the central area flowed to the west, south, and east; and streams in the area of the Elko basin likely flowed to the south and, possibly, east and north (Figs. 12A and 14). As such, the greater Carlin-Tuscarora area was a paleotopographic upland, and other areas were relatively lower. In addition, a drainage divide was present along the Nevada-Idaho border (Beranek et al., 2006). These patterns generally continued into the middle Miocene (Figs. 12B and 14), although the structural development of the Carlin basin and Pine Valley induced southerly streamflow in those areas. Drainage in the three western basins ultimately merged and produced a westerly flow system. The Elko basin began to drain to the west after ca. 9 Ma and merged with these drainage systems (Fig. 12D).

The strath terrace patterns indicate that all streams in the region had integrated into the west-flowing Humboldt River system by the
Late Miocene. Pine Valley, which drained to the south in the middle Miocene and drained internally in the Pliocene and Pleistocene, eventually was captured by the Humboldt River during downcutting and began to drain to the north (Fig. 14; Gordon and Heller, 1993; Reheis, 1999a, 1999b; Wallace, 2005). In the Pliocene, the Reese River began to flow north to the Humboldt River from south of Austin, traversing what was a series of 16–12 Ma lacustrine and fluvial sedimentary basins (Fig. 14; Colgan et al., 2008). Similarly, Pliocene and Pleistocene lakes and streams in the region between central Nevada and the Humboldt River became integrated into an overall drainage system that flowed north to the river and then to the west (Fig. 13; Reheis et al., 2002). All of these north-draining streams indicate that central Nevada, at the latitude of Austin and Eureka (Fig. 14), was higher than the Humboldt River by at least the Pliocene. Fault-controlled grabens in the lower reaches of the Humboldt River, roughly downstream from Beowawe (Fig. 1), received the sediments eroded from upstream areas.

The Yellowstone mantle plume migrated east-northeastward from the area of the McDermitt eruptive center west of the Oregon-Idaho border at ca. 16 Ma to the Twin Falls eruptive center north of the Utah-Nevada border by ca. 10.5 Ma (Fig. 14; Perkins and Nash, 2002; Pierce et al., 2002). This migration path was due north of the four basins described in this paper, and it progressed eastward at the same time that sedimentation was taking place in the basins. As evidenced by the bimodal volcanism along the northern Nevada rift, middle Miocene plume-related crustal heating extended south into central Nevada. Pierce et al. (2002) proposed a broad thermal bulge related to the plume head, with a rain shadow on its east (leeward) side and an elevational decline in its wake (Fig. 14). Within the present study area, the entire western half of the transect has drained to the west since at least the middle Miocene (Fig. 14), and the Ivanhoe and Chimney basins connected across the northern Nevada rift, which was the proposed early axis of the bulge (Pierce et al., 2002). These relations do not support an early thermal bulge in that area, and the widespread streams and lakes in the basins to the east similarly do not support a significant rain shadow (Pierce et al., 2002). However, even without a bulge, the progressive change to an entirely westerly flow direction for all four of the basins by the time the plume was east of the study area (Fig. 14) may in part reflect a progressive decrease in elevation as the plume-related halo of crustal heating migrated to the east. At the same time, normal faulting and horst-basin development in northwestern Nevada after ca. 10 Ma (Colgan et al., 2004, 2006) also contributed to the observed decrease in the base level of the Humboldt River drainage system.

Collapse of central Nevada after the middle Miocene is not supported by: (1) the progression from middle Miocene, southward-and-westward flow to late Miocene and younger, westward flow along the entire Humboldt River; (2) the reversal of flow in Pine Valley from south to north; and (3) the integration of middle Miocene basins into the north-flowing Reese River in the Pliocene. Instead, and perhaps overly simplistically, the pattern suggests that northern Nevada has decreased in elevation relative to central Nevada, and that the westward flow of the Humboldt River between Wells and Winnemucca since the late Miocene reflects both that relative topographic change and a base-level decrease in...
northwestern Nevada (Wallace, 2005). If central Nevada did decrease in elevation, and if the downstream tectonic activity and possible mantle plume wake did produce the drainage-pattern changes in northern Nevada, then the effects of the latter two exceeded the rate of decline of central Nevada.

Horton et al. (2004) used isotopic data from Miocene sediments in northern Pine Valley to suggest an elevation decrease of 1–1.5 km to the west or southwest of the area, and thus a reduced orographic effect on rainfall in the Pine Valley area, between ca. 15 and 1 Ma. If central Nevada remained high during this period, then the elevation decrease must have been to the west. Indeed, the progressive decrease in the downstream base level of the Humboldt River, as shown by the strath terraces and downstream basin filling, indicates that the proposed elevation decrease may have been in northwestern Nevada.

**IMPLICATIONS FOR MINERAL DEPOSITS**

North-central Nevada contains numerous large, late Eocene, Carlin-type and middle Miocene, epithermal mineral deposits (John, 2001; Cline et al., 2005), and the geologic history of the area described in this paper has had a major impact on the formation, modification, and concealment of those deposits (Cline et al., 2005; Wallace, 2005). The Eocene deposits were faulted, exposed, and weathered in the middle Tertiary. Some of the deposits were buried beneath sediments in the middle Miocene, and many of the deposits or mineral districts were faulted, partially exhumed, and weathered for a second time after the middle Miocene.

The weathering, erosion, and related supergene processes had an important influence on the degree of oxidation of sulfides and carbon (thus influencing the mineral processing methods needed) and the amount of the original deposit that was preserved. The Gold Quarry deposit was subjected to all three stages of the landscape evolution described above, leading to a complex geologic setting and degree of oxidation. Some deposits, such as the combined Paleozoic, Mesozoic, and Eocene gold- and base metal-rich deposits at the Mike property west of Gold Quarry, were subjected to the middle Tertiary and middle Miocene processes but less so the later events, leaving a partially modified but entirely concealed orebody (Norby and Orobona, 2002). There, pre-middle Miocene supergene weathering redistributed copper and zinc, and the zinc was concentrated at the boundary between oxidized and primary reduced ores; the Miocene cover preserved that interface (Bawden et al., 2003). Miocene sediments likely did not cover the Rain deposit in the northern Piñon Range, but various stages of faulting, weathering, and erosion affected the deposit as it sat in a topographically high position for several tens of millions of years.

The middle Miocene epithermal deposits formed at or near the paleosurface during the formation of the sedimentary basins. The resulting deposits, such as those in the Ivanhoe district, included sinter deposits at the paleosurface, stratiform replacement bodies in underlying sedimentary units, and veins along faults in deeper volcanic and pre-Tertiary units (Bartlett et al., 1991; Wallace, 2003c). The epithermal deposits owe their origins to a combination of volcanism, which provided the heat; faulting, which created deep conduits for fluid circulation and served as important ore hosts; and the lakes, which contributed abundant water and also produced sedimentary aquifers that, in places, were silicified and mineralized (John et al., 2003; Wallace, 2003c; Vikre, 2007). As shown on Figure 1, many epithermal systems formed where these three components overlapped. Depending on the combination of events at any place and time, the deposits formed both within and along the margins of the presently exposed basins. Thus, exploration for epithermal deposits in this region requires a knowledge of the middle Miocene paleogeography, faulting, volcanism, and hydrothermal processes to target areas where optimal conditions existed at that time. The absence of even one component could have reduced or eliminated the chance of forming an epithermal mineral deposit.

Because the epithermal deposits formed near the paleosurface, destruction of that surface could have eliminated parts or all of a mineral deposit (Wallace et al., 2004b). As shown in this study, post-middle Miocene erosion and faulting along the eastern half of the transect has removed or modified many parts of the middle Miocene paleosurface. In contrast, much of the paleosurface and many of the related epithermal systems have been preserved in the western half of the transect, such as in the Ivanhoe district and the Chimney basin. At the Hollister gold deposit in the Ivanhoe district, the water table has remained high and supergene oxidation extends only 30–150 m beneath the present surface (Bartlett et al., 1991). The tops of some epithermal systems, such as at Midas, were domed and eroded after mineralization, leaving only the underlying fault-controlled veins (although it might be argued that those veins are extremely high grade and the loss of the top of the system may not have been economically significant).

Therefore, exploration for epithermal deposits along the transect, as well as in other areas in the region that may contain epithermal deposits, must consider both the environment during mineralization and the events that followed mineralization.

The late Miocene and younger transport of sediments eroded from northeastern Nevada to downstream basins in northwestern Nevada has covered areas that previously were exposed, including old pediments along range fronts and any Miocene sedimentary units. Some of these buried bedrock areas contain economic mineral deposits, including the Twin Creeks gold deposit and drilled resources northwest of the Lone Tree mine west of Battle Mountain (Fig. 1). Similar to the mineral deposits much farther upstream, the geologic history at any one location affected the amount of pre-sedimentation erosion, weathering, enrichment, and faulting, not to mention the paleotopography and thus the amount of later burial in the downstream areas.

Exploration for mineral deposits of all types in the region, as well as the use and economics of various mineral processing methods needed to extract the metals, must consider the post-mineralization events that affected the deposits. This pertains to the Eocene Carlin-type and Miocene epithermal deposits described in this paper, as well as the many other types of mineral deposits of various ages that formed in the region (Wallace et al., 2004b). As shown in this paper, no one landscape evolution model applies to the region as a whole, and regional subdomains that combine time, faulting, erosion, and sedimentation must be developed to evaluate specific areas. For example, the major porphyry copper deposits at Yerington in western Nevada, Ely in eastern Nevada, and Copper Canyon near Battle Mountain formed in the Jurassic, Cretaceous, and Eocene, respectively. Each has been subjected to different post-mineralization histories that have affected their present morphologies and degrees of oxidation and preservation (Proffett, 1977; Dilles and Gans, 1995; Seedorff et al., 1996; Theodore, 2000). The post-mineralization histories of even small areas, such as the one that includes the Gold Quarry and Mike gold deposits described above, can be sufficiently varied to produce different degrees of weathering, oxidation, and amount of preservation.

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Chris Henry (Nevada Bureau of Mines and Geology) provided a 40Ar/39Ar date on a Chimney basin rhyolite (sample ChB-5) and, along with Joe Colgan, Keith Howard, and David John, stimulating discussions on regional geochronology and paleogeography. Steve Williams obtained tephra correlation ages from three samples in the Carlin basin and southern Independence Valley, as cited in Table 1. Unpublished supergene alunite dates in Table 2 from the Gold Quarry and Lone Tree mine areas came courtesy of Peter
APPENDIX 1. SAMPLE DESCRIPTIONS

Table 1 includes numerous samples collected and dated for the present study. These are listed as “This study” in the Reference column of that table, and the sample number in parentheses refers to brief location and sample descriptions of those samples in this Appendix. Not included in this Appendix are descriptions of dated samples that have been published elsewhere or were collected and dated by other workers and used by permission. The latitude and longitude for each sample are in Table 1. All samples are of fresh rocks that are free of diagenetic or hydrothermal alteration minerals.

Chimney Basin

CB-1: One-meter–thick, poorly consolidated, silvery-gray, fine-grained, unworked tephra bed (unit Tsl of Wallace, 1993). Varied reworked to unworked, leucocratic ash-rich lacustrine strata enclose the tephra bed. Located ~2 m below the contact between these lacustrine units and overlying fluvial units. Exposure is on the hillside on the north side of Twentynine Creek, Humboldt County. Tephra correlation age only.

CB-2: Flow unit of the Little Humboldt rhyolite (Wallace, 1993). Sample also listed in the Snowstorm Mountains section of Table 1. Flow-folded, devitrified, reddish-brown rhyolite with sanidine, plagioclase, and lesser quartz and iron oxide phenocrysts. Sample collected from outcrop southeast of the Little Humboldt Ranch (Fig. 3), Elko County. 4Ar/3Ar date on sanidine.

CB-3: Massive, nearly aphyric andesite flow unit enclosed in tuffaceous sedimentary units along Spring Creek north of the Little Humboldt River, Humboldt County. 4Ar/3Ar date on whole-rock date.

CB-4: Welded, phenocryst-rich, rhyolitic ash-flow tuff with abundant quartz, sanidine, and plagioclase phenocrysts, with lesser hornblende and biotite. Exposed along Spring Creek north of sample CB-3 and directly underlies that unit. 4Ar/3Ar date on sanidine.

CB-5: Flow-folded and -banded, moderately phenocrystic rhyolite flow unit exposed at Chimney Dam along the Little Humboldt River, Humboldt County. Underlies tuffaceous sedimentary units of Chimney basin; base not exposed but likely overlies 22 Ma andesite flow units to south. 4Ar/3Ar date on sanidine.

Carlin Basin

CB-1: Three-meter–thick, silver-gray, planar-bedded, fine-grained ash bed. Outcrop exposure a few hundred meters east of NV 278 between Cole and Ferndale Creeks, Eureka County. Stratigraphically above distal breccia related to the Palisade Canyon rhyolite (unit B-3, Section 3 of Regnier, 1960). Tephra correlation age only.

CB-2: Upper part of 4-m-thick, fine-grained, laminated, silver-gray, ash-rich bed in old road cut west of the Gold Quarry mine, Eureka County. Tephra correlation age only.

CB-3: Lower part of 4-m-thick, fine-grained, laminated, silver-gray, ash-rich bed in old road cut west of the Gold Quarry mine, Eureka County. Tephra correlation age only.

CB-4: Several-meter–thick, fine-grained, laminated, silver-gray, ash-rich bed in roadcut along eastbound I-80 east of Emigrant Pass, Eureka County. Tephra correlates to the Obliterator Tuff (see Perkins and Nash, 2002).

CB-5: Several-meter–thick, fine-grained, laminated, silver-gray, ash-rich bed in direct depositional contact with the Palisade Canyon rhyolite south of I-80 between Emigrant Pass and Carlin, Eureka County. 4Ar/3Ar date on sanidine; tephra is the same unit as CB-4 and correlates to the Obliterator Tuff (see Perkins and Nash, 2002).

CB-6: Silver-gray tephra bed, less than 1 m thick, within planar-bedded, epiclastic siltstone strata. Units are near the contact of the Humboldt Formation with underlying Paleozoic rocks. Excavated subcrop sample collected west-northwest of the Hunter exit along I-80 on the south side of the Adobe Range, Elko County. Tephra correlation age only.

CB-7: Silver-gray, structureless to weakly laminated tephra bed, 2 m thick, within off-white ash-rich lacustrine and mixed lacustrine and epipelic fluvial units. Units are near the contact of the Humboldt Formation with underlying Paleozoic rocks. Outcrop sample collected just north of the Hunter exit along I-80 on the south side of the Adobe Range, Elko County. Tephra correlation age only.

CB-8: Silver-gray, laminated, moderately well indurated, 20-cm–thick tephra bed within other ash-rich lacustrine strata. About 1 m above sample CB-9. Collected from outcrop in the uppermost Welches Canyon drainage basin, Eureka County. Tephra correlation age only.

CB-9: Silver-gray, laminated, moderately well indurated, 20-cm–thick tephra bed within other ash-rich lacustrine strata. About 1 m below sample CB-9. Collected from outcrop in the uppermost Welches Canyon drainage basin, Eureka County. Tephra correlation age only.

CB-10: Silver-gray, laminated 1-m–thick tephra bed within tan, fine-grained epiclastic siltstone beds. Bed is exposed in the overflow excavation at a reservoir in the upper end of the Gold Quarry mine, Elko County. Bed is ~1.5 m stratigraphically above sample CB-11. Tephra correlation age only.

CB-11: Silver-gray, laminated 1-m–thick tephra bed within tan, fine-grained epiclastic siltstone beds. Bed is exposed in the overflow excavation at a reservoir northeast of the Gold Quarry mine, Elko County. Tephra correlation age only.


CB-13: Palisade Canyon rhyolite. Massive flow unit directly above basal flow unit, which was deuterally altered and not suitable for dating. Rhyolite package overlies late Eocene Indian Wells Formation. Collected from base of cliff along railroad and Humboldt River north of Palisade. 4Ar/3Ar date on sanidine.

Elko Basin

EB-1: Andesite flow unit within epiclastic fluvial sediments of the Humboldt Formation, southwest flank of the Pekel Hills, Elko County. Flow contains 1-cm plagioclase phenocrysts and is very similar to undated flow units exposed in the Carlin basin and west of Twin Buttes between the North Fork Humboldt River and Marys River (Fig. 10). 4Ar/3Ar date on plagioclase.

EB-2: Silver-gray, laminated, fine-grained ash bed. Unit is the upper member of the Sedimentary and volcanic rocks of Threemile Spring unit of Thor- man et al. (2003), which is part of the Humboldt Formation (Sharp, 1939; Mueller and Smoke, 1993). Sample collected from cuts along railroad just west of Wells, Elko County. 4Ar/3Ar date on plagioclase.

Independence Valley

IV-1: 627-6 Silver-gray to off-white, fine-grained, poorly lithified ash bed, ~50–70 cm thick, enclosed within thick-to-thinly bedded, ash and epiclastic siltstone units. Excavated from roadcut along NV 226. Bedding nearly horizontal. About 3 m stratigraphically above sample IV-2. Tephra correlation age only.

IV-2: 627-5 Silver-gray, fine-grained, moderately lithified ash bed, ~1.5 m thick, enclosed within thick-to-thinly bedded, ash and epiclastic siltstone units. Bedding nearly horizontal. Collected from borrow pit just south of NV 226. Tephra correlation age only.

IV-3: 627-4 Thin (20 cm), silver-gray, fine-grained, poorly lithified ash bed enclosed within thick-to-thinly bedded, ash and epiclastic siltstone units. Bed dips gently to east. Excavated from roadcut along NV 226. Stratigraphically the lowest of IV-1, 2, and 3, and upsection from Eocene volcanic tuff units. Tephra correlation age only.

IV-4: 627-7 Coarse, silver-gray ash bed; thin, planar bedding. Enclosed within structureless, tan epiclastic to pumiceous sandstone beds, with burrows extending from overlying sandstone into ash bed, and from ash bed down into sandstone. Sample selected was evenly bedded and not near burrows. Collected from roadcut along NV 226 ~1.5 km west of intersection with NV 225. Ash bed is stratigraphically above Eocene volcanic bedrock, and Humboldt Formation dips more steeply here than in western part of basin. Tephra correlation age only.

IV-5: Unbedded to bedded, moderately consolidated, silver-gray ash bed, ~1.5 m thick, within ash-rich epiclastic siltstone and sandstone beds. Sample collected from outcrop on hillside south of Camp Creek, Elko County. Tephra correlation age only.

IV-6: Very thinly bedded and laminated, silver-gray ash bed, ~4 m thick, within epiclastic sandstone beds on hillside south of Camp Creek. Base of bed fills paleodepression or channel; upper part of bed partially incised prior to deposition of overlying sandstone bed. Some evidence of reworking and sorting in middle third of bed; sample collected from thinly laminated lower part of bed. Stratigraphically ~30 m above sample IV-5. Tephra correlation age only. These two tephra beds (IV-5, 6; 15.5–15.3 Ma) are in the same sequence from which Van Houton (1956) collected vertebrate fossil remains. Identification of these fossils led to an age assignment of late Miocene to early Pliocene; at the time, the Miocene-Pliocene boundary was 10 Ma. Thus, the vertebrate fossils are middle Miocene in age.


