Cenozoic extension superimposed on a Mesozoic thrust belt, central Nevada

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CENOZOIC EXTENSION SUPERIMPOSED ON A MESOZOIC THRUST BELT,
CENTRAL NEVADA

by

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ABSTRACT

Cenozoic Extension Superimposed on a Mesozoic Thrust Belt, Central Nevada

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The central Nevada thrust belt (CNTB) is a north-south trending belt of contractile structures in the hinterland of the Sevier orogenic belt. This study tests the hypothesis that analysis and regional correlation of CNTB structures is possible after retrodeformation of younger extensional faults.

The central Pancake Range underwent multiple episodes of Cenozoic extensional faulting. Fault sets generally strike N-S or NE-SW; one synvolcanic set strikes E-W. These normal fault sets: (1) form five temporally distinct episodes that correspond to regional tectonic events, (2) record a rotation in the stress field through time; and (3) locally developed within a three dimensional strain field as opposed to plane strain.

The Sand Spring, Ike Spring, and Indian Spring thrusts (previously undocumented) regionally correlate to the Duckwater and White Pine thrusts. These thrust faults have steeply-dipping ramps and formed out-of-sequence. These structures represent the CNTB in east-central Nevada, a single contractional belt with at least 27 km of shortening.
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CHAPTER 1

INTRODUCTION

The east-central portion of the Great Basin underwent contraction mainly in the Mesozoic followed by episodic Cenozoic extensional deformation and volcanism. Despite numerous studies of the extent and timing of these events, the number of episodes of deformation in the Great Basin makes it difficult to completely understand the evolution of both contraction and extension through time. This paper focuses on unraveling the tectonic development during both deformations through a detailed study of the Mesozoic central Nevada thrust belt and Cenozoic extensional structures exposed in a part of east-central Nevada. A major objective is to determine the original along strike continuity of folds and thrusts of the central Nevada thrust belt. The goals in examining the Cenozoic faults are to (1) understand them in detail to allow accurate retrodeformation of these structures and thus, allow analysis of the central Nevada thrust belt and (2) determine whether significant changes in extension directions occurred through time.

The Mesozoic central Nevada thrust belt is a north-south trending fold and thrust belt associated with deformation in the hinterland of the Sevier orogenic belt (Figure 1). The structures within it were originally defined as minor hinterland structures of the Sevier belt by Armstrong (1968). Later these structures were defined as a belt of
contractional structures, the Eureka belt, connected to the Sevier belt by a large basal decollement (Speed, 1983). Many contractional structures have since been documented in east-central Nevada, indicating that the region has gone through significant shortening during mainly Mesozoic time (Bartley and Gleason, 1990; Camilleri, 1992; Armstrong and Bartley, 1993; Carpenter et al., 1993; Taylor et al., 1993; Dobbs et al., 1994; Langrock, 1995; Cole and Cashman, 1999; Taylor et al., 2000). East-vergent folds and thrust faults with long, steeply dipping ramps dominate the types of structures exposed within the belt (Carpenter et al., 1993; Taylor et al., 1993, 2000). Current workers correlate structures within the southern central Nevada thrust belt directly with structures in the Sevier orogenic belt. Thus, they demonstrate that the central Nevada thrust belt is an internal branch of the Sevier fold and thrust belt (Taylor et al., 2000).

Extensional overprinting in the eastern Great Basin began in the late Mesozoic and is coeval with the last phases of Sevier contraction (Wells, 1997). During the principal period of Cenozoic extension, different regions in the Great Basin experienced different episodes of extension, with probable spatial and temporal variations in the regional stress field (Bartley, 1989; Taylor et al., 1989; Switzer and Taylor, 2001). Several episodes of Eocene to Quaternary extension are recognized in the Pancake and nearby ranges (Taylor et al., 1989; Camilleri, 1992; Axen et al., 1993; Liberty et al., 1994; Langrock, 1995; Gans et al., 2001).

This study focuses on structures exposed within the central Pancake Range, Nye County, Nevada and areas directly to the east across Railroad Valley (Figure 2). Within the central Pancake Range, both contractional and extensional structures were mapped. Detailed mapping, stereonet, and retro-deformable cross section analyses delineate
contractional structures related to the central Nevada thrust belt, (2) document up to five sets of extensional structures within the central Pancake Range, (3) provide constraints on the relative timing of events that created these structures, and (4) ultimately allow interpretations about the tectonic development of the area in the Mesozoic and Cenozoic eras.

The structures in the central Pancake Range are important because the area lies between previously mapped contractional structures of the central Nevada thrust belt to the southeast within the Grant and Quinn Canyon Ranges (e.g., Taylor et al., 2000; Fryxell, 1998) and to the north in the northern Pancake and Diamond Ranges (e.g., Perry and Dixon, 1993; Carpenter et al., 1993; Dobbs et al., 1994; Langrock, 1995; French and Schalla, 1998). Along strike correlation of contractional structures reveals the sequence of thrusting and the genetic relationships of central Nevada thrust belt structures to ultimately provide interpretations of the age, amounts of shortening, and tectonic development of the central Nevada thrust belt at the latitude of the study area. Prior to this study, central Nevada thrust belt structures exposed to the north and west of Railroad Valley and those exposed south and east of the valley were analyzed separately. In this study, the thrust belt is extended across Railroad Valley.

Despite its importance for correlating local and regional structures, detailed structural mapping in the central Pancake Range had not been performed prior to this study. Two significant contractual structures exposed within the northern Pancake Range are the McClure Spring syncline and the Pancake thrust (Perry, 1991; Carpenter et al., 1993; Perry and Dixon, 1993; Langrock, 1995), but their extent and relationship to the regional tectonic framework was unknown. The Sand Spring, Ike Spring, and Indian
Spring thrusts (named in this paper) and other contractional structures mapped in the region extend the known length of the central Nevada thrust belt to over 1500 km.
CHAPTER 2

GEOLOGIC BACKGROUND

Contractional Deformation

Several contractional events influenced the tectonic development of the central and eastern Great Basin since the Devonian. These events include the Late Devonian to Mississippian Antler orogeny, a late Pennsylvanian to Permian event, and the Cretaceous to Eocene Sevier orogeny (Speed and Sleep, 1982; Speed, 1983; Miller et al., 1992; Cashman et al., 2000).

The Devonian to Mississippian Antler orogeny was the earliest of these events in the vicinity of the study area. Basinal sedimentary rocks that formed in a back-arc setting or within an accretionary prism were thrust over passive margin Paleozoic sedimentary rocks (Figure 1) (Armstrong 1968; Speed, 1983; Miller et al., 1992).

Although the Antler orogeny accommodated a substantial amount of shortening, structures associated with it lie approximately 50 km west of the Pancake Range (Figure 1). It is, however, responsible for the formation of a foredeep basin represented by the Mississippian Chainman Shale, Diamond Peak Formation, and Joana Limestone (Miller et al., 1992).

The Late Pennsylvanian to Early Permian contractional event is represented by both deformed strata equivalent to the Ely Limestone and by a regional-scale
unconformity within Pennsylvanian/Permian Ely Limestone equivalent rocks to the north and west of the study area (Smith and Ketner, 1978; Cashman et al., 2000). Little is known about this event and Pennsylvanian rocks exposed within the study area appear to be unaffected by it. No Permian rocks are known to crop out within the study area.

The Mesozoic Sevier orogenic belt is a generally continuous belt of east-vergent contractional structures that extends from southern California and Nevada to Idaho, Montana, and Wyoming (Figure 1) (Armstrong, 1968; Speed, 1983; Speed et al., 1988; Allmendinger, 1992; Yonkee, 1992). Its major features include thin-skinned style, east-vergent folds and thrust faults. In many areas, the structurally highest thrust sheets are also the oldest, indicating that the belt generally formed by successively younger thrust faults sequentially from west to east, with stacking of older thrust sheets behind the youngest fault (Axen et al., 1990; Cowan and Bruhn, 1992; Decelles and Mitra, 1995). Basement involved thrust sheets are not common in the eastern portion of the belt, but are documented in the Wasatch Range in Utah, in southeastern California, and near the Idaho-Montana border (Allmendinger, 1992; Miller et al., 1992; Yonkee, 1992).

Central Nevada Thrust Belt

Contractional structures mapped in central and east-central Nevada delineate a significant hinterland fold and thrust belt, the central Nevada fold and thrust belt, bracketed between Permian and Cretaceous in age (Figures 1 and 3) (Armstrong, 1968; Speed, 1983; Speed et al., 1988; Bartley and Gleason, 1990; Perry, 1991; Perry and Dixon 1993; Armstrong and Bartley, 1993; Taylor et al., 1993; Fryxell, 1998; Taylor et
This belt incorporates folds and thrust faults with significant offset in at least nine ranges and is interpreted to be an internal branch within the Sevier hinterland, possibly representing an early contractional pulse in the Sevier orogeny (Taylor et al., 1993, 2000). These structures include east-vergent thrust faults; steeply dipping thrust ramps; large, open and upright to overturned folds; and local backthrusts (Taylor et al., 1993, 2000).

Speed (1983) described the Eureka belt, a fold and thrust belt that partly coincides with the central Nevada thrust belt. The Eureka belt contains east-vergent folds and thrusts (Speed, 1983; Speed et al., 1988). Since the original designation of the Eureka belt, the central Nevada thrust belt was defined as pre-Tertiary contractional structures east of the Fencemaker allochthon, including structures of the Eureka belt and additional structures not originally placed in the Eureka belt (Taylor et al., 1989) (Figure 1).

The structures exposed in the central Nevada thrust belt were assigned various tectonic significances in past interpretations. Armstrong (1968) interpreted many of the structures in this area as broad, gentle folds that accommodated only a small amount of shortening. Large stratigraphic separation faults were interpreted to have formed during Cenozoic extension. Speed (1983; Speed et al., 1988) described the hinterland structures as a north-south trending belt of folds and thrusts, the Eureka belt. These structures, interpreted as Mesozoic in age, were related to the Sevier orogenic belt via a large subsurface decollement (Speed et al., 1988). Bartley and Gleason (1990) restored the extension, placing the fold and thrust belts of the Eureka belt within 150 km of the Sevier thrust front, suggesting that contractional structures in central Nevada are in the western portion of the Sevier belt (Bartley and Gleason, 1990) (Figure 4). Taylor et al. (1993)
later reinterpreted some structures of the Eureka belt; added additional structures that together form a north-south trending suite of contractional structures that extends from Eureka to Alamo, Nevada; and named the belt the central Nevada thrust belt (Figure 1).

The direct correlation of central Nevada thrust belt structures to the north of the Grant Range, and especially across Railroad Valley, is complicated by the tectonic overprinting by Cenozoic extension and cover of Tertiary and Quaternary alluvium (Figure 2). However, Mesozoic-age folds and thrust faults have been mapped in the individual ranges (Cameron, 1986; Perry, 1991; Carpenter et al., 1993; Perry and Dixon, 1993). Langrock (1995) documented folds and thrust faults in the White Pine Range and related them in cross section to Mesozoic structures in the northern Pancake Range.

North of the study area, a large overturned syncline (the McClure Spring syncline) and at least 3 thrust faults (the Pancake, Duckwater, and Easy Ridge thrusts) were previously mapped (Quinlivan et al., 1974; Perry, 1991; Carpenter et al., 1993; Perry and Dixon, 1993).

Extensional deformation

The structural style as well as number the of episodes of differs temporally and spatially throughout the Great Basin (Taylor et al., 1989; Axen et al., 1993; Switzer and Taylor, 2001). Both high- and low-angle normal faulting have occurred. Normal faulting in the Pancake Range may have begun as early as late Cretaceous to mid Tertiary time, with much of the extension occurring between the late Eocene and Miocene (Taylor et al., 1989; Camilleri, 1992; Axen et al., 1993; Liberty et al., 1994; Langrock, 1995).
In the Basin and Range province near the Pancake Range, extension was episodic. Distinct, crosscutting normal fault sets are documented and the number of temporally distinct fault sets varies locally from two to four (Quinlivan et al., 1974; Taylor et al., 1989; Camilleri, 1992; Axen et al., 1993; Liberty et al., 1994; Langrock, 1995). An extensional episode is represented by a set of faults that have similar geometry and show consistent cross cutting relationships. These cross cutting relationships indicate that each set of faults occurred at different times. Extensional episodes occur within a protracted extensional period from the late Cretaceous to present time. To the east in the White Pine Range, Langrock (1995) distinguished four extensional events, a pre-volcanic episode and three sets of post-Oligocene faults (Figure 2). The pre-volcanic episode is indicated by normal faults overlapped by Tertiary volcanic rocks (Langrock, 1995). Post-Oligocene fault ages are determined by crosscutting relationships with rhyolitic ash-flow tuffs, such as the Stone Cabin Formation (Langrock, 1995). In contrast, Liberty et al. (1994) suggested that the rocks in the Pancake Range and Railroad Valley underwent two post-Oligocene phases of extensional faulting that dictate the general Basin and Range geometry. Although the model proposed by Liberty et al. (1994) addresses the geometry of the normal faults that created the basins, it does not take into account the existence of low-angle normal faults. Liberty et al.’s (1994) and Langrock’s (1995) interpretations may be compatible because local areas of extension may not continue across the entire basin. Indeed, Langrock (1995) suggests the possibility that faults in the White Pine Range may represent two different extensional styles. The older post-volcanic set of low-angle normal faults and the associated high-angle faults are followed by a high-angle set of basin-bounding structures.
CHAPTER 3

STRATIGRAPHY

The stratigraphy within the central Great Basin consists of a thick section of Paleozoic passive margin (miogeoclinal) sedimentary rocks overlain by Tertiary tuffs and flows of mainly intermediate to silicic composition (Figures 5, 6 and 7) (Kleinhaml and Ziony, 1985; Elrick, 1996; French and Schalla, 1998; Montgomery et al., 1999; Trexler and Cashman, personal comm., 2000).

Paleozoic passive margin rocks that crop out within the study area are described on Plate 1. Cambrian to Devonian strata not exposed within the study area that are shown at depth in cross sections are assumed from previous work (Nolan et al., 1956; Moores et al., 1968; Smith and Ketner, 1975; Hose et al., 1982; Kleinhaml and Ziony, 1985; Perry and Dixon, 1993; Elrick, 1996; French and Schalla, 1998; Montgomery et al., 1999) (Figure 5). Thicknesses for these units are compiled from Nolan et al. (1956), Smith and Ketner (1975), Kleinhaml and Ziony (1985), Elrick (1996), and Montgomery et al. (1999).

Unconformably overlying Paleozoic passive margin rocks is the Sheep Pass Formation, a Cretaceous to Eocene, laterally discontinuous and possibly syn-orogenic, conglomerate, lacustrine limestone, and siltstone (Winfrey, 1958; Fouch et al., 1979). The Sheep Pass Formation is recognized by a distinct red paleosol, previously described.
as a "Terra Rosa" by Perry and Dixon (1993) and Langrock (1995), overlying a poorly outcropping limestone clast conglomerate (Winfrey, 1958; Fouch et al., 1979; Perry and Dixon, 1993). Regionally, the Cretaceous Newark Canyon Formation underlies the Sheep Pass Formation, however this unit does not extend into the study area (Fouch et al., 1979; Vandervoort and Schmitt, 1990; Perry and Dixon, 1993).

Tertiary volcanic rocks unconformably overlie both Paleozoic and Eocene sedimentary rocks. Three units, the Lower rhyolite flow (Tl), the Stone Cabin Formation (Tsl, Tsm, Tss, Tsub, Tsu), and the Windous Butte Formation (Tw) crop out in the study area. These three units are separated by several laterally discontinuous lava flows and rhyolitic tuffs (Radke, 1992) (Figure 7; Plate 1). Radke (1992) reports a $^{40}$Ar/$^{39}$Ar age on sanidine of 35.34 ± 0.07 Ma for the middle member of the Stone Cabin Formation. The age of 31.3 Ma for the Windous Butte Formation is based upon K/Ar analysis of biotite phenocrysts (Phillips, 1989; Best et al., 1993).
CHAPTER 4

METHODS

Using standard geologic mapping techniques, detailed field mapping was performed at 1:24,000 scale (Plate 1, Figure 8). Mapping was completed on four U.S. Geological Survey topographic maps (Bradshaw Spring, Meteorite Crater, Portuguese Mountain, and Sand Spring) with the aid of 1:24,000 scale aerial photographs. Approximately 40 km² were mapped, including areas that overlap with maps produced by Perry and Dixon (1993) and Quinlivan et al. (1974).

Field map data were digitized, georeferenced, and integrated into ArcView GIS for analyses and data presentation. Separate overlays were created for each component of geologic data (i.e., faults, strike and dip data, rock units, and fold locations). Topographic USGS DRG (Digital Raster Graphic) files were utilized as a topographic base for geologic data. Four GIS extensions were used in conjunction with ARCview to manipulate DRG files, and place geology accurately. They are: ArcView GIS image analyst (scanning and digitization of base maps), DRG clipper 3.0 (removal of DRG collars and collar data), DRGtools by (color modification and edge matching of DRG files), and GEOclassifier (strike and dip data rotation and symbols) (Patterson, 1999; University of Kansas Structure and Tectonics GIS laboratory, 1999; Swigart, 2000).
Deformed state and retrodeformed cross-sections were constructed to constrain geometries of normal faults, oblique-slip faults, folds, and thrust faults. Both deformed-state and retrodeformed cross sections were constructed using standard cross section construction techniques (Dahlstrom, 1969; Woodward et al., 1985; Crane, 1987; Groshong, 1989; Nunns, 1991; White, 1992). Geophysical data from the Exxon Wildhorse unit #1 well (spudded in 1989) were used to constrain depth to units within the cross sections. Well data was provided by Riley Electric Logs, Inc. of Oklahoma City, Oklahoma.

2-D seismic data from Anschutz/Digicon line #106 were used to help constrain geometries and depths of structures within the central Pancake Range, as well as structures within Sand Spring Valley west of the study area. Seismic data were provided by Veritas Land Surveys of Houston, Texas.

Stereonet analyses were completed using GEORient 8.0 (Holcome, 2001). Stereonets were used to define geometric normal fault sets, compare contractional fold orientations, and compare thrust fault and fold geometries within the study area. The three point method was employed to delineate normal and thrust fault geometries used in stereonet analyses and cross-section construction, unless orientations were measured in the field.
CHAPTER 5

SEISMIC ANALYSIS

Seismic data in the form of Anschutz/Digicon Line 106 were provided courtesy of Veritas Land Surveys, Houston, Texas and was used to constrain units and structural geology within Big Sand Springs Valley (Figures 9, 10 and 11). Line 106 was previously published by Liberty et al. (1994) and crosses a part of the Pancake Range mapped by Quinlivian et al. (1974). Cross section b-b’ by Quinlivian et al. (1974) appears to run parallel to and over the trace of Line 106, and therefore identification of Paleozoic units within the seismic section were inferred from that cross section and map.

The data was processed using standard processing techniques with the addition of FK (Frequency-wavenumber) power spectrum filters to the raw and refraction corrected CMP (Common Midpoint) gathers. FK power filters are designed to remove steeply dipping noise (i.e. incoherent shot noise or groundroll). These processes should preserve most steeply dipping data (e.g., faults or steeply dipping units), but fault surfaces or other moderate to steeply dipping reflectors may be lost due to the smoothing effects of the process. These processes were considered during interpretation of the data.

West-dipping reflectors are discernable between receiver stations 2780 and 2720 and coincide with Paleozoic strata that crop out at the surface in that location, near Wood Canyon (Quinlivian et al., 1974) (Figures 9 and 11). The depth of these reflectors was calculated using the RMS (Root Mean Squared) velocities at the top of the section. A

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strong reflection occurs at 3600 m. Wavelet breaks appear to be compressional (positive) within the section, indicating low over high reflection coefficients. These data support the interpretation of soft shale over harder limestone, therefore the 3600 m reflection is interpreted to be the contact between the Cambrian Dunderberg Shale and the Cambrian Lincoln Peak Formation.

Big Sand Springs Valley lies west of receiver station 2780 (Figures 9 and 11). Coherent reflections in this region are interpreted to be either contacts between valley fill and volcanic units and/or contacts between individual volcanic units (Quinlivan et al., 1974; Liberty et al., 1994; French and Schalla, 1998). The transition between outcropping Paleozoic units east of receiver station 2780 and volcanic rocks and basin fill west of receiver station 2780 appears to be a series of two or perhaps three steeply west-dipping normal faults. Coherent reflections deeper than 2 seconds within the basin are difficult to discern.

The seismic data constrained the normal fault geometries and locations of normal faults within Big Sand Springs Valley, which further helped in cross-section construction. Paleozoic strata and the geometry of those strata in relation to the Sand Spring thrust fault were not imaged within the section.
CHAPTER 6

STRUCTURAL DESCRIPTIONS

Up to four sets of both high- and low-angle normal faults as well as thrust faults and folds are exposed in different portions of the study area. The study area is divided into 5 structural domains for the purpose of describing the spatial relationships among these structures (Figure 12). Contained within domain 5 is a north-south trending ~3.6 km long ridge composed of mainly Devonian Guillmette Formation that, for the purposes of describing the locations of structures, will be referred to as the “central ridge” (Figure 12; Plate 1). Another landmark used for describing the location of structures is Portuguese Mountain, which lies mostly within domain 4 (Figure 12; Plate 1).

Extensional faults

Four sets of normal and oblique slip faults are exposed. These faults were identified based on offset units, fault surfaces, and/or the presence of breccia and slickenlines. Faults can be grouped into sets based on their cross-cutting relationships with strata as well as their geometric similarities to each other. These fault sets include one pre-volcanic or pre 35.34 ± 0.07 Ma set, here named set A, one syn-volcanic set (set
B), and two post-volcanic or post-35.34± 0.07 Ma sets. The post-volcanic fault sets are named C and D in chronological order (Figure 12; Plate 1). Volcanic unit ages are from Radke (1992).

Nine pre-volcanic faults (fault set A) cut Paleozoic strata and overlying Late Eocene and younger volcanic rocks in domains 3, 4, and 5 (Figure 12). Fault set A includes some faults that are not exposed near Tertiary units, but are included in this set based upon geometric similarities to faults directly overlapped by Tertiary volcanic strata. Strikes within this set range from N39°E to N18°W (Figure 12). Dips range from 11° to 80° E and 11° to 68° W. A low-angle fault included in fault set A is exposed east of the central ridge (domain 5), cutting the Devonian Guillmette Formation. This low-angle fault is parallel to bedding in its easternmost outcrop.

Set A faults are cut by E-W striking normal and oblique-slip faults (fault set B) in domains 2 and 5 north of the central ridge (Figure 12; Plate 1). Also, set A faults do not cut the Sheep Pass Formation nor any of the Tertiary or younger units. For these reasons, set A faults must have formed prior to other normal faults and must be older than 35.34 ± 0.07 Ma, the age of the oldest dated volcanic unit that overlaps set A faults.

Stratigraphic separations along set A faults range from 33 m to 1530 m. The total amount of horizontal extension for set A faults is at least 1721 m in cross section B-B' and approximately 1363 m across cross section A-A' (Plates 2 and 3).

Seven E-W striking faults compose fault set B. These faults generally have E-W strikes between N54°W and N43°E (Figure 12). Steep dips typify these faults; dips range from 48° to 73° E and 53° to 78° W. Rakes of slickenlines measured on faults exposed
on the southern flank of Portuguese Mountain range from 78°, S30°W to 83°, S29°W. Units juxtaposed by these faults suggest apparent normal and apparent left slip.

Fault set B is the only set that formed syn-volcanically, after the deposition of Oligocene rhyolite flows and the Stone Cabin Formation but before the deposition of the Windous Butte Formation. This relationship can be best observed in domains 2 and 3 (Figures 8 and 12; Plate 1). In domain 2, set B faults cut members of the Stone Cabin Formation. Within domain 3, a set B fault cuts the Devonian Guillmette Formation and is overlapped by an unnamed andesite lava and the Windous Butte Formation. Therefore, set B faulting began after 35.34 Ma and ended by 31.3 Ma.

Cross-cutting relationships among faults indicate set B was the second extensional fault set. Near the top of Portuguese Mountain in domain 4, a set B fault offsets a N-S striking fault of set A. Set B faults appear to truncate faults of set C in some areas and are cut by a fault of set C along the same fault in domain 2 (Figure 13). This relationship between faults of set B and set C implies reactivation of at least some set B faults during fault set C time.

One set B fault in the southeast corner of domain 4 was originally designated as a thrust fault by Quinlivian et al. (1974) (Figures 8 and 12; Plate 1). However, this fault is exposed at the surface and has a strike and dip of N77°E, 70°SE. Rakes of slickenlines were measured along this fault plane and had a plunge and trend of 73°, N77°E. This fault also has Mississippian Chainman Shale in its hangingwall and Devonian Guillmette Formation in its footwall. Given these data, this fault is a normal fault and not a thrust fault.
Total stratigraphic separations across set B faults range from 242 m to 1273 m. North-south directed extension for set B faults is at least 333 m in cross section D-D' and 606 m in cross section C-C'. Set B faults also have a component of east-west directed extension and accommodate up to 121 m of horizontal extension in cross section B-B'. East-west directed extension along set B faults occurs because set B faults in cross section B-B' do not strike exactly east-west. Amounts of horizontal extension appear to increase toward the west on set B faults.

The 35 faults of set C are exposed in domains 2 and 3. These faults strike both NW and NE (Figures 8 and 12; Plate 1). Northwest strikes are between N18°W and N52°W, and northeast strikes are between N10°E and N54°E. Set C faults dip between 78° to 26° E and 28° to 68° W. These NW- and NE-striking faults are not a conjugate set, because conjugate normal faults have sub-parallel strikes (e.g., Anderson, 1951). Instead, NW striking set C faults appear to be cut by other faults within the same set, suggesting that these faults formed synchronously and represent a three dimensional (3-D) strain field. All NW striking faults are associated with three dimensional strain accommodation in set C (See Chapter 7).

Fault set C formed after 31.3 Ma but before set D faults. Within domain 3, faults of set C cut the 31.3 Ma Windous Butte Formation and are cut by N-S striking faults of set D (Figures 8 and 12 Plate 1). Also within domain 3, a set D fault terminates at a set C fault, which implies that set C faults must have been in place to provide a barrier to further fault propagation (Figure 14; Plate 1). No set C or D faults offset units younger than Oligocene.
Stratigraphic separations across set C faults range from 33 m to 1455 m. Total east-west directed horizontal extension is at least 455 m across cross section B-B' and 2848 m across cross section A-A'.

Fault set D contains nine north-south striking faults that are exposed in domain 3 and cut the youngest volcanic unit, the 31.3 Ma Windous Butte Formation. Set D faults strike between N66°W and N55°E and dip steeply, between 45° to 87° E and 65° to 85° W. Set D faults also cut faults of set C in domain 3.

Stratigraphic separations for set D faults range from 121 m to 1576 m. Set D faults also accommodate at least 1182 m of horizontal extension in cross section B-B' (Plate 3).

Set D faults cut both set C faults and the youngest volcanic unit. However, no set D faults cut Tertiary or Quaternary alluvium. This suggests that set D faults are younger than 31.3 Ma. Therefore, this episode of extension must predate the most recent Basin and Range extension because Quaternary faults, including active faults, are known to cut alluvium in Railroad Valley (Dohrenwend et al., 1991; Liberty et al., 1994; Langrock, 1995; Williams 2000).

Thrust faults and folds

Contractional structures are exposed deforming Paleozoic strata. No folds or thrust faults are exposed within Cenozoic units. Thrust faults and folds occur at both outcrop and map scale. Tertiary units are exposed overlying an angular unconformity with units in the western limb of the McClure Spring syncline (Plates 1 and 2).
A thrust fault exposed along the eastern margin of Big Sand Springs Valley in the west side of the study area is here named the Sand Spring thrust (Figures 8 and 12; Plate 1). The Sand Spring thrust places Devonian Guillmette Formation over Mississippian Chainman Shale and has a strike and dip of N23°E, 46°W in domain 1 (Figures 8 and 12; Plate 1). In domain 1, the thrust fault is pervaded by a north-south trending outcrop of jasperoid. Though covered by alluvium in the west-central study area, the Sand Spring thrust is also exposed just north and west of Portuguese Mountain in domain 4 and has a calculated strike and dip of N27°E, 38°W (Figures 8 and 12; Plate 1). The jasperoid in domain 1 is not exposed in the fault zone in domain 4. Stratigraphic separation across the Sand Springs thrust fault is up to 1272 m.

In domain 4, another thrust fault is exposed and here named the Ike Spring thrust (Figures 8 and 12; Plate 1). The Ike Spring thrust is exposed at the base of the eastern flank of Portuguese Mountain where it is dissected by east-west striking normal faults of set B (Figures 8 and 12; Plate 1). This fault places Devonian Guillmette Formation over poorly outcropping Mississippian Joana Limestone and Mississippian Chainman Shale. It has a calculated strike and dip of N5°E, 11°W. Stratigraphic separation across the Ike Spring thrust is up to 727 m (Plate 5).

In domain 5, a thrust fault that crops out on the southeastern flank of the central ridge is here named the Indian Spring thrust. The Indian Spring thrust fault places Devonian Guillmette Formation over Mississippian Chainman Shale (Figures 8 and 12; Plate 1). It has a calculated strike and dip of N32°E, 24°W. Stratigraphic separation across the Indian Spring thrust is up to 363 m (Plate 5).
Two thrust klippe are exposed on the central ridge in domain 5 (Figures 8 and 12; Plate 1). These klippe place Devonian Guillmette Formation over Devonian Guillmette Formation. The thrust fault that underlies both of these klippe is interpreted to be the same fault. This thrust fault has a measured strike and dip of N12°W, 29°W.

Two large folds are exposed within the study area. The McClure Spring syncline and a large anticline on the central ridge (Figure 15; Plates 1 and 2). The central part of the McClure Spring syncline is exposed in domain 5 and is heavily dissected by multiple high-angle normal faults of sets A and C (Figure 12). It is a steeply inclined, east vergent, open, gently north-plunging syncline. The plunge and trend of the fold axis are 23°, N2°E (Figure 15). The large anticline on the central ridge in domain 5 appears to be in the footwall of the previously described thrust klippe and within the Indian Spring thrust plate (Figures 8 and 12; Plate 1). This anticline is steeply inclined, east-vergent, open, and gently north plunging with the fold axis oriented 8°, N1°E (Figure 15).

Several small outcrop scale, possibly parasitic folds deform Paleozoic strata in domain 5 on the central ridge (Figure 12). An outcrop scale anticline is exposed on the western flank of the central ridge in domain 5. It is steeply inclined, open, gentle, and gently southeast-plunging with a fold axis oriented 2°, S62°E. This outcrop scale anticline appears to be cut by the above mentioned thrust klippe. Another set of folds is exposed in domain 2 (Figures 8 and 12; Plate 1). In outcrop, these folds deform the Lower Pennsylvanian Ely Limestone. They are all anticlines and are open, steeply inclined, and plunge gently north and south; 8°, N15°E and 5°, S5°E.

An outcrop scale anticline-syncline pair is exposed in domain 3. These two folds deform beds within the Guillmette Formation and are steeply inclined, open, and
southwest-plunging. The plunges and trends of the fold axes are $12^\circ$, S$38^\circ$W for the anticline and $12^\circ$, S$31^\circ$W for the syncline.
CHAPTER 7

DISCUSSION

Several episodes of normal faulting occurred within the Great Basin: the number of episodes varies locally (Moores et al., 1968; Camilleri, 1992; Langrock, 1995; Mueller et al., 1999; Williams, 2000; Gans et al., 2001). Within the areas immediately surrounding the study area, the number of faulting episodes varies between two and four (e.g., Liberty et al., 1994; Langrock, 1995). Within the study area, these normal fault sets denude central Nevada thrust belt structures. Therefore, understanding the geometry and tectonic development of Cenozoic normal faults is necessary to unravel the structural and tectonic development of central Nevada thrust belt structures and to determine whether major changes in stress field orientation occurred during the Cenozoic.

Onset of Extension

The age of faults in set A, the oldest extensional set, are poorly constrained between the Pennsylvanian Ely Limestone, which is cut, and the ~35 Ma Tertiary volcanic rocks that overlap the faults. The Late Cretaceous to Eocene Sheep Pass Formation crops out within the study area (Winfrey, 1960; Kellogg, 1964; Vandervoot and Schmitt, 1990). However, a lack of proximity between exposures of Sheep Pass
Formation and set A faults and the lateral discontinuity of the Sheep Pass Formation renders difficult any interpretations about the onset of normal faulting in set A relative to the timing of deposition of the Sheep Pass Formation. Locally (e.g. Egan Range – type section), sediments of the Sheep Pass Formation accumulated in a basin controlled by a Late Cretaceous to Eocene normal fault (Kellogg, 1964) (Figure 2). Furthermore, late Cretaceous extension that resulted from the extensional collapse of the Sevier hinterland has been well documented in northern Nevada and northwestern Utah (Wells et al., 1990; Livaccari, 1991; Wells, 1997). Therefore, a late Cretaceous to Eocene age of initial extension is possible.

In contrast to a late Cretaceous age of extension, some workers (e.g., Gans et al., 2001) suggest that the onset of normal faulting at the latitude of the study area occurred in late Eocene time and that no significant extension near this latitude began then. Indeed, Taylor et al. (1989) concluded that minor late Cretaceous extension occurred at this latitude, but that significant extension did not begin until the late Eocene or early Oligocene. Together, these interpretations suggest that normal faulting within the study area may have initiated as early as late Cretaceous or as late as late Eocene.

Transverse Faults

Cross section construction and geometric analysis of set B faults allow interpretations to be made about their geometry and kinematics. No linear features were observed offset by set B faults, but oblique slip motion is geometrically required to explain the outcrop patterns and to construct viable cross sections. Therefore,
interpretations regarding set B faults should be considered to be estimates of motion along these faults.

Set B faults are normal in domain 4 and on the southern flank of the central ridge in domain 5 because slickenline data indicate dip slip (see chapter 6). Additionally, oblique slip along these southern faults is not required to construct a viable cross section (Plate 4). However, two transverse faults, X and Y, north of the central ridge in domain 5 are interpreted to be normal left-lateral oblique-slip faults (Figures 12 and 13; Plates 1, 2, and 4). Map relations between offset units, in addition to cross section analysis, suggest that the southern fault, Y, accommodates a minimum of 454 m of left-lateral motion and 1363 m of normal dip slip. The northern transverse fault, X, has a minimum of 300 m of left-lateral slip and 1600 m of normal dip slip. The cumulative motion on these faults is a minimum of 754 m of left-lateral slip.

Cross sections C-C' and D-D' are nearly normal to the transport direction of all set B faults. Therefore, the majority of fault bounded blocks moved in and out of the plane of section. This means that the cross sections should not and can't be retrodeformed in the plane of section. In the central portion of cross section C-C', faults X and Y place the footwall of the Indian Spring thrust fault (the western limb of the McClure Spring syncline) on the north next to the hanging wall of the Indian Spring thrust fault on the south (Plate 3). Oblique-slip faults X and Y (cross section C-C') place Guillmette Formation, Joana Formation, and Chainman Shale from the Indian Spring thrust plate, next to Middle Devonian and older units in the Indian Spring thrust footwall.
Other set B transverse faults are exposed in domains 4 and 5 south of Portuguese Mountain and the central ridge. These faults have a significant amount of normal offset with a minor component of left-lateral movement based on stratigraphic offset of east dipping Paleozoic units and slickenlines (Plate 3).

3-D strain

Set C faults strike both NE and NW in domain 2 (Figures 8, 12, and 17; Plate 1). These complex fault patterns are commonly interpreted to arise from multiple episodes of faulting, because classical fault theory (plane strain) suggests that faults in a single set have parallel strikes (e.g., Anderson, 1951). However, more recent workers suggested that multiple fault orientations, called polygonal or orthorhombic fault sets, can form in a single episode of normal faulting (Reches and Dieterich, 1983; Krantz, 1989; Langrock, 1995; Neito-Samaniego and Alaniz-Alvarez, 1997; Lonergan and Cartwright, 1999). These polygonal fault patterns form in three dimensional (3-D) strain rather than in 2-D (plane) strain in which conjugate fault sets occur (Krantz, 1989). In a 3-D strain field, strain can be accommodated by extension in several directions simultaneously. Conversely, in a 2-D or plane strain field, strain is accommodated in one direction only (parallel to the minimum principal stress direction in extended regions).

Polygonal fault sets can form within a given domain in several ways. These include, but are not limited to, compaction of sediments during lithification, fracture along preexisting planes of weakness within rock units, and juxtaposition of differently shaped blocks across low-angle normal faults (Reches and Dieterich, 1983; Krantz, 1989;

A model suggested by Langrock (1995) and Taylor and Novak (in review) shows that irregular, corrugated, or non-planar faults can't simply translate upper-plate rocks, but they induce a 3-D strain field within the hangingwall. Langrock (1995) showed that orthorhombic fault patterns, characteristic of 3-D strain, existed in the hangingwall of a non-planar low-angle normal fault. Specifically, the orthorhombic faults lie above a fault corrugation that changes shape in the direction of transport. The mismatch in shape of the upper and lower plates required 3-D strain.

Similarly, a set C fault, labeled L, has several other set C faults in its hanging wall (Figure 17a). The upper plate faults labeled M, N, and O represent 3-D strain because they display a polygonal fault geometry (Figure 17a). Faults M, N, and O are interpreted to have moved synchronously with fault L based on the cross-cutting relationships among these faults. Three point calculation of the attitude of fault L shows that the dip of the fault changes along strike from 26°NW along the southern portion to 50°NW along the northern exposure of the fault (Figure 17a; Plate 1). These two sections connect across a primary fault bend. Faults M, N, and O lie above the change in dip or bend in fault L. Thus, it appears that motion down the bend necessitated 3-D strain in the hanging wall. Smaller corrugations may exist along fault L that also may have affected hanging wall faults. However, the determination of which bends within fault L caused upper-plate deformation would require linear kinematic data, which were unavailable.
In addition to fault L, another set of faults, shown in figure 17b, appear to have formed in 3-D strain (Figures 8, 12, and 17b). Faults in figure 17b display a polygonal geometry, however no single corrugated fault exists to explain the orthorhombic fault patterns. Therefore, the corrugated fault model does not explain the polygonal faults in this area (Figure 17b).

A model that would explain the fault patterns in Figure 17b was proposed by Neito-Samaniego and Alaniz-Alvarez (1997). This model suggests that polygonal fault patterns form in 2-D (plane) strain where strain is accommodated on pre-existing, non-interacting planes of weakness in the rocks, rather than by forming new fractures. These pre-existing planes might be joint sets, other fractures, or areas of weakness that would be preferred places for faults to break. The faults in Figure 17b do not strike symmetrically around a vertical axis, which may suggest that older fractures within the volcanic rocks may have caused a less systematic distribution of faults within this region.

Rotation of the stress field through time

The four temporally and geometrically distinct normal fault sets record a change in the orientation of the regional stress field with time from; (1) east-west extension and minimum principal stress direction to (2) north-south directed extension and minimum principal stress direction, and eventually back to (3) east-west extension and minimum principal stress direction. Fault strike is not a sensitive or exact indicator of principal stress direction, but large changes in strike between fault sets suggest a regional change of stress directions through time (Taylor and Switzer, 2001). This change in the orientation of the principal stresses may be related to differences in the causes (dynamics)
of extension at different times. Late Mesozoic and Cenozoic extension within the Basin and Range province was controlled by a combination of extensional collapse of the Sevier hinterland, plate boundary forces, and a southward sweep of Tertiary volcanism through time (Bartley, 1989; Wells et al., 1990; Axen et al., 1993; Taylor and Switzer, 2001).

Changes in normal fault orientation may correspond with the southward migration of Tertiary volcanism (the “ignimbrite flareup”) through the Great Basin from Late-Eocene to Miocene time (Best, 1988; Bartley, 1989; Best and Christensen, 1991; Taylor and Switzer, 2001). The ignimbrite flareup is represented by generally WNW trending belts of volcanic centers that young southward. In addition, syn-volcanic extension in parts of the Great Basin is documented along east-west striking normal faults (Bartley, 1989; Best and Christensen, 1991; Overtoom and Bartley, 1996; Taylor and Switzer, 2001).

Prior to the Tertiary ignimbrite flareup, extension in the Great Basin region was controlled by a combination of plate boundary forces and extensional collapse of the Sevier hinterland (Bartley, 1989; Wells et al., 1990; Best and Christensen, 1991; Taylor and Switzer, 2001). When the magmatism migrated into an area, it added a thermally controlled stress field with a N-S directed minimum principal stress direction to the already existing stress field associated with E-W directed extension (Bartley, 1989). The ignimbrite flareup occurred in belts of volcanic centers oriented WNW. Thus, stresses associated with active spreading were directed in a NNE direction (Bartley, 1989). When these stresses were significantly increased relative to the initial stresses promoting E-W extension, extension occurred in a N-S direction. When the belts of volcanic centers
migrated south past a given point, the original E-W directed extension was restored (Bartley, 1989; Taylor and Switzer, 2001).

Syn-volcanic fault sets within the study area strike east-west, nearly normal to faults of other extensional episodes within the study area, and accommodate a minimum of 545 m north-south directed extension in cross sections C-C' and D-D'. Additionally, the timing of east-west striking normal faults corresponds with the timing of rhyolitic volcanism in the region. Therefore, east-west striking faults in set B are interpreted to have formed in a similar manner to other east-west striking normal faults within the Great Basin (e.g., Taylor and Switzer, 2001). The control on the stress regimes switched from plate boundary forces during Late Cretaceous and/or Eocene time to a local southward-sweeping magmatic-volcanic belt during the Oligocene.

Another possible explanation for the syn-volcanic transverse faults within the study area, is that they may be directly related to an appropriate aged caldera, either as radial faults fanning away from a caldera or as concentric ring faults around a caldera. However, this interpretation is not favored for several reasons. Transverse faults within the study area are sub-parallel and do not display a divergent pattern away from, nor are they concentrically curved around a vertical axis. Though several workers suggest a close proximity for a caldera that was the source for the Stone Cabin Formation, transverse faults within the study area are at oblique angles to the proposed location (Radke, 1992; Best et al., 1993) (Figure 2). The proposed location of the source caldera is in close proximity to the central Pancake Range; however, in the study area, no stratigraphic indicators for a caldera are present. For example, dramatic thicknesses of exposed tuffs and intercalated breccias are not present. Therefore, the caldera is not
extremely close. Radke (1992) favors an interpretation of a caldera in northern Railroad Valley as a more likely location. Likewise, caldera related faults would also be located further north and east of the study area, which is consistent with the map data.

Oblique-Slip Transverse Faults and the Prichards Station Lineament

The geometry and kinematics of two set B faults, X and Y, resemble those of the Currant Summit fault (Figure 16) (Williams, 2000). The Currant Summit fault and set B faults X and Y lie along the Prichards Station Lineament (Ekren et al., 1976). Lineaments in the Great Basin (including the Prichards Station Lineament) consist of zones of east-west trending basins and/or ridges and geophysical anomalies that commonly are associated with east-west striking, or transverse, faults (Ekren et al., 1976). These transverse faults are commonly thought to have strike-slip or oblique slip components and can separate regions with different amounts of extension (Faulds and Varga, 1998; Williams, 2000).

The Currant Summit fault separates strata of the Horse Range from rocks of the White Pine Range (Figure 16). This normal-left oblique slip fault has a strike parallel offset of 2700 m (Williams, 2000). Williams (2000) demonstrated that the Currant Summit fault is a barrier to fault propagation. This fault separates a more highly extended region in the White Pine Range in the north from a less extended region to the south. Because faults do not propagate across the Currant Summit Fault normal slip from N/S striking normal faults in the White Pine Range is accommodated by strike-slip motion on the Currant Summit Fault (Williams, 2000).
Though set B faults X and Y accommodate less offset than the Currant Summit fault, all three faults are kinematically similar. Set C faults terminate at set B faults X and Y, suggesting that faults X and Y are barriers to fault propagation. In addition, these faults may separate a more highly extended northern portion of the study area from a less extended area to the south, because total amounts of extension on set C faults are greater on cross section A-A' (mainly north of faults X and Y) than on cross section B-B' (south of faults X and Y). These observations support the speculative interpretation that faults X and Y have a significant amount of strike-parallel offset relative to their net slip as shown in cross sections A-A', C-C', and D-D', despite the lack of piercing points across the faults (Plates 3 and 4).

Faults X and Y appear to have been active during a different time interval than the Currant Summit fault. At least some of the normal-left oblique motion along the Currant Summit Fault occurred in Quaternary time, but initial slip on the Currant Summit Fault may have begun as early as Oligocene. Within the study area, the transverse faults of set B that lie along the Prichards Station Lineament are cut by later sets (C and D) and are overlapped by Oligocene rocks, implying that left-lateral motion along these faults has been inactive since the mid-Oligocene. Other left-lateral oblique slip faults mapped approximately 11 km south of the study area in the Pancake Lineament were also active during Oligocene time, and show no Quaternary offset (Ekren et al., 1976) (Figure 16).

The data of Ekren et al. (1976), Williams (2000), and this study suggest that faults in the Prichards Station Lineament share similar kinematic styles along strike. These faults may separate areas with different amounts of extension, and faults within this lineament can be active during different times (Taylor et al., 1999).
Regional Extension

Construction of regional cross section B*-B”

Using data from this study and maps of the Horse and White Pine Ranges by Moores et al. (1968), the eastern portion of the Pancake Range by Radke (1992), and the south central Horse Range by J.E. Fryxell (unpublished map data), a regional cross section was constructed (Plate 5). Cross section B*-B” is line-length balanced and restorable to within a line-length error of approximately 5% (Plate 5). Structures are not constructed to show minimum offset, rather they reflect the realistic and geometrically viable amounts of offset on all faults.

Data from this study and the above listed previous workers constrain the geometry of structures within the Pancake and Horse ranges, however limited data were available for Railroad Valley. Therefore, the structure of Railroad Valley has been simplified. The single fault on the west side of Railroad Valley represents several normal faults that may lie along the western edge of the valley. These faults were active both prior to and after volcanism within the region (Liberty et al., 1994; Langrock, 1995). The Ragged Ridge fault mapped by Moores et al. (1968) is interpreted to be a low angle normal fault that was active prior to range bounding faults on the east side of Railroad Valley. This interpretation is partially based on a model for the development of the Horse Camp Basin by Horton and Schmitt (1998).

Episodes of extension

Different episodes of extension occurred during similar time periods (Late Cretaceous to present day) in the Great Basin, but there appears to be no regionally
consistent model for the number of extensional episodes nor the structural style of the
extensional faults (Taylor et al., 1989; Taylor, 1990; Lund et al., 1993; Liberty et al.,
1994; Langrock, 1995; Lewis et al., 1999; Miller et al., 1999; Mueller et al., 1999;
Williams 2000). For example, Taylor et al. (1989) demonstrated up to four episodes of
both pre- and post-volcanic normal faults mainly occurring between late Eocene and
Holocene time between the Grant Range and the Ely Springs Range, approximately 100
km southeast of the study area (Figure 2). In contrast, Liberty et al. (1994) document
only two periods of significant extension in the Pancake Range approximately 15-20 km
south of the study area. Gans et al. (2001) reported one period of significant extension
during the same time period in the central Egan Range 80 km east of the study area. Two
extensional periods are reported approximately 150 km northeast of the study area in the
Snake Range (Lewis et al., 1999; Miller et al., 1999). Therefore, the number of fault
episodes, timing, and their relationship to volcanism cannot be universally applied to all
data sets within this region (Wernicke, 1992; Gans et al., 2001).

None of the fault sets in the study area offset units younger than Oligocene.
However, Holocene faults do exist within Railroad Valley to the east and within Big
Sand Springs Valley to the west (Moores et al., 1968; Liberty et al., 1994; Langrock,
1995; Williams 2000) (Figures 2 and 16). Therefore, despite evidence for only four
episodes of extension within the study area, five episodes of normal faulting may typify
the central Pancake Range and the surrounding valleys, if basin bounding faults represent
a separate episode of faulting.

During the five periods of faulting, approximately 18.5 km of horizontal
extension was accommodated along the 65.5 km length of cross section B*-B”.

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amount of extension is comparable to Langrock's (1995) calculation of approximately 13.8 km of horizontal extension in a regional cross section of similar length approximately 32 km north of cross section B*-B**.

Plate boundary conditions (i.e., variations in movements along the transform plate margin) appear to control periods of east-west extension within the Basin and Range Province (Dickinson and Wernicke, 1997). Extension after 31.3 Ma that occurs across sets C and D faults may relate to divergent motion during the development and evolution of the San Andreas transform margin. Additionally, the strikes of sets C and D faults are approximately normal to the minimum principal stress direction for less than 10 Ma extensional stress patterns in the Great Basin (Zoback, 1989; Dixon et al., 1995). However, the strikes of these faults have a significant range, which may reflect pre-existing zones of weakness or lateral variations within lithospheric structure (Zoback, 1989). Therefore, plate boundary conditions and resulting stress patterns in the Great Basin generally are consistent with the strikes of sets C and D faults.

Contraction

Three previously unmapped thrust faults, associated folds and two thrust klippe were delineated within the study area: the Sand Spring thrust, the Ike Spring thrust, and the Indian Spring thrust. In addition to thrust faults, the McClure Spring syncline previously mapped by Kleinhampl and Ziony (1985), Perry (1991), and Perry and Dixon (1993) is geometrically required to exist within the study area (Plate 3). These structures can be used to link known central Nevada thrust belt structures in the north to structures in the south. Linking these structures along strike is important to understanding the
tectonic development of the Mesozoic central Nevada thrust belt as a whole. The Mesozoic thrust faults and folds are described in chronological order below.

Ike Spring and Indian Spring thrust faults

The Ike Spring and the Indian Spring thrust faults share a similar geometry (Plates 2 and 3). The orientation of Paleozoic units along with measurements of fault planes indicate that these faults have steeply dipping ramps near the present surface with a curved geometry at depth (Plates 2 and 3). The steep thrust ramps are also geometrically required to preserve constant offset along the trace of the thrust faults at depth. In a line length balanced cross section, such as cross section B-B’, constant offset along all faults is required to maintain constant line lengths between loose lines (Dahlstrom, 1969).

The Ike Spring and Indian Spring thrust faults are interpreted to sole into the base of the Ordovician Ninemile Formation. The Ninemile Formation is the middle unit of the Pogonip Group. It consists of argillaceous shale, calcareous shale, siltstone, and shaley and silty limestone (Kleinhampl and Ziony, 1985). The Ninemile Formation is mechanically weak, and therefore it provides an ideal slip surface for fault propagation.

The angle between the thrust ramps and footwall strata are interpreted to be the maximum dip of the thrust faults at the end of time of fault motion (Plates 2 and 3). The ramp of the Indian Springs thrust cuts footwall strata near the surface at an angle of almost 70°. The Ike Spring thrust cuts footwall strata at an angle of 80° near the present surface. These angles appear unusually steep, even for hinterland thrusts. However, these thrusts are deformed by the McClure Spring syncline. To estimate the pre-folding dip of the thrusts, the McClure Spring syncline was retrodeformed in cross section B-B’ using a flexural slip folding model. From this, the pre-folding dips of the two thrust
faults can be estimated. It was assumed that flexural slip occurred parallel to bedding in the limbs of the syncline only up to the footwall cut off of each thrust fault (Plates 3e and 3f). Line lengths were measured along unit contacts from the trace of the axial surface of the McClure Spring syncline to each thrust fault trace. A line was then drawn between the tips of the two unfolded line lengths. This line represents the pre-folding configuration of the thrust faults. From this line, the angle of the thrust faults relative to the footwall beds can be calculated. Assuming that the Paleozoic strata were flat lying prior to deformation along the thrust faults, the Indian and Ike Springs thrust faults had minimum pre-folding dips of 39° and 55° near the present day surface, respectively.

Within the study area, the Ike Springs and Indian Springs thrust probably do not coalesce into a single fault strand along strike (Plate 2). To the north, these thrust faults must remain separate strands separated by a normal fault to explain the thickness of the Chainman Shale exposed at the surface in domain 1 (Plates 1 and 2).

An anticline exposed on the central ridge is within the hanging wall of the Indian Spring thrust fault (Figure 8, Plates 1, 2, and 3). This fold may be a fault propagation anticline because bedding in the western limb of the fold is not parallel to the Indian Spring thrust ramp.

Also within the hanging wall of the Indian Spring thrust fault are two thrust klippe that cut the fault propagation anticline on the central ridge. These klippe are situated between the Ike Spring and the Indian Spring thrust faults. This location and the fact that these klippe cut a fault propagation fold suggests that this deformation formed out of sequence with movement on the Indian Spring thrust fault. Because these thrust klippe do not show significant offset, they may represent internal deformation within the
Guillmette Formation. These klippe formed during thrusting or perhaps during the formation of the McClure Spring syncline.

McClure Spring syncline

The McClure Spring syncline is an east-vergent syncline that crops out for approximately 18 km along trend just north of the study area (Perry and Dixon, 1993). Perry and Dixon (1993) suggested that stratigraphic and structural relationships in the study area could be explained by a low-angle normal fault. This low-angle normal fault was not observed. In contrast, this study shows that the McClure Spring syncline crops out within the northern and east-central portion of the study area and is geometrically required in the subsurface throughout much of the area (Plates 2 and 3). Radke (1992) showed that Paleozoic rocks dip west in the Pancake Range. In the study area, Paleozoic rocks dip east (Plate 1). The simplest interpretation for this relationship is a syncline; other interpretations would require the existence of curved faults with significant offset where no faults are exposed. This observation expands the known length of the fold to 24 km.

In the region north of the study area, the McClure Spring syncline is a recumbent fold (Perry and Dixon, 1993). Within the study area, however, the syncline is a steeply inclined, open fold. This suggests that the fold dies out to the south of the study area. Alternatively, the McClure Spring syncline may continue to the south, underneath Railroad Valley.

The McClure Spring syncline lies in the footwall of the Sand Spring thrust fault and dips of the Mississippian Chainman Shale in the western limb of the syncline.
increase in proximity to the Sand Spring thrust fault (Plate 1). These observations are consistent with a genetic relationship between a thrust fault and a footwall fold.

**Identification and Correlation of the Sand Spring Thrust**

The Sand Spring thrust, the westernmost thrust, was previously suggested to be a normal fault by Perry and Dixon (1993) and later portrayed as a west-dipping normal fault by French and Schalla (1998). French and Schalla's (1998) published map shows that this fault places Mississippian Joana Limestone on the west side of the fault next to Mississippian Chainman Shale on the east. Though dolostone on the west side of the fault is brecciated, it was identified as Devonian Guillmette Formation (not Joana Limestone) based on fresh and weathered surface color, lack of crinoid fossils, and thickness of discernable beds. This fault is exposed at the surface and dips 41° west. The Chainman Shale dips to the east in the footwall. Given the west dip of the fault, the bed dips, and the Devonian over Mississippian juxtaposition, this fault has reverse sense offset.

The Sand Spring thrust fault may be out-of-sequence with the Ike Spring and Indian Spring thrust faults. In cross sections A-A' and B-B' the Ike Spring and Indian Spring thrust faults sole into the east-dipping Ninemile Formation, and the deeper and shallower parts of the thrusts dip in opposite directions. Therefore, these faults must have been tilted and their ramps were steepened relative to bedding after they formed. Tilting of these two thrust faults is interpreted to have been caused by the formation of the McClure Spring syncline.

The Sand Spring thrust is geometrically similar to the Pancake thrust, another central Nevada thrust belt structure. The Sand Spring thrust has stratigraphic separation
and hangingwall/footwall geometry similar to the Pancake thrust. The Pancake thrust lies along strike of the Sand Springs thrust approximately 35 km north of the study area (Carpenter et al., 1993; Langrock, 1995). Langrock (1995) suggested that the McClure Spring syncline is a footwall syncline that formed during movement along the Pancake thrust. The McClure Spring syncline within the study area is a footwall syncline to the Sand Spring thrust. Perry and Dixon (1993), working approximately 15-25 km north of the study area, suggested that a thrust fault formed the McClure Spring syncline. This thrust fault must exist just to the west of the syncline, under Big Sand Springs Valley 12 km north of, and nearly along strike of the Sand Spring thrust. This study suggests that the Sand Spring thrust must also be a controlling factor in the formation of the McClure Spring syncline based on the tilted geometry of the Ike Spring and Indian Spring thrust faults. Thus, movement on the Sand Spring thrust caused the formation of the McClure Spring syncline, a footwall syncline. In addition, Carpenter et al. (1993) demonstrated that the Pancake thrust cuts footwall folds, suggesting that the Pancake thrust, like the Sand Spring thrust, is an out-of-sequence thrust fault. Given the along strike relationships, the geometric similarities, the out-of-sequence development, and the same fold in the footwall of both the Pancake and Sand Spring thrusts, it is my interpretation that these two thrust faults are the same fault.

Regional Contraction

Contractional structures mapped within the study area can be tectonically correlated to other contractional structures mapped both to the north and south (Moores et al., 1968; Carpenter et al., 1993; Perry, 1991; Perry and Dixon, 1993; Langrock, 1995;
Cross section B*-B” shows from west to east, the Sand Spring thrust, Ike Spring thrust, Indian Spring thrust, McClure Spring syncline, Duckwater thrust, and the White Pine thrust. These structures formed in a normal sequence except for the Sand Spring thrust and the White Pine thrust.

The Duckwater thrust crops out in the Duckwater Hills approximately 30 km north of the study area (Carpenter et al., 1993; Langrock, 1995). Within cross section B*-B”, the Duckwater thrust has at least 13.6 km of stratigraphic separation, which is greater than that calculated for the fault (4.9 km) 20 km north of the study area (Langrock, 1995). Such a change is consistent with a thrust that dies out or decreases slip to the north or along which the slip is transferred onto a branch thrust northward. Most data available from the surrounding ranges suggest that the Paleozoic rocks under Railroad Valley dip to the west (J.E. Fryxell, unpublished mapping; this study). Therefore, a fault propagation anticline may have existed structurally above the present day Railroad Valley. This geometry requires significant offset on the Duckwater thrust.

The Ike Spring and Indian Spring thrusts lie within the upper plate of the Duckwater thrust. These thrust faults have less stratigraphic separation than the Duckwater thrust. Thus, they may represent hangingwall imbrications that formed coeval with the Duckwater thrust. Hangingwall imbricate thrusts are known within the Duckwater thrust plate. Carpenter et al. (1993) document back-limb imbrications within the Duckwater thrust plate. Alternatively, it is possible that the Ike Spring and Indian Spring thrusts formed independently of the Duckwater thrust.

The Easy Ridge thrust documented in the Pancake Range approximately 27 km north of the north edge of the study area and approximately 8 km north of the Duckwater...
Hills (Carpenter et al., 1993; Langrock 1995), does not exist within cross section B*-B". Langrock (1995) demonstrated that this thrust exists in the area beneath Railroad Valley, approximately 32 km north of cross section B*-B". However, the Easy Ridge thrust is geometrically inadmissible in cross section B*-B" because up to 4000 m of Paleozoic strata would need to be repeated within consecutive thrust plates under Railroad Valley between the Duckwater thrust and the White Pine thrust. The Ragged Ridge fault must accommodate 3000 m of offset, and therefore, this provides too little space under Railroad Valley to repeat the Paleozoic section between thrust plates. Therefore, the Easy Ridge thrust must either join the Duckwater thrust north of B*-B" or its southern terminus must lie north of cross section B*-B". A simple and geometrically consistent explanation for the change in amounts of stratigraphic separation along the Duckwater thrust from B*-B" and where it crops out north of B*-B" is to allow the Easy Ridge thrust to combine with the Duckwater thrust just north of cross section B*-B".

Another thrust fault, here interpreted as the White Pine thrust, crops out in the central Horse Range (J. E. Fryxell, unpublished mapping) (Plate 5). The White Pine thrust underlies part of the western White Pine Range and does not crop out in that range south of 38°55'N latitude (Langrock, 1995). A low-angle normal fault, the Blackrock fault, is interpreted to reactivate a portion of the White Pine thrust. The Blackrock fault appears to have contractional folds in its footwall. Additionally, it juxtaposes younger over older rocks, which suggests that the stratigraphic separation of the Blackrock fault is greater than that of the White Pine thrust that it reactivated (Langrock, 1995). Because the White Pine thrust may underlie part of the White Pine Range, it is likely that movement on the Currant Summit fault displaced the thrust fault to the east in the Horse.
Range (cross section B*-B") where a thrust crops out (Williams, 2000; J.E. Fryxell, unpublished mapping).

The White Pine thrust fault cuts an open anticline in its lower plate in the Horse Range (Plate 5). Therefore this thrust fault formed out-of-sequence with an older thrust fault in the White Pine thrust footwall.

A fault-bend anticline lies in the upper plate of the White Pine thrust (Plate 5). This anticline is steeply-inclined, parallel, and east-vergent. The geometry of the fold does not match classic models of fault-bend folds (e.g., Woodward et al., 1985), but it does match fault-bend fold models by Salvini et al. (2001). Salvini et al. (2001) showed that fault-bend folds forming above rigid thrust ramps dipping up to 60° do not always preserve the initial fault style. Instead, the roundness of a parallel fault-bend fold is dictated by the amount of offset on the through-going thrust fault; the more offset on the fault, the tighter the fold. The White Pine thrust in cross section B*-B” dips 56°W in relation to the footwall Paleozoic strata that it cuts, which is consistent with the interpretation of a parallel fault-bend fold (Plate 5).

Central Nevada Thrust Belt

All of the contractional structures mapped within the study area lie within and can be correlated to other structures of the central Nevada thrust belt. Central Nevada thrust belt faults typically have long, steeply dipping thrust ramps (Taylor et al., 1993, 2000; Langrock, 1995). Thrust faults within regional cross section B*-B” share this geometry, with long footwall ramps dipping from 39° to 56° in relation to the footwall Paleozoic strata. Amounts of horizontal shortening across central Nevada thrust belt thrust faults are small when compared to foreland Sevier-type thrust faults (Taylor et al., 1993, 2000;
Langrock, 1995). In addition, the Sand Spring, Indian Spring, and Ike Spring thrust faults all have between 1.2 and 2.0 km of stratigraphic separation on them. This is comparable to other central Nevada thrust faults (Langrock, 1995; Taylor et al., 2000).

Up to 27 km of horizontal shortening occurred at the latitude of the study area across the 65.5 km length of cross section B*-B". Amounts of shortening across other transects within the central Nevada thrust belt are comparable to structures within and near the study area (Armstrong and Bartley, 1993; Carpenter et al., 1993; Perry and Dixon, 1993; Taylor et al., 1993, 2000; Langrock, 1995).

Fault propagation and fault-bend folds have been mapped in other parts of the central Nevada thrust belt (Armstrong and Bartley, 1993; Carpenter et al., 1993; Cole and Cashman, 1999; Taylor et al., 2000). Folds such as the McClure Spring syncline and the fault-bend fold in the upper plate of the White Pine thrust are steeply-inclined and are similar to a steeply-inclined fault propagation/fault growth fold in the Golden Gate Range (Armstrong and Bartley, 1993). Steeply inclined to upright folds within the central Nevada thrust belt are also documented by Taylor et al. (2000) and Langrock (1995). Central Nevada thrust faults commonly have long steep ramps and the inclination of the axial surface of the fold is strongly influenced by the dip of the thrust fault. Steep thrust ramps create more upright folds (Salvini and Storti, 2001; Salvini et al., 2001). Therefore the geometry of folds and thrust faults within cross section B*-B" is consistent with central Nevada thrust belt structures.

Structures in cross section B*-B" can be regionally correlated to central Nevada thrust belt structures both to the north and south. The easternmost thrust fault in cross section B*-B" is the White Pine thrust. This thrust fault may lie structurally higher than
the Sawmill thrust, the structurally highest thrust fault in the Grant Range in the southern part of the central Nevada thrust belt (Taylor et al., 2000; J.E. Fryxell, personal comm. 2001). The Sawmill thrust fault places Cambrian and Ordovician rocks over Devonian units. Therefore, the White Pine thrust is probably not the northern equivalent of the Sawmill thrust unless a lateral thrust ramp cuts down section southward.

The Sand Spring thrust correlates to the Pancake thrust to the north of the study area. The McClure Spring Syncline, Indian Spring and Ike Spring thrusts, and the Duckwater thrust all lie in thrust plates between the Sand Spring thrust and the White Pine thrust. They represent the central Nevada thrust belt in east-central Nevada.

Approximately 60-100 km north of the study area near Eureka, Nevada, other central Nevada thrust belt structures such as the Diamond Mountains syncline and the Hoosac thrust are exposed (Carpenter et al., 1993). The Hoosac thrust and Diamond Mountains syncline lie to the north and west of central Nevada thrust belt structures in cross section B*-B". Therefore, if they continue that far south, may lie west of the study area, possibly in the upper plate of the Sand Spring thrust.

The central Nevada thrust belt structures mapped in this study and the regional cross section allow connection of the northern and southern parts of the central Nevada thrust belt. Together, these structures form a continuous belt that is at least 1500 km long along strike. Except where it is disrupted by later extensional fault sets, the central Nevada thrust belt is continuous along this length, but individual thrust faults continue for only parts of that length.
CHAPTER 8

CONCLUSIONS

Field mapping and cross section analysis delineate contractional structures that are related to the central Nevada thrust belt. Thrusting and folding were followed by up to five periods of normal faulting. Both the contractional and extensional events relate in time and space to the regional tectonic development of the east-central Great Basin.

Extension

This study documents up to four periods of normal and oblique slip faulting that occurred within the central Pancake Range, beginning in late Cretaceous to Eocene time. The region near the central Pancake Range experienced up to five temporally distinct periods of normal faulting. The first episode is pre-volcanic or Late-Cretaceous to Eocene. The second episode is syn-volcanic or late Eocene to Oligocene. Up to three episodes are post-volcanic or younger than Oligocene. The youngest set includes Holocene faults in Railroad Valley and Big Sand Springs Valley.

Though fault strike is not an exact indicator of principal stress directions, this study suggests that a rotation in the regional stress field occurred during Oligocene time. The rotation of the stress field from an east-west minimum principal stress to north-south
minimum principal stress correlates in time and space with a southern migration of volcanism. These relationships are explained by a model suggested by Bartley (1989) and modified by Taylor and Switzer (2001), in which tectonic stresses controlling extension changed from regional to localized stresses during mid-Tertiary volcanism in the Great Basin.

Some east-west striking normal and oblique-slip faults of set B lie along the Prichards Station lineament. Normal-left oblique motion along these faults is kinematically similar to other transverse faults mapped along the Prichards Station lineament, namely the Currant Summit fault. However, those set B faults were active during Oligocene time, while the Currant Summit Fault is as young as Holocene.

Within the study area, some faults of set C formed within a 3-D strain field, creating a polygonal fault set. In the west-central portion of domain 2, 3-D strain was created during faulting along a normal fault with markedly different dips across a fault bend. The primary corrugation on this fault created volume disparities within the hanging wall, creating dilatation in its upper plate. In the eastern portion of domain 2, fault cross-cutting relationships that suggest that 3-D strain can be explained by pre-existing planes of weakness within Tertiary volcanic units. These weaknesses provided pre-existing cracks along which 3-D strain was accommodated.

Contraction

Several previously unmapped thrust faults and associated folds were delineated. The Sand Spring, Ike Spring, and Indian Spring thrust faults all have long and steeply
dipping thrust ramps, similar to other thrust faults mapped in east-central Nevada. The Sand Spring thrust fault regionally correlates to the Pancake thrust of the central Nevada thrust belt based upon its geometry and structural position along strike and within a stacked sequence of thrust faults. This group of thrust faults and associated folds formed both normal- and out-of-sequence, a common occurrence in hinterland terranes.

The core of the north-south trending McClure Spring syncline lies in the subsurface of the study area and continues to the south, where it may die out. This interpretation contrasts with that of Perry and Dixon (1993), who suggested outcrop patterns could be explained by a low-angle detachment fault.

Regional cross section analysis shows that thrust faults and folds are structurally linked to the Duckwater thrust and the White Pine thrust. The Duckwater thrust is exposed approximately 25 km north, and is geometrically required to exist under Railroad Valley (Plate 5). In the lower plate of the Duckwater thrust, the White Pine thrust formed in a normal sequence. In the hanging wall of the White Pine thrust fault, a parallel fault-bend anticline formed at the top of the steeply dipping footwall ramp. The White Pine thrust fault formed out-of-sequence with another structure in its footwall. Together, these structures represent the Central Nevada thrust belt in east-central Nevada, a single contractional belt with at least 27 km of horizontal shortening.

This study correlates central Nevada thrust belt structures mapped to the south and east, to structures north and west, across Railroad Valley. This correlation extends the along strike length of the central Nevada thrust belt to at least 1500 km.
Figure 1. Map of thrust belts in relation to the Pancake Range and the central Nevada thrust belt. A = Alamo, E = Eureka AOB = Antler Orogenic Belt, CNTB = central Nevada thrust belt, LFB = Luning/Fencemaker Belt, (modified from Langrock, 1995).
Figure 2. Location of study area in relation to other contractional structures of the central Nevada thrust belt. GGR = Golden Gate Range, TR = Timpahute Range, ESR = Ely Springs Range, DT = Duckwater Thrust, ERT, Easy Ridge Thrust, GGT = Golden Gate Thrust, LT = Lincoln Thrust, MIT = Mount Irish Thrust, MS = McClure Spring Syncline, PT = Pancake Thrust, RRT = Rimrock Thrust, S = Proposed location of Stone Cabin Caldera (Best et al., 1993) SST = Sand Springs, SMT = Sawmill Thrust, ST = Schofield Thrust, WPT = White Pint Thrust. Gray delineates ranges and white shows basins that are filled with Tertiary and Quaternary alluvium.
Figure 3. Diagrammatic cross section of the Cretaceous arc showing the location of the central Nevada thrust belt in relation to the Sevier Orogenic belt and other tectonic elements.
Figure 4. Regional map showing an area of restored and unrestored Cenozoic extension. Note the removal of Cenozoic extension moves the CNTB much closer to the Sevier foreland fold and thrust belt (modified from Bartley and Gleason, 1990).
Figure 5. Paleozoic stratigraphy assumed in the subsurface of the study area based on regional geology. Compiled from Nolan et al. (1956); Smith and Ketner (1975); Kleinhampl and Ziony (1985); Elrick (1996); French and Schalla (1998); Montgomery et al. (1999).
Figure 6. Upper Paleozoic units that crop out within the study area. Unit thicknesses were calculated from map data and compiled from Nolan et al., (1956), Smith and Ketner, (1975), Kleinhampl and Ziony, (1985), Elrick, (1996), French and Schalla, (1998), Montgomery et al., (1999). See Plate 1 for unit descriptions.
Figure 7. Tertiary rock units exposed within the study area. Pancake Summit Tuff, andesite lava, and basaltic andesite lava units are not laterally continuous throughout the study area. Age of Windous Butte Formation is 31.3 Ma. Age of Stone Cabin Formation upper member is 35.3 Ma. Unit thicknesses and ages compiled from Cook (1965), Phillips (1989), Radke (1992), and Best et al. (1993).
Figure 8. Simplified geologic map of the study area. Cross sections are shown in Plates 2, 3, and 4. Upper volcanic rocks include the Windous Butte Formation, Pancake Summit Tuff, andesite, and upper rhyolite. Lower volcanic rocks include the Stone Cabin Formation, lower rhyolite, and basaltic andesite. Upper Paleozoic exposed rocks include the Ely Limestone, Diamond Peak Formation, Chainman Shale, and Joana Limestone. Lower Paleozoic exposed rocks include the Guillmette Formation, Simonson Dolomite, and Sevy Dolomite. See Figures 5, 6, 7 and Plate 1 for descriptions.
Figure 9. Location of regional cross section B*-B" in relation to the study area and surrounding ranges. DT = Duckwater thrust, ERT = Easy Ridge thrust, IFB = Illipah Fold Belt, PT = Pancake thrust, SST = Sand Spring thrust. RS = receiver station.
Figure 10. Uninterpreted portion of seismic line #106. Data Courtesy of Veritas Land Surveys, Houston, Texas.
Figure 11. Interpreted eastern portion of Big Sand Springs Valley in seismic line #106. Contact between Pzu and Pzl may be the Cambrian Dunderberg Shale. QTal = Tertiary and Quaternary alluvium; Tv = Tertiary volcanic rocks; Pzu = Upper portion of Paleozoic stratigraphy; Pzl = Lower portion of Paleozoic stratigraphy; Pz = Paleozoic rocks, undifferentiated; pCu? = possibly Precambrian rocks. Data Courtesy of Veritas Land Surveys, Houston, Texas.
Figure 12. Faults within the study area. (A) Fault timing map that includes all of the faults mapped at a scale of 1:24,000. IT = Ike Springs Thrust, IST = Indian Springs Thrust, MS = McClure Spring Syncline, SST = Sand Springs Thrust. (B) Five structural domains distinguished in the text.
Figure 13. Transverse faults of set B (labeled X and Y). Set C faults are also shown (See also Plate 1).
Figure 14. (A) Outline map showing the location of set C and set D fault, and their cross cutting relationships. (B) Detail map of cross cutting relationships. Set D faults are in gray. Set C faults are black. See text for discussion. (See also Plate 1).
Figure 15. Equal area stereoplots of poles to beds. (A) The McClure Spring syncline. Plunge and trend of the fold axis is 23°, N2°E. (B) The anticline on the central ridge in domain 5. Plunge and trend of the fold axis is 8°, N1°E. Cross symbols indicate the plunge and trend of the fold axes.
Figure 16. Location of Pancake and Pritchards Station lineaments in relation to the study area and surrounding ranges. PL = Pancake Lineament, PSL = Pritchards Station Lineament. Note location of Currant Summit Fault. Inset highlights major lineaments that exist in the region. Lineaments from Williams (2000) and Ekren et al. (1976).
Figure 17. (A) Detailed fault map of three faults of fault set C (M, N, and O) showing orthorhombic symmetry in the hanging wall of low angle fault L. (B) Detailed fault map of set C faults in an orthorhombic pattern in domain 2. Figure locations shown on figure 12.
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Oversize maps and charts are microfilmed in sections in the following manner:

LEFT TO RIGHT, TOP TO BOTTOM, WITH SMALL OVERLAPS

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Lake Range, Nye County, Nevada.
Nevada.
Quaternary Alluvium
Unconsolidated sand and gravel. Clasts range from fine sand to boulders and are composed of limestone and volcanic rock. Occurs in active washes and on fan surfaces. Unconformably overlies bedrock and older alluvium within study area. Thickness is 1-5 m.

Tertiary/Quaternary Alluvium
Poorly to highly indurated alluvium. Medium to fine grained sandstone and pebble to boulder conglomerate. Lithic clasts consist of limestone, dolomite, tuff, and rhyolite lava. Commonly dissected by active washes and Quaternary alluvium. Contact with underlying bedrock is sharp. Thickness is 5-15 m.

Tertiary(?) Jasperoid
Jasperoid. Light red to rust-red brown weathered. Light red-orange to red fresh. Concoidal fracture pattern. Small lenticular and amorphous bodies occupy fault zones within the study area.

Tertiary(?) Joana Jasperoid Replacing Limestone
Jasperoid and jasperoid cemented conglomerate. Red to rust-red brown weathered red and red brown fresh. Conglomerate clasts are light green to gray and red weathered and fresh. Conglomerate clasts composed of quartz and limestone. Sub-rounded to sub-angular clasts. Clasts are 3 - 5 cm in diameter. Concoidal fracture pattern. Amorphous to semi-lenticular. Jasperoid may include portions of the upper Guillmette Formation.

Tertiary Windous Butte Formation
Rhyolitic tuff. Red to brick red weathered, red to light red-brown fresh. Distinct black dense vitrophyre at base. Forms cliffs and steep slopes. Phenocrysts 16-28% of rock, quartz 25-30%, sanidine 40-44%, plagioclase 23-31%, biotite 1-4%, hornblende 0-0.4%, olivine 0.6-1.6%. Age is 31.3 Ma (Radke, 1992). Thickness is 50 - 200 m.

Tertiary Pancake Summit Tuff
Rhyolite Tuff. White to white pink weathered, white fresh. Moderate to highly devitrified. White to gray-white pumice 1-3 cm across. Phenocrysts 38% of rock; quartz 45%, sanidine 35%, plagioclase 18%, biotite 2%, olivine 0.5% (Radke, 1992). Forms slopes and low areas. Contact with overlying Windous Butte Formation is sharp. Age is 34.8 Ma (Radke, 1992). Thickness is 15 m (Radke, 1992).
Tertiary Stone Cabin Formation
The Stone Cabin Formation consists of four members as defined by Radke (1992). Phenocryst percent is percentage of whole rock, minerals are listed as percent of total phenocrysts (modal percent). Contacts between members are sharp. Contact with overlying rhyolite flow is not exposed.

Stone Cabin Formation Upper Member
Compound cooling unit of rhyolitic tuff. Brown to tan yellow weathered, red-brown to gray fresh. Large 2-4 cm long twisted and flattened pumice clasts characteristic. White to yellow dusty gray, 1-2 cm limestone lithic clasts are less common. Weathers to steep slopes to cliffs. Large quartz and sanidine phenocrysts, with smaller biotite in books. Crystal rich. Phenocrysts 34-44%; quartz 43-54%, sanidine 27-38%, plagioclase 10-20%, biotite 2-5%, trace olivine 0.5% (Radke, 1992). 35.31 Ma (Best et al., 1993; Radke, 1992). Locally not present in unfaulted sections. Maximum thickness is 160-400 m.

Stone Cabin Formation Base of Upper Member
Highly devitrified base of upper member, gray-white to yellow white weathered, white to light gray fresh. Less abundant phenocrysts than upper member. Forms slopes and soils within the study area. Thickness is 20-30 m.

Stone Cabin Formation Salmon Member
Single cooling unit of rhyolitic tuff. Distinct orange-pink to brown weathered, light pink fresh. Moderate to poorly welded. Characteristic large 5-6 mm smoky quartz phenocrysts abundant. Lithic fragments consist of white to chalky limestone 1-2 cm across and large (up to 10-15 cm) rhyolitic fragments. Forms slopes and rounded hills. Phenocrysts 28-32%; quartz 38-42%. sanidine 25-28%, plagioclase 26-31%, biotite 4-6%, hornblende trace, olivine 0.3-0.4% (Radke, 1992). Thickness is 60-120 m.

Stone Cabin Formation Middle Member
Compound cooling unit rhyolitic tuff. Moderate to densely welded. Red-brown to brown-black and gray weathered light brown-red to red-gray fresh. Forms cliffs and ledges. Gray to light brown and glassy black pumice. Pumice clasts common, 1-4 cm across. Flattened pumice can form a slotted weathering pattern. Contains up to 6 vitrophyric zones (Radke, 1992). Forms cliffs and rounded steep slopes. Phenocrysts appear equigranular. Biotite exists in books. Phenocrysts 29-45%; quartz 33-47%, sanidine 3-21%, plagioclase 29-52%, biotite 4-9%, hornblende 0-1%, olivine trace (Radke, 1992). Age is 35.34 ± 0.07 Ma (Radke, 1992). Thickness is 100-300 m.

Stone Cabin Formation Lower Member
Poorly welded tuff and tuffaceous sandstone and conglomerate. White to yellow white weathered, white fresh. Abundant rhyolitic lithic fragments 1-30 cm across. Forms low hills and slopes. Relatively phenocryst poor. Biotite occurs in books. Phenocrysts total 23%; quartz 30-33%, sanidine 5-6%, plagioclase 51-54%, biotite 7-10%, olivine 1-2% (Radke, 1992). Unit pinches out or was not deposited in northern and western parts of study area. Thickness is 50-100 m.

Tertiary Lower Rhyolitic Lava Flows
Pink to red-brown and gray-black weathered, gray to brown gray and black fresh. Flow banded. White to cloudy white sandstone 0.5-3 mm across. Forms cliffs to massive.

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<table>
<thead>
<tr>
<th>Formation</th>
<th>Description</th>
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<tr>
<td><strong>Ely Limestone Upper Member</strong></td>
<td>Limestone. Gray to brown-gray weathered, medium to light gray and white.</td>
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<tr>
<td></td>
<td>Fine grained limestone containing significant clastic material. Incremental thickness is 1-3 m thick.</td>
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<td></td>
<td>Cross beds occur in clastic rich beds. Small chert pebbles 1 cm crinoids common. Forms cliffs to steep slopes. Contact with D Formation is sharp. Thickness is 457 m (Kleinhamp and Ziony, 1985).</td>
</tr>
<tr>
<td><strong>Ely Limestone Lower Member</strong></td>
<td>Limestone. Gray to light gray-white weathered, light gray-white fresh.</td>
</tr>
<tr>
<td></td>
<td>Beds 2-4 m thick. Abundant red to brown bedding parallel nodular clast assemblage consists of small 5-10 mm disarticulated crinoids, brachiopods, and coral. Local beds of large 10-17 cm long rugose corals and plecopods. Forms cliffs and steep slopes. Contact with upper strata is dramatic increase in clastic cobbles and sand within the limestone. Contact with upper strata is sharp. Thickness is 610 m (Kleinhamp and Ziony, 1985).</td>
</tr>
<tr>
<td><strong>Mississippian Diamond Peak Formation</strong></td>
<td>Sandstone and sandstone grading to green and black chert clast conglomerate.</td>
</tr>
<tr>
<td></td>
<td>Lighter cream-brown to tan brown weathered. Clean quartz sandstone grading to 2-5 cm chert pebble conglomerate. Commonly silicified, exposed in isolated outcrops. Forms cliffs where silicified, elsewhere, slopes and steep slopes. Contact with upper limestone and Chainman Shale is sharp. Thickness is 305 - 427 m (Kleinhamp and Ziony, 1985).</td>
</tr>
<tr>
<td><strong>Mississippian Chainman Shale</strong></td>
<td>Distinct black shale with interbedded lenticular and tabular sandstone and conglomerates.</td>
</tr>
<tr>
<td></td>
<td>Shale is black to dark rust brown to black weathered. Quarter-brown fresh. Fissile. Sandstone and conglomerate are tan-brown to tan fresh. Sandstone is medium to fine grained, thin bedded with no apparent layering. Contacts between sandstone beds is sharp. Forms slopes, low hills and occurs in most valleys with the underlying unit. Fossils found in sandstone consist of rod-like sphenopliyta? 2-5 cm long with Joana Limestone is sharp. Thickness is 1067-1220 m (Kleinhamp and Ziony, 1985).</td>
</tr>
<tr>
<td><strong>Mississippian Joana Limestone</strong></td>
<td>Limestone and limestone grading to chert clast conglomerate in upper portion of unit.</td>
</tr>
<tr>
<td></td>
<td>Blue-gray to light and medium to light-brown limestone.</td>
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<tr>
<td></td>
<td>Limestone is very fine grained, creating distinct plumose marks when viewed under the microscope. The portion of unit can be highly recrystallized, and consist almost entirely of Abundant crinoid fossils 0.5-3 cm in diameter. Formed in rounded reefal mound. Contact with underlying Guillmette Formation is sharp. Thickness is 2957 m (Kleinhamp and Ziony, 1985).</td>
</tr>
<tr>
<td><strong>Devonian Guillmette Formation Upper</strong></td>
<td>Limestone with interbeds or local regions of dolostone. Gray and light gray-white fresh on limestone surface. Light to deep.</td>
</tr>
</tbody>
</table>
Ely Limestone Upper Member
Limestone. Gray to brown-gray weathered, medium to light gray and gray-brown fresh. Fine grained limestone containing significant clastic material, increases upsection. Beds 1-3 m thick. Cross beds occur in clastic rich beds. Small chert pebbles within beds. Small 1 cm criniods common. Forms cliffs to steep slopes. Contact with Diamond Peak Formation is sharp. Thickness is 457 m (Kleinhampl and Ziony, 1985).

Ely Limestone Lower Member
Limestone. Gray to light gray-white weathered, light gray-white fresh. Fine grained. Beds 2-4 m thick. Abundant red to brown bedding parallel nodular chert. Fossil assemblage consists of small 5-10 mm disarticulated crinoids, brachiopods, and rugose corals. Local beds of large 10-17 cm long rugose corals and plecypods. Blocky to massive outcrop. Forms cliffs and steep slopes. Contact with upper member is marked by dramatic increase in clastic cobbles and sand within the limestone. Contact with Diamond Peak Formation is sharp. Thickness is 610 m (Kleinhampl and Ziony, 1985).

Mississippian Diamond Peak Formation
Sandstone and sandstone grading to green and black chert clast conglomerate. Brown to dark brown-black weathered. Lighter carmel-brown to tan brown fresh. Beds are 0.25 - 2 m thick and locally cross bedded. Clean quartz sandstone medium grained, grading to 2-5 cm chert pebble conglomerate. Commonly silicified, makes very resistant outcrops. Forms cliffs where silicified, elsewhere, slopes and steep hills. Contact with Chainman Shale is sharp. Thickness is 305 - 427 m (Kleinhampl and Ziony, 1985).

Mississippian Chainman Shale
Distinct black shale with interbedded lenticular and tabular sandstones and conglomerates. Shale is black to dark rust brown to black weathered and black to dark-brown fresh. Fissile. Sandstone and conglomerate are tan-brown weathered to light gray-tan fresh. Sandstone is medium to fine grained, thin bedded with no observed cross beds. Clasts within the conglomerate consist of green and black chert with less common dark gray to black, medium to fine grained limestone. Contacts between sandstone and shale beds is sharp. Forms slopes, low hills and occurs in most valleys within the study area. Fossils found in sandstone consist of rod-like sphenophyta? 2-5 cm in length. Contact with Joana Limestone is sharp. Thickness is 1067-1220 m (Kleinhampl and Ziony, 1985).

Mississippian Joana Limestone
Limestone and limestone grading to chert clast conglomerate in upper portion. Medium blue-gray to light and medium to light-brown limestone. Beds 15-45 cm thick. Limestone is very fine grained, creating distinct plumose marks when fractured. Upper portion of unit can be highly recrystallized, and consist almost entirely of jasperoid. Abundant criniod fossils 0.5-3 cm in diameter. Forms low rounded hills and slopes. Contact with underlying Guillmette Formation is sharp. Thickness is 23-55 m (Kleinhampl and Ziony, 1985).

Devonian Guillmette Formation Upper
Limestone with interbeds or local regions of dolostone. Gray and light to medium olive gray weathered, light olive gray fresh on limestone surface. Light to medium gray...
Tertiary Pancake Summit Tuff
Rhyolite Tuff. White to white pink weathered, white fresh. Moderate to highly devitrified. White to gray-white pumice 1-3 cm across. Phenocrysts 38% of rock; quartz 45%, sanidine 35%, plagioclase 18%, biotite 2%, olivine 0.5% (Radke, 1992). Forms slopes and low areas. Contact with overlying Windows Butte Formation is sharp. Age is 34.8 Ma (Radke, 1992). Thickness is 15 m (Radke, 1992).

Tertiary Andesite Lava Flow
Andesite lava flow. Distinct red to black fresh, black weathered. Locally vesicular. Dense and glassy groundmass with characteristic plagioclase phenocrysts. Forms low hills to step-like cliffs. Phenocrysts 35% of rock; plagioclase 40%, clinopyroxene 40%, orthopyroxene 20%, olivine 1% (Radke, 1992). Forms slopes and small hills. Contact with Pancake Summit Tuff is not exposed. Contact with underlying rhyolite flows, Stone Cabin Formation is unconformable. Thickness is up to 400 m (Radke, 1992).

Tertiary Upper Rhyolitic Lava Flows
Rhyolite lava flows. Brown to tan and red-brown weathered, red to tan and dark brown fresh. Distinct flow banding. Locally brecciated. Quartz phenocrysts are locally smokey. Forms low hills to step-like cliffs. Phenocrysts 15-40% of rock; quartz 20-40%, sanidine 10-30%, plagioclase 25-60%, biotite 1-10%, hornblende 0-5%, olivine 0-3% (Radke, 1992). Contact with underlying Stone Cabin Formation is sharp. Stratigraphic position is defined by Radke (1992). Thickness 100-120 m.

Location of Exxon Wildhorse Unit #1 oil well (dry hole).
Tertiary Lower Rhyolitic Lava Flows
Pink to red-brown and gray-black weathered, gray to brown gray and black fresh. Flow banded. White to cloudy white sanidine 0.5-3 mm across. Forms cliffs to massive rounded hills. Lower contact with Paleozoic rocks is unconformable, contacts with overlying and underlying Tertiary volcanic units are sharp. Thickness is 150-200 m (Radke, 1992).

Tertiary Andesite Flow
Andesite lava flow. Black to brown weathered, black fresh. Phenocrysts less than 1%. Plagioclase and pyroxene phenocrysts, 0.5 mm to 1 mm across. Forms low, rounded hills. Contact with Paleozoic rocks is unconformable. Thickness is 125 m (Radke, 1992).

Tertiary Dike
Rhyolite dike. White to rust brown and gray weathered, white to light gray fresh. Commonly silicified in the western portion of the study area. Forms linear ridges and small hills. Plagioclase altered to clay minerals. Phenocrysts 23% of rock; quartz 33%, sanidine 38%, plagioclase 27%, biotite 2% (Radke, 1992). Dike rock is not exposed cutting volcanic units, but cuts Paleozoic units. Therefore, Tertiary age is based on similarity to other volcanic rock units exposed in the area.

Tertiary (?) Dike rock, unknown

Tertiary/Cretaceous Sheep Pass Formation
Interbedded conglomerate and coarse grained sandstone; well rounded and matrix supported limestone clasts. Distinct red paleosol overlies the conglomerate. Poor outcrop, consists of distinct red soil with local exposure of conglomerate. Outcrops usually smaller than 3 m wide. Contact with Paleozoic units is sharp and unconformable. Thickness is 30-70 m (Perry and Dixon, 1993).

Symbols and Contacts

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Devonian Guillmette Formation Upper

Limestone with interbeds or local regions of dolostone. Gray and light to medium gray weathered, light olive gray fresh on limestone surface. Light to medium gray weathered, medium gray fresh dolostone beds. Medium to fine grained. Beds 1 m thick. Unit is distinguished by abundant medium to coarsely recrystallized fossils and less abundant crinoid columnals. Contact with the rest of the Guillmette Formation is sharp. Thickness is 23-55 ft (Ziony, 1985).

Devonian Guillmette Formation

Limestone with interbeds or local regions of dolostone. Base of unit consists of a colored band of light gray to gray-yellow limestone and silty limestone (Kleinhampl and Ziony, 1985). Gray and light to medium olive gray weathered, light olive gray fresh on limestone surface. Medium to dark gray weathered, dark gray to gray-black dolostone beds. Medium grained. Beds 0.5 - 3 m thick. "Spaghetti" and stromatoporoids found within some beds. Top of unit contains abundant small crinoid columnals across and some 0.5 - 1 cm across gastropod fossils. Forms step-like cliffs to cliffs. Contact with Simonson Dolomite is sharp. The Eureka District lithologic equivalent of the Guillmette is the Devils Gate Limestone. Thickness is 545 ft (Kleinhampl and Ziony, 1985).

Devonian Simonson Dolomite

Dolomite. Interbedded light and medium to dark gray-brown weathered, medium gray fresh. Locally, beds have burrow mottling. Bedding from 10 cm to 2 m thick. Medium to coarse grained. Intrabed laminae common. Local calcite filled vugs and cavities. Beds contain abundant "spaghetti" fossils. Forms cliffs to slopes. Contact with Simonson Dolomite is sharp. The Eureka District equivalent of the Simonson Dolomite is the Devils Gate Limestone. Thickness is 201 m (Kleinhampl and Ziony, 1985).

Devonian Sevy Dolomite

Dolomite. Light gray to yellowish-white weathered, light to medium brown fresh. Fine grained. Beds are 5-15 cm thick. Laminae and desiccation cracks. Fractured conchoidally. Poor exposures within study area. Step to rounded cliff forms with underlying units are not exposed within the study area. The Eureka District lithologic equivalent of the Sevy Dolomite is the Beacon Peak Dolomite of the Provo Group. Thickness is 305 m (Kleinhampl and Ziony, 1985).

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Tie line. Used to identify areas of where the same formation is exposed, but where the area is too small to be labeled.
portion of unit can be highly recrystallized, and consist almost entirely of jasperoid. Abundant crinoid fossils 0.5-3 cm in diameter. Forms low rounded hills and slopes. Contact with underlying Guillmette Formation is sharp. Thickness is 23-55 m (Kleinhampfl and Ziony, 1983).

**Devonian Guillmette Formation Upper**
Limestone with interbeds or local regions of dolostone. Gray and light to medium olive gray weathered, light olive gray fresh on limestone surface. Light to medium gray weathered, medium gray fresh dolostone beds. Medium to fine grained. Beds are 0.5 to 1 m thick. Unit is distinguished by abundant medium to coarsely recrystallized gastropod fossils and less abundant crinoid columnals. Contact with the rest of the Guillmette Formation marked by the absence of gastropod fossils and darker fresh colored dolomite. Locally present within study area. Minimum thickness is 36 m.

**Devonian Guillmette Formation**
Limestone with interbeds or regions of dolostone. Base of unit consists of a yellow colored band of light gray to gray-yellow limestone and silty limestone (Kleinhampfl and Ziony, 1985). Gray and light to medium olive gray weathered, light olive gray fresh on limestone surface. Medium to dark gray weathered, dark gray to gray-black fresh dolostone beds. Medium grained. Beds 0.5 - 3 m thick. "Spaghetti" and stromatoporoid fossils found within some beds. Top of unit contains abundant small crinoids 1-3 mm across and some 0.5 - 1 cm across gastropod fossils. Forms step-like cliffs to massive cliffs. Contact with Simonson Dolomite is sharp. The Eureka District lithologic equivalent of the Guillmette is the Devils Gate Limestone. Thickness is 549-762 m (Kleinhampfl and Ziony, 1985).

**Devonian Simonson Dolomite**
Dolomite, interbedded light and medium to dark gray-brown weathered, medium brown-gray fresh. Locally, beds have burrow mottling. Bedding from 10 cm to 2-3 m. Medium to coarse grained. Intrabed laminae common. Local calcite filled vugs. Some beds contain abundant "spaghetti" fossils. Forms cliffs to slopes. Contact with Sevy Dolomite is sharp. The Eureka District equivalent of the Simonson Dolomite consists of the Sentinel Mountain Dolomite, the Woodpecker Limestone and the Bay State Dolomite. Thickness is 201 m (Kleinhampfl and Ziony, 1985).

**Devonian Sevy Dolomite**
Dolomite. Light gray to yellowish-white weathered, light to medium brown-gray fresh. Fine grained. Beds are 5-15 cm thick. Laminae and desiccation cracks. Fractures conchoidally. Poor exposures within study area. Step to rounded cliff former. Contact with underlying units are not exposed within the study area. The Eureka District lithologic equivalent of the Sevy Dolomite is the Beacon Peak Dolomite of the Nevada Group. Thickness is 305 m (Kleinhampfl and Ziony, 1985).
Contour interval on Sand Spring and Portuguese Mtn. quadrangles: 40 feet

Contour interval on Bradshaw Spring and Meteorite Crater quadrangles: 20 feet

Scale 1:24,000

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mapping completed on Bradshaw Spring, Portuguese Mtn., Sand Spring, and Meteorite Crater USGS 7.5' quadrangles

Magnetic declination: 16.5 degrees East

mapping by Joseph J. Gilbert
June, July, and August 2000
Location of Exxon Wild Unit #1 oil well (dry hole)

Strike and dip of bedding

Strike and dip of compaction

Anticline. Positioned on the axial surface. Arrows show trace points in the direction of dip.

Syncline. Positioned on the axial surface. Arrows show trace points in the direction of dip.

Stratigraphic contact. Dashed line shows location of contact that is approximately located and concealed under alluvium.
Tertiary/Cretaceous Sheep Pass Form
Interbedded conglomerate and coarse grain supported limestone clasts. Distinct red outcrop, consists of distinct red soil with usually smaller than 3 m wide. Contact width is 30-70 m (Perry and Dixon, 1

Symbols an

Location of Exxon Wildhorse
Unit #1 oil well (dry hole).

Stike and dip of bedding

Strike and dip of compaction foliation

Anticline. Positioned on the trace of the axial surface. Arrow on axial trace points in the direction of plunge.

Syncline. Positioned on the trace of the axial surface. Arrow on axial trace points in the direction of plunge.

Stratigraphic contact. Dashed where approximately located and dotted where concealed under alluvium.
Triassic/Cretaceous Sheep Pass Formation
Embeded conglomerate and coarse grained sandstone; well rounded and matrix
ported limestone clasts. Distinct red paleosol overlies the conglomerate. Poor
crop, consists of distinct red soil with local exposure of conglomerate. Outcrops
ally smaller than 3 m wide. Contact with Paleozoic units is sharp and unconformable.
ickness is 30-70 m (Perry and Dixon, 1993).

Symbols and Contacts
Dolomite is sharp. The Eureka District equivalent of it of the Sentinel Mountain Dolomite, the Woodpecker Limestone. Thickness is 201 m (Kleinhampl and Ziony, 1985).

**Devonian Sevy Dolomite**

Dolomite. Light gray to yellowish-white weathered, light gray to yellowish-white weathered, light gray to yellowish-white weathered, light gray to yellowish-white weathered, light gray to yellowish-white weathered, light gray to yellowish-white weathered, light gray to yellowish-white weathered, light gray to yellowish-white weathered, light gray to yellowish-white weathered. Beds are 5-15 cm thick. Laminae and dense bands are conchoidally. Poor exposures within study area. Slopes with underlying units are not exposed within the study area. In the study area, the lithologic equivalent of the Sevy Dolomite is the Beacon Group. Thickness is 305 m (Kleinhampl and Ziony, 1985).

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Tie line. Used to identify areas of where the same formation is exposed, but where the area is too small to be labeled.

Normal fault. Dashed where approximately located and dotted where concealed under alluvium. Ball and bar are on the hanging wall.

Low-angle normal fault. Dashed where approximately located and dotted where concealed under alluvium. Parallel tick marks are on the hanging wall.

Thrust fault. Dashed where approximately located and dotted where concealed under alluvium. Barbs are on the hanging wall.
Devonian Sevy Dolomite

Dolomite. Light gray to yellowish-white weathered, light to medium brown-gray fresh. Fine grained. Beds are 5-15 cm thick. Laminae and desiccation cracks. Fractures conchoidally. Poor exposures within study area. Step to rounded cliff former. Contact with underlying units are not exposed within the study area. The Eureka District lithologic equivalent of the Sevy Dolomite is the Beacon Peak Dolomite of the Nevada Group. Thickness is 305 m (Kleinhampl and Ziony, 1985).
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Plate 2. Deformed
ed state and retro-def
deformed cross section

Intersection with Cross Section C-C'
Intersection with Cross Section D-D'

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Unit Key

Qal
Alluvium

QTal
Inactive alluvium

Fj
Jasperoid

Ji
Joana jasperoid replacing limestone

Tw
Windous Butte formation

Tp
Pancake Summit tuff

I
Andesite lava

Tr
Upper Rhyolite lava flow

Str
Upper member

Sm
Base of upper member

Sm
Salmon member

Sm
Middle member

Sm
Lower member

Stone Cabin Formation

Pennsylvania

Mississippi

Devonia

Silurian

Ordovician
a. Present day cross section A-A'. Cross section lo
section location shown on Plate 1. Scale is 1:24000. No vertic...
No vertical exaggeration.
Lower Rhyolite lava
Andesite lava
Dike rock
Dike rock, unknown
Sheep Pass Formation

* Unit colors used in cross section construction may differ slightly from unit colors in Plate 1
♀ Units not present in this cross section
Precambrian units, undivided

Pioche Shale/Prospect Mountain Quartzite

Pole Canyon Limestone

Lincoln Peak Formation

Dunderberg Shale

Windfall Formation

Goodwin Limestone

Ninemile Formation

Pogonip Group

Cambrian
b. Partially restored cross section A-A'. Faults of
Faults of set C restored. No Faults of set D intersect the plane.
At the plane of cross section A-A'.

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c. Cross Section A-A' with all normal faults restored. Set
faults restored. Set B faults B1 and B2 in cross section A-A'.
and B2 in cross section A-A' are not restorable due to movement.
restorable due to movement in and out of the plane of section.
due to movement in and out of the plane of section.
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UMI
Deformed state (present)

Intersection with cross section C-C'
tro-deformed cross section

5° bend in section to accommodate regional cross section B-B"
section B-B'
a. Present day cross section B-B'. Cross section location shown on Figure 1.
Reproduction shown on plate 1. Scale is 1:24,000. No vertical exaggeration.
c. Cross section B-B' with set D (post-volcanic) and set C (p
and set C (post-volcanic) normal faults restored. Cross section location shown on Plate 1.
The formation shown on Plate 1. Scale is 1:24,000. No vertical exaggeration.
e. Cross section
ross section B-B' with set D (post-volcanic), set C (post-volcanic), set B (syn-volcanic), and
volcanic), and set A (pre-volcanic) normal faults restored. Cross-section location shown on p.
on plate 1. Scale is 1:24,000. No vertical exaggeration
Cretaceous

- Zba: Andesite lava
- Tdi: Dike rock
- Tdu: Dike rock, unknown
- Ksp: Sheep Pass Formation

* Unit colors used in cross section construction may differ slightly from unit colors in Plate 1
○ Units not present in this cross section
Cambrian

- Dunderberg Shale
- Lincoln Peak Formation
- Pole Canyon Limestone
- Pioche Shale/Prospect Mountain Quartzite

Precambrian

- Precambrian units, undivided
Cambrian

- Dunderberg Shale
- Lincoln Peak Formation
- Pole Canyon Limestone
- Pioche Shale/Prospect Mountain Quartzite

Precambrian

- Precambrian units, undivided
b. Cross section B-B' with set D (post-volcanic) normal faults
normal faults restored. Cross section location shown on plate 1. Scale is 1:24,000. No verti
No vertical exaggeration
d. Cross section B-B’ with set D (post-volcanic), set C (post-volcanic), and set B
(post-volcanic), and set B (syn-volcanic) normal faults restored. Cross section location shown...
loss section location shown on plate 1. Scale is 1:24,000. No vertical exaggeration.
volcanic), set B (syn-volcanic), set A (pre-volcanic) normal faults, Sand Spring thrust, and McC.

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ust, and McClure Spring syncline restored. Cross section location shown on plate 1. Scale is
n on plate 1. Scale is 1:24,000. No vertical exaggeration.
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UMI
Plate 4. Deformed structure

Intersection with cross section D-D'

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Cross section C-
C-C’ and Cross sect

Unit Key*

Tertiary

Qal  Alluvium
QTal  Inactive alluvium
J  Jasperoid
J  Joana jasperoid replacing limestone
Tw  Windous Butte formation
P  Pancake Summit tuff
E  Andesite lava
I  Upper Rhyolite lava flow

Upper member
Base of upper member

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Key*

Pennsylvanian

Mississippian

Devonian

Silurian

Ely Limestone

Upper member

Lower member

Diamond Peak Formation

Chainman Shale

Joana Limestone

Guillmette Formation, upper unit

Guillmette Formation

Simonson Dolomite

Sevy Dolomite

Laketown Dolomite

Fisheaven Dolomite

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a. Cross section C-C'. This cross section is not retrod. Unit thickness may appear thicker, or may not match units displaced westward, into the plane of section by on Plate 1. Scale is 1:24,000. No vertical exaggeration.
not retrodeformable because of movement in and out of the plane of section. It does not match across the cross section from north to south. This is due to more steeply dipping beds by normal faults that strike parallel to this cross section. Cross section location is slightly exaggerated.
The steeply dipping section location is shown.
* Unit colors used in cross section construction may differ slightly from unit colors in Plate 1

Units not present in this cross section
b. Deformed state cross section D-D’. This cross section is not retrofitted. Cross section location shown on plate 1. Scale is 1:24,000. No vertical
Cross section is not retrodeformable because of movement in and out of the plane of section. Scale is 1:24,000. No vertical exaggeration.
of the plane of section.
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UMI
5. Deformed state (p
(present day) and re
Retrodeformed cross section
Cross section B*-B’’
a. Deformed state cross section B-B". Scale is 1:72000. No vertical exag
vertical exaggeration. Location of cross section shown on figure 9.
Unit Key*

QTal  Quaternary/Tertiary alluvium, undivided
TKu   Tertiary/Cretaceous units, undivided
Peu  Pennsylvanian units undivided
M  Mississippian units undivided
Du   Devonian units undivided
Silurian units undivided
Ou   Ordovician units undivided
Oub2 Ordovician units brecciated,
Eu   Upper Cambrian units, undivided
Cub  Cambrian units brecciated
Cl   Lower Cambrian units undivided
b. Retrodeformed cross section B-B². Scale is 1:72000. No vertical exaggeration.
Ical exaggeration. Location of cross section shown on figure 9.
Upper Cambrian units, undivided
Cambrian units brecciated
Lower Cambrian units undivided

Sub-Tertiary Unconformity