RECONSTRUCTIONS OF CENOZOIC EXTENSIONAL FAULTING WITHIN THE SOUTHERN RUBY MOUNTAINS METAMORPHIC CORE COMPLEX AND ADJACENT AREAS, NORTHEASTERN NEVADA

by

James Russell Pape

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James Pege
(author's signature) 9-Nov-2010
(date)

APPROVAL BY RESEARCH COMMITTEE

As members of the Research Committee, we recommend that this thesis be accepted as fulfilling the research requirement for the degree of Master of Science.

Eric Seedorff
Major Advisor (type name) 25 Oct 2010
(signature) (date)

Mark Barton
(type name) 21-Oct-2010
(signature) (date)

Peter DeCelles
(type name) 9-Nov-2010
(signature) (date)
Abstract

This study presents five detailed cross-sectional restorations of Cenozoic upper-crustal extensional faulting in a ~150 km transect across northeastern Nevada in the vicinity of the Ruby Mountains-East Humboldt Range core complex, Eocene Carlin-type gold deposits, and Tertiary petroleum accumulations. The transect includes, from west to east, the Emigrant Pass area, the Piñon Range/Pine Valley area, Huntington Valley and the southern Ruby Mountains, the Medicine Range, and Spruce Mountain. These detailed cross-sectional restorations integrate data from new field observations and preexisting surface geologic maps, seismic reflection profiles, and exploratory oil wells to provide new insights into the Cenozoic structural evolution of the southern Ruby Mountains and the surrounding ranges. Upper crustal extension within the study area has been partitioned into zones of high and low extensional strain, and local extensional strains calculated from the rigorously reconstructed cross sections range from ~10% to greater than 100%. Synthesis of these restorations into a more schematic regional cross section leads to a regional estimate of ~32 km or ~50% upper crustal extension across the study area. Most of this regional extensional strain is accommodated by the west-dipping fault systems that exhumed the central and southern Ruby Mountains.

Extensional strain at Emigrant Pass, the Medicine Range, and Spruce Mountain has been accommodated by numerous closely-spaced “domino style” normal faults. One to two generations of domino-style faulting accommodate low to moderate (13 to 20%) extensional strains at Emigrant Pass and the Medicine Range. However, at Spruce Mountain, as many as six superimposed sets of domino-style normal faults have generated locally extreme extensional strains (~132%). The central Piñon Range appears to be the least extended of all areas examined.
in this study (~10% extension) and is primarily deformed by small- to moderate-offset, high-angle normal faults that have been accompanied by little overall fault block tilting.

The southern Ruby Mountains were primarily exhumed by at least two crosscutting systems of west-dipping normal faults that initiated at moderate to high-angles but were tilted to low angles by footwall flexure and east-directed tilting of the southern Ruby Mountains as extension progressed. The younger of these two fault systems formed modern day Huntington Valley. Differential tilting of the southern Ruby Mountains relative to the adjacent Piñon Range and northern Maverick Springs Range during active extension on these major west-dipping faults is interpreted to have been accommodated by kilometer-scale synformal folding in their footwalls and antiformal folding in their hanging walls. The southern Ruby Mountains and Huntington Valley area are estimated to have experienced 22.1 km or 94% extension since the late Eocene. As a consequence, the southern end of the Carlin trend is currently about twice the distance from the Ruby Mountains as it was at the time the gold deposits formed during the Eocene. The ages of extension are still uncertain in many places within the study area; however, it is clear that multiple phases of extension have affected northeastern Nevada beginning in early Tertiary time. The most significant episode of regional extension appears to have occurred during the middle Miocene. Extension and active faulting has also continued throughout the region well into Quaternary time.

In addition to having implications for the magnitude and kinematics of extension within the southern Ruby Mountains and surrounding areas, these cross sectional reconstructions provide new insights into both the present-day and pre-extensional structure of this region and may have implications for future exploration for mineral and petroleum resources.
Introduction

Cenozoic extension has played a critical role in shaping the geology of the Great Basin. Upper-crustal extensional strain has been heterogeneously distributed throughout the Great Basin, and zones that have experienced large magnitudes of upper-crustal extension are commonly juxtaposed against areas that have experienced comparatively little extension (e.g., Gans, 1987; Seedorff, 1991a). The Ruby Mountains East-Humboldt (RM-EH) metamorphic core complex located in northeastern Nevada is a key example of a domain of extreme extensional strain bordered by considerably less extended areas (Colgan and Henry, 2009). The present-day surface exposures of tectonically exhumed mid-crustal rocks within the RM-EH core complex juxtaposed against supracrustal rocks provide an excellent opportunity to investigate extensional processes at both shallow and mid-crustal levels. Moreover, the presence of world-class Carlin-type gold deposits (e.g., Cline et al, 2005) and important petroleum accumulations (e.g., Flanigan et al., 1990) in the Elko-Carlin region immediately west of the RM-EH core complex make this portion of northeastern Nevada an area of great geologic and economic interest.

Numerous studies have contributed to a better understanding of the timing, distribution, and style of extension within the RM-EH core-complex and adjacent areas of northeastern Nevada (e.g., Snoke, 1980; Snoke and Lush, 1984; Snoke and Miller, 1988; Snoke et al., 1990; Hurlow et al, 1991; McGrew and Snee, 1994; MacCready et al., 1997; McGrew et al., 2000; Satarugsa and Johnson, 2000; Howard, 2003; Sullivan and Snoke, 2007; Colgan and Henry, 2009). However, fundamental questions remain regarding both the pre-extensional structure of northeastern Nevada and the nature of upper-crustal Cenozoic extension within the RM-EH core-complex and surrounding areas. A topic of particular interest examined in this paper is how the
highly extended RM-EH core-complex relates to nearby areas that have experienced lower magnitudes of upper-crustal extension.

This study integrates existing geologic maps, seismic data, and oil well data in conjunction with selected field checking and mapping in order to develop a new geologic synthesis, resulting in a series of five cross-sectional restorations of Cenozoic upper-crustal extensional faulting in a ~150 km transect. This transect includes, from west to east, the Piñon Range/Pine Valley area, the southern portion of the RM-EH core complex, the Medicine Range, and Spruce Mountain. The results from the five rigorous reconstructions are then synthesized into a single, more schematic regional-scale reconstruction across the study area. These reconstructed cross sections provide insights into the structural architecture of this section of northeastern Nevada prior to the onset of Cenozoic extension and may have implications for hydrocarbon and mineral exploration in the region.

Geologic Setting

Northeastern Nevada underwent a complicated pre-Cenozoic geologic history that includes the deposition of a thick miogeoclinal sedimentary sequence, multiple Paleozoic and Mesozoic orogenic events, and Jurassic back-arc magmatism (e.g., Dickinson, 2006). By the end of the Late Cretaceous Sevier orogeny, the crust of northeastern Nevada had most likely been substantially thickened (e.g., Coney and Harms, 1984; Snoke and Miller, 1988; DeCelles, 2004). Potentially beginning as early as latest Cretaceous to earliest Tertiary time, this region of thickened crust began to undergo extensional collapse (Coney and Harms, 1984; Hodges et al., 1992; Camilleri and Chamberlain, 1997; McGrew et al., 2000; Sullivan and Snoke, 2007). This collapse was characterized by multiple extensional episodes, culminating in a period of
significant extension during the mid-Miocene (Colgan and Henry, 2009). Extensional faulting has continued in northeastern Nevada into Quaternary time, as evidenced by Quaternary fault scarps and historical seismic activity (e.g., Zoback et al., 1981; Dohrenwend et al., 1991; Wesnousky and Willoughby, 2003; dePolo, 2008).

Howard (2003) subdivided the region examined in this study into three broad tectonic domains. This subdivision provides a useful framework for considering the geologic history of the region and is adopted herein. From west to east, these domains are the Elko-Carlin domain, the RM-EH metamorphic core complex, and the Sevier hinterland domain (Fig. 1).

The Elko-Carlin Domain

The Elko-Carlin domain, as defined in this study, encompasses the western part of the study area including the Piñon Range/Pine Valley area, Emigrant Pass, as well as the Tuscarora Mountains and Adobe Range north of the primary study area (Fig. 1). The Elko-Carlin domain is an area of both geologic and economic significance. The main Carlin trend of Eocene (~42 to 36 Ma) disseminated gold deposits occurs within this domain (Cline et al., 2005), and important petroleum accumulations are present in Pine Valley (e.g., Flanigan et al., 1990).

After a long period of passive-margin sedimentation from Neoproterozoic to Late Devonian time, the Elko-Carlin domain was involved in the Late Devonian-Early Mississippian Antler orogeny (e.g., Speed and Sleep, 1982), and the leading edge of the Roberts Mountain allochthon and the western margin of the Antler foreland basin are present within the domain (Poole, 1974; Stewart, 1980). Following the Antler orogeny, the Elko-Carlin domain was affected by multiple episodes of contractional deformation between middle Mississippian and Cretaceous time (e.g., Dott, 1955; Vandervoort and Schmidt, 1990; Carpenter et al., 1993;
Trexler et al., 2003, 2004). Mesozoic volcanism occurred in the western portion of the Elko-Carlin domain during the Jurassic (Muffler, 1964), and Cenozoic volcanism occurred during the Eocene and Miocene (e.g., Smith and Ketner, 1976; Ressel and Henry, 2006). This area has also been affected by multiple episodes of Cenozoic extension beginning in the Eocene (Henry et al., 2001; Haynes, 2003; Colgan and Henry, 2009). However, despite multiple phases of extensional faulting, the Elko-Carlin domain appears to have experienced relatively little Cenozoic extensional strain compared to the RM-EH core complex domain immediately to the east (C. Postlethwaite, written comm., 2008; Colgan and Henry, 2009)

Ruby Mountains-East Humboldt Core Complex Domain

The core complex domain includes the RM-EH metamorphic core complex and the adjacent extensional basins (Fig 1). The RM-EH Range is a classic example of a Cordilleran metamorphic core complex (Snoke, 1980; Armstrong, 1982). From south to north, the RM-EH core complex exposes rocks of gradually increasing metamorphic grades and paleo-burial depths (Snoke et al., 1990; Hudec, 1992; Wright and Snoke, 1993; Howard, 2003; Sullivan and Snoke, 2007). In the southern Ruby Mountains, an east-tilted section of unmetamorphosed to lower greenschist-facies metamorphosed Precambrian through lower Mississippian miogeoclinal strata is exposed as a largely intact horst (Sharp, 1942; Willden and Kistler, 1979; Burton, 1997). In the central Ruby Mountains, the ~36 Ma Harrison Pass pluton intrudes metasedimentary rocks in a zone of rapidly increasing metamorphic grade (Wright and Snoke, 1993; Burton, 1997; Barnes et al., 2001), and amphibolite-grade metamorphosed rocks immediately north of the Harrison Pass pluton record penetrative deformation and metamorphism contemporaneous with the intrusion of Jurassic two-mica granites (Hudec, 1992). The northern Ruby Mountains and East Humboldt
Range expose high-grade metamorphic rocks characterized by migmatization and recumbent folding that are prominently overprinted on the western side of the range by a gently west-dipping mylonitic shear zone (Snoke, 1980; Howard, 1980; Snoke and Lush, 1984).

The RM-EH core complex experienced a complex, polyphase Mesozoic-Cenozoic deformational history. Multiple episodes of penetrative deformation, numerous Jurassic and Cretaceous intrusions, and at least two distinct metamorphic events affected the RM-EH core complex during Mesozoic crustal thickening and tectonic burial (e.g., Howard, 1980; Snoke; 1980; Snoke and Miller, 1988; Hudec, 1992; Camilleri and Chamberlain, 1997; McGrew et al., 2000; Sullivan and Snoke, 2007). Peak metamorphic conditions were reached in the Late Cretaceous (Hodges et al., 1992; Camilleri and Chamberlain, 1997; McGrew et al., 2000), suggesting that the most recent crustal thickening event in northeastern Nevada may have occurred at this time (Colgan and Henry, 2009). Clockwise P-T-t paths record exhumation in the northern part of the range as early as the Late Cretaceous, and ~7 km of tectonic denudation is inferred to have occurred in the East Humboldt Range from the Late Cretaceous to early Tertiary (Hodges et al., 1992; McGrew and Snee, 1994; McGrew et al., 2000). The initial formation of the extensional mylonitic shear zone in the northern Ruby Mountains likely occurred during the early to middle Eocene (Mueller and Snoke, 1993; McGrew and Snee, 1994; Sullivan and Snoke, 2007), and $^{40}$Ar/$^{39}$Ar and K/Ar cooling ages of biotite, muscovite, and hornblende have been interpreted to indicate that significant tectonic denudation of the RM-EH core complex occurred during mid- to late-Oligocene time (Kistler et al., 1981; Dallmeyer et al., 1986; McGrew and Snee, 1994). Plutonism occurred in the late Eocene (~40 to 36 Ma) and again in the Oligocene (~29 Ma) (Wright and Snoke, 1993; Barnes et al., 2001). Middle Miocene apatite fission-track and (U-Th)/He ages from the Harrison Pass pluton and the deposition of significant thicknesses
of Miocene coarse clastic rocks in basins adjacent to the RM-EH core complex suggest a phase of major upper-crustal extension occurred within the core complex domain between 17 and 10 Ma (Colgan and Metcalf, 2006; Colgan and Henry, 2009).

The RM-EH core complex is flanked by fault-controlled sedimentary basins that were formed during Tertiary extension. These basins are Ruby Valley, Huntington Valley, and Lamoille Valley, which are respectively situated to the east, west, and northwest of the Ruby Mountains (Fig. 1). Interpretations of seismic reflection profiles demonstrate that Huntington and Lamoille Valleys are half-grabens controlled by shallowly west-dipping (20°-26°) normal faults on their eastern margins (Satarugsa and Johnson, 2000). Conversely, seismic reflection profiles across Ruby Valley north of the Medicine Range show that it is a full-graben, flanked by a high-angle (~60°) east-dipping normal fault system on the west and a moderately west dipping (~38°) normal fault on the east (Satarugsa and Johnson, 2000).

**Sevier Hinterland Domain**

The eastern portion of the study area, including the Spruce Mountain area and Medicine Range, is part of the much more extensive region of eastern Nevada and western Utah that comprises the hinterland of the Sevier fold and thrust belt (Armstrong, 1968; DeCelles, 2004). The Sevier hinterland domain experienced passive margin (dominantly carbonate) sedimentation from Neoproterozoic through Devonian time (e.g., Stewart, 1980) and was located within the Antler foreland basin during the Mississippian (e.g., Poole, 1974). Carbonate sedimentation resumed in the Pennsylvanian, and a thick sequence of Pennsylvanian and Permian carbonate rocks were deposited in the Sevier hinterland domain during this time (e.g., Snyder et al., 1991). Marine sedimentation ended in the Triassic, and the Sevier hinterland domain underwent
compressional deformation, structural thickening, and intrusion during the Mesozoic (e.g.,
DeCelles, 2004; Dickinson, 2006; du Bray, 2007). Volcanism occurred within the Sevier
hinterland domain during the Eocene (Brooks et al., 1995), and extensional faulting affected the
area at various times during the Cenozoic (e.g., Seedorff, 1991a).

**Stratigraphic Framework**

Rocks that crop out within the study area include Proterozoic through Triassic
miogeoclinal sedimentary rocks, limited exposures of Mesozoic volcanic and lacustrine rocks,
Tertiary to recent sedimentary and volcanic rocks, and Jurassic, Cretaceous, and Tertiary
intrusions (e.g., Stewart, 1980; Coats, 1987). The general stratigraphic relationships from west to
east across the study area are summarized in Figure 2.

In the Elko-Carlin domain, Paleozoic stratigraphic relationships are complicated by
effects of the Antler orogeny and subsequent Paleozoic deformation events (Trexler et al., 2003,
2004). In general, however, the stratigraphy in the southern portion of the Elko-Carlin domain
consists of a thick (>11km) Neoproterozoic through Upper Permian sedimentary sequence
overlain unconformably by Cenozoic sedimentary and volcanic rocks. Mississippian clastic
strata deposited in the Antler foreland basin attain significant thicknesses (>2 km) in the Piñon
Range (Smith and Ketner, 1975) but are absent further west where Ordovician to Devonian rocks
of the Roberts Mountains allochthon have been thrust over Neoproterozoic to Devonian strata
(e.g., Stewart, 1980). Jurassic volcanic rocks of the Pony Trail Group (Muffler, 1964) and
Cretaceous nonmarine sedimentary rocks unconformably overlie older Paleozoic strata in the
westernmost parts of the study area but are only locally present further east (Smith and Ketner,
1976). Throughout much of the Elko-Carlin domain, Cenozoic strata unconformably overlie
older sedimentary and volcanic rocks and include Eocene lacustrine sediments of the Elko Formation, Eocene-Oligocene volcanic and tuffaceous sedimentary rocks of the Indian Well Formation, Miocene lacustrine, volcanioclastic, and alluvial sedimentary rocks of the Humboldt Formation, and Pliocene and Pleistocene sedimentary rocks of the Hay Ranch Formation (Smith and Ketner, 1976).

A thick (~5 km) section of unmetamorphosed to lower greenschist facies Neoproterozoic to early Mississippian aged strata is exposed within the southern Ruby Mountains (Sharp, 1942; Willden and Kistler, 1979; Burton, 1997). Paleozoic rocks in the southern Ruby Mountains generally display conformable relationships; however, several workers have recognized an unconformity between the Ordovician Pogonip Group and the Silurian Lone Mountain Dolomite (Sharp, 1942; Willden and Kistler, 1979; Burton, 1997). Late Mississippian strata of the Diamond Peak Formation, ~ 1 km thick, are preserved in the Mitchell Creek klippe (Fig. 1) and are in fault contact with older Paleozoic strata on the western side of the southern Ruby Mountains (Sharp, 1942; Willden et al., 1967; Willden and Kistler, 1979). Although Pennsylvanian and younger strata are absent within the southern Ruby Mountains (Willden and Kistler; 1979), a 3- to 4-km thick section of upper Paleozoic rocks was presumably present above Mississippian strata prior to Mesozoic compression and Tertiary extension within the southern Ruby Mountains (Burton, 1997).

In the Sevier hinterland domain, a considerable thickness (>12 km) of miogeoclinal sedimentary rocks was deposited between Neoproterozoic and Lower Triassic time (e.g. Stewart, 1980; Camilleri and Chamberlain, 1997). Upper Paleozoic and Mesozoic rocks are unconformably overlain by Eocene volcanic rocks (Brooks et al., 1995) and/or Miocene clastic and volcanioclastic strata of the Humboldt Formation (Coats, 1987). Mesozoic and Tertiary
intrusive rocks crop out in numerous localities within the Sevier hinterland domain, and interpretations from aeromagnetic and magnetotelluric data suggest that several large plutonic bodies are present in the subsurface (Fig. 1; Coats, 1987; Grauch, 1996; Wannamaker and Doerner, 2002; du Bray, 2007).

**Restorations – West to East Transect**

In the following section, a series of five detailed cross-sectional restorations of extensional faulting across individual ranges and basins within the study area are presented in a west-to-east transect (Fig. 1). Each of these restorations emphasizes: 1) the magnitude of local extensional strain, 2) the timing and style of extension, and 3) the pre- and post-extensional structure of each area. These retrodeformable geologic cross sections in part are based on new geologic mapping (Appendix 1) and on field observations at selected sites across the transect during two months of field work, but they integrate data from published surface geologic maps, published seismic reflection profiles, and exploratory oil wells to provide new constraints on the style, timing, and magnitude of crustal extension within the region.

**Emigrant Pass**

Emigrant Pass is located south of the Carlin trend within the Elko-Carlin domain (Fig. 1, 3). The presence of an extensive, well-dated Tertiary volcanic and sedimentary section within the Emigrant Pass quadrangle combined with detailed surface geologic mapping (Henry and Faulds, 1999) make it an area of interest for evaluating the timing and magnitude of local Tertiary extension that has occurred in the vicinity of the southern portion of the main Carlin trend (Henry et al., 2001).
Cross Section A to A’

To illustrate the Cenozoic structural evolution of the Emigrant Pass area, a cross-sectional restoration of extensional faulting was performed for a line of section (A to A’ in Fig. 3) across the southern portion of Emigrant Pass (Fig. 4). The primary marker beds employed in this restoration are Oligocene (~25 Ma) and Eocene (~38 to 36 Ma) volcanic rocks and the mid-Eocene Elko Formation (Fig. 4, Henry and Faulds, 1999). Thickness estimates of formations are based on measurements by Henry and Faulds (1999) and the map patterns of geologic units. The presence of a positive magnetic anomaly in the northern portion of the Emigrant Pass area (Hildenbrand and Kucks, 1988; Ressel and Henry, 2006) combined with the abundance of lava flows and subvolcanic intrusions that crop out in the area suggests that a large intrusive body may be present in the subsurface at relatively shallow depths (≤3 km) (Ressel and Henry, 2006). However, the locations of any shallow intrusive bodies that may project into cross section A to A’ are uncertain, and no intrusions are shown in Figure 4.

The Emigrant Pass area is deformed by numerous north-south striking, predominantly west-dipping, normal faults (Fig. 3). The surface traces of the normal faults with respect to topography in section A to A’ indicate that the faults are high-angle structures. However, there are few direct dip measurements (Fig. 3; Henry and Faulds, 1999), and actual dips of the faults may vary as much as ±10° from the dips shown in Figure 4a. The effect of increasing the dips of major faults would be to reduce the estimated amount of extension, whereas decreasing the dips would increase the amount of extension. In general, the attitudes of bedding at Emigrant Pass are similar in the hanging wall and footwall of faults, implying that faults have dominantly planar or curviplanar subsurface geometries.
Cross section A to A’ was restored as a series of rigid fault blocks. Faults are simplified as planar structures, and fault blocks are treated as perfectly rigid bodies. These simplifying assumptions lead to small gaps/overlaps (Fig. 4), which are assumed to be compensated by small-scale internal deformation of blocks (e.g., Proffett, 1977; Colgan et al., 2008).

*Age of Extension at Emigrant Pass*

The well dated Tertiary volcanic and sedimentary sequence at Emigrant Pass allows for excellent constraints on the age of Tertiary extension in the vicinity of the southern Carlin trend. Two separate episodes of Tertiary extension are well defined at Emigrant Pass, and additional Tertiary extensional episodes may have also affected the area (Henry et al., 2001). The Elko Formation may have been deposited in an erosional basin at Emigrant Pass (Henry, 2008), but if the middle Eocene Elko Formation was instead deposited in an extensional basin, then the earliest episode of Tertiary extension may have occurred prior to or during the deposition of the Elko Formation (Henry et al., 2001; Haynes, 2003).

The earlier of the two known phases of extension at Emigrant Pass is estimated to have occurred between ~40 Ma and 38 Ma. This episode of extension is evident from tilting of the middle Eocene Elko Formation, which strikes consistently northeast, typically dips between ~15° and ~35° southeast, and is in slight (~10°-15°) angular unconformity with overlying Eocene volcanic rocks (Henry and Faulds, 1999; Henry et al., 2001). Based on the southeast dip of bedding of the Elko Formation, it is inferred that this earlier phase of extension probably had a northwest-southeast orientation (Henry et al., 2001). However, no northeast-southwest striking, northwest-dipping normal faults that clearly belong to this earlier episode of Eocene extension have been mapped within the Emigrant Pass area or in the Piñon Range/Pine Valley area to the
southeast (Smith and Ketner; 1978; Henry and Faulds, 1999), and it is possible that these middle Eocene normal faults have been covered by younger volcanic and sedimentary deposits (Fig. 4).

A later Tertiary extensional episode, which initiated between 25 and 15.2 Ma, resulted in eastward tilting of the Eocene-Oligocene volcanic rocks in the Emigrant Pass area by numerous closely spaced, north-south striking, west-dipping normal faults (Henry et al., 2001). Middle Miocene rhyolite lava flows and tuffaceous sedimentary rocks dated at ~15 Ma also are gently east tilted (although less so than the underlying Eocene-Oligocene volcanic rocks) and are cut by north-south striking normal faults, indicating that Tertiary extension younger than 15 Ma has also occurred at Emigrant Pass (Fig. 3, Henry and Faulds, 1999; Henry et al., 2001). This post-15 Ma extension may represent a continuation of the extensional episode that initiated between 25 and 15.2 Ma or may have been a discrete, later phase of Tertiary extension at Emigrant Pass (Henry et al., 2001). In addition to these episodes of Tertiary faulting and associated tilting, the Bob’s Flat fault is mapped as cutting Quaternary deposits (Henry and Faulds, 1999), implying that this fault has been active into latest Cenozoic time.

Present Day Structure

The present-day subsurface structure of the southern Emigrant Pass area along section A to A’ is presented in Figure 3a. The north-south striking west-dipping normal faults that are mapped at the surface in the Emigrant Pass area are younger than 25 Ma, and most of these faults were probably active during regional middle Miocene extension that occurred between ~17 and 10 Ma (Colgan and Henry, 2009). Individual faults generally have offsets of 10’s to 100’s of meters. The faults with the largest interpreted offsets in section A to A’ are the Bob’s Flat fault and the Primeaux Springs fault, which have offsets of ~460 m and ~420 m, respectively.
Buried middle Eocene aged faults that caused the southeast-oriented tilting of the Elko Formation are interpreted to be present in the subsurface in the southeastern portion of the Emigrant Pass area; however, the locations of these faults are uncertain, and their traces are dashed and queried in Figure 4. Additional buried middle Eocene aged faults may also be present in the western portions of Emigrant Pass, but no data are available to constrain the potential subsurface locations of these faults, and no middle Eocene faults are shown in the western segment of section A to A’.

Subsurface Distribution of the Elko Formation and Eocene Normal Faults

Emigrant Pass is the westernmost location of known outcrops of the Elko Formation (Henry et al., 2001; Haynes, 2003), and a middle Eocene paleotopographic high is interpreted to have been present in the approximate location of the modern day northern Piñon Range during deposition of the Elko Formation (Haynes, 2003). The map pattern of the Elko Formation suggests that it underlies younger Eocene volcanic rocks throughout the southeastern portion of the Emigrant Pass area, dipping between ~15° and 35° to the southeast (Fig. 3). If the contact between the Elko Formation and underlying Paleozoic rocks was originally subhorizontal (i.e., the Elko Formation was originally deposited in an area of little topographic relief), projecting the Elko Formation into section A to A’ as an intact southeast-dipping wedge would result in it attaining a cross sectional thickness of >1 km in the southeastern portion of the section. This thickness significantly exceeds measured thicknesses of the Elko Formation at Emigrant Pass and regionally (Smith and Ketner, 1976; Haynes, 2003). Henry and Faulds (1999) give a minimum thickness estimate of the Elko Formation at Emigrant pass of ~125 m, whereas Haynes (2003) measured a thickness of only 56 m of Elko Formation at Emigrant Pass and estimated that
the formation may reach a maximum thickness of ~130 m. Regionally, the Elko Formation is generally \( \leq 600 \) m thick (Henry, 2008). Thus, it seems unlikely that the Elko Formation attains a substantial thickness (e.g., >500 m) in the subsurface at Emigrant Pass.

In light of the inferred location of Eocene paleotopographic highs east of Emigrant Pass and the multiphase extensional history at Emigrant Pass (Henry et al., 2001; Haynes, 2003), two subsurface geometries can reasonably account for the map pattern of the Elko Formation at Emigrant Pass without requiring substantial thickening of the formation in the subsurface, either 1) the Elko Formation is cut by west-dipping normal faults associated with the ~40 to 38 Ma phase of extension that are presently covered by younger strata, or 2) the original depositional contact of the Elko Formation with underlying Paleozoic rocks was not subhorizontal but rather northwest-dipping (i.e., when it was deposited, the Elko Formation onlapped onto an area of higher topographic elevation to the southeast). At Emigrant Pass, both subsurface geometries appear to be reasonable assumptions, and both buried west-dipping faults and a flattening of the Elko-Paleozoic contact are shown in the reconstructed cross section in order to minimize the subsurface thickness of the Elko Formation (Fig. 4).

**Restored Pre-Extensional Structure**

The first restored panel of cross section A to A’ (Fig. 2b) restores the post-25 Ma north-south striking, west dipping faults. Based on this restoration, the total amount of post-25 Ma extension is estimated to be approximately 1 km, or about 13% extension, across section A to A’. An average of 10° of tilting has been restored in Figure 2b, leading to initial fault dips of 65° to 75°. The reconstructed geometry is consistent with deposition of the Eocene to Oligocene volcanic sequence in a shallow basin of relatively low topographic relief with volcanic strata.
pinching out against a paleotopographic high located to the southeast. Although the thin Oligocene ash-flow tuff that comprises the uppermost unit of the Eocene-Oligocene volcanic sequence is represented as a continuous, thin layer in Figure 2b, Henry and Faulds (1999) interpreted this unit as having been deposited in modest paleovalleys in the Emigrant Pass area, and the original distribution of this formation may have been more discontinuous.

A final reconstructed panel (Fig. 2c) shows the possible structural configuration of the eastern portion of section A to A’ prior to the mid-Eocene extensional episode. Given the uncertainties regarding the original distribution of the Elko Formation as well as the subsurface locations of normal faults related to this earlier phase of extension, this final restored panel is less certain. An additional 10° of tilting has been applied to this final reconstructed panel, consistent with the 10° to 15° angular unconformity present between the Elko Formation and overlying Eocene volcanic sequence at Emigrant Pass (Henry and Faulds, 1999). This restoration suggests that middle Eocene faulting may have accommodated ~10% extension in the Emigrant Pass area, although this estimate is highly uncertain. Nonetheless, the small magnitude of the angular unconformity between the Elko Formation and Eocene volcanic sequence is consistent with relatively modest extensional strains during middle Eocene extension in the Emigrant Pass area (Henry et al., 2001).

It is likely that some uplift and erosion of the Elko Formation took place during the ~40 Ma period of extension and tilting of the formation in the Emigrant Pass area (Fig. 3c). However, unless this uplift and erosion were fairly significant prior to the deposition of the Eocene-Oligocene volcanic sequence beginning at ~38 Ma, then it is likely that the Eocene paleogeographic distribution of the Elko Formation in the Emigrant Pass area was similar to its modern distribution. If this is the case, it would imply that the sedimentary basin into which the
Elko Formation at Emigrant Pass was deposited was of a relatively restricted areal extent, and that the Elko Formation likely pinches out both to the east and west in the vicinity of Emigrant Pass. This type of basin geometry is consistent with the hypothesis of Henry (2008) that the Eocene Elko Basin was made up of numerous discrete extensional or erosional basins, as opposed to a regionally extensive, contiguous lacustrine basin (e.g., Solomon, 1992).

**Pine Valley/Central Piñon Range**

The Piñon Range is located on the border of Elko and Eureka Counties, south of the town of Carlin (Fig. 1, 5). The Piñon Range experienced a complicated pre-Cenozoic deformational history characterized by numerous Paleozoic and Mesozoic contractional events (e.g., Dott, 1955; Johnson and Pendergast, 1981; Speed and Sleep, 1982; Vandervoort and Schmidt, 1990; Carpenter et al., 1993; Trexler et al., 2003, 2004). These contractional events have produced upright to east-vergent folds and west-dipping thrust faults that exhibit eastward hanging wall translation in the Piñon Range (Carpenter et al., 1993). Subsequently, low magnitude Cenozoic extensional faulting (~10% extension) has been superimposed on older contractional structures within the Piñon Range, and has resulted in the formation of the Pine Valley basin to the west of the Piñon Range (Gordon and Heller, 1993; Wallace et al., 2008).

**Cross Section B to B’**

To illustrate the Cenozoic structural evolution of the Piñon Range area, a cross-sectional restoration of extensional faulting was performed for a line of section (B to B’ in Fig. 5) that traverses the central Piñon Range and the adjacent Pine Valley and Dixie Flats areas (Fig. 6). The primary marker beds employed for restoring fault offsets in this reconstruction are Devonian
through Permian Paleozoic strata and the Paleogene Elko and Indian Well Formations.

Thicknesses of Paleozoic units are based on measurements reported by Smith and Ketner (1975).

Published seismic interpretations and geologic cross sections through the Piñon
Range/Pine Valley area based on borehole, seismic, and surface geologic data have interpreted
normal faults in the Piñon Range/Pine Valley area as having planar to slightly listric subsurface
geometries (Carpenter et al., 1993; Gordon and Heller, 1993; Hansen et al., 1994a, b; Flanigan,
1994). These interpretations are consistent with the observation that strikes and dips of bedding
typically do not change substantially between the footwall and hanging wall blocks across
normal faults within the Piñon Range, suggesting that these faults are planar or curviplanar in the
subsurface.

Pine Valley

The western portion of cross section B to B’ traverses Pine Valley, a half-graben
controlled by the west-dipping Pine Valley fault that bounds the eastern side of the basin
(Gordon and Heller, 1993). Two oil wells in Pine Valley were projected into section B to B’,
Tomera Ranch South No. 9-1 and Evans Flat No. 11-1 (Table 1; Fig. 6).

Constraints on the subsurface geology of Pine Valley on the western end of section B to
B’ are primarily based on data from the Tomera Ranch South No. 9-1 well that was spudded
~400 m north of section B to B’ (Fig. 5; Smith and Ketner, 1978; Hess, 2004). The stratigraphic
tops chosen for the Tomera Ranch South No. 9-1 well in this study (Table 1) are based primarily
on lithologic logs and differ somewhat from the original tops picks by the operator in that: 1) a
sequence of interbedded tuffaceous sedimentary rocks and gravel between 567 and 683 m depth
originally assigned to the Indian Well Formation has been assigned to the overlying Humboldt
Formation, 2) a sequence of volcanic tuffs encountered between 695 and 1112 m has been assigned to the Indian Well Formation, and 3) a sequence of siltstone, sandstone, and minor mafic igneous rocks encountered between 1112 and 1216 m is interpreted as western facies rocks in the upper plate of the Roberts Mountain allochthon. Additional constraints on the subsurface structure beneath Pine Valley are based on structural interpretations of the Tomera Ranch oil Field located ~2.5 km north of section B to B’ (Figs. 5 and 6; Hansen et al., 1994a). Hansen et al. (1994a) interpreted three Neogene normal faults in the subsurface below the Tomera Ranch field, which have west, southwest, and south dips. The west-dipping fault is interpreted to project southward into section B to B’ and is labeled as the Tomera Ranch fault in Figure 6.

The dip of the Pine Valley fault is constrained by data from Evans Flat No. 11-1 well. This well was spudded ~1.7 km north of the trace of section B to B’ and approximately 900 m from the nearest outcrops of Mississippian siliciclastic rocks in the Piñon Range (Fig. 5; Smith and Ketner, 1978; Hess, 2004). The well was drilled to a total depth of 1288 m without intersecting the west-dipping Pine Valley fault (Table 1; Hess, 2004). The fault must be present between the surface outcrops of Mississippian siliciclastics and the Evans Flat No. 11-1 well, and cross-sectional analysis indicates that the fault must dip more steeply than 55°; otherwise, it would have been intersected by the well.

Central Piñon Range

Most of the high-angle faults that are present within the central Piñon Range are widely interpreted to be Tertiary normal faults. One exception is the approximately north-south striking Willow Creek Fault that runs along the crest of the central Piñon Range (Fig. 5), which has been subject to varying interpretations by different workers. The Willow Creek fault was originally
interpreted as a high-angle, east-dipping normal fault by Smith and Ketner (1978). However, subsequent field studies by Carpenter et al. (1993) revealed that this fault dips 50° to 60° west in the vicinity of Willow Creek, ~10 km south of cross section B to B’, and accommodates a reverse sense of offset. Carpenter et al. (1993) interpret the Willow Creek fault as an east-vergent thrust fault that initiated in Latest Devonian to earliest Mississippian time, just prior to the emplacement of Roberts Mountain allochthon in the Piñon Range, and subsequently experienced east-vergent compressional reactivation during the middle to late Pennsylvanian, and again during the Cretaceous. Alternatively, the Willow Creek fault has been interpreted as a right-lateral transpressional fault of probable Cretaceous age (Ransom and Hansen, 1993). Although the structural history of the Willow Creek fault remains uncertain, it is here interpreted as pre-extensional reverse fault of compressional or transpressional origin.

In the Railroad mining district ~4 km north of section B to B’, the Eocene (~37 Ma) Bullion stock and numerous porphyritic rhyolite dikes intrude Paleozoic strata (Fig. 5; Smith and Ketner, 1978). Based on outcropping intrusive rocks and a positive magnetic anomaly ~10 km in diameter centered 1 to 2 km southwest of the Bullion stock, a large Eocene plutonic body is interpreted to underlie the northern and central Piñon Range (Smith and Ketner, 1976; Hildenbrand and Kucks, 1988; Grauch, 1996; Ressel and Henry, 2006). A zone of low resistivity on a magnetotelluric profile that crosses the central Piñon Range nearly coincident with the trace of section B to B’ also suggests the presence of a large plutonic body beneath the range (Wannamaker and Doerner, 2002). This plutonic body is likely present beneath the Railroad district at relatively shallow depths (≤ 3 km) (Wannamaker and Doerner, 2002; Ressel and Henry, 2006). However, section B to B’ crosses the Piñon Range just south of the positive aeromagnetic anomaly, so shallow intrusive bodies are not depicted in the cross section (Fig. 6).
Dixie Flats

East of the central Piñon Range, a gently west dipping (10° to 20°) sequence of Eocene-Oligocene Indian Well Formation crops out over much of the southern Dixie Flats area (Smith and Ketner, 1978; Palmer et al., 1991). The east-dipping Robinson Mountain fault has been projected ~1.5 km north from mapped outcrops (Smith and Ketner, 1978) and is interpreted to bound the eastern side of the central Piñon Range where it is traversed by section B to B’ (Figs. 5, 6) Although the thickness of the Indian Well Formation in the Dixie Flats area is not well constrained, it may be quite thick. West of the northern part of Cedar Ridge, ~10 km north of cross section B to B’, Smith and Ketner (1976) measured a thickness of 1015 m of Indian Well Formation. A similar thickness of ~900 m was estimated for the Indian Well Formation by Palmer et al. (1991) in the Hackwood Ranch area ~4.5 km north of cross section B to B’.

Isolated outcrops of Elko Formation are also exposed in the Dixie Flats area (Smith and Ketner, 1978), and the Elko Formation may underlie the Indian Well Formation throughout much of the area between the Piñon Range and Cedar Ridge (Haynes, 2003). Exploratory oil wells near Jiggs encountered thicknesses of Elko Formation in excess of 500 m (Schalla, 1992; Haynes, 2003), and similar subsurface thicknesses of Elko Formation may be present in the Dixie Flats area. Paleozoic rocks underlying the Dixie Flats area most likely are Mississippian through Permian siliciclastic rocks that are assumed to be deformed in upright folds of a similar structural style to Paleozoic rocks cropping out in the Piñon Range to west.

Present Day Structure
The present-day subsurface structure of the Piñon Range area along section B to B’ is presented in Figure 5a. The central Piñon Range is interpreted as a horst block composed of folded Paleozoic strata bound on the west by the Pine Valley fault and on the east by the Robinson Mountain fault. Individual faults generally have offsets of 10’s to 100’s of meters, and the Pine Valley fault has the greatest amount of interpreted offset (~1150 m). Although the Pine Valley fault dips west at high angles (>55°) in the northern part of the Pine Valley basin, the fault(s) that bound the eastern margin of southern Pine Valley apparently dip much more shallowly. Based on borehole data, Flanigan (1994) interpreted the basin-bounding fault of southern Pine Valley as dipping ~25° west in the Blackburn Oil Field, located ~30 km to the south of section B to B’. Similarly, a published interpretation of a shallow seismic reflection profile from southern Pine Valley immediately to the north of the Blackburn Oil Field shows the basin bounding fault dipping 15° to 20° west (Gordon and Heller, 1993). This apparent southward shoaling of the west-dipping, basin-bounding fault(s) within Pine Valley may reflect a fundamental along-strike change in the style of extensional faulting and/or magnitude of extensional strain that has occurred within Pine Valley. Northern Pine Valley and the central Piñon Range are interpreted to have experienced relatively small extensional strains and little fault block tilting during Tertiary extension (see discussion below). Conversely, the presence of shallowly dipping normal faults in southern Pine Valley may indicate that larger extensional strains and/or greater magnitudes of fault block tilting occurred during Tertiary extension in this area.

Fault Block Tilting
The central Piñon Range does not appear to have experienced significant amounts of fault block tilting during extensional faulting. Rocks within the Piñon Range dip only ~5° east. Steeply east-dipping conglomerates on the east side of the southern Piñon Range, originally mapped as early Eocene (Smith and Ketner, 1978), may, in fact, be Cretaceous, and no gently dipping normal faults have been mapped cutting Paleozoic rocks within the Piñon Range. Also, Paleozoic rocks are present in upright folds within the range instead of showing preferential tilts to the east or west (Colgan and Henry, 2009). Taken together, these observations suggest that only a minimal amount of net block tilting has accompanied extensional faulting in the central Piñon Range.

It is, however, worth noting that both east and west dipping, north-south striking, high-angle normal faults with similar magnitudes of offset are present within the Piñon Range/Pine Valley area. These east- and west-dipping faults would produce opposite senses of block tilting. Thus, the tilting associated with movement on east dipping faults, for example, would tend to reverse the tilting associated with movement on west dipping faults. Therefore, if the respective magnitudes of block tilting associated with slip on the east and west dipping faults are similar, the net interaction of these oppositely dipping normal faults would produce a block-faulted terrain that appears largely untilted (e.g., Axen, 1986; Maher, 2008).

**Restored Pre-Extensional Structure**

The restored pre-extensional structural interpretation of cross section B to B’ is presented in Figure 6b. Extensional faults postdate the deposition of the Eocene-Oligocene Indian Well Formation, and late Eocene and older rocks are preserved in the restored cross section. The central Piñon Range has experienced only a modest amount of Cenozoic extension (~1.6 km or
Cenozoic extensional faulting has primarily been responsible for formation of the Pine Valley basin west of the Piñon Range and for down-dropping Indian Well Formation and older rocks east of the Piñon Range in the Dixie Flats area. Meanwhile, the Piñon Range itself has behaved as a largely intact, uplifted horst block.

The geometry of the Indian Well Formation in the restored cross section suggests that it may have originally covered the central Piñon Range prior to extensional faulting (Fig. 6b). However, in the present day cross section (Fig. 6a), the Indian Well Formation beneath Pine Valley appears to thin westward toward the Piñon Range. If this thinning is reflective of the original depositional distribution of the formation, then the Piñon Range may have been a paleotopographic high during the deposition of the Indian Well Formation. Further evidence that the Piñon Range may have been an Eocene paleotopographic high is based on the distribution of mid-Eocene sedimentary rocks. The Elko Formation is absent in Pine Valley and the western Piñon Range, whereas considerable thicknesses of Elko Formation have been mapped immediately east of the northern Piñon Range and encountered in boreholes in Huntington Valley (Smith and Ketner, 1976; Schalla, 1992; Solomon, 1992; Haynes 2003). The Elko Formation is also present in the Dixie Flats area but is absent further west (Fig. 5, 6), possibly indicating that the Piñon Range and Pine Valley area were topographically elevated relative to the Dixie Flats area during the mid-Eocene.

**Age of Extensional Faulting**

Normal faults within the central Piñon Range area generally cut, and are, therefore, younger than, the 38-33 Ma Indian Well Formation (Smith and Ketner, 1978). In the Dixie Flats area, Palmer et al. (1991) noted an angular unconformity between 35.8 - 37.0 Ma tuffs belonging
to the Indian Well Formation and an overlying, areally restricted tuff dated at 30.8 Ma, which they interpreted to be related to Late Eocene/Early Oligocene extensional faulting. Dates on supergene alunite from the Rain mine ~15 km north of section B to B’ range from 22 to 18 Ma (Williams, 1992) which are interpreted to indicate that uplift and erosion, possibly related to extensional faulting, was occurring in the northern Piñon Range at that time (Wallace et al., 2008). In Pine Valley, Wallace et al. (2008) note that an angular unconformity is present between the moderately (35°-20°) east-dipping Indian Well Formation and the more shallowly (20°-10°) east-dipping middle Miocene Humboldt formation in northeastern Pine Valley (Smith and Ketner, 1978), implying a period of middle Miocene extension on west dipping faults in northern Pine Valley. Additionally, Pine Valley basin was actively subsiding during deposition Pliocene-Pleistocene Hay Ranch Formation, presumably as a result of slip on the Pine Valley fault (Gordon and Heller, 1993). The above constraints suggest that multiple phases of extension have occurred within the central Piñon Range area since late Eocene time. However, with the exception of the recently active Pine Valley fault, it is difficult to constrain the age(s) of slip on individual faults. In general, extensional faults in the Piñon Range area are best dated as post-late Eocene, and the majority of normal faults are probably of middle Miocene age or younger.

Southern Ruby Mountains

A moderately east-tilted (~20° to 50°) section of Proterozoic through lower Mississippian miogeoclinal strata is exposed in the southern Ruby Mountains as a structurally intact horst block (Figs. 1 and 7; Sharp, 1942; Willden and Kistler, 1979; Burton, 1997). These Precambrian and Paleozoic rocks are locally intruded by outlying dikes and apophyses of granodiorite and monzogranite of the late Eocene (36 Ma) Harrison Pass pluton (Wright and Snoke, 1993; Barnes
et al., 2001), the main body of which is exposed immediately to the north (Fig. 7a). The southern Ruby Mountains underwent significant extension during Tertiary time, and major gently west-dipping, normal faults are exposed beneath extensional klippen on the western side of the range (Sharp, 1942; Willden et al., 1967; Willden and Kistler, 1979; Snoke and Lush, 1984; Burton, 1997).

Cross Section C to C’

To illustrate the Cenozoic structural evolution of the southern Ruby Mountains, a cross-sectional restoration of upper-crustal extensional faulting was performed for a line of section (C to C’ in Fig. 7a) that traverses Huntington Valley, the southern Ruby Mountains, southern Ruby Valley, and the northern Maverick Springs Range (Fig. 8). The primary marker beds employed for restoring fault offsets in this reconstruction are Cambrian through Permian strata and Eocene tuffs and sedimentary rocks. Thicknesses of Paleozoic rocks in the southern Ruby Mountains are primarily based on measurements by Willden and Kistler (1979) and Burton (1997), as well as the map patterns of geologic units.

Huntington Valley

Constraints on the subsurface structure of Huntington Valley are primarily provided by exploratory oil wells (Hess, 2004), published seismic reflection profiles (Reese, 1986; Satarugsa and Johnson, 2000), and Bouguer gravity data (Ponce et al., 1996). Additionally, copies of both published and unpublished seismic reflection profiles from Huntington and Ruby Valleys originally interpreted by Satarugsa and Johnson (2000) were made available for this study. The locations of key seismic profiles from southern Huntington Valley are shown in Figure 7a.
Two exploratory oil wells located in Huntington Valley, Federal No. 16-5 and Aspen
Unit No. 1 (Hess, 2004), are projected into section C to C’ (Fig. 7; Table 2). Federal No. 16-5
was spudded ~3.7 km west of the southern Ruby Mountains and drilled to a total depth of 1,267
m (Hess, 2004). A sequence of ~500 m of volcaniclastic rocks and tuffaceous siltstone and
claystone underlain by ~100 m of cherty limestone encountered between 518 and 1128 m depth
in Federal No. 16-5 has previously been assigned to the Elko Formation (46 – 39 Ma) based on
lower Eocene age dates from palynological analyses of well cuttings between 768 and 1122 m
depth (TH Geological Services, 1994; Haynes, 2003). However, the thick oil shale sequence and
basal conglomerates that are generally characteristic of the Elko Formation east of the Piñon
Range (Smith and Ketner, 1976; Solomon, 1992; Haynes, 2003) are absent in Federal No. 16-5.
The Eocene strata encountered by Federal No. 16-5 more closely resemble the sequence of
limestone and volcanic rocks characteristic of the middle Eocene (~43 – 39 Ma) Northern
Nevada Volcanic field (Brooks et al., 1995); hence, these lower to middle(?) Eocene strata are
not assigned to a formal stratigraphic unit. No direct age constraints are available for the
volcaniclastic rocks encountered between 518 and 768 m depth. The late Eocene-early Oligocene
(38-33 Ma) Indian Well Formation, which crops out in the Piñon Range to the west (Smith and
Ketner, 1976) and is encountered in exploratory oil wells ~16 km northwest of the Federal No.
16-5 well (Reese, 1986; Schalla, 1992), appears to be absent in the Federal No. 16-5 well.
Tertiary rocks encountered in the Federal No. 16-5 well unconformably overlie Paleozoic cherty
limestones that, on the basis of palynological ages, have been assigned to the Pennsylvanian Ely
Formation (TH Geological Services, 1994).

The Aspen Unit No. 1 well was spudded at the western edge of Huntington Valley ~7.5
km east of the southern Piñon Range (Figs. 7 and 8) and drilled to a total depth of 3,743 m (Hess,
2004). The stratigraphic tops chosen for the Aspen Unit No. 1 well in the present study (Table 2) were determined based on lithologic and geophysical logs and are similar to the original stratigraphic picks reported by the operator (Hess, 2004). Mississippian clastic rocks corresponding to the Diamond Peak and Chainman Formations have been structurally duplicated in the Aspen Unit No. 1 well and are present both structurally above and below Silurian(?)-Devonian carbonate strata. Based on this duplication of stratigraphic units, the Aspen well is interpreted to have intersected a thrust fault that placed Silurian(?)-Devonian carbonate rocks structurally above siliclastic Mississippian strata (Fig. 8). The subsurface orientation and transport direction of this inferred thrust fault are uncertain. However, east-vergent thrusts and reverse faults have been documented in the Piñon Range immediately west of cross section C to C’ (Carpenter et al., 1993), as well as in the southern Ruby Mountains to the east (e.g., Willden and Kistler, 1979). Therefore, the inferred fault intersected by the Aspen Unit No. 1 well is tentatively interpreted in cross section C to C’ to be an east-vergent thrust, although other interpretations are possible.

Published depth-migrated seismic reflection profiles across Huntington Valley west of Harrison Pass (Fig. 7) show that the general structure of the Huntington Valley basin is a half-graben bounded on its eastern margin by the gently west-dipping Huntington Valley fault (Fig. 8; Satarugsa and Johnson, 2000). Interpretations of the subsurface geometry of the Huntington Valley fault based on these seismic profiles suggest that it maintains a relatively constant 20°-26° west dip from near the surface to a depth of ~9 km (Satarugsa and Johnson, 2000).

Southern Ruby Mountains
West-transported extensional klippen crop out discontinuously along the western side of the southern and central Ruby Mountains and are underlain by gently west-dipping normal faults (e.g., Snoke and Lush, 1984). The largest of these klippen is the Mitchell Creek klippe, which preserves Mississippian clastic rocks in the hanging wall of the Mitchell Creek fault that rest on unmetamorphosed Upper Cambrian carbonate rocks in the footwall of the fault (Fig. 7a,b; Sharp, 1942; Willden and Kistler, 1979; Burton, 1997). The Mitchell Creek klippe crops out ~1.5 km south of section C to C’, and the Mitchell Creek fault is projected north into the cross section (Fig. 8b). A structure contour map (Fig. 7b) of the Mitchell Creek fault, which is based on the intersection of the fault trace with topography, indicates that the fault dips between 12° and 17° west and suggests that the fault has a relatively planar three-dimensional form beneath the klippe. Willden and Kistler (1979) also mapped several additional extensional klippen immediately east of the Mitchell Creek klippe that preserve Devonian carbonate rocks in fault contact with underlying Cambrian and Ordovician limestones (Fig. 7b).

The Rattlesnake Mountain antiform (Fig. 7) is a doubly plunging antiform with fold axes oriented 28, 095° and 09, 170° (Burton, 1997). The south-trending hinge of the antiform is overturned and east-vergent, whereas the east-trending hinge of the antiform is an upright, open fold (Willden and Kistler, 1979; Burton, 1997). A west-dipping thrust fault (the Rattlesnake Mountain thrust) that places Middle Cambrian rocks in its hanging wall on Upper Cambrian and Lower Ordovician rocks in its footwall is present along strike to the south of the antiform (Fig. 7a; Willden and Kistler, 1979; Coats, 1987). Northward projection of the trace of this thrust fault suggests that it cores, and probably is responsible for generating, the south-trending folding of the Rattlesnake Mountain antiform.
The Rattlesnake Mountain antiform is interpreted to have a polyphase history of folding. Burton (1997) interpreted the east-trending hinge of the Rattlesnake Mountain antiform to have formed during the emplacement of the Harrison Pass pluton, based on its orientation parallel to metamorphic foliation in the contact metamorphosed aureole of the Harrison Pass pluton. Conversely, Burton (1997) interpreted the south-trending hinge of the antiform, which parallels the axes of contractional folds further south in the Ruby Mountains (Willden and Kistler, 1967), to have been formed by contractional deformation that occurred prior to intrusion of the Harrison Pass pluton. Additionally, mapping by Willden and Kistler (1979) shows that the Rattlesnake Mountain thrust is cut by a dike related to the Harrison Pass pluton, implying that the thrust fault and related folding are older than the pluton.

The three-dimensional subsurface geometry of the Harrison Pass pluton is poorly constrained. Surface geologic maps show that dikes and apophyses of Harrison Pass granodiorite and monzogranite intrude Paleozoic sedimentary rocks up to 9 km south of the primary exposure of the Harrison Pass pluton (Willden and Kistler, 1979; Howard et al., 1979; Burton, 1997), suggesting that the pluton extends southward beneath these sedimentary rocks. Additionally, geologic interpretations of the crustal velocity structure of the Ruby Mountains based on vertical-incidence to wide-angle multicomponent seismic data collected along the eastern flank of the range suggest that the Harrison Pass pluton may underlie the southern Ruby Mountains at upper crustal levels as far south as Overland Pass (Satarugsa and Johnson, 1998). The southern contact of the Harrison Pass pluton and the Paleozoic country rocks dips between 25° and 35° to the south (Burton, 1997). Projection of this contact southward indicates that the pluton may underlie the sedimentary section beneath cross section C to C’ at a depth of 3 to 5 km.
Southern Ruby Valley/Northern Maverick Springs Range

The eastern portion of cross section C to C’ traverses the southern end of Ruby Valley and the northeastern portion of the Maverick Springs Range. The northeastern Maverick Springs Range was mapped in detail as part of this study (see Appendix 1) and consists of gently to moderately west-dipping Permian carbonate rocks that have been deformed by a series of north-south striking, east-dipping high-angle normal faults (Fig. 7a). No evidence was observed in the field indicating that the northeastern Maverick Springs Range is bounded by a west-dipping fault on its western side, and southern Ruby Valley is interpreted to be a half-graben controlled by the east-dipping Ruby Valley fault that bounds the eastern side of the southern Ruby Mountains. Where the Ruby Valley fault has been imaged by seismic data in Northern Ruby Valley, it dips >60° to the east (Sataruga and Johnson, 2000).

Subsurface Structure of Huntington Valley

Huntington Valley is interpreted to be a half-graben formed by a series of west-dipping normal faults where the valley is traversed by cross section C to C’ (Fig. 8a,b). Although Cenozoic strata are only ~1 km thick in the Federal No. 16-5, the basin is inferred to be deeper further west. Borehole and seismic reflection data show that Tertiary strata are nearly 3 km thick below central Huntington Valley east of Cedar Ridge (Reese, 1986; Schalla, 1992; Sataruga and Johnson, 2000), and Sataruga and Johnson (2000) interpreted the Tertiary-Paleozoic unconformity to be present at a depth of ~2.6 km in their southernmost cross-line seismic reflection profile across (CT 11) that traverses Huntington Valley ~5 km north of section C to C’ (Fig. 7a). Where seismic and borehole data are available in Huntington Valley, the portions of the basin containing the thickest sequences of Tertiary strata generally correspond with a north-
south trending negative Bouguer gravity anomaly (Ponce et al., 1996). This anomaly extends into southern Huntington Valley, suggesting that the thickness of Cenozoic strata is similar in both the central and southern portions of the basin (Ponce et al., 1996). Moreover, a reflector on the CT 19 long-line seismic reflection profile (Fig. 7a) interpreted to be the Paleozoic-Cenozoic unconformity appears to become only slightly shallower (~1.7 s to ~1.5 s two way travel time) from north to south along the profile.

Additionally, constraints based on the restored pre-extensional geology of cross section C to C’ suggest that buried west-dipping normal faults may be present within Huntington Valley. Because Mississippian clastic rocks were encountered at the bottom of the Aspen Unit No. 1 well at a depth of 3,743 m (Table 2), this provides a minimum constraint on the depth of the paleo-regional stratigraphic elevation of Mississippian strata prior to Mesozoic structural thickening and subsequent Tertiary extensional faulting in the western Piñon Range. Restoring the offsets of only the Huntington Valley and Mitchell Creek faults in section C to C’ is insufficient to return the Mississippian strata at the bottom of the Aspen Unit No. 1 well to a similar regional stratigraphic elevation as the Mississippian strata underlying the northern Maverick Springs Range east of the southern Ruby Mountains, with the result that marker beds on the eastern side of the restored cross section are uplifted several kilometers above marker beds at similar stratigraphic levels on the western side of the restored cross section. This leads to two alternative end-member possibilities: 1) either additional west-dipping normal faults are present beneath Huntington Valley that further down-drop rocks in the Piñon Range relative to rocks at the same stratigraphic levels in the southern Ruby Mountains and northern Maverick Springs Range, or 2) strata presently cropping out in the southern Ruby Mountains and northern Maverick Springs Range were uplifted several kilometers above regional stratigraphic elevations prior to the onset
of extension. Although both scenarios are possible, the former scenario is preferred here because there are no thrust faults exposed in the ranges east of the southern Ruby Mountains that could have produced substantial amounts of stratigraphic relief within the southern Ruby Mountains or northern Maverick Springs Range (Hose and Blake, 1976; Coats, 1987; Howard 2003), so. Nonetheless, the possible presence of tectonic wedges below the southern Ruby Mountains, which have been suggested by several workers as a potential mechanism of Mesozoic structural thickening of the Paleozoic section above the high grade metamorphic rocks exposed further north in the Ruby Mountains (e.g., Snoke and Miller, 1988; Howard, 2003), would allow for strata in the southern Ruby Mountains to have been uplifted above regional stratigraphic elevations without requiring the surface exposure of large-scale upper crustal thrusts in the Sevier hinterland to the east.

Taken together, the above constraints suggest that buried, west-dipping normal faults that collectively accommodate several kilometers of extension are present beneath Huntington Valley. These faults have been represented in cross section C to C’ by a major west-dipping fault located to the west of the Federal No. 16-5 well that increases the depth of the basin and by two additional west-dipping faults within the deepest portion of the basin that approximately correspond to the location of the negative Bouguer gravity anomaly (Ponce et al., 1996). These inferred west-dipping normal faults are shown as dashed lines in Figure 8.

Crosscutting Fault Relationships

The Mitchell Creek fault, including potentially associated splays represented by the small extensional klippen that occur in its footwall, is interpreted to be the oldest extensional fault in cross section C to C’, and this fault is inferred to be cut and offset by younger normal faults
within Huntington Valley, including the Huntington Valley fault (Fig. 8). The strongest evidence for this crosscutting relationship is based on the Pennsylvanian carbonate rocks encountered at the bottom of the Federal No. 16-5 well. The Mitchell Creek fault carries Mississippian clastic strata in its hanging wall, whereas Middle and Upper Cambrian rocks in its footwall crop out in the southern Ruby Mountains to the west of the Federal No. 16-5 well. The simplest explanation for the presence of Pennsylvanian rocks at the bottom of the Federal No. 16-5 well is that the Huntington Valley fault cuts and offsets the Mitchell Creek fault, down-dropping upper Paleozoic rocks in the hanging wall of the Mitchell Creek fault against Cambrian and Precambrian strata in its footwall (Fig. 8b,c). Additionally, where the Mitchell Creek fault is exposed, it dips at a shallower angle than the Huntington Valley fault, implying that the Mitchell Creek fault experienced greater amounts of tilting during Tertiary extension than the Huntington Valley fault and is therefore an older structure.

No crosscutting relationships directly constrain the relative age of the Ruby Valley fault, which dips at a high angle (Satarugsa and Johnson, 2000) and appears to have experienced relatively little tilting compared to the Huntington Valley and Mitchell Creek faults, implying that it may have initiated later than both of these structures. In any case, both the Huntington Valley and Ruby Valley faults exhibit evidence of latest Quaternary (<130 ka) rupture and have probably been slipping broadly concurrently at least since then (dePolo, 2008).

Nature of the Mitchell Creek Fault System

As mentioned above, Willden and Kister (1979) mapped several small extensional klippen just to the east of the Mitchell Creek klippe, and these klippen carry Devonian carbonate rocks in their hanging walls (Fig. 7). If the faults underlying these subsidiary klippen are the
eastward continuation of the Mitchell Creek fault, then they may simply represent deeply eroded remnants of a formerly more extensive Mitchell Creek kippe. A structure contour map of the Mitchell Creek fault, however, suggests that this fault projects above these outlying klippen and the faults that underlie them (Fig. 7b). If this is the case, then (1) the small klippen may represent an earlier fault (or set of faults) that is cut and offset by the Mitchell Creek fault, or (2) the small klippen and the Mitchell Creek fault both may be part of a system of closely-spaced normal faults that slipped broadly concurrently and collectively accommodated large amounts of west-directed normal sense offset. In any case, the Mitchell Creek fault is shown as a single, large-offset normal fault at the scale of the reconstructed cross section (Fig. 8), but either of the above alternative scenarios might decrease the amount of slip inferred on the Mitchell Creek fault *sensu stricto*.

*Fault Block Tilting in Cross Section C to C’*

Several lines of evidence suggest that the southern and central Ruby Mountains are east-tilted between 30° and 40° as a result of block tilting about a roughly north-south horizontal axis during Cenozoic extension. Paleozoic strata in the southern Ruby Mountains dip dominantly to the east (Sharp, 1942; Willden and Kistler, 1979), and tabular monzogranite sills within the Harrison Pass pluton dip 28° to 38° east-southeast, which is consistent with 28° to 38° of post-36 Ma eastward tilting of the Ruby Mountains, assuming that these tabular monzogranites were originally emplaced with subhorizontal dips (Burton, 1997). Biotite K–Ar dates young westwards across the Harrison Pass pluton from 36 to 20 Ma, which suggests that the deepest structural levels of the pluton are present on the western side of the range (Kistler et al., 1981; Burton 1997). Synformal reflectors from the Miocene Humboldt Formation and younger
sedimentary rocks near the Ruby Mountains in Ruby Valley have also been cited as possible evidence of eastward rotation of the southern and central Ruby Mountains (Satarugsa and Johnson, 2000). Additionally, the moderate- to low-angle west dips of the Mitchell Creek and Huntington Valley faults require that eastward tilting of the range occurred if these faults initiated at higher angles.

Although there is considerable evidence that the southern Ruby Mountains have been tilted eastward during Tertiary extension, the Piñon Range to the west of the southern Ruby Mountains and the northern Maverick Springs Range to the east do not appear to have been affected by similar amounts of eastward tilting. Colgan and Henry (2009) concluded that the Piñon Range most likely was not significantly tilted in the Tertiary based on shallow (5°) eastward dips of Eocene volcanic rocks, the absence of gently dipping normal faults, and the upright folding of Paleozoic rocks within the range. Additionally, restored cross sections of the central Piñon Range (see above) also suggest that that net amount of block tilting associated with high-angle extensional faulting in the Piñon Range has been minimal. Permian carbonate rocks in the northern Maverick Springs Range generally dip between 15° and 30° to the west and may have experienced moderate amounts of west-directed block tilting by east dipping, north trending high-angle normal faults (Fig. 7a). However, unless these Permian carbonate rocks were steeply west-tilted prior to the onset of Tertiary extension, it appears unlikely that they have been subject to significant eastward block tilting.

Based on the above observations, the southern Ruby Mountains appears to be a moderately east-tilted horst domain bordered to the east and west by untilted to gently west tilted domains. The differential tilting of the southern Ruby Mountains relative to these adjacent domains must be accommodated by some form of upper-crustal deformation, such as by large-
scale antiformal folding between the southern Ruby Mountains and the Piñon Range and synformal folding between the southern Ruby Mountains and the northern Maverick Springs range (Fig. 8). In fact, the west dips of Permian carbonate rocks in the northern Maverick Springs Range require that the Paleozoic rocks beneath Ruby Valley are folded into a syncline, regardless of whether or not it is of a Tertiary extensional origin. Eastward tilting of the southern Ruby Mountains was most likely synchronous with movement on the Mitchell Creek and Huntington Valley faults, and thus folding required to accommodate differential tilting of the southern Ruby Mountains and adjacent areas must also be synchronous with movement on these faults. The high-angle dip of the Ruby Valley fault suggests that it is responsible for relatively modest (if any) footwall tilting within the southern Ruby Mountains. However, slip on the Ruby Valley fault would produce westward tilting that and would decrease the net amount of eastward tilting generated by offset on the Mitchell Creek and Huntington Valley faults.

Thus, eastward footwall tilting of the southern Ruby Mountains during active extension on the west-dipping Mitchell Creek and Huntington Valley faults is interpreted to have been accommodated by synformal folding in the footwalls of these fault systems, and antiformal folding in their hanging walls. Similar synformal-footwall and antiformal-hanging wall folding is predicted by rolling-hinge models of extension (e.g., Buck, 1988; Wernicke and Axen, 1988), as well as by the structural cantilever model of extension (e.g., Kusznir and Ziegler, 1992). In both cases, isostatic uplift is thought to be the driving force behind fault block tilting and associated hanging wall and folding. It is worth noting that, although antiformal hanging-wall folding in normal fault systems is considered to be diagnostic of a listric normal fault geometry (e.g., Hamblin, 1965; Xiao and Suppe, 1992), antiforms also can develop in the hanging walls of
planar normal faults during isostatic uplift (e.g., Kusznir and Ziegler, 1992; Roberts and Yielding, 1994).

**Restored Structure of Cross Section C to C’**

The restoration of extensional faulting in cross section C to C’ is presented in two stages (Fig. 8c, d). The first stage (Fig. 8c) shows the interpreted structural geometry of the central Ruby Mountains and eastern Huntington Valley area prior to the initial slip on the system of west-dipping faults that form the present-day Huntington Valley half-graben. The final stage (Fig. 8d) presents a line-balanced restoration of all extensional faults in section C to C’. Folding related to eastward tilting of the southern Ruby Mountains is also restored in this final stage. Restoration of this folding using area- and line-balancing techniques results in loose lines at both ends of the cross section that are slanted at a ~65° angle, the significance of which is addressed in a later section (see Discussion). The total amount of Tertiary extensional strain accommodated by normal faulting across section C to C’ is estimated to be ~22.1 km or ~94% extension. The faults that accommodate the largest amounts of extensional strain are the Huntington Valley and the Mitchell Creek faults, which are, respectively, estimated to have accumulated ~8.9 and ~8.2 km of slip, given the simplifications or assumptions discussed above.

As shown in Figure 8d, the southern Ruby Mountains are interpreted to have been deformed by east-vergent folds and thrust faults prior to Tertiary extension. The dominant structure in the restored cross section is a large fault-propagation fold cored by the Rattlesnake Mountain thrust. The age of thrust faulting and folding is directly constrained only as post-Mississippian in the southern Ruby Mountains. However, Triassic rocks are present in the cores of north-south trending synclines in the Maverick Springs Range and southern Pequop
Mountains to the east of the Ruby Mountains (Hose and Blake, 1976; Coats, 1987), and several episodes of Mesozoic contractional deformation and metamorphism are well documented in the central and northern Ruby Mountains (e.g., Snoke and Lush, 1984; Hudec, 1992; Hodges et al., 1992; Camilleri and Chamberlain, 1997; McGrew et al., 2000). Thus, regional relationships suggest that major folding and thrusting in the southern Ruby Mountains most likely occurred during the Mesozoic. The Harrison Pass pluton is restored to a depth of ~10.6 km, which is consistent with paleodepth estimates of the emplacement of the pluton (Burton, 1997; Barnes et al. 2001).

In the final reconstructed panel (Fig. 8d), the lower to middle(?) Eocene lacustrine sedimentary and volcaniclastic rocks encountered in Federal No. 16-5 are shown pinching out eastward, implying that these rocks were deposited in an erosional paleovalley, as opposed to an extensional basin. Paleovalleys filled with middle Eocene lacustrine sedimentary rocks and volcanic tuffs have been documented north of the Ruby Mountains (Henry, 2008), and similar middle Eocene paleovalleys may have been present further south as well. Nonetheless, it is equally possible that these rocks were deposited in an extensional basin given the present uncertainties in the age of Tertiary extension within the southern Ruby Mountains (of which there may have been multiple phases) and in the distribution of lower to middle Eocene aged rocks within Huntington Valley. If these lower to middle(?) Eocene rocks were deposited in an extensional basin, the magnitude of this extension need not have been large in order to accommodate the moderate thickness (~600 m) of lower to middle(?) Eocene rocks encountered in Federal No. 16-5.

Age of Extension in the Southern Ruby Mountains
If the Eocene volcaniclastic rocks and lacustrine strata encountered in Federal No. 16-5 were deposited in an extensional basin, then the earliest phase of extensional faulting within the southern Ruby Mountains may have occurred in the early to middle Eocene. This early episode of extension might have been part of a more widespread phase of early to middle Eocene extension that is interpreted to have formed the Eocene Elko basin, which may have extended as far south as the southern Ruby Mountains in middle Eocene time (Solomon, 1992; Satarugsa and Johnson, 2000; Haynes, 2003). Alternatively, extension may not have begun in the southern Ruby Mountains until considerably later in Tertiary time if the Eocene volcaniclastic and lacustrine rocks were deposited in a paleovalley (Henry, 2008; Colgan and Henry, 2009).

Major tectonic exhumation of the southern Ruby Mountains appears to postdate the intrusion of the 36 Ma Harrison Pass pluton. Paleodepth estimates for the roof of the pluton range from 6 to 11 km (Burton, 1997; Barnes et al., 2001; Howard, 2003). Thus, a minimum of 6 to 11 km of tectonic denudation must have occurred within the central/southern Ruby Mountains since the emplacement Harrison Pass pluton. Additionally, the Harrison Pass pluton is cut by a major gently west-dipping (~5° to 10°) cataclastic normal fault underlying the Cedar Mountain klippe (Fig. 7a; Willden et al., 1967; Snoke and Lush, 1984; Burton, 1997) that is likely part of the same system of low-angle normal faults as the Mitchell Creek fault, suggesting that these early, gently-dipping faults postdate the emplacement of the pluton.

Muscovite and biotite $^{40}\text{Ar}/^{39}\text{Ar}$ and K-Ar cooling ages young westwards from 36 to 25 Ma across the Harrison Pass pluton and from 36 to 20 Ma across the entire Ruby Mountains-East Humboldt Range core complex (Kistler et al., 1981; Dallmeyer et al., 1986; McGrew and Snee, 1994), and these dates have been interpreted to record the age of slip on a west-dipping normal fault during that time (Dallmeyer et al., 1986; Dokka et al., 1986; McGrew and Snee, 1994).
However, no sedimentary or volcanic rocks that date between 31 and 16 Ma have been encountered either in outcrop or in drill holes in the vicinity of the southern Ruby Mountains; therefore, there is no evidence for the development of an Oligocene hanging-wall sedimentary basin during this inferred middle Tertiary period of west-directed extensional faulting (Wallace et al., 2008; Colgan and Henry, 2009). Additionally, no significant angular unconformity is present beneath the Humboldt Formation (16 to 9 Ma) where it rests on the Indian Well Formation (38 to 33 Ma) on the eastern side of the Piñon Range, suggesting that the 36 to 25 Ma mica $^{40}$Ar/$^{39}$Ar and K-Ar cooling ages may record something other than the age of slip, such as partial argon loss at elevated temperatures (Colgan and Henry, 2009). Recent apatite fission track and (U-Th)/He ages from the Harrison Pass pluton range from 18 to 10 Ma and yield weighted-mean ages of $14.6 \pm 1.1$ Ma and $14.8 \pm 1.5$ Ma, respectively (Colgan and Metcalf, 2006). These apatite fission track and (U-Th)/He ages do not exhibit an east to west change across the pluton and imply that the southern Ruby Mountains experienced a period of rapid exhumation during the middle Miocene (Colgan and Metcalf, 2006; Sullivan and Snoke, 2007; Colgan and Henry, 2009). Additionally, thick sequences of middle Miocene coarse clastic sedimentary rocks were deposited in Huntington and Ruby Valleys beginning $\sim$16 Ma, consistent with a major period of middle Miocene extension in the southern Ruby Mountains (Colgan and Henry, 2009).

Given the above constraints, extensional faulting could have initiated within the southern Ruby Mountains as early as the early to middle Eocene. Mica $^{40}$Ar/$^{39}$Ar and K-Ar cooling ages from the Harrison Pass pluton suggest a period of west-northwest directed Oligocene exhumation of the southern Ruby Mountains (e.g., McGrew and Snee, 1994). However, the apparent absence of a large sedimentary basin of Oligocene age west of the core complex is anomalous (Wallace et al., 2008; Colgan and Henry, 2009), and whether these cooling ages correspond to a period of
major extensional faulting is uncertain. A more definite period of major extensional faulting occurred during the middle Miocene and may have been contemporaneous with a period of major regional extension between 17 to 10 Ma (Colgan and Henry, 2009). Additionally, Quaternary fault scarps are present along the eastern and western flanks of the southern Ruby Mountains, indicating that active extension continues within this area (Sharp, 1939; Dohrenwend et al., 1991; dePolo, 2008).

Medicine Range

The Medicine Range is located ~30 km east of the southern Ruby Mountains (Fig. 1) and has been relatively little studied compared to the ranges west of the Ruby Mountains. Rocks exposed in the Medicine Range dominantly consist of east-dipping Permian carbonate rocks, although Lower Triassic rocks also crop out in the center of the range (Fig. 9; Collinson, 1966; Collinson, 1968). Restricted outcrops of Tertiary volcanic rocks of probable Eocene age are present in the eastern portions of the range (Collinson, 1966; Coats, 1987). However, most of the Tertiary rocks mapped within the Medicine Range area consist of fanglomerate and tuffaceous sedimentary rocks that most likely correlate with the middle Miocene Humboldt Formation west of the Ruby Mountains (Collinson, 1966; Coats, 1987). Areally restricted outcrops of Cretaceous(?) granitic intrusive rocks are also present in the western portion of the Medicine Range (Coats, 1987), and geologic interpretations of aeromagnetic and magnetotelluric data suggest that the southern and western portions of the range are underlain by one or more plutonic bodies at relatively shallow depth (Hildenbrand and Kucks, 1988; Grauch, 1996; Wannamaker and Doerner, 2002). Lead-silver occurrences of the Mud Springs mining district occur in the northern and western parts of the range (LaPointe et al., 1991).
Cross Section M to M’

To illustrate the Cenozoic structural evolution of the Medicine Range, a cross-sectional restoration of extensional faulting was performed for a line of section that crosses the central Medicine Range (M to M’ in Fig. 9). Paleozoic, Mesozoic, and Tertiary rocks within the Medicine Range are deformed by a series of closely spaced, north-south to northeast-southwest striking, dominantly west-dipping, high-angle normal faults (Collinson, 1966). No direct measurements of fault dips were available, and faults in cross section M to M’ are assumed to dip ~50° west, which is consistent with their traces with respect to topography. In general, the attitudes of bedding are similar in the hanging wall and footwall of faults, implying that these faults have dominantly planar or curviplanar subsurface geometries. Thickness estimates of Paleozoic formations are based on measurements reported by Collinson (1966) and compiled regional stratigraphic data (Fig. 2, Appendix 2).

Present-Day Cross Section

The Medicine Range is interpreted to have a relatively simple structure, consisting of a series of “domino style” east-tilted fault blocks bounded by numerous, closely-spaced, west dipping high-angle faults, as shown in Figure 10a. Maximum measured offset on individual faults is ~1.3 km; however, the average estimated individual fault offset is ~0.4 km. A Cretaceous(?) intrusion is interpreted to be present at shallow depths on the western end of cross section M to M’, consistent with the presence of several surface outcrops of Cretaceous(?) granitic rocks and a positive magnetic anomaly beneath the southwestern Medicine Range (Coats, 1987; Hildenbrand
and Kucks, 1988; Grauch, 1996). However, neither the subsurface geometry nor the age of plutonic rocks beneath the Medicine Range is well defined.

Seismic reflection profiles across Ruby Valley ~15 km north of section M to M’ show that northern Ruby Valley is a graben bounded by the high-angle, east-dipping Ruby Valley fault on its western side, and a moderately west-dipping (~38°) normal fault on its eastern side near Delcer Buttes (Satarugsa and Johnson, 2000). This fault may curve around the west side of the Medicine Range (Satarugsa and Johnson, 2000) or may tip out along strike north of the Medicine Range. The latter alternative is favored here because the southern portion of Ruby Valley narrows considerably to south of Delcer Buttes (Fig.1), suggesting that it is less extended than northern Ruby Valley. Moreover, field studies also have not provided evidence that the Medicine Range is bounded by a major west-dipping fault on its western side (Fig. 1; Collinson, 1966; Coats, 1987). An intriguing possibility is that the numerous, closely spaced, small-offset faults distributed throughout the Medicine Range represent the along-strike expression of the large-offset, west-dipping fault that bounds the eastern side of northern Ruby Valley.

**Fault Block Tilting**

The Medicine Range appears to have experienced east-directed fault block tilting during Tertiary extension. Where bedding attitudes have been measured on Tertiary sedimentary rocks, they generally dip between 15° and 25° to the east. Additionally, the Permian and Triassic strata that crop out in the range display dominant eastward dips of between 15° and 45°. Thus, although these rocks had been deformed by folding prior to Tertiary extension, their dips are consistent with east-directed fault block tilting (Collinson, 1966). East-directed fault block tilting
is also consistent with the westerly dips of normal faults in cross section M to M’ (Fig. 10a), and
~20° of tilting have been applied to the reconstructed cross section (Fig. 10c).

Age of Extension and Restored Structure

The primary evidence for the age of faulting in the Medicine Range is based upon the
relationships between faults and Tertiary sedimentary and volcanic rocks, which have not been
well dated or mapped in detail in the Medicine Range. Thus, the age of extension can only be
tentatively constrained pending additional data on the age and distribution of Tertiary sediments
and volcanic rocks within the Medicine Range/Maverick Springs Range area. The current
interpretations of the ages of Tertiary units suggest that much of the extensional faulting within
the Medicine Range occurred during the Miocene. Coarse fanglomerate deposits that are
presumably Miocene are present in the immediate hanging walls of high-angle normal faults
(Collinson, 1966) and are likely synextensional deposits. Strata of probable Miocene age dip as
much at 25° to the east and in places are cut by west-dipping normal faults (Collinson, 1966;
Coats, 1987). Nonetheless, Tertiary strata are also mapped covering normal faults, and may
indicate that multiple episodes of faulting occurred during Tertiary extension. Normal faulting
has continued in the Medicine Range area into Quaternary time, as evidenced by <130,000 ka
fault scarps offsetting alluvial deposits in Butte Valley southeast of the Medicine Range and a
<1.8 Ma fault along the southwestern edge of the range (Collinson, 1966; Dohrenwend et al.,
1991; dePolo, 2008).

Restoration of cross section M to M’ shows that normal faulting has accommodated ~3.2
km of extensional strain or 20% extension across the cross section (Fig. 10b, c). Thin wedges of
Tertiary sedimentary rocks in the hanging walls of normal faults in the western portion of section
M to M’ are presumed to be synextensional deposits that filled half grabens as extension progressed. Paleozoic and Mesozoic rocks in the restored cross section are folded into a broad, open anticline, and the Medicine Range appears to have been relatively deformed during Mesozoic compression.

Spruce Mountain

The Spruce Mountain area (Fig. 1, 11a), exposes a dominantly east-dipping section of Ordovician to Permian miogeoclinal strata in numerous fault-bounded blocks. All of the exposed faults, including those that dip at low angles, are normal faults, as they place younger rocks on older rocks. Limited outcrops of Tertiary volcanic and sedimentary rocks unconformably overlie Paleozoic rocks at the Spruce Mountain and generally display gentle to moderate east-northeast dips (Harlow, 1956; Hope, 1972). Although no Mesozoic rocks crop out at Spruce Mountain proper, a considerable thickness (~1.0-1.2 km) of Lower Triassic marine shale and limestone is exposed within the core of the Pequop syncline ~10 km west of Spruce Mountain (Fraser et al., 1986; Swenson, 1991). Surface exposures of intrusive rocks in the Spruce Mountain area are limited to a stock of hornblende diorite that crops out north of Spruce Mountain Ridge and a granite porphyry dike that cuts across the northern side of Spruce Mountain (Hope, 1972). Skarn and carbonate replacement deposits are spatially associated with the granite porphyry dike and have been historically mined for lead and silver, with subordinate zinc, copper, and gold. Additionally, a resource of ~80 Mt of low-grade porphyry copper-molybdenum mineralized rock is estimated to be present in the southwestern part of the Spruce Mountain area (LaPointe et al., 1991). There are no published radiometric dates on the deposits or associated dikes, but similar compositions of igneous rocks and types of ore deposits in Elko and White Pine Counties.
generally are late Eocene or Oligocene in age (Sheet 1 of Stewart and Carlson, 1976; Seedorff, 1991a), for which the Hunter district is but one example (~36 Ma, Gans, 1982; Gans and Miller, 1983; Gans et al., 1989).

**Plan-View Analysis of Faulting at Spruce Mountain**

Given the complexity of faulting within the Spruce Mountain area, a plan-view analysis of extensional faulting was performed to constrain the relative timing of crosscutting faults and to determine the present-day orientations of low-angle faults (Fig. 11). The primary resource for this analysis was the surface geologic map of the Spruce Mountain Quadrangle of Hope (1972). Additional data were also incorporated from geologic maps of the southern Pequop Range by Fraser et al., (1986) and Swenson (1991), previous geologic mapping of the Spruce Mountain area by Harlow (1956), the Elko County geologic map (Coats, 1987), and field checks of key locations by the author. Names of faults follow previous workers where possible; certain previously unnamed faults have been given names here for ease of reference.

**Crosscutting Fault Relationships**

Numerous crosscutting high- and low-angle normal faults have accommodated extensional strain at Spruce Mountain. These crosscutting faults are grouped into six sets of similar relative age and orientation, numbered 1 through 6, from oldest to youngest (Fig. 11a). In some cases, particularly in the southeastern portion of the study area where numerous faults intersect, the crosscutting relationships among faults are uncertain. In these cases, faults have still been assigned to a fault set in Figure 11a, but the fault traces are dashed and queried to reflect this uncertainty.
The two youngest fault sets (5 and 6 in Fig. 11a) consist of a set of southwest to southeast striking, northwest to southwest dipping, high- to moderate-angle normal faults on the western side of the study area and a set of northwest and northeast striking, northeast and southeast dipping high-angle normal faults of small offset in the easternmost part of the study area. No faults from either of these youngest sets intersect within the study area, and thus their relative ages cannot be established. However, both of these fault sets cut faults grouped into fault set 4. Fault set 4 consists of generally north striking, east dipping, moderate- to high-angle normal faults that crop out throughout the study area. Fault set 3 consists of a set of nearly east-west striking, north dipping high-angle faults that traverse the central portion of the study area. Fault set 2 consists of generally south to southwest striking, west to northwest dipping, high- to moderate-angle normal faults that crop out throughout the study area. Certain west-dipping faults grouped within fault set 2 may in fact crosscut one another, especially in the southeastern part of the study area where numerous faults intersect and crosscutting relationships are difficult to determine. Because of this complexity and because these faults have similar orientations and display similar crosscutting relationships to relative faults from other fault sets, all south to southwest striking, west to northwest dipping high- to moderate-angle normal faults mapped within the eastern part of the study area that are crosscut by faults belonging to sets 3 and 4 have been assigned to fault set 2. The oldest fault set includes all low-angle normal faults that are present within the study area. Where they crop out, these low-angle faults are always cut by high- to moderate-angle faults. However, like faults grouped in fault set 2, the individual low-angle faults within fault set 1 appear to crosscut one another in some cases. Collectively, however, these low-angle faults are the oldest extensional structures that crop out at Spruce Mountain and have, therefore, been included within the same fault set.
Structure Contour Maps of Low-Angle Faults

Three major low-angle faults (the North Peak, South Peak, and Spruce Spring faults) crop out on the peak of Spruce Mountain (Fig. 11a). Structure contour maps of these faults were generated where they intersect topography to better constrain their three-dimensional geometries (Fig. 11b). All three low-angle faults appear to have relatively planar geometries where they crop out at the surface. Dips of faults calculated from these structure contour maps show that the Spruce Spring fault dips 15 - 18° to the southwest, the South Peak fault dips south at ~20°, and the North Peak fault dips northwest 13- 17°. In addition to these major low-angle faults, a subsidiary low-angle fault was mapped by Hope (1972) structurally above the North Peak fault on the North Peak of Spruce Mountain that juxtaposed the Ely Limestone with the Diamond Peak Formation/Chainman Shale, undivided. However, field observations conducted by the author found no evidence of this subsidiary fault (See Appendix 1), and the contact between the Diamond Peak Formation and overlying Ely Limestone is here interpreted to be depositional.

The direction of slip on these low-angle faults has not been directly determined, but it is likely that the hanging walls of these faults have moved westward relative to the footwall. Strata in the upper and lower plates of these faults dip fairly uniformly eastward, and stratigraphic section is always missing between the hanging walls and footwalls of the faults, which place younger strata on older strata. A westward, normal-sense, hanging-wall transport direction can explain these observations without requiring large amounts of slip on these faults (Hope, 1972).

Mapped relationships among these faults indicate that the Spruce Spring fault lies structurally above and truncates the South Peak fault (Hope, 1972). The relationship between the North Peak fault and the Spruce Spring and South Peak faults is less certain. The North Peak fault may be
truncated by the northward projections of the Spruce Spring and South Peak faults (Hope, 1972). Alternatively, the South Peak fault and North Peak fault may be the along-strike continuation of the same fault if this fault is bowed over the top of Spruce Mountain.

Restored Cross Section D to D’

To illustrate the Cenozoic structural evolution of the Spruce Mountain area, a stepwise cross-sectional restoration of extensional faulting was performed for a line of section (D to D’ in Fig. 11a) that traverses the central portion of the Spruce Mountain area (Fig. 12). The primary marker beds employed for restoring fault offsets in this reconstruction are Ordovician through Permian miogeoclinal strata that are assumed to maintain relatively constant thicknesses. Formation thicknesses used in this reconstruction are based on measurements reported by Hope (1972).

Strikes and dips of bedding typically do not change substantially between the footwall and hanging wall blocks across faults, suggesting that these faults are planar or curviplanar in the subsurface. Few measurements of fault dips in the Spruce Mountain area have been reported, but fault dips can be constrained in some cases. Hope (1972) reports that the East fault and West fault (Fig. 11a) dip ~45° to the east and west, respectively, where they are intersected by section D to D’. Where section D to D’ intersects the surface exposure of North Peak fault, the fault projects into the cross section with a ~5° apparent northwestly dip.

It is uncertain how the Spruce Spring fault and South Peak fault may project into section D to D’. Hope (1972) interpreted both of these faults as steepening northward and truncating the North Peak fault south of section D to D’. Alternatively, as stated above, the South Peak fault could represent the along-strike continuation of the North Peak fault if the fault is bowed over the peak of Spruce Mountain. The Spruce Spring fault may project into cross section D to D’ at
high structural and stratigraphic levels. However, given the uncertainties in its three-dimensional geometry, the Spruce Spring fault was not projected into the restored cross section. Any additional low-angle faults projected into the reconstructed section, such as the Spruce Spring fault, would increase the reported estimate of extension across the Spruce Mountain area. Hence, the amount of extension estimated based on the restoration presented in Figure 12 may be closer to a minimum value.

Faults belonging to sets 5, 4, 2, and 1 project into section D to D’ (Fig. 11a), and these sets are restored in a stepwise fashion based on their relative ages from youngest to oldest (Fig. 12b-f). The cross section is restored as a series of rigid fault blocks, ignoring internal deformation (folding) within individual fault blocks. These assumptions can lead to space problems where fault blocks are differentially tilted. However, these space problems are small relative to the uncertainties of subsurface fault dips and are assumed to be accommodated by curvature of faults and/or internal deformation within fault blocks.

Present-Day Structural Interpretation

The interpreted present-day subsurface structure of the Spruce Mountain area along section D to D’ is presented in Figure 12a. A critical aspect of the subsurface structure in this cross section is the geometry of the low-angle North Peak fault, which is cut and down-dropped on either side of the peak of Spruce Mountain by the East and West faults. In Figure 12a, the North Peak fault is inferred to maintain a shallow west-dipping orientation in the subsurface and to root westward in the apparent hanging wall transport direction. The North Peak fault is interpreted to daylight in the eastern part of the cross section where it is cut by the small-offset Coyote fault and to project above the southern Pequop Range east of Spruce Mountain.
This interpretation differs substantially from the previous structural interpretation along the same line of section by Hope (1972), who depicted the North Peak fault as abruptly changing orientations from a subhorizontal dip on the peak of Spruce Mountain to dipping \( \sim 40^\circ \) E in the eastern portion of the cross section after being down-dropped by the East fault. Hope’s (1972) structural interpretation results in the North Peak fault projecting beneath the southern Pequop Range, where no surface exposures of major extensional faults are present (Fraser et al., 1986; Swenson, 1991), and, therefore, requires the entire southern Pequop Range to be underlain by the North Peak fault. Because the structural interpretation presented in Figure 12a does not require a substantial change in the orientation of the North Peak fault, nor does it necessitate that the southern Pequop Range be west translated as a detached block in the hanging wall of the North Peak fault, it is here preferred to the previous interpretation of Hope (1972).

**Evidence for Fault Block Tilting**

Several lines of evidence suggest that extensional faulting was accompanied by net eastward tilting: (1) Outcrops of Tertiary (Miocene?) sedimentary rocks of within the southern portion of the study area display generally eastward dips of between 17 and 25\(^\circ\) (2) Paleozoic strata within the Spruce Mountain area dominantly dip eastward, which is consistent with eastward extensional tilting if these strata were approximately upright prior to extension. That the Paleozoic strata in the Spruce Mountain area were approximately upright prior to extension may be a reasonable assumption if the pre-extensional structure in the Spruce Mountain area was characterized by broad open folding of a style similar to the adjacent Pequop syncline immediately to the east. (3) East-dipping Paleozoic strata in the footwalls and hanging walls of the normal faults that presently dip at low angles generally display moderate to high (\(>40^\circ\)) cut-
off angles with these faults where they are exposed at the surface, suggesting that these faults may have initiated at high to moderate angles. (4) The inferred westward hanging wall transport directions of the presently low-angle faults is consistent with net eastward tilting if these faults initiated as higher angle, west-dipping normal faults.

Given the above evidence for tilting of fault blocks during extensional faulting in the Spruce Mountain area, rotation consistent with the sense of slip of each restored fault set is applied to the stepwise cross-sectional restoration (Fig. 12c-f). However, the magnitude of tilting associated with each fault set is unknown. Dips of Tertiary (Miocene?) sedimentary rocks suggest that fault blocks have experienced a net eastward tilting of at least 25° during extension in the Spruce Mountain area, but this tilting could have been greater if extension and had begun prior to their deposition. Because cut-off angle relationships suggest that the currently low-angle faults in the Spruce Mountain area may have initiated at high to moderate angles, the final reconstructed panel (Fig. 12f) is shown tilted 42° west relative to the present day cross section, restoring the North Peak fault to an initial orientation of ~50° (Fig. 12b).

Restored Pre-Extensional Structure of the Spruce Mountain Area

The extensional restoration shows that cross section D to D’ has been extended by ~7 km or ~132% (Fig. 12). Although the total amount of extension across the section is large, it has been distributed among numerous faults, and the largest amount of slip on any one fault, the Banner Hill fault, is only 1.9 km.

The restored pre-extensional structural interpretation of cross section D to D’ is presented in Figure 12e. Miogeoclinal strata in the restored section are characterized by west-vergent folding and gentle westward dips (15-20°). The overall westward dip of beds in the
restored section is interpreted to reflect that the Spruce Mountain area was involved in large-scale open folding of a character similar to the Pequop syncline immediately east of the Spruce Mountain area. The Spruce Mountain area may have formed the western limb of an anticline that connected to the Pequop syncline prior to being dismembered by extensional faulting and may have connected to another upright syncline further west. The kilometer-scale, west-vergent anticline shown in the restored cross section is consistent with the scale and character of west-vergent folds and thrust faulting that occurred in the southern Pequop Mountains during the development of the Pequop syncline (Swenson, 1991) and may indicate that the Spruce Mountain area was similarly deformed during the same contractional deformational event.

Regional studies of Mesozoic deformation and metamorphism in the Pequop Mountains–Wood Hills–East Humboldt Range region to the north of Spruce Mountain and in the Ruby Mountains to the east have demonstrated that significant crustal thickening must have occurred in northeastern Nevada during the Mesozoic (e.g., Snoke and Miller, 1988; Camilleri and Chamberlain, 1997). Sheared fabrics locally present in limestones of the Pogonip Group on the southern side of Spruce Mountain have been interpreted to indicate that these rocks reached lower greenschist facies metamorphic conditions during the Mesozoic and may indicate that significant structural thickening occurred in the Spruce Mountain area during the Mesozoic (Camilleri and Chamberlain, 1997). No thrust faults have been mapped in the Spruce Mountain area, and the reconstructed cross section (Fig. 12) suggests that significant structural duplication of stratigraphic section has not occurred at Spruce Mountain between Ordovician and Lower Permian strata. Thus, if significant structural thickening and overthrusting did occur in the Spruce Mountain area during the Mesozoic, it likely occurred at higher structural and stratigraphic levels than are presently exposed at Spruce Mountain.
Age of extensional faulting

Normal faults in the Spruce Mountain area deform Paleozoic rocks and dismember older Mesozoic contractional structures and are most likely Tertiary structures, but the absolute age of extensional faulting is difficult to establish. Moreover, because there are six crosscutting sets of faults at Spruce Mountain, there may be multiple discrete extensional episodes. Granite porphyry dikes on the north peak of Spruce Mountain, which are undated but likely are of Eocene or Oligocene age, crosscut or intrude the North Peak fault, a member of the earliest set of faults; this indicates that normal faulting may have initiated here in the Eocene or Oligocene. Tertiary sedimentary rocks that crop out in the southern portion of the study area, which probably correlate with the middle Miocene Humboldt Formation (Harlow, 1956; Coats, 1987), dip eastward and appear to display fanning dips, suggesting that at least some extensional faulting within the Spruce Mountain area is Miocene in age. The ages of young faults have been determined, mostly on the basis of photogeologic evidence (e.g., Dohrenwend et al., 1991, 1996). As shown by dePolo (2008), several normal faults mapped on the eastern sides of Spruce Mountain Ridge and Spruce Mountain ruptured in the Quaternary and the latest slip on normal faults mapped along the western sides of Spruce Mountain Ridge and Spruce Mountain probably is even more recent (Latest Quaternary, i.e., <130,000 yr).

Regional Structural Development of Tertiary Extension

To summarize the regional structural development of Tertiary extensional faulting in the entire study area, a schematic restoration is presented of a cross section from Pine Valley to the Medicine Range, crossing the Ruby Mountains just north of Harrison Pass (Figs. 1 and 13).
The style of Tertiary extensional faulting in the central Ruby Mountains appears to be similar to that in the southern Ruby Mountains ~15 km further south (Fig. 8), and much of the upper-crustal extensional strain is interpreted to have been accommodated by at least two presently gently west-dipping, crosscutting normal fault systems (Figs. 8, 13). Approximately 22 km of Tertiary extension occurred across the southern Ruby Mountains (Fig. 8), which was primarily accommodated by the west-dipping low-angle faults that flank the western part of the range. Given that paleoburial depths of rocks exposed within the Ruby Mountains increase to the north (Snoke et al., 1990; Hudec, 1992; Wright and Snoke, 1993; Howard, 2003; Sullivan and Snoke, 2007), it is likely that an even greater amount of extension was accommodated by faulting within the central and northern Ruby Mountains, and ~27 km of extension was restored across the central Ruby Mountains in the final panel of Figure 13. This estimate of extension is primarily driven by paleodepth estimates for the Harrison Pass pluton, which indicate that the roof of the pluton was present at a depth of between 6 and 11 km in late Eocene time (Burton, 1997). The roof of the Harrison Pass pluton is restored to a depth of ~7.5 km in Figure 13. Restoring the pluton to deeper crustal levels would result in concomitantly larger estimates of Tertiary extension. Thus, the amount of extension across the central Ruby Mountains shown in the schematic reconstruction is probably close to a minimum estimate. Combining the amount of extension estimated across the central Ruby Mountains with the extension calculated across the Piñon Range and Medicine Range (Figs. 6, 10) leads to a total of ~32 km or ~50% upper crustal extension across the length of the entire schematic cross section. This regional estimate of extension is similar to the 55 to 66 km (40% to 50%) of Cenozoic extension estimated by Colgan and Henry (2009) over a 200-km long east-west transect across northeastern Nevada from the East Range to Ruby Valley, which included much of the western portion of the study area.
The final restored panel (Fig. 13) is a schematic cross-sectional representation of the regional structural configuration of the study area during the late Eocene, prior to the onset of major upper crustal extension and just after the intrusion of the Harrison Pass pluton. As shown in the restored panel (Fig. 13), the Harrison Pass pluton was much closer to the location of the modern day Piñon Range during the late Eocene, which itself is inferred to be underlain by a large volume of Eocene (~38 Ma) intrusive rocks (Grauch, 1996; Wannamaker and Doerner, 2002; Ressel and Henry, 2006). The estimates for the late Eocene depth of emplacement of the Harrison Pass pluton, 6 to 11 km, are approximately 2 to 7 km greater than the stratigraphic thickness between the Upper Ordovician strata that are present in the roof of the Harrison Pass pluton and the Pennsylvanian strata encountered in wells in Huntington Valley that restore above it. This implies that the Paleozoic section was structurally thickened prior to intrusion of the pluton. That structural thickening occurred within the core complex domain during the Mesozoic is well established from thermobarometric data and structural studies of the recumbently folded infrastructure of the Ruby Mountains core complex (Howard, 1980; Hodges et al., 1992; Hudec, 1992; Camilleri and Chamberlain, 1997; McGrew et al., 2000). However, the style of tectonic thickening remains largely uncertain. The reconstructed pre-Tertiary geology of northeastern Nevada suggests that the upper crust was generally deformed by broad folds prior to Tertiary extension (Colgan and Henry, 2009). Major contractional structures appear to have had little surface expression within the study area (Snoke and Miller, 1988; Howard, 2003; Colgan and Henry, 2009), which is consistent with models of Mesozoic thickening in northeastern Nevada that involve blind structures, such as tectonic wedging (e.g., Snoke and Miller, 1988).

**Discussion**
Constraints on Structural Reconstructions

Section balancing techniques were originally developed in contractions (Dahlstrom, 1969) and have since been widely applied in the structural analysis of fold and thrust belts (e.g., Hossack, 1979; Mitra, 1992; Wilkerson and Dicken, 2001). Cross-sectional reconstructions are less commonly attempted in extensional environments but have seen increasing application in settings of continental extension, particularly in the Basin and Range province (e.g., Proffett, 1977; Wernicke and Axen, 1988; Smith et al., 1991; Seedorff, 1991a; McGrew, 1993; Brady et al., 2000; Colgan et al., 2008). Settings of large-magnitude continental extension present unique challenges for constructing balanced cross sections, and fundamental questions remain regarding the mechanisms of continental extension, thus adding to the uncertainty of structural reconstructions in these environments. Particular challenges that complicate the construction of restored geologic cross sections in northeastern Nevada are the presence of large sedimentary basins that commonly obscure key structural relationships and the complex pre-extensional history of the region.

Where seismic reflection and borehole data are available, synextensional sedimentary basins can provide potentially useful constraints on both the age and kinematics of extensional faulting (e.g., Schlische, 1991) but can also introduce considerable uncertainties. Even where abundant geophysical and borehole data are available to constrain basin geometries, multiple structural interpretations of those data commonly are permissible, and thick accumulations of basin-fill strata typically obscure the structure of the pre-extensional “basement” rocks. The large sedimentary basins within this study area, such as Huntington and Ruby Valleys, introduce uncertainty into the cross-sectional reconstructions presented here, particularly regarding the structure of pre-extensional strata buried beneath these basins.
The polyphase deformational history of northeastern Nevada further complicates the construction of restored geologic cross sections of extensional faulting within the study area. In most of the reconstructions presented in this study, the primary structural markers used for restoring fault offsets are Paleozoic passive margin strata. Prior to experiencing Cenozoic extension, these strata experienced an earlier history of contractional deformation during the Mesozoic and in some cases, particularly in the western portion of the study, during the Paleozoic as well (Snyder et al., 1991; Carpenter et al., 1993; Camilleri and Chamberlain, 1997; Howard, 2003; Trexler et al., 2003, 2004; Dickinson, 2006). In many cases, Cenozoic extension has dissected and obscured the earlier contractional structures; this is particularly true in zones of large magnitude extension, such as at Spruce Mountain and within the core complex domain. In these complexly extended areas, uncertainties typically exist regarding both the nature of extensional faulting and the pre-extensional structure. Consequently, an iterative approach was used to produce viable cross sectional restorations through such complicated areas, wherein the pre-extensional model is progressively modified to produce a restored extensional structural geometry that satisfies all of the available geologic constraints. This approach, while necessary to restore complexly faulted areas, results in a degree of internal circularity within the retrodeformable cross section because the extensional structure and the pre-extensional structural model are not mutually independent, i.e., uncertainties regarding the structure of both the extended and restored states of the cross sections effectively increase the number of potentially valid retrodeformable cross sectional models. If the retrodeformable cross section honors the available geologic constraints in both the extended and restored states, then it can be considered a valid reconstructed section if it also is consistent with other relationships in the surrounding region.
Relationships between Areas of Differential Tilting

A key consideration, particularly for regional-scale reconstructed cross sections through extensional areas, is to determine how extended and tilted areas relate to adjacent untilted or differentially tilted areas. One simple geometric solution involves a breakaway zone where a listric fault soles into a subhorizontal master detachment. Planar or listric normal faults in the hanging wall of the detachment rotate to lower angles based on the template constraint imposed by master detachment fault (e.g., Wernicke and Burchfiel, 1982). This model works well where differential tilting has occurred between tilted hanging wall fault blocks and untilted footwall blocks. However, a common observation in extended terrains is that the footwall of the breakaway zone itself has experienced uplift and tilting, necessitating more complex models of extensional faulting and cross sectional restoration (e.g., Buck, 1988; Wernicke and Axen, 1988).

The southern Ruby Mountains (Figs. 7 and 8) appears to be one such situation where large-scale footwall tilting has occurred. West-dipping strata in the Maverick Springs Range and the apparent absence of large-offset west-dipping extensional faults imply that a major west-directed extensional breakaway zone is not present to the east of the Ruby Mountains. Meanwhile, several lines of evidence suggest that the southern Ruby Mountains has been tilted eastward between 30° and 40°, whereas the Piñon Range to the west and the northern Maverick Springs Range to the east appear to be largely untilted or slightly west tilted. If this is the case, then the eastward tilting of the southern Ruby Mountains relative to the northern Maverick Springs Range and Piñon Range must be accommodated by synformal folding in the footwall of the major west-dipping fault system that bounds the western side of southern Ruby Mountains and antiformal folding in the hanging wall (or equivalent upper-crustal structures), regardless of
whether these major west-dipping faults remain planar or become increasingly listric at depth (Fig. 8).

Restoration of the upper-crustal extensional faulting and inferred folding related to the eastward tilting of the southern Ruby Mountains relative to the northern Maverick Springs Range and Piñon Range results in loose lines in the restored cross section that are slanted at ~65° (Fig. 8d). This apparent balance problem is the inevitable result of restoring the west-dipping low-angle faults to higher angles using line balance techniques to maintain bed thicknesses, cross-sectional area, and cut-off angles between faults and stratigraphic horizons, and it may have several alternative geologic implications. One possibility is that the southern Ruby Mountains have not, in fact, experienced significant eastward tilting during extensional faulting, although this appears to contradict the available geologic and thermochronologic data. Alternatively, these slanted loose lines could largely be returned to vertical orientations by thickening the stratigraphic horizons in the restored section, possibly indicating that ductile attenuation of stratigraphic horizons occurred during extensional faulting in the southern Ruby Mountains. Plastic flow and attenuation of rocks at mid-crustal levels during Tertiary extension has been well documented in the northern Ruby Mountains (e.g., MacCready et al., 1997) and can also be inferred on a regional scale from the relatively uniform crustal thickness of the modern day Basin and Range province despite the heterogeneous distribution of upper-crustal extension (e.g., Gans, 1987; Wernicke, 1992). Precambrian and lower Paleozoic strata in the southern Ruby Mountains were present at considerable depths at the onset of Tertiary extension and commonly display sheared fabrics (Burton, 1997), suggesting that they may have experienced ductile attenuation during extension. The slanted loose lines may also be balanced in the upper crust by cryptic structures such as closely spaced antithetic faults or small-scale extensional folds.
Simple two-dimensional, line balanced cross-sectional reconstructions such as those presented in this study have obvious limitations when applied to complex settings of continental extension such as the Basin and Range province, especially because there are strong indications in certain regions of significant flow of material into and out of the plane of section at the crustal scale (e.g., Gans, 1987; MacCready et al., 1997; Maher, 2008). Nonetheless, these reconstructions can provide relatively rigorous estimates of the magnitude of upper-crustal extension, important constraints on the timing and kinematics of normal faulting, and useful insights for the development of more advanced, three-dimensional structural models of extended regions.

**Distribution and Style of Upper-Crustal Extension**

Cenozoic upper crustal extensional strain was generally directed east-west to northwest-southeast within the study area and is distributed heterogeneously, with areas of high extensional strain separated by relatively low-strain areas (Colgan and Henry, 2009). This heterogeneous distribution of extensional strain is a characteristic feature of extension in the Great Basin (e.g., Gans and Miller, 1983), and, combined with a relatively uniform crustal thickness of 30 to 35 km across the eastern Great Basin, is inferred to reflect large scale mid-crustal flow of rocks from domains that have experienced little upper-crustal extension into regions of large upper-crustal extension (e.g., Gans, 1987; Smith et al., 1991).

High strain areas examined in detail as a part of this study include the southern portion of the Ruby Mountains-East Humboldt Range core complex and the Spruce Mountain area. Both of these areas are characterized by multiple sets of crosscutting normal faults, and fault block tilting of initially high-angle normal faults to gentler dips. However, these areas differ significantly in
terms of both the magnitude and style of extension. Although the Spruce Mountain area is highly extended (~132%) this extensional strain is quite localized, and ~7 km of upper-crustal extension are inferred to have been accommodated by faulting at Spruce Mountain (Fig. 12). Moreover, the large extensional strain is distributed among numerous faults in the Spruce Mountain area that do not appear to have experienced very large amounts (>2 km) of individual offset. By contrast, extension in the southern Ruby Mountains appears to have been largely accommodated by large magnitude offset on individual faults, and the absolute amount of extension (~22 km) is far greater. Why large extensional strains are in some areas distributed among numerous faults and elsewhere concentrated by large-offset faults is not well known. However, geophysical models suggest that the style of extension may be strongly affected by the thermal structure of the lithosphere, and that normal faults that experience large amounts of hydrothermal cooling during extension have a tendency to accumulate greater amounts of offset (Lavier and Buck, 2002).

In low- to moderate-strain areas, including Emigrant Pass, the Medicine Range, and the central Piñon Range, gently dipping normal faults are absent, and the magnitude of fault block tilting is generally less than that in more extended areas. The central Piñon Range appears to be the least extended of all areas examined in this study and is primarily deformed by small to moderate offset, high-angle normal faults that have been accompanied by little overall fault block tilting. At both Emigrant Pass and in the Medicine Range, extension is accommodated by numerous closely spaced normal faults that individually lack significant offsets. This structural style is similar to the style of normal faulting at Spruce Mountain. However, while both the Medicine Range and Emigrant Pass areas have been deformed by one or two sets of closely spaced normal faults, Spruce Mountain has been deformed by as many as six crosscutting sets of faults. Thus, all of these areas appear to have been deformed by a similar style of closely spaced
normal faults, with Spruce Mountain representing a more extreme example of this style of extension.

*Ages of Extension within the Regional Study Area*

Like the Great Basin as a whole, the history of extensional deformation in northeastern Nevada has been protracted and polyphase, and though many areas have experienced large magnitude extension, the locus of active extension has appeared to have varied through time (e.g., Seedorff, 1991a; Axen et al., 1993). Thermochronologic data suggest that tectonic unroofing and major extension may have begun as early as the Late Cretaceous in the northern portion of the Ruby Mountains-East Humboldt Range core complex (Hodges et al., 1992; McGrew and Snee, 1994; Camilleri and Chamberlain, 1997; McGrew et al., 2000). Middle to late Eocene extension has been inferred based on thermochronologic data and/or the tilts of sedimentary strata in the northern part of the study area in the Elko Hills (Haynes, 2003), at Emigrant Pass (Henry et al., 2001), and in the northern part of the Ruby Mountains-East Humboldt Range core complex (Mueller and Snoke, 1993; McGrew and Snee, 1994; Camilleri and Chamberlain, 1997). Although the magnitude of Eocene extension may have been quite large in the northern part of the core complex domain (Mueller and Snoke, 1993), the magnitude of middle and late Eocene extension appears to have been relatively minor compared to later phases of extension elsewhere within the study area (Colgan and Henry, 2009). Oligocene muscovite and biotite $^{40}$Ar/$^{39}$Ar and K-Ar cooling ages have been interpreted to indicate that major extension also occurred in the Ruby Mountains-East Humboldt Range core complex between 36 and 25 Ma (Kistler et al., 1981; Dallmeyer et al., 1986; McGrew and Snee, 1994); however, based on the absence of Oligocene basin-fill sediments and a major unconformity between Eocene and Miocene strata...
west of the Ruby Mountains, recent studies have questioned whether these Oligocene cooling ages reflect upper-crustal extension in the core complex at that time (Colgan and Henry, 2009).

Particularly in the southern portion of the study area, much of the upper-crustal extension that affected northeastern Nevada appears to have occurred during the middle Miocene (Henry et al., 2001; Colgan and Henry, 2009). A major phase of middle Miocene extension is especially well documented west of the Ruby Mountains-East Humboldt Range core complex, where large thicknesses of middle Miocene coarse-grained clastic sediments have accumulated within Huntington Valley (Wallace et al., 2008), and middle Miocene cooling ages of apatite fission track and (U-Th)/He from the Harrison Pass pluton (Colgan and Metcalf, 2006) suggest rapid exhumation at that time. East of the core complex, Tertiary strata are not as well mapped or dated, and the age of faulting is largely uncertain. However, the available constraints on the age of extension are consistent with extensional faulting occurring during middle Miocene time both in Medicine Range and Spruce Mountain area, although the oldest extensional faults in the Spruce Mountain area may have initiated as early as the Eocene. Throughout the study area, extensional faulting has continued into recent times, and faults that have been demonstrably active during the late Miocene and Quaternary time are widely spaced and generally appear to dip at high angles (Colgan and Henry, 2009); however, Quaternary fault scarps are present along western flank of the Ruby Mountains (Sharp, 1939; Wesnousky and Willoughby, 2003) and may represent continued movement on the present gently dipping fault system that bounds Huntington Valley (Howard, 1992).

Implications for Petroleum Exploration in Northeastern Nevada
Although northeastern Nevada is better known for its world-class mineral deposits, the petroleum reserves in Pine Valley are an important economic asset (Flanigan et al., 1990), and it is likely that other areas within northeastern Nevada host undiscovered economic petroleum accumulations. Cenozoic extension has played a key role in the structural development of northeastern Nevada and has resulted in both increased potential and increased risk for petroleum exploration.

Cenozoic extensional faulting and the development of extensional basins resulted in the generation and retention of several known petroleum deposits in northeastern Nevada. Organic-rich shales such as the Mississippian Chainman Shale and the Eocene Elko Formation are the primary source rocks for petroleum generation in northeastern Nevada (van de Kamp and Samoun, 1992; Inan and Davis, 1994). Oil was generated in Pine Valley during late Miocene and Quaternary time as a result of the progressive burial of organic-rich lithologies of the Mississippian Shale beneath synextensional deposits (Flanigan, 1994). Oil encountered in exploratory wells in the Jiggs area was sourced from both the Chainman and Elko Formations (Schalla, 1992). Cenozoic normal faulting has also resulted in the development of structural traps at both the Tomera Ranch and North Willow Creek oil fields in Pine Valley (Hansen et al., 1994a, b). Thus, where known petroleum source rocks have been buried beneath extensional basins in northeastern Nevada, the potential for hydrocarbon generation and retention exists. Nonetheless, the polyphase structural history of northeastern Nevada can result in significant structural complexity in these prospective areas, which can complicate the exploration process. Section balancing techniques, combined with subsurface geophysical and borehole data, can help to produce robust structural models of extensional basins that can aid in the identification of
prospective hydrocarbon reservoirs and reduce the uncertainty associated with petroleum exploration in structurally complex settings like northeastern Nevada.

**Implications for Formation and Deformation of Mineral Deposits**

The Great Basin is host to considerable mineral wealth, and the abundance of Carlin-type gold deposits within northeastern Nevada make it one of the premier areas of gold production in the world (Teal and Jackson, 2002). Extensional faults commonly serve as conduits for hydrothermal fluids, and west- to northwest-directed Eocene extension and normal faulting played an important role in localizing fluid flow during the formation of Carlin-type deposits (Cline et al., 2005). Extensional faulting that occurs subsequent to mineralization can tilt and dismember ore deposits and can obscure the original regional geologic relationships that may be necessary for clarifying the genesis of the deposits. Numerous deposits in the Great Basin have been tilted and dismembered, including porphyry copper deposits in the Yerington district (e.g., Proffett, 1977; Dilles et al., 2000), the Buffalo Valley gold mine (Seedorff et al., 1991), the Hall porphyry molybdenum deposit (Shaver and McWilliams, 1987), the Mount Hope porphyry molybdenum deposit (Westra and Riedell, 1996), the Royston porphyry copper prospect (Seedorff, 1991b), and the Robinson porphyry copper district (Seedorff et al., 1996; Gans et al., 2001). Post-mineral extensional deformation can substantially aid scientific understanding of the mineralized system and can both facilitate and complicate the discovery and development of such deposits (Seedorff et al., 2008).

Within the study area, the highly extended Spruce Mountain district is an example of a locality where mineralization has been localized by pre-existing normal faults (LaPointe et al., 1991), and, depending on the age of major extension relative to mineralization, may also have
subsequently dismembered the deposit. Restoring late Eocene and younger extension across the central Ruby Mountains (Fig. 13) brings the position of the southern Carlin trend much closer to the restored positions of the Ruby Mountains and the Medicine Range (Cline et al., 2005). However, the reconstructed sections across the central Píñon Range and the Emigrant Pass areas nearest to the Carlin trend suggest that the Elko-Carlin domain has experienced only modest amounts of Tertiary extension, and ore deposits within this area are likely to be found largely intact. Particularly in structurally complex areas, cross sectional restorations such as the ones presented in this study can be useful to constrain interpretations of the subsurface structure in extended areas and can aid in the discovery of new mineral resources.

Conclusions

Detailed cross-sectional restorations of upper-crustal extensional faulting along a ~150 km transect of northeastern Nevada provide new insights into the Cenozoic structural evolution of the southern Ruby Mountains and the surrounding ranges. Upper crustal extension within the study area has been partitioned into zones of high and low extensional strain. From west to east, estimates of total extensional strain from each of the five rigorously reconstructed cross-sections presented in this study are: ~1.0 km, or about 13% extension, across Emigrant Pass; ~1.6 km, or about 10% extension, across the central Píñon Range; ~22.1 km, or about ~94% extension, across the southern Ruby Mountains and Huntington Valley; ~3.2 km, or about 20% extension, across the Medicine Range; and ~7 km or, about 132% extension, within the Spruce Mountain area. Synthesis of these rigorous restorations into a more schematic regional cross section leads to a regional estimate of ~32 km, or about ~50%, upper crustal extension across the study area.
Most of this regional extensional strain is accommodated by the fault systems that exhumed the central and southern Ruby Mountains.

The southern Ruby Mountains were primarily exhumed by at least two crosscutting systems of west-dipping normal faults that initiated at moderate to high-angles but were tilted to low angles by footwall flexure and east-directed tilting of the southern Ruby Mountains as extension progressed. Differential tilting of the southern Ruby Mountains relative to the adjacent Piñon Range and northern Maverick Springs Range during active extension on these major west-dipping faults is interpreted to have been accommodated by kilometer-scale synformal folding in their footwalls and antiformal folding in their hanging walls. Extensional strain in the areas surrounding the RM-EH core complex has frequently been accommodated by numerous closely spaced “domino style” normal faults. This style of faulting can generate extreme amounts of extension where multiple sets of crosscutting normal faults have overprinted one another, such as in the Spruce Mountain area.

Although the ages of extension are still uncertain in many places within the study area, it is clear that multiple phases of extension have affected northeastern Nevada since at least the Eocene. The most significant episode of regional extension appears to have occurred during the middle Miocene. Extension and active faulting also continues throughout the region today.

Structural complexities can complicate exploration efforts for petroleum and mineral resources in highly extended areas. Cross sectional restorations such as the ones presented in this study can be useful to constrain interpretations of the subsurface structure and fault kinematics in extended areas and, therefore, can aid in the discovery of new mineral and petroleum resources.
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References


dePolo, C.M., 2008, Quaternary faults in Nevada: Nevada Bureau of Mines and Geology Map 167, scale 1:1,000,000.


Dohrenwend, J.C., Schell, B.A., Menges, C.M., Moring, B.C., and McKittrick, M.A., 1996, Reconnaissance photogeologic map of young (Quaternary and late Tertiary) faults in Nevada, in Singer, D.A., ed., An analysis of Nevada's metal-bearing mineral resources:


Hildenbrand, T. G., and Kucks, R. P., 1988, Residual total magnetic field reduced to the North Magnetic Pole: Nevada Bureau of Mines and Geology Map 93B, scale 1:1,000,000.


McGrew, A. J., and Snee, L. W., 1994, $^{40}$Ar/$^{39}$Ar thermochronologic constraints on the
tectonothermal evolution of the northern East Humboldt Range metamorphic core

McGrew, A. J., Peters, M. T., and Wright, J. E., 2000, Thermobarometric constraints on the
tectonothermal evolution of the East Humboldt Range metamorphic core complex,

Mitra, S., 1992, Balanced structural interpretations in fold and thrust belts, in Mitra, S., and
Fisher, G. W., eds., Structural geology of fold and thrust belts: Baltimore, Johns Hopkins
University Press, p. 53-77.

Mueller, K. J., and Snoke, A. W., 1993, Progressive overprinting of normal fault systems and
their role in Tertiary exhumation of the East Humboldt-Wood Hills metamorphic


Nolan, T. B., Merriam, C. W., and Williams, J. S., 1956, The stratigraphic section in the vicinity


Struhsacker, E., eds., Geology and ore deposits of the American Cordillera, Geological Society of Nevada Field Trip Guidebook Compendium, 1995, Reno/Sparks, p. 87-91.


Stewart, J.H., and Carlson, J.E., 1976, Cenozoic rocks of Nevada--Four maps and brief
description of distribution, lithology, age, and centers of volcanism: Nevada Bureau of
Mines and Geology Map 52, scale 1:1,000,000, 4 sheets, text, 5 p.

Stewart, J. H., 1980, Geology of Nevada: A discussion to accompany Geologic Map of Nevada:

Sullivan, W. A., and Snoke, A. W., 2007, Comparative anatomy of core-complex development in

Swenson, R. F., 1991, Analysis of a fault-fold system in the eastern Pequop Mountains, Elko
County, Nevada [M.S. thesis]: Laramie, University of Wyoming, 131 p.

Teal, L., and Jackson, M.R., 2002, Geologic overview of the Carlin trend gold deposits, in
Thompson, T.B., Teal, L., and Meeuwig, R.O., eds., Gold deposits of the Carlin trend:

TH Geological Services, 1994, 7 Well Cutting Samples, Frontier Federal #10-5 (SE SE Sec. 5, T.
27N., R. 56E.), Lindsay Prospect WC, Elko County, NV: TH Geological Services Inc.,
Sand Springs, Oklahoma, Interoffice Memorandum on file at Nevada Bureau of Mines
and Geology, 3 p.


**Figure Captions**

Figure 1: Simplified geologic and shaded relief map of the study area showing the locations of the main Carlin trend, major geographic features, and tectonic domains, in addition to the locations of maps and the regional cross section discussed in text (modified from Raines et al., 2003). Inset: Index map of Nevada, Utah, and southern Idaho showing location of the study area (gray box) relative to metamorphic core complexes (black), and eastern limits of the Sevier fold and thrust belt and allochthonous western facies rocks of the Antler orogenic belt (modified from Howard, 2003).
Figure 2: Generalized composite stratigraphic columns of Neoproterozoic to Tertiary stratigraphy for the western, central, and eastern portions of the study area. Average stratigraphic thicknesses and depths are shown for Neoproterozoic to Mesozoic strata. These columns were constructed using data from Nolan et al. (1956); Woodward (1963); Collinson (1966); Thorman (1970); Hope (1972); Smith and Ketner (1975); Smith and Ketner (1976); Willden and Kistler (1979); Stewart (1980); Coats (1987); Swenson (1991); Brooks et al. (1995); and Burton (1997).

Figure 3: Geologic map of southern Emigrant Pass area showing line of section A to A’. Simplified from Henry and Faulds (1999).

Figure 4: a) Present day structure along section A to A’ across southern Emigrant Pass (Fig. 3) with bedding attitude data shown. Faults and key stratigraphic horizons have been projected above the topographic profile. Faults shown with dashed traces in the subsurface are inferred. b) Reconstruction of cross section A to A’ with extensional faulting younger than 25 Ma and ~10° of eastward tilting restored. c) Eastern half of cross section A to A’ with inferred middle Eocene extensional faults and ~20° of eastward tilting restored.

Figure 5: Geologic map of the central Piñon Range showing locations of oil wells in Pine Valley and line of section B to B’. Modified from Smith and Ketner (1978).

Figure 6: a) Present day structure along section B to B’ across the central Piñon Range (Fig. 5) with borehole and bedding added shown. Faults and key stratigraphic horizons have been
projected above the topographic profile. b) Late Eocene reconstruction of cross section A to A’ with extensional faulting restored.

Figure 7: a) Geologic map of southern Ruby Mountains and Huntington Valley showing locations of oil wells and available seismic reflection profiles (Satarugsa and Johnson, 2000) as well as line of section C to C’. This map was compiled and modified from Smith and Ketner (1978), Howard et al. (1979), Willden and Kistler (1979), Coats (1987), Burton (1997), and new field mapping by the author. b) Structure contour map of the Mitchell Creek fault (contour interval of 200 ft) from the intersection of the mapped fault trace with topography. Based on mapping by Willden and Kistler (1979).

Figure 8: a) Present day structure along section C to C’ across Huntington Valley and the southern Ruby Mountains (Fig. 7a) with borehole and bedding attitude data shown. Faults shown with dashed traces in the subsurface are inferred. b) Present day structure along section C to C’ with faults and key stratigraphic horizons projected above the topographic profile. c) Intermediate stage of reconstruction with the west dipping faults that form present day Huntington valley restored; ~15° of eastward tilting have been restored in this section (the untilted eastern and westernmost ends of the cross section are not shown in this panel). d) Pre-extensional state of section C to C’ after a line-balanced restoration of extensional faulting, folding, and ~30° of eastward tilting within the southern Ruby Mountains. Dashed line at the top of the restored section demarcates the approximate location of the middle Eocene paleosurface based on the restored structural level of lower to middle(?) Eocene lacustrine sedimentary and volcanioclastic deposits.
Figure 9: Geologic map of the Medicine Range area showing line of section M to M’. Modified from Collinson (1966) and Coats (1987).

Figure 10: a) Present day structure along section M to M’ (Fig. 9) across the Medicine Range with bedding attitudes shown. b) Present day structure along section M to M’ with faults and key stratigraphic horizons projected above the topographic profile. c) Reconstruction of cross section M to M’ with extensional faulting and ~20° of eastward tilting restored.

Figure 11: a) Geologic map of the Medicine Range area showing line of section D to D’ and crosscutting fault sets. This map was compiled and modified from Hope (1972), Fraser et al. (1986), and Swenson (1991). b) Structure contour map (contour interval of 200 ft) of major low-angle faults from the intersection of the mapped fault trace with topography. Based on mapping by Hope (1972).

Figure 12: a) Present day structure along section D to D’ (Fig. 11a) across the Spruce Mountain area with measurements of bedding attitudes shown. b) Present day structure along section D to D’ with faults and key stratigraphic horizons projected above the topographic profile. c-f) Progressive restoration of crosscutting normal faults; 42° of eastward tilting have been restored in the final panel.
Figure 13) Schematic cross sectional restoration of upper-crustal extensional faulting along a regional line of section (Fig. 1) that traverses the Piñon Range, central Ruby Mountains, and the Medicine Range.
### Table 1: Stratigraphic Units Encountered in Oil Wells Projected into Cross Section B to B’

<table>
<thead>
<tr>
<th>Well</th>
<th>Stratigraphic Units Encountered</th>
<th>Depth Interval (m)</th>
<th>Drilled Thickness (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Evan’s Flat No. 11</strong></td>
<td>Hay Ranch Fm. and Humboldt Fm.</td>
<td>0 - 933</td>
<td>933</td>
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<td></td>
<td>Indian Well Fm.</td>
<td>933 - 1177</td>
<td>244</td>
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<tr>
<td></td>
<td>Mississippian clastic rocks</td>
<td>1177 - 1288 (T.D.)</td>
<td>111</td>
</tr>
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<td><strong>Tomera Ranch South No. 9-1</strong></td>
<td>Hay Ranch Fm. and Humboldt Fm.</td>
<td>0 - 695</td>
<td>695</td>
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<td></td>
<td>Indian Well Fm.</td>
<td>695 - 1112</td>
<td>417</td>
</tr>
<tr>
<td></td>
<td>Roberts Mountain allochthon</td>
<td>1112 - 1216</td>
<td>104</td>
</tr>
<tr>
<td></td>
<td>Devils Gate Limestone</td>
<td>1216 - 1288 (T.D.)</td>
<td>72</td>
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</table>

### Table 2: Stratigraphic Units Encountered in Oil Wells Projected into Cross Section C to C’

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<th>Well</th>
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<td>Mississippian clastic rocks (Diamond Peak and Chainman Fms.)</td>
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<td></td>
<td>Silurian-Devonian carbonates</td>
<td>1273 - 2305</td>
<td>1032</td>
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<td></td>
<td>Mississippian clastic rocks, undivided</td>
<td>2305 - 3743 (T. D.)</td>
<td>1438</td>
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<td><strong>Federal No. 16-5</strong></td>
<td>Quaternary and Tertiary alluvial fill (including Humboldt Fm.)</td>
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<td></td>
<td>Eocene volcanic rocks and lacustrine sedimentary rocks</td>
<td>518 - 1128</td>
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<tr>
<td></td>
<td>Pennsylvanian Ely Limestone</td>
<td>1128 - 1267 (T.D.)</td>
<td>139</td>
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Appendix 1: Field Mapping and Observations

Geologic maps of the northern Maverick Springs Range and the Cedar Mountain extensional klippe within the central Ruby Mountains are presented with brief descriptions of mapped lithologic units and key references. Critical field observations from the Spruce Mountain area are also summarized below.

Section 1: Geologic Map of the Northern Maverick Springs Range

Field mapping of the northern Maverick Springs Range was undertaken during the fall of 2009 to better define the structural relationships of upper Paleozoic rocks that crop out immediately east of the southern Ruby Mountains. The only previously published geologic map of this area is the 1:250,000 Elko county map (Coats, 1987). The map presented here (Plate 1) is considerably more detailed (1:24,000 scale). The following section provides brief descriptions of mapped units and structural features within the map area.

Mapped Units

Rocks cropping out in the northern Maverick Springs Range consist predominantly of Permian carbonate strata. The youngest Permian units in the mapped area belong to the Upper Permian Park City Group, which includes the Kaibab, Plympton, and Gerster Formations (Hose and Repenning, 1959). In the western portion of the mapped area, the Kaibab Formation is mapped separately, whereas in the eastern part of the study, Upper Permian units above the Loray Formation, including the Kaibab Formation, are mapped as the Park City Group, undivided. At least two igneous dikes and scattered outcrops of rhyolite tuff are also present within the mapped area.

Permian Pequop Formation: Dark gray, abundantly fossiliferous limestone. Crinoid columnals are locally common, and in some places within the field area the Pequop Formation is composed almost entirely of fusulinid forams. Limestone is bioclastic to micritic and has 10-cm to 1-m scale bedding, and appears to be sandier in the upper part. Freshly broken rocks have a distinctive hydrocarbon odor. The Pequop Formation commonly crops out with a ledge-and-bench expression within the mapped area but locally can be cliff-forming. The bottom of the formation is not observed within the mapped area, but the Pequop Formation is at least several hundred meters thick in the northern Maverick Springs Range.

Permian Loray Formation: Thin-bedded, calcareous siltstone, medium to coarse sandstone, and silty limestone. The siltstone and sandstone are sometimes laminated. The Loray Formation is generally poorly exposed below cliff-forming Kaibab Limestone and above the Pequop Formation and is probably less than 35 m thick in the field area.

Permian Park City Group: The Park City Group consists of three Late Permian carbonate formations, the Kaibab Formation, the Plympton Formation, and the Gerster Formation (Hose and Repenning, 1959). Upper Permian units in the eastern part of the field area are mapped as the Park City Group, undivided. Lithologically, these rocks consist of light gray to light brown cherty limestone and dolomite and are generally medium to thick bedded, although in some placed massively bedded. Chert within the limestone is both layered and concentric. Locally, the limestones are quite fossiliferous and contain crinoid columnals and abundant brachiopods.
Permian Kaibab Formation: Light brown to light gray dolomitic limestone; 1-3 m bedding common, with some thinner bedded intervals. Massive cliff former, commonly ledgy with light gray weathered surfaces; at least ~125 m thick in the map area. The upper part consists of light gray to whitishgray, chert-rich, silty micritic limestone and minor dolomite. Chert is brown, light gray, and white. Chert locally forms up to 60% of the formation. Bedding is thick to massive, and the formation is commonly prominently exposed.

Jurassic (?) quartz monzodiorite: A dike of pink-hued quartz monzodiorite crops out in the eastern portion of the mapped area. It has a seriate to equigranular texture with the largest phenocrysts reaching ~0.5 cm in diameter. Phenocrysts include plagioclase, potassium feldspar, amphibole (altered to chlorite) and biotite (altered to chlorite + sericite). The dike has been propylitically altered; biotite phenocrysts are altered to chlorite + sericite, amphibole is altered to chlorite, and plagioclase is dusted with sericite. Potassium feldspar phenocrysts appear fresh. Quartz is present in the groundmass, but no quartz phenocrysts are observed. This dike is undated, but is compositionally similar to Jurassic aged rocks in northeastern Nevada (du Bray, 2007), and is, therefore, tentatively assigned a Jurassic age here.

Tertiary (?) granodiorite: A dike of biotite granodiorite crops out in the northern part of the mapped area. It has a seriate to equigranular texture with the largest pheoncrysts reaching ~0.5 cm in diameter. Feldspars are white and dominantly consist of plagioclase. Quartz is present in the groundmass and may comprise as much as 15% of the rock. The rock contains 20 to 25% total mafic minerals, which are dominantly biotite. Amphoboles appear to have been biotitized, indicating possible potassic alteration. Although undated, this dike is tentatively assigned a Tertiary age (Coats, 1987).

Tertiary rhyodacite tuff: Crystal-rich, moderately to densely welded devitrified tuff with a pinkish-violet hue. Outcrops of this unit are locally present within the mapped area. The tuff is composed of ~40% crystals which include quartz, sanidine, plagioclase, and biotite. Biotite is the primary mafic phase. Some quartz crystals are embayed or rounded, and fiamme up to ~5 cm in length are locally present within the tuff. Although the tuff is undated in the mapped area, it is most likely Tertiary in age and is lithologically similar to probable late Miocene tuff units described by Collinson (1966) in the Medicine Range 25 km to the northeast of the study area.

Quaternary deposits: Quaternary deposits within the field area consist of unconsolidated gravels, sands, silts, thick soils talus on slopes, and occasional caliche deposits. These deposits were not mapped in detail for this study but are widely present within the valleys between ridges and form a thin veneer over older Paleozoic and Tertiary rocks.

Structure

The northern Maverick Springs Range is cut by a several east-dipping, north-striking, normal faults, and a north-dipping normal fault is present in the eastern part of the mapped area. Faults are generally covered by colluvium, and no direct measurements of fault dips were made. Nonetheless, the map traces of the faults demonstrate that they largely ignore topography, so the faults most likely dip at relatively steep angles. Permian strata cropping out within the range generally display 10° to 40° west to southwesterly dips. Westward fault-block tilting of the Permian strata is consistent with the eastward dips of most of the high-angle normal faults, and much of this tilting likely occurred.
during extensional faulting. Southwestward dips of the Pequop Formation in the northwestern part of the mapped area suggest that the Paleozoic rocks beneath Ruby Valley to the west may be folded into a syncline. No evidence was observed in the field for a major normal fault bounding the western side of the range.

**Section 2: Geologic Map of the Cedar Mountain Klippe**

Cedar Mountain klippe is one of several extensional klippen present on the western side of the Ruby Mountains and is among the largest and best-exposed examples. These extensional klippen are underlain by gently dipping normal faults that played an important role in the exhumation of the Ruby Mountains core complex (e.g., Snoke and Lush, 1984). Because of the significance of these low-angle fault systems in the deformational history the Ruby Mountains, the Cedar Mountain klippe was mapped in detail (1:12,000 scale) during the Fall of 2009 as part of this study (Plate 2). The Cedar Mountain klippe has previously been mapped by several workers, including Willden et al. (1967), Willden and Kistler (1969), and Burton (1997). The map presented here is more detailed than previously published maps and differs regarding the locations of certain small-offset high-angle normal faults, the areal extent of the klippe, and locations of stratigraphic contacts. The following section provides brief descriptions of mapped units and structural features within the map area.

**Mapped Units**

**Tertiary monzogranite of the Harrison Pass pluton**: The Harrison Pass pluton has been mapped in considerable detail by Burton (1997). Based on mapping and descriptions by Burton (1997), the plutonic rock in fault contact with the Cedar Mountain klippe consists of biotite and two-mica monzogranite. These rocks are coarse to fine equigranular to porphyritic rocks that are compositionally and texturally layered. In general, monzogranites are composed of quartz, plagioclase, alkali feldspar, biotite, and muscovite, with rare garnet. Modal abundance of quartz varies between 24% and 45%, and the plagioclase to total feldspar ratio \( (P/A+P) \) varies between 0.30 and 0.73 (Burton, 1997).

**Devonian Devils Gate Formation**: Gray to light gray, medium- to thick-bedded limestone. The Devils Gate Formation forms conspicuous outcrops at the top of Cedar Mountain. Willden et al. (1967) report the presence of *Amphipora* and other stromatoporoids, which are the basis on which the limestone has been assigned a Late Devonian age.

**Devonian Nevada Formation**: Gray to dark gray, laminated to thick-bedded mottled dolomite. Bedding is commonly indistinct. The dolomite has been heavily brecciated and recemented near the low-angle fault surface. The dolomite is largely covered at Cedar Mountain and commonly must be mapped from float.

**Structure**

The Cedar Mountain klippe exposes unmetamorphosed Devonian carbonate rocks in low-angle fault contact with Tertiary monzogranite of the Harrison Pass pluton. Although this is a older-on-younger relationship (Willden et al., 1967), it reflects the fact that major extension in the central Ruby Mountains postdates intrusion of the late Eocene (36 Ma) pluton (Barnes et al., 2001). Similar extensional klippen further south within the Ruby Mountains, such as the Mitchell Creek klippe, clearly place younger rocks on older rocks, and paleodepth estimates suggest that the western
portion of the Harrison Pass pluton was emplaced at a depth of between 11 and 15 km (Burton, 1997). The fact that the Devonian carbonates exposed within the hanging wall of the Cedar Mountain are unmetamorphosed provides additional evidence that hanging wall rocks from relatively high structural levels (<10 km) have been juxtaposed against deeper rocks in the footwall in a normal fault relationship.

The Devonian rocks in the hanging wall of the klippe were folded prior to extensional faulting, most likely during Mesozoic compression which has been well documented in the central Ruby Mountains (e.g. Hudec, 1992). These strata are broadly folded into an anticline, and smaller-scale parasitic folds are present in outcrops on the peak of the klippe. The klippe is cut by at least one, small-offset, northwest-dipping, high-angle fault that appears to postdate the gently dipping normal fault.

**Section 3: Low-Angle Normal Faulting on the Peak of Spruce Mountain**

On the north peak of Spruce Mountain, Hope (1972) mapped a subsidiary gently-dipping normal fault structurally above the North Peak fault that juxtaposes the upper portion of the Ely Limestone with the Diamond Peak Formation/Chainman Shale, undivided (Fig. A1). However, field observations conducted as part of this study during the summer of 2010 found no evidence of this fault.

Limestone containing concentrically banded chert nodules characteristic of the lower member of the Ely Limestone (Hope, 1972) was encountered on the north peak of Spruce Mountain, structurally and stratigraphically above outcrops of conglomerate of the Diamond Peak Formation (Fig. A2a,b). Additionally, limestone interbedded with conglomerate was encountered at the contact between the Diamond Peak Formation/Chainman Shale, undivided, and the Ely Limestone on the north peak of Spruce Mountain, suggesting that the contact between the Diamond Peak Formation and overlying rocks, the lower member of the Ely Limestone, is depositional rather than a low-angle fault (Fig. A2c).

**References:**


Figure A1: Geologic map by Hope (1972) showing a subsidiary gently dipping normal fault on the north peak of Spruce Mountain.
Figure A2: a, b) Photographs of limestone containing concentrically banded chert nodules characteristic of the lower member of the Ely Limestone on the north peak of Spruce Mountain. Hand lens and rock hammer for scale. c) Limestone interbedded with conglomerate was encountered at the contact between the Diamond Peak Formation/Chainman Shale, undivided, and the Ely Limestone on the north peak of Spruce Mountain. Geologist for scale.
Geologic Map of the Northern Maverick Springs Range

Plate 1
James Pape, 2010

Stratigraphic Units

Quaternary
- Qd: Quaternary deposits

Tertiary
- Tt: Tertiary rhyodacite tuff
- T(?)g: Tertiary(?); granodiorite

Jurassic
- J(?)m: Jurassic(?); quartz monzodiorite

Permian Park City Group
- Ppc: Permian Park City Group (includes Kaibab, Plympton, and Gerster Fms.)
- Pk: Permian Kaibab Formation
- Pl: Permian Loray Formation
- Pp: Permian Pequop Formation

Map Symbols

25\ expression: Strike and dip of bedding
Contact
- Dashed where approximately located

High-angle normal fault
- Dashed where approximately located, dotted where covered. Ball and bar on downthrown side.

Scale
1:24,000

Base Maps
U. S. Geological Survey
7.5 Minute Quadrangle Topographic Maps
Ruby Lake NW and Franklin Lake SW
Geologic Map of the Cedar Mountain Extensional Klippe

Plate 2
James Pape, 2010

Stratigraphic Units

- Tertiary
  - Thp: Tertiary monzogranite of the Harrison Pass pluton

- Devonian
  - Ddg: Devonian Devils Gate Limestone
  - Dn: Devonian Nevada Formation

Map Symbols

- Strike and dip of bedding

- Contact

- High-angle normal fault
  - Dashed where approximately located.
  - Ball and bar on downthrown side.

- Low-angle normal fault
  - Hatchures on hanging wall

Scale
1:12,000

0 0.5 1 kilometers

0 0.5 1 miles

Base Map
U. S. Geological Survey
7.5 Minute Quadrangle Topographic Map
Harrison Pass
Appendix 2: Pre-Cenozoic Stratigraphic Summary for Regional Study Area

This appendix summarizes the pre-Cenozoic stratigraphy of the regional study area. The primary purpose of this summary is to establish a stratigraphic framework for reconstructed cross sections. Therefore, special emphasis is given to the thicknesses of sedimentary strata and the presence of major unconformities in the stratigraphic section, whereas formation lithologies have been generalized. For more detailed stratigraphic information, the reader is directed to the references cited herein. The stratigraphic thicknesses and relationships shown in the reconstructed cross sections are based on the information contained in this summary.

Precambrian to Devonian Strata

The oldest significant sedimentary sequence in northeastern Nevada is the Neoproterozoic to Devonian portion of the Cordilleran miogeocline, which was deposited in a shallow-water, passive margin setting after Neoproterozoic rifting and attains substantial thickness within the regional study area (e.g., Stewart, 1980; Poole et al., 1992; Dickinson, 2006).

Neoproterozoic and Early Cambrian Strata

Neoproterozoic metasedimentary rocks only crop out in a few places in northeastern Nevada, and with the exception of the intruded, highly metamorphosed, and polydeformed rocks of the central and northern Ruby Mountains (e.g., Howard et al., 1979; Snoke, 1980), Neoproterozoic rocks do not crop out within the study area. The nearest significant exposure of Precambrian metasedimentary rocks occur in the Egan Range north of Ely and ~100 km southeast of the southern Ruby Mountains. There, Woodward (1963) measured a thickness of ~2576 m (8450 ft) of Precambrian quartzite, slate, phyllite, and argillite assigned to the McCoy Creek Group, which is the oldest sedimentary sequence that crops out in northeastern Nevada. The base of the section is not exposed in the Egan Range (Woodward, 1963), and, therefore, the measured thickness there represents a minimum. Approximately 24 km east of the Egan Range in the Schell Creek Range, Misch and Hazzard (1962) measured a similar thickness (8800 ft; 2682 m) of McCoy Creek strata, the base of which is also not exposed.

Directly overlying the McCoy Creek Group in northeastern Nevada is the Late Proterozoic to Early Cambrian Prospect Mountain Quartzite (Misch and Hazzard, 1962; Woodward, 1963). The base of the Prospect Mountain Quartzite is considered time transgressive in the Great Basin (Coats, 1987), and its contact with the underlying the McCoy Creek Group may represent an unconformity (Woodward, 1963). Within the study area, the Prospect Mountain Quartzite crops out in the Ruby Mountains (Willden and Kistler, 1979; Burton, 1997). Lithologically, the Prospect Mountain Quartzite is comprised of medium-bedded quartzite with minor interbeds of phyllite or schist (Willden and Kistler, 1979). A 1220-m (4000-ft) thick, partial section of Prospect Mountain Quartzite was measured by Willden and Kistler (1979) in the central Ruby Mountains that they considered to be close to a maximum thickness for the formation.

Based on interpretation of COCORP seismic data and surface mapping along a regional transect of central and eastern Nevada that passes ~60 km south of the regional study area, Smith et al. (1991) estimated a thickness of ~6 km of late Precambrian to Early Cambrian clastic strata (shown on their regional cross sections) from the Egan Range to the Sulphur Spring Range. This portion of the section of Smith et al. (1991) corresponds roughly to the east-to-west position of the Medicine Range to the Piñon Range further north. The thickness of Precambrian to Early
Cambrian clastic strata shown on the cross section of Smith et al. (1991) thins to ~4.5 km beneath the Roberts Mountains, which corresponds to western Pine Valley and the Cortez Range further north. Although not explicitly stated by Smith et al. (1991), their regional cross sections and maps imply that their late Precambrian to Early Cambrian clastic unit includes the Prospect Mountain Quartzite and all older miogeoclinal rocks.

Given the above constraints, a 5- to 6-km stratigraphic thickness for Neoproterozoic through Lower Cambrian strata seems a reasonable estimate for these units in the regional study area.

**Middle and Upper Cambrian Strata**

Within the regional study area, Middle and Upper Cambrian units crop out only within the Ruby Mountains. A complete sequence of variably metamorphosed, carbonate and fine-grained siliciclastic Middle and Upper Cambrian rocks in the southern Ruby Mountains conformably overlies the Prospect Mountain Quartzite (Sharp, 1942; Willden and Kistler, 1979; Burton, 1997). Although these rocks are temporally equivalent to regionally correlated Cambrian formations, correlations are complicated by the effects of regional metamorphism, thermal metamorphism adjacent to Mesozoic and Cenozoic intrusive rocks, a polyphase deformational history, and apparent facies changes (Willden and Kistler, 1979, Burton, 1997). Because of this, the only regional formation names that have been applied to these strata are the Pioche Shale and the Windfall Formation. The Pioche Shale is a ~150-m thick sequence of silty and sandy recrystallized limestone and sandstone underlain by a 7-m thick basal phyllite member that forms the lowest Middle Cambrian Unit, whereas the Windfall Formation is a ~300-m thick section of cherty, shaly, and sandy limestone that is the uppermost Upper Cambrian unit (Willden and Kistler, 1979; Burton, 1997). Thickness estimates of Middle and Upper Cambrian strata in the southern Ruby Mountains have been made by several previous workers, and, likely as a result of the structural complexity in this area, have been somewhat variable. Total thickness estimates of Upper and Middle Cambrian strata in the southern Ruby Mountains fall between ~1700 and 2200 m (Sharp, 1942; Willden and Kistler; 1979; Hudec, 1990; Burton, 1997). Burton (1997) measured a total thickness of between 1980 and 2120 m of Middle and Upper Cambrian strata using a Jacobs staff and clinometer in the southern Ruby Mountains, and his thickness estimate is here considered to be the highest confidence estimate available.

Cambrian rocks do not crop out immediately east of the Ruby Mountains within the regional study area; however, where the nearest complete Upper and Middle Cambrian stratigraphic section is exposed in the northern Egan Range, approximately 40 km southeast of the Medicine Range and 70 km south of Spruce Mountain. The total thickness of Middle and Upper Cambrian strata in the Egan Range is ~2400 m (Fritz, 1968). Similarly, the Middle and Upper Cambrian section does not crop out in the vicinity of the Piñon Range. However, a complete Middle and Upper Cambrian section, ~1800 m thick, has been well documented in the Eureka district, approximately 90 km south of the Piñon Range by Nolan et al. (1956).

Based on the above stratigraphic constraints, Middle and Upper Cambrian strata appear to thicken within the regional study area from west to east, from ~1800 m in the Pine Valley/Piñon Range area, to ~2000 to 2100 m in the southern Ruby Mountains, to a maximum of ~2400 m in the Spruce Mountain area.
Ordovician Strata

Ordovician strata crop out within the regional study area in multiple locations, including the Piñon Range, Ruby Mountains, and Spruce Mountain. Ordovician strata in northeastern Nevada are typically divided, in ascending stratigraphic order, into the Pogonip Group, the Eureka Quartzite, and the Fish Haven Dolomite or equivalent units (Stewart, 1980). Ordovician strata within the regional study area that underlie the Eureka Quartzite are assigned to the Pogonip Group and are considered to be Lower and Middle Ordovician (e.g., Coats, 1987). At Spruce Mountain, Hope (1972) reported a minimum thickness of 762 m (2500 ft) of Ordovician shaly limestone and dolomite that he assigned to the Pogonip Group; however, the base of the group was not exposed. Hope (1972) also recognized a very thin (~2 m thickness) quartzite at the top of the Ordovician sequence that he assigned as the Eureka Quartzite. Thorman (1970) reported a thickness of ~680 m (2240 ft) of Ordovician strata in the northern Pequop Range, but the base of the section was also not exposed. In the northern Egan Range, approximately 70 km south of Spruce Mountain, the Ordovician section is over 1220 m (4000 ft) thick (Fritz, 1968). However, regional isopach maps (Stewart, 1980) show the total thickness of Ordovician strata in the eastern portion of the study area thickening substantially from north to south; hence the maximum thickness of Ordovician strata at Spruce Mountain may not be significantly greater than the 762 m reported by Hope (1972).

In the southern Ruby Mountains, Sharp (1942) originally measured a total thickness of 1112 m (3650 ft) of gray, massive, cherty limestone, platy argillaceous limestone, and shale, topped by limy quartzite, that he assigned to the Pogonip Group. Willden and Kistler (1979) broke the Pogonip Group into four informal members in the southern Ruby Mountains and measured a thickness of between 825 and 885 m of Ordovician strata. Based on the local absence of the Eureka Quartzite, Willden and Kistler (1979) recognized an unconformity between the upper Pogonip Group and the overlying Silurian Lone Mountain Dolomite. Willden and Kistler (1979) interpreted this unconformity as a major regional hiatus indicative of an Ordovician orogeny, and they associated folding in the southern Ruby Mountains with this event. Alternatively, the folding and thrust faulting in the southern Ruby Mountains may be related to the forced emplacement of the Harrison Pass pluton and/or later Paleozoic or Mesozoic thrusting (Reese, 1986; Burton, 1997). Due to structural complexity within the southern Ruby Mountains as well as the unconformity present at the top of the Ordovician section, reported thicknesses of Ordovician strata vary considerably from 808 m to 1579 (Sharp, 1942; Willden and Kistler, 1979; Hudec, 1990; Burton, 1997). The upper thickness estimates were measured in folded areas near the contact with the Harrison Pass pluton, and the apparent stratigraphic thicknesses in those areas may have been structurally thickened. Given the above constraints, a stratigraphic thickness of between 810 and 1100 m is considered a reasonable estimate of the total stratigraphic thickness of Ordovician strata in the southern Ruby Mountains.

In the Piñon Range, a complete Ordovician section is not exposed. However, Smith and Ketner (1975) measured 107 m (350 ft) of dark gray dolomite that they correlated to the uppermost portion of the Pogonip Group. The base of the section was not exposed. Above this dolomite unit, Smith and Ketner (1975) measured an additional thickness of 67 m (220 ft) of Middle and Late Ordovician strata including the Eureka Quartzite. South of the Piñon Range in the Eureka district, a complete Ordovician section has a total measured thickness of 731 m (2400 ft) (Nolan et al., 1956).

Given the above constraints, Ordovician strata generally appear to maintain a fairly constant total thickness of between 700 and 800 m from west to east across the regional study area.
area, possibly thickening somewhat to the east. However, as both measured sections in the southern Ruby Mountains and regional isopach maps show (Stewart, 1980), the Ordovician strata thicken substantially toward the southern portion of the study area, and a thickness of up to 1100 meters of Ordovician strata may be present in the southern Ruby Mountains and underlying the Medicine Range.

Silurian and Devonian Strata

Silurian and Devonian rocks crop out in multiple locations throughout the study area. East of the Ruby Mountains at Spruce Mountain, Hope (1972) measured a minimum thickness of 915 m (3000 ft) of Silurian and Devonian limestone, dolomite, sandy limestone and minor shaly dolomite. Hope (1972) did not assign a formal name to the Silurian strata, but he assigned the Devonian strata to the Simonson Dolomite and Guilmette Formation. In the southern Cherry Creek Range, Fritz (1968) measured a total thickness of 1400 m (4580 ft) of Silurian and Devonian dolomite, limestone, and shale that included an unnamed Silurian dolomite, the Sevy Dolomite, the Simonson Dolomite, and the Guilmette Formation. In the northern Pequop Range, Thorman (1970) reported a thickness of 1230 m (4030 ft) of Silurian and Devonian dolomite and limestone that includes the Laketown Dolomite, the Simonson Dolomite, and the Guilmette Formation.

In the southern Ruby Mountains, the Silurian to Early Devonian dolomite are assigned to the Lone Mountain Dolomite. The basal contact of the Lone Mountain Dolomite with the underlying Pogonip Group in the southern Ruby Mountains has been interpreted as an unconformity based on the absence of the Eureka Quartzite and an angular discordance between the Lone Mountain Dolomite and underlying strata (Sharp, 1942; Willden and Kistler, 1979; Burton, 1997). Thickness estimates of the Lone Mountain Dolomite in the southern Ruby Mountains vary between 442 and 750 m (Sharp, 1942; Willden and Kistler, 1979; Hudec, 1990; Burton, 1997). This wide range of thickness estimates is likely the result of structural and stratigraphic complications in the southern Ruby Mountains, and a thickness of between 650 and 750 m is here considered the most reasonable estimate for the Lone Mountain Dolomite in the southern Ruby Mountains. Overlying the Lone Mountain Dolomite in the southern Ruby Mountains are the Devonian dolomites and limestones that are assigned to the Nevada and Devils Gate Formations, which have an estimated total thickness of between 641 and 686 m (Willden and Kistler, 1979). Analysis of the map patterns of Devonian strata, however, shows that these units may be thicker in the southern Ruby Mountains (as much as 870 m thick). Thus, based on the above constraints, the total stratigraphic thickness of Silurian-Devonian units in the southern Ruby Mountains varies between 1300 and 1700 m.

In the Piñon Range, Smith and Ketner (1975) measured 237 m of Silurian-Devonian Dolomite that they assigned to the Lone Mountain Dolomite, but the top of this section was not exposed. Based on the map pattern of the Lone Mountain Dolomite south of Raven’s Nest, Smith and Ketner (1975) estimated a thickness of 426 meters, but where this estimate was made, the section is bounded by faults and the total stratigraphic thickness of the formation may be greater. In the field area, the Lone Mountain Dolomite is thought to be Late Silurian, and the basal contact with the underlying Upper Ordovician Hanson Creek Dolomite may represent a considerable depositional hiatus. Stratigraphically above the Lone Mountain Dolomite in the Piñon Range, Smith and Ketner (1975) measured an additional thickness of 773 to 1260 m of Devonian dolomite and limestone, which includes the Nevada Formation and the Devil’s Gate
Limestone. Thus, the total thickness of Silurian and Devonian strata in the Carlin-Piñon Range area ranges between ~1200 and 1700 m.

Based on the above constraints, the total thickness of Silurian and Devonian strata range between a minimum value of ~1000 m in the easternmost portion of the study area and a maximum value ~1700 m in the southern Piñon Range and southern Ruby Mountains.

**Antler Orogeny**

In Late Devonian-Early Mississippian time, northeastern Nevada experienced a major contractional deformation event known as the Antler orogeny that was most likely caused by the collision of an island arc complex with the North American continent (e.g., Speed and Sleep, 1982; Dickinson, 2006). During the Antler orogeny, deep-water siliciclastic rocks were thrust eastward over coeval shelfal carbonate rocks along the Roberts Mountain thrust (e.g., Stewart, 1980; Coats, 1987). East of the Roberts Mountains allochthon, a thick wedge of dominantly siliciclastic sediments was deposited in the Antler foreland basin throughout Mississippian time (Poole, 1974).

**Allochthonous Lower Paleozoic Strata**

Allochthonous Ordovician to Devonian siliciclastics transported eastward in the upper plate of the Roberts Mountains thrust during the Late Devonian to Early Mississippian Antler orogeny (e.g., Speed and Sleep, 1982; Coats, 1987) crop out in the westernmost part of the study area at Emigrant Pass and in the Piñon Range. Lithologically, these allochthonous upper-plate rocks consist of a mix of siliceous siltstone and shale with occasional interbeds of limestone, dolomite, and chert, as well as mafic lavas or sills. These rocks were probably deposited in a deep marine setting (e.g., Smith and Ketner, 1975; Henry and Faulds, 1999). The allochthonous rocks are typically highly internally deformed and of variable thickness. Because of this structural complexity, stratigraphic thickness estimates are not given here.

**Mississippian Strata**

Mississippian strata crop out over much of the study area and are dominantly composed of a thick sequence of fine- to coarse-grained siliciclastic rocks deposited in the foreland basin of the Antler orogen (e.g., Poole, 1974). Mississippian strata thicken substantially from east to west across the regional study area, consistent with the inferred location of the Antler foreland keel in the vicinity of the modern-day Piñon Range (e.g., Poole, 1974; Speed and Sleep, 1982). In this report, the Late Devonian – Early Mississippian Pilot Shale and the Diamond Peak Formation (which is predominately Late Mississippian, but as it is mapped in some places also includes strata of very Early Pennsylvanian age) are included within the Mississippian Group.

East of the Ruby Mountains, Hope (1972) mapped the Mississippian sequence at Spruce Mountain as an undivided unit of Chainman Shale and Diamond Peak Formation. The base of the section was not exposed, and he estimated the thickness to be over 760 m (2500 feet). Approximately 35 km north of Spruce Mountain in the northern Pequop Range, Thorman (1970) estimated a total thickness of ~1310 m (~4300 ft) of Mississippian strata that include the Joana Limestone, Chainman Formation, and Diamond Peak Formation. Thorman’s (1970) thickness estimate of 950 m (2970 ft) of Chainman Formation in the northern Pequop Mountains differs significantly from an earlier thickness estimate of 305 to 365 m (1000 to 1200 ft) by Bissell (1967). In the southern Cherry Creek range, approximately 70 km south of Spruce Mountain,
Fritz (1968) reports a thickness of 532 m (1745 ft) of Mississippian shale and limestone, including the Late Devonian-Early Mississippian Pilot Shale, the Joana Limestone, and the Chainman Shale.

Mississippian rocks that are exposed in the southern Ruby Mountains include the Late Devonian-Mississippian Pilot Shale, the Joana Limestone, Chainman Shale, and the Mississippian-Early Pennsylvanian Diamond Peak Formation (Willden and Kistler, 1979). A complete section of Mississippian rocks does not crop out in the southern Ruby Mountains, and therefore, the original stratigraphic thickness of the Mississippian section is somewhat uncertain. Complete sections of both the Pilot Shale and the Joana Limestone crop out in the southern Ruby Mountains, with a total thickness of ~165 m (Willden and Kistler, 1979). Only ~10 m of Chainman Shale crops out in the southern Ruby Mountains (Willden and Kistler 1979); however, a thickness of 300+ m of Chainman Shale crops out at the Big Bald Mountain 30 km to the south of the Ruby Mountains (Nutt and Hart, 2004). In both locations, a complete thickness of the Chainman Shale is not exposed. A substantial thickness (~945 m) of conglomerate assigned to the Diamond Peak Formation crops out in the hanging wall of Mitchell Creek klippe, but neither the base nor the top of the formation is exposed (Willden and Kistler, 1979). Reconstructed isopach maps by Poole (1974) and Stewart (1980) show a considerable original thickness of Mississippian strata in the southern Ruby Mountains of ~2400 m and ~1850 m, respectively, reflective of the proximal position of the southern Ruby Mountains to the Antler thrust front within the foreland basin of the Antler orogen. Burton (1997) estimated an original thickness of ~2 km for Mississippian rocks within the southern Ruby Mountains. The above stratigraphic constraints appear to require a minimum original thickness that is on the order of 1300 m for Mississippian strata in the southern Ruby Mountains, and this thickness may have been as much as 2400 m. Given these constraints and uncertainties, a thickness of ~1850 m is assumed in this study for the original thickness of Mississippian strata deposited within the southern Ruby Mountains.

The Mississippian stratigraphy of the Piñon Range is complicated by the effects of the Antler orogeny, and, particularly in the northern Piñon Range, has been the subject of considerable study and reinterpretation by numerous workers (e.g., Dott, 1955, Smith and Ketner, 1975; Johnson and Pendergast, 1981; Iverson, 1992; Carpenter et al., 1993; Trexler et al., 2003). One critical observation is that, in several locations within the Piñon Range, upper plate strata of the Roberts Mountain thrust have been overthrust onto Mississippian strata that are interpreted to have been deposited within the Antler foreland basin (Iverson, 1992). This relationship indicates that either the Roberts Mountain thrust overrode its own foreland basin strata during the Antler orogeny (e.g., Johnson and Pendergast, 1981), that the Roberts Mountain thrust was reactivated during subsequent east-vergent compressional events (e.g., Trexler et al., 2003), or that both processes occurred. Additionally, the presence of angular unconformities within the Mississippian sequence in the northern Piñon Range indicates that Mississippian deformation occurred within the area after the onset of the Antler orogeny. Addressing much of the Mississippian stratigraphic complexity within the Piñon Range in detail is beyond the scope of this study, and Mississippian strata are generally treated as a single unit in the restored cross sections. The maximum combined thickness of the Chainman and Diamond Peak Formations within the Piñon Range is estimated by Smith and Ketner (1975) to be between 1850 and 2150 m.

Regional palinspastic isopach maps (e.g., Poole, 1974; Stewart, 1980) have been useful for estimating stratigraphic thickness variations of Mississippian strata across the study area. The
isopach map of Stewart (1980) shows a maximum thickness of ~2440 m (~8000 ft) in the vicinity of the southern Ruby Mountains, thinning eastward to ~915 m (~3000 ft) in the vicinity of the Medicine Range. Isopachs presented by Poole (1974) show a similar eastward thinning trend to those of Stewart (1980) but generally appear to show a somewhat lower total thickness of Mississippian strata (as much as 1000 ft less), especially in the eastern part of the study area.

**Pennsylvanian and Permian Strata**

Pennsylvanian and Permian strata crop out widely over the study area both east and west of the Ruby Mountains. Especially east of the Ruby Mountains, these strata are commonly the youngest rocks that crop out below the Tertiary strata. With the exception of an isolated outcrop of Ely Limestone just east of Harrison Pass, no outcrops of Pennsylvanian and Permian rocks are present within the southern and central Ruby Mountains; however, Pennsylvanian and Permian units have been mapped in fault-bounded slices within the northern Ruby Mountains and East Humboldt Range (e.g., Howard, et al., 1979; Snoke and Lush, 1984).

East of the Ruby Mountains at Spruce Mountain, Hope (1972) measured a total thickness of ~915 m of Pennsylvanian cherty limestone, including the Pennsylvanian Ely Limestone, and the Pennsylvanian/Lower Permian Riepe Spring Limestone. Hope (1972) reported an additional thickness of 2050+ m of Permian cherty limestone, siltstone, silty limestone, and cherty dolomite in units that include the Rib Hill Formation, the Pequop Formation, the Loray Formation, the Kaibab Limestone, and the Pympton Formation. Immediately east of Spruce Mountain, in the southern Pequop Mountains, Fraser et al. (1986) report a wide range in thickness of between ~550 and ~1550 m of Pennsylvanian cherty limestone, including the Pennsylvanian Ely Limestone and the Pennsylvanian/Lower Permian Riepe Spring Limestone. Overlying these Pennsylvanian formations, Fraser et al. (1986) report a thickness of between ~1060 and ~2000 m of Permian cherty limestone, siltstone, silty limestone, and cherty dolomite that include the Rib Hill Formation, the Pequop Formation, the Loray Formation, and the Park City Group, which consists of the Kaibab Limestone, the Pympton Formation, and the Gerster Formation. In the southern Cherry Creek Range, Fritz, (1968) reports a thickness of ~650 m of Pennsylvanian cherty limestone and siltstone mapped as Ely Limestone. Overlying this formation is 601 m of Permian siltstone, shale, and limestone mapped as the Arcturus Formation; the top of the formation is eroded in the Cherry Creek Range. In the northern Pequop Mountains Thorman (1970) reports a thickness of ~460 to ~610 m of Pennsylvanian cherty limestone, siltstone, and shale that includes the Ely Limestone and the Hogan Formation. Thorman (1970) reports an additional thickness of 690+ m of Permian cherty and silty limestone that included the Ferguson Mountain Formation and the Pequop Formation. In the Medicine Range, Collinson (1966) reports a thickness of ~1930 m of Permian strata that include the Pequop Formation and the Park City Group.

West of the Ruby Mountains, in the Carlin-Piñon Range area, Pennsylvanian strata equivalent to the Ely Limestone are mapped as the Moleen and Tomera Formations (Smith and Ketner, 1975; Coats, 1987). The Moleen Formation ranges from ~365 to 490 m in thickness and is composed of limestone that is in some places silty or sandy and contains chert-pebble conglomerate lenses. The Tomera Formation commonly overlies the Moleen Formation. Lithologically, the Tomera Formation is composed of interfingerling limestone and siliceous clast conglomerate that has a thickness of between 520 and 610 m. Unconformably above the Moleen and Tomera Formations is a sequence of up to 1525 m of Upper Pennsylvanian limestone, conglomerate, calcareous sandstone and siltstone, and dolomite that includes the Strathearn
Formation in the northern portion of the Carlin-Piñon Range area but otherwise has been mapped as an undivided unit. The thickness of these Upper Pennsylvanian and Permian units is highly variable in the Carlin-Piñon Range area because both the upper and lower contacts are unconformities (Smith and Ketner, 1975).

Regional reconstructed isopach maps (Stewart, 1980) show an original stratigraphic thickness of up to ~2500 m of Pennsylvanian and Permian strata in the southeasternmost portion of the study area within the Medicine Range, which thins westward to ~1850 m beneath Pine Valley (Stewart, 1980).

Given the above constraints, the thickness of Pennsylvanian and Lower Permian strata underlying the Medicine Range and northern Maverick Springs Range, including the Ely Limestone and the Riepe Spring Limestone or equivalent strata, is estimated to be between ~600 and 1000 m. The original thickness of Permian strata east of the southern Ruby Mountains, including the Rib Hill Formation/Riepetown Sandstone, Pequop Formation, and the Park City Group, is estimated to be between ~2100 and 2200 m. West of the Ruby Mountains in the Carlin-Piñon Range area, the total original stratigraphic thickness of Pennsylvanian and Permian may have been as great as ~2600 m in some areas, but the presence of unconformities within the Pennsylvanian/Permian stratigraphic section implies that original sedimentary thicknesses of these units may have been variable.

**Mesozoic Strata**

Mesozoic strata are exposed primarily at the eastern and western extremes of the study area and are absent in the immediate vicinity of the Ruby Mountains (Coats, 1987).

**Triassic Stratigraphy**

Triassic sedimentary rocks are exposed in the easternmost portion of the study area within the Medicine Range and just east of Spruce Mountain in the core of the Pequop Syncline; Triassic rocks are also present north of the study area within the East Humboldt Range (Coats, 1987). The most complete Lower Triassic section in northeastern Nevada crops out in the southern Pequop Range. Swenson (1991) measured a thickness of between ~1150 and 1200 m of marine shale and thin-bedded limestone that he assigned to the Early Triassic Thaynes Formation. Fraser et al., (1986) measured a similar thickness of ~1000 m of shale and limestone that they assigned to the Lower Triassic Thaynes Formation in the southern Pequop Range. Overlying the Thaynes Formation, Swenson (1991) mapped a thickness of 50+ m of non-marine Upper Triassic(?) sandstone, mudstone, and conglomerate. In the Medicine Range, Collinson (1968) reports a thickness of ~160 m of Lower Triassic calcareous sandstone, sandy limestone, conglomerate, and calcareous siltstone that he assigned to the Thaynes Formation. The top of the Triassic section within the Medicine Range is eroded. Immediately south of the southern Pequop Mountains near Currie, Nelson (1956) measured ~670 m of Thaynes Formation, which may be structurally attenuated in this area (Swenson, 1991). Overlying the Thaynes Formation in the Currie area, Nelson (1956) described an additional thickness of ~220 m of Upper Triassic(?) sandstone, conglomerate, and shale tentatively assigned to the Timothy and Chinle Formations.

**Jurassic Stratigraphy**

Stratified Jurassic rocks crop out in the westernmost extreme of the study area within the Cortez Range/Pine Valley area and also just outside the southeasternmost portion of the study area in the Currie area. In the Currie area, Coats (1987) reports a thickness of ~850 m of fine-
medium-grained non-marine sandstone tentatively correlated with the Nugget Sandstone of southwestern Wyoming. In the Cortez Range/Pine Valley area, the Pony Trail Group consists of a sequence of Jurassic volcanic and volcaniclastic rocks that may be as much as 2750 m thick (Muffler, 1964). With the exception of a potential Jurassic rhyolite occurrence in the southern Piñon Range (Smith and Ketner, 1976), these rocks dominantly crop out in the Cortez Range west of the study area (Muffler, 1964).

**Cretaceous Stratigraphy**

Cretaceous sedimentary rocks crop out in limited exposures within the Piñon Range and more extensively at the western edge of Pine Valley near the Cortez Range. These rocks are assigned to the Newark Canyon Formation, which consist of mudstone, siltstone, conglomerate, shale, and limestone and attain a maximum thickness of ~2200 to 2400 m on the western side of Pine Valley near the Piñon Range (Smith and Ketner, 1976).

**References**


Nutt, C. J., and Hart, K. S., 2004, Geologic map of the Big Bald Mountain quadrangle and part of the Tognini Spring Quadrangle, White Pine County, Nevada: Nevada Bureau of Mines and Geology Map 145, scale 1:24,000, text, 8 p..


