An upper-crustal fold province in the hinterland of the Sevier orogenic belt, eastern Nevada, U.S.A.: A Cordilleran Valley and Ridge in the Basin and Range

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ABSTRACT

Overprinting Cenozoic extension hinders analysis of Cordilleran contractional deformation in the hinterland of the Sevier thrust belt in Nevada. In this study, a 1:250,000 scale paleogeologic map of eastern Nevada, showing spatial distributions of Paleozoic–Mesozoic rocks beneath a Paleogene unconformity, is combined with dip magnitude maps for Paleozoic–Mesozoic and Tertiary rocks, published sedimentary thickness records, and a published reconstruction of extension, in order to define and regionally correlate thrust faults and folds, and estimate the pre-extensional amplitude, wavelength, and limb dips of folds. A new structural province, the Eastern Nevada fold belt, is defined, consisting of a 100-km-wide region containing five first-order folds that can be traced for map distances between 100 and 250 km and have amplitudes of 2–4 km, wavelengths of 20–40 km, pre-extensional limb dips typically between 10° and 30°, and deform rocks as young as Jurassic and Early Cretaceous. No regional-scale thrust faults or décollement horizons breach modern exposure levels in the Eastern Nevada fold belt. First-order folds of the Eastern Nevada fold belt are interpreted to have formed above a deep (≥10 km below the Paleogene unconformity), blind décollement or shear zone, perhaps the westward projection of the master décollement of the Sevier thrust belt.

Three hinterland structural provinces, the Central Nevada thrust belt, Western Utah thrust belt, and the intervening Eastern Nevada fold belt, collectively record low-magnitude (a few tens of kilometers), upper-crustal shortening that accompanied Cretaceous translation of the Cordilleran passive-margin basin ~220 km eastward during the Sevier orogeny. Low deformation magnitudes in the hinterland are attributed to the rheological competence of this thick basin.

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INTRODUCTION

The Jurassic–Paleogene Cordilleran orogenic belt is the result of over 100 m.y. of contractional deformation of the North American plate above an Andean-style subduction zone (Allmendinger, 1992; Burchfiel et al., 1992; DeCelles, 2004; Dickinson, 2004). In the western interior United States, the majority of Cordilleran crustal shortening was accommodated in the Sevier thrust belt in western Utah (Fig. 1; e.g., Armstrong, 1968; Royse et al., 1975; Villien and Kligfield, 1986; DeCelles and Coogan, 2006). Decades of study in the Sevier thrust belt leave the geometry, kinematics, magnitude, and timing of deformation well documented (e.g., Jordan, 1981; Allmendinger et al., 1983, 1987; Royse, 1993a; DeCelles et al., 1995; Mitra, 1997; DeCelles and Coogan, 2006). However, in the hinterland of the thrust belt in eastern Nevada, fundamental questions remain regarding the style, geometry, and magnitude of Cordilleran deformation, and how it relates temporally to deformation in the frontal thrust belt (e.g., Armstrong, 1972; Miller et al., 1988; Miller and Hoisch, 1995; Camilleri and Chamberlain, 1997; Taylor et al., 2000; Long, 2012; Greene, 2014; Long et al., 2014). This can be attributed in part to sparse preservation of synorogenic sedimentary rocks, but it is primarily due to the complex dismemberment of the region by Cenozoic extension (e.g., Gans and Miller, 1983; Coney and Harms, 1984; Dickinson, 2002).

Two zones of thrust faults, the Central Nevada thrust belt and Western Utah thrust belt (Fig. 1), have been identified in the Sevier hinterland, and they are interpreted as contemporary, interior components of the Sevier thrust system (Taylor et al., 2000; Long, 2012; Greene, 2014; Long et al., 2014). The region of eastern Nevada between the Central Nevada thrust belt and Western Utah thrust belt is interpreted to have experienced relatively little upper-crustal Cordilleran contractional deformation (e.g., Armstrong, 1972; Speed et al., 1988). This is primarily based on compilation maps of stratigraphic levels exposed beneath a regional unconformity that places Paleogene rocks over Paleozoic–Mesozoic rocks, which show that the total postorogenic structural relief of this region was low (2–4 km; Armstrong, 1972; Gans and Miller, 1983; Long, 2012). In addition, early mapping studies in eastern Nevada documented that angularity across the unconformity is typically ≤10°–15° (Young, 1960; Kellogg, 1964; Moores et al., 1968), which has led to a prevailing view that any pre-Tertiary deformation was gentle and minor.

However, despite the low structural relief of this region, the map patterns of regionally traceable (≥100 km) folds are defined on subcrop maps (Gans and Miller, 1983; Long, 2012), and detailed geologic maps (Humphrey, 1960; Brokaw and Barosh, 1968; Nolan et al., 1971; Nutt, 2000) show that many more folds are present, although to date they have not been regionally correlated or described in detail. Recent work near Eureka that documents 60°–70° of angularity across the Paleogene unconformity (Long et al., 2014), and geologic maps showing fold limb dips ranging from 30° to overturned (Humphrey, 1960; Larson and Riva, 1963; Brokaw and Barosh, 1968; Nolan et al., 1971) both challenge the interpretation of minimal pre-Tertiary deformation and justify a structural synthesis of this region.

In this study, the primary goal is to describe folding in the region between the Central Nevada thrust belt and Western Utah thrust belt; this exercise leads to definition of a new structural province, the Eastern Nevada fold belt.

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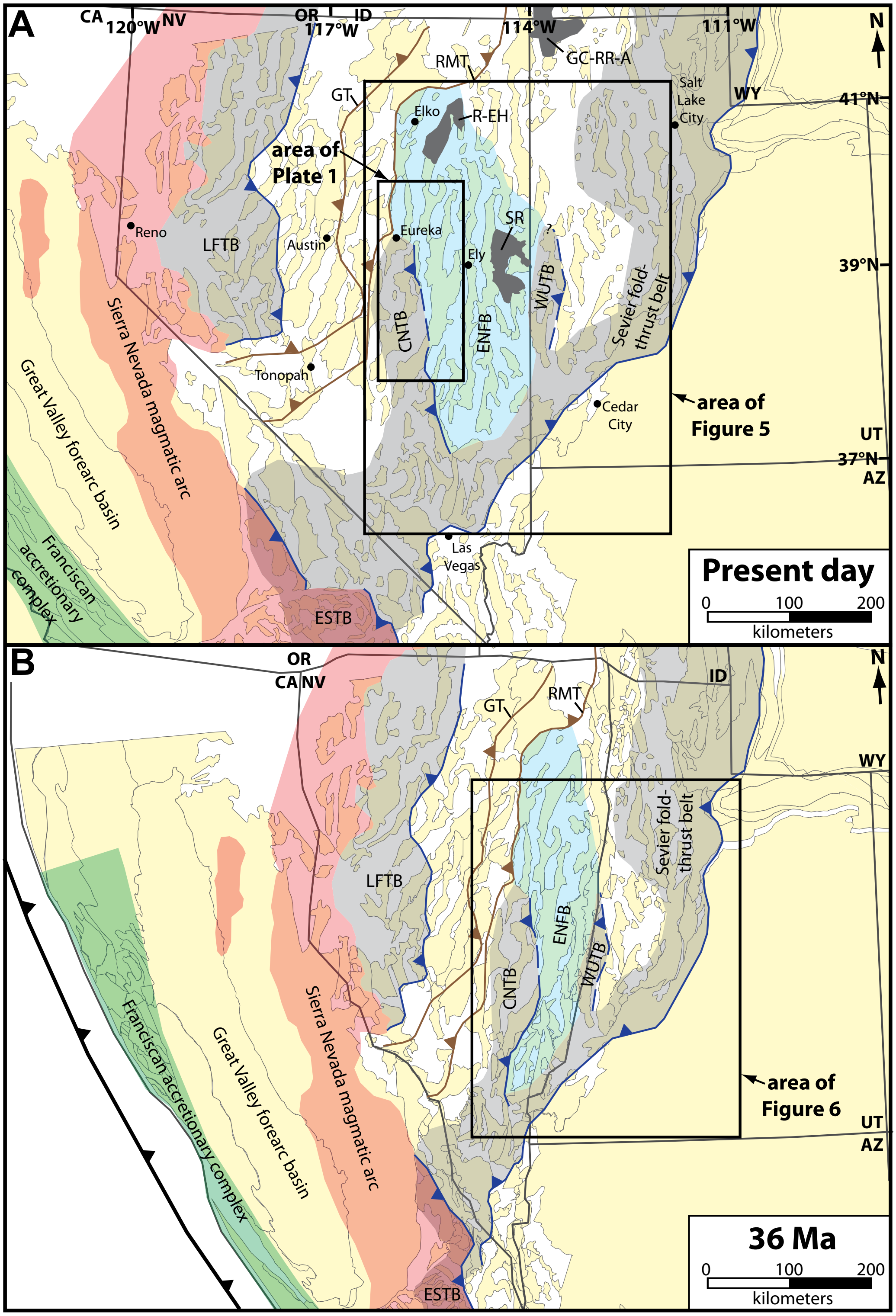


Figure 1. (A) Map of the Cordilleran retroarc region in Nevada, Utah, and eastern California, superimposed over a base map from McQuarrie and Wernicke (2005, their Fig. 7A), which shows present-day dimensions. Deformation fronts of Cordilleran thrust systems are shown, and spatial extents are shaded gray; locations are modified from Long et al. (2014), with additions from Greene (2014) and this study. Area of Eastern Nevada fold belt (ENFB) is shaded blue. Paleozoic–early Mesozoic thrust systems are shown in brown. Location of Sierra Nevada magmatic arc is from Van Buer et al. (2009) and DeCelles (2004). (B) Map of same area as A, but superimposed over a base map from McQuarrie and Wernicke (2005, their Fig. 9E) which shows a 36 Ma tectonic reconstruction. The map shows approximate pre-extensional dimensions of the Cordilleran retroarc region. Abbreviations: GT—Golconda thrust, RMT—Roberts Mountains thrust, LFTB—Luning-Fencemaker thrust belt; ESTB—Eastern Sierra thrust belt; CNTB—Central Nevada thrust belt; ENFB—Eastern Nevada fold belt; WUTB—Western Utah thrust belt; R-EH—Ruby–East Humboldt core complex; SR—Snake Range core complex; GC-RR-A—Grouse Creek–Raft River–Albion core complex. State abbreviations: CA—California, OR—Oregon, NV—Nevada, ID—Idaho, UT—Utah, WY—Wyoming, AZ—Arizona.

A 1:250,000 scale paleogeologic map of east-central Nevada, constructed by compiling formation-scale divisions of Paleozoic–Mesozoic rocks exposed beneath the Paleogene unconformity, is presented. Subcrop patterns defined on the map allow regional correlation of fold axes, and these are combined with published sedimentary thicknesses (Stewart, 1980) to estimate fold amplitude. The map is combined with dip magnitude maps for Paleozoic–Mesozoic and Tertiary rocks, which corroborate the correlation of fold axes and allow estimation of pre-Paleogene limb dips. Fold axes are superimposed over a pre-extensional tectonic reconstruction (McQuarrie and Wernicke, 2005), which illustrates the synorogenic map dimensions of individual structures and structural provinces and allows estimation of fold wavelength.

In addition, the subcrop map provides a detailed view of the map patterns of thrust faults and folds in the northern Central Nevada thrust belt, including the recently defined Eureka culmination (Long et al., 2014). These observations are combined with the structural synthesis of the Eastern Nevada fold belt and published cross sections of the Western Utah thrust belt (Greene, 2014) and Sevier thrust belt (DeCelles and Coogan, 2006) to present a model for contractional deformation in the Sevier hinterland.

TECTONIC SETTING

The Sevier thrust belt represents part of the Cordilleran retroarc thrust belt system, which extends from Mexico to Alaska, and formed during Jurassic to Paleogene contractional deformation of the North American plate above subducting oceanic plates (e.g., Allmendinger, 1992; DeCelles, 2004; Dickinson, 2004). The Cordilleran magmatic arc at the latitude of Nevada is represented by the Sierra Nevada batholith (Fig. 1; e.g., Coleman and Glazner, 1998; Ducea, 2001). East of the magmatic arc, the Luning-Fencemaker thrust belt (Fig. 1) is interpreted to record Jurassic closure of a backarc basin (e.g., Oldow, 1984; Wyld, 2002). The genetic relationship between the Luning-Fencemaker and Sevier thrust belts is debated; Wyld (2002) argued that these two systems represent deformation during distinct Jurassic and Cretaceous tectonic events, while DeCelles (2004) interpreted the Luning-Fencemaker thrust belt as an older, hinterland component of the Sevier orogenic wedge.

To the east of the Luning-Fencemaker thrust belt, there lie two zones affected by Paleozoic to early Mesozoic orogenic events. The oldest is the Mississippian Antler orogeny, in which Ordovician to Devonian, deep marine, sedimentary and volcanic rocks were emplaced eastward over shelf rocks across the Roberts Mountains thrust and related structures (Fig. 1; e.g., Burchfiel and Royden, 1991; Speed and Sleep, 1982; Dickinson, 2000), and the youngest is the Triassic Sonoma orogeny, where Devonian to Permian, deep marine, sedimentary and volcanic rocks were translated eastward over the Golconda thrust (Fig. 1; e.g., Oldow, 1984; Miller et al., 1992; Dickinson, 2000).

Farther to the east, the Central Nevada thrust belt and Western Utah thrust belt (Fig. 1) are zones of north-striking, dominantly east-vergent thrust faults and folds that branch off of the Sevier thrust belt, and each accommodated ~10 km of shortening (Bartley and Gleason, 1990; Taylor et al., 1993, 2000; Long, 2012; Greene, 2014; Long et al., 2014). Though deformation timing constraints are broad, the Central Nevada thrust belt and Western Utah thrust belt are interpreted to represent contemporary, interior components of the Sevier thrust system (Taylor et al., 2000; Greene, 2014; Long et al., 2014). The region of eastern Nevada between these two thrust belts, which is the subject of this paper, exhibits upper-crustal folding but lacks regional-scale thrust faults (e.g., Armstrong, 1972; Gans and Miller, 1983). However, this region did experience Cordilleran ductile deformation and metamorphism at midcrustal levels, as recorded by Late Cretaceous peak metamorphism and ductile fabrics in rocks now exposed in metamorphic core complexes and highly extended ranges (e.g., Miller et al., 1988; Miller and Gans, 1989; McGrew et al., 2000; Wells and Hoisch, 2008).

On the east end of the Cordilleran deformation system, the Sevier thrust belt accommodated ~220 km of upper-crustal shortening in western Utah (Currie, 2002; DeCelles and Coogan, 2006). The timing of initial deformation in the Sevier thrust belt is debated (e.g., Jordan, 1981; Wiltschko and Dorr, 1983; Heller et al., 1986; DeCelles, 2004; DeCelles and Coogan, 2006), but most agree that the onset of subsidence in the adjacent foreland basin, which is attributed to crustal thickening in the Sevier thrust belt, occurred by at least ca. 125 Ma (Early Cretaceous; Jordan, 1981). However, initial motion on the westernmost fault of the thrust belt has been argued to be as old as ca. 145 Ma (Jurassic-Cretaceous boundary; DeCelles, 2004; DeCelles and Coogan, 2006). Deformation in the Sevier thrust belt at this latitude continued through the Late Cretaceous and Paleocene (Lawton and Trexler, 1991; Lawton et al., 1993, 1997; DeCelles et al., 1995; DeCelles and Coogan, 2006), on the basis of geochronology, biostratigraphy, and structural relationships of foreland basin strata.

The hinterland of the Sevier thrust belt in western Utah and eastern Nevada is the hypothesized site of a relict orogenic plateau (e.g., Coney and Harms, 1984; Allmendinger, 1992; Jones et al., 1998; Dilek and Moores, 1999; DeCelles, 2004; Best et al., 2009). The existence of this plateau, termed the “Nevadaplano” after comparison to the modern Andean Altiplano-Puna (Allmendinger, 1992; Jones et al., 1998; DeCelles, 2004), is supported by restoration of Tertiary extension (Coney and Harms, 1984) and estimation of crustal shortening accommodated in the Sevier thrust belt (DeCelles and Coogan, 2006), which both suggest that crustal thicknesses of at least 50 km were achieved in the hinterland region by the end of shortening. These crustal thicknesses may have supported surface elevations as high as 3 km (DeCelles and Coogan, 2006; Snell et al., 2014).

METHODS

Paleogeologic Map

Plate 1 presents a paleogeologic map of east-central Nevada that shows the distribution of Paleozoic and Mesozoic sedimentary rocks exposed under a regional unconformity beneath Paleogene volcanic and sedimentary rocks (see Fig. 2 for a guide to range, valley, and other geographic names). This form of map removes the vertical effects of tectonism that postdate the unconformity (e.g., Armstrong, 1968), including differential uplift and subsidence associated with Neogene normal faulting. When combined with stratigraphic thickness data (Stewart, 1980) and structural data from published geologic maps, this map can be used to illustrate the map patterns of pre-Paleogene structures, and to estimate Paleogene structural relief, the amplitude of erosionally beveled folds, and stratigraphic throw on thrust faults at Paleogene erosion levels.

The map was constructed by first georeferencing 50 published geologic maps, the majority of which are between 1:24,000 and 1:62,500 in scale, and four 1:250,000 scale Nevada county geologic maps (see Plate 1 for guide to source maps). Then, following methods outlined in Long (2012), all locations of exposures of the Paleogene unconformity were compiled, and the ages of Paleozoic–Mesozoic sedimentary rocks that directly underlie the unconformity at each locality were divided out (where possible) at the formation scale. Across the majority of the map area, the oldest preserved Paleogene rocks are late Eocene to Oligocene volcanic rocks of the Great Basin ignimbrite flare-up (e.g., Armstrong and Ward, 1991; Best and Christiansen, 1991; Best et al., 2009). Less commonly, alluvial and lacustrine rocks of the Sheep Pass Formation, which have been interpreted to represent deposition within isolated extensional basins (Druschke et al., 2009a), are preserved beneath ignimbrite flare-up rocks. Within the map area, the Sheep Pass Formation is interpreted as Paleocene to Eocene in age (Hose and Blake, 1976; Kleinhampl and Ziony, 1985; Lund et al., 1988; Vandervoort and Schmitt, 1990; Fouch et al., 1991), although farther east in Nevada, it has been interpreted to be as old as Maastrichtian (Druschke et al., 2009b). Localities where late Eocene to Oligocene volcanic rocks versus Paleocene to Eocene sedimentary rocks overlie the unconformity are not differentiated on Plate 1, and therefore the total age range of unconformities compiled spans the Paleogene.

In total, 1024 individual locations of surface exposures of the Paleogene unconformity were compiled. In addition, in 809 places where the unconformity was not exposed, the age of the

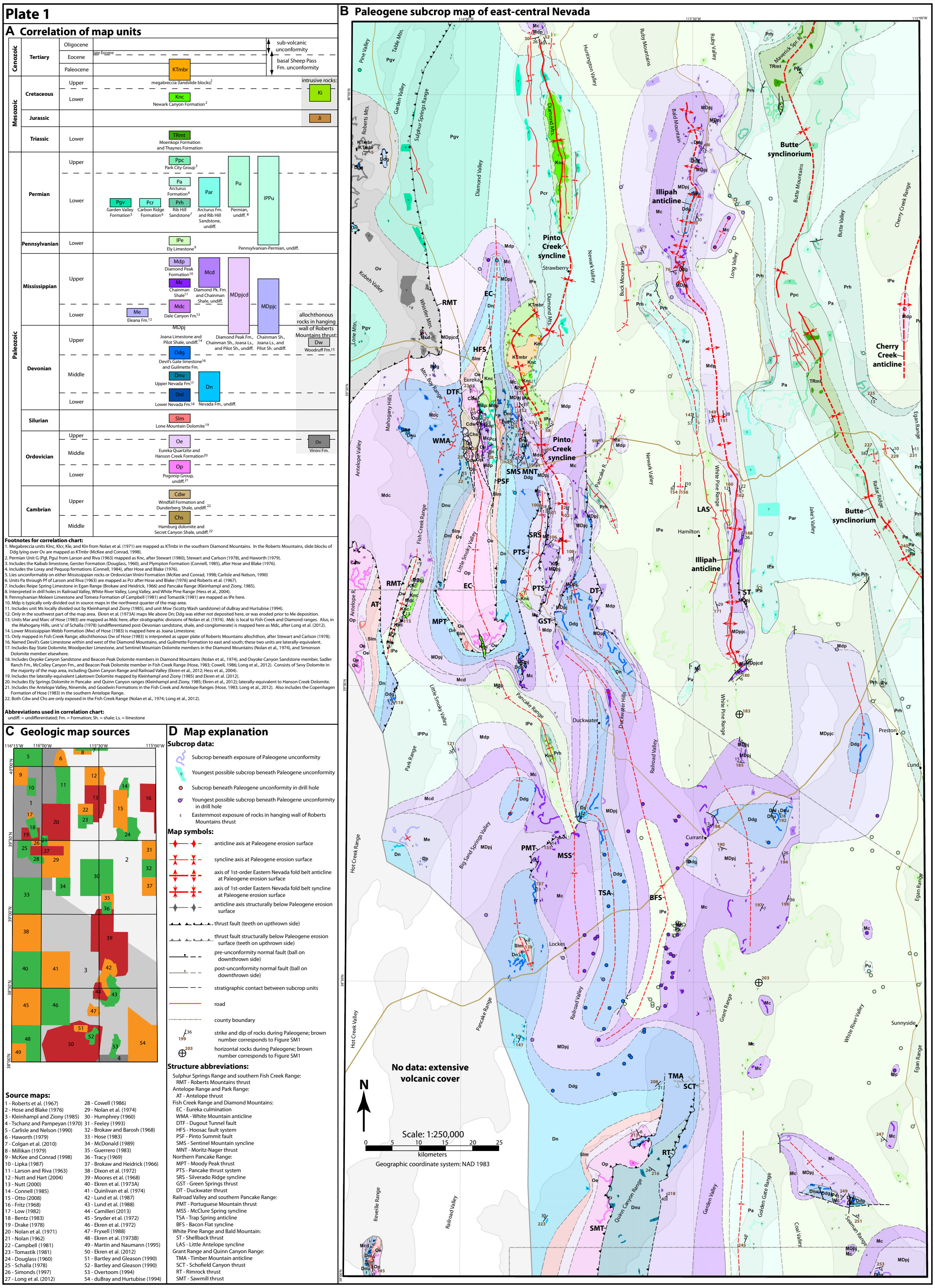


Plate 1. Paleogene subcrop map of east-central Nevada at 1:250,000 scale, with a correlation chart of map units, an index to geologic map sources, and an explanation of subcrop data, map symbology, and structure abbreviations. Strike and dip symbols show attitudes of rocks during the Paleogene, after retrodeformation of tilts of Tertiary rocks; brown numbers next to strike and dip symbols correspond to stereoplots in Figure SM1 in the supplemental file (see text footnote 1). For a full-sized PDF file of Plate 1, please visit http://dx.doi.org/10.1130/GES01102.S1 or the full-text article on www.gsapubs.org.

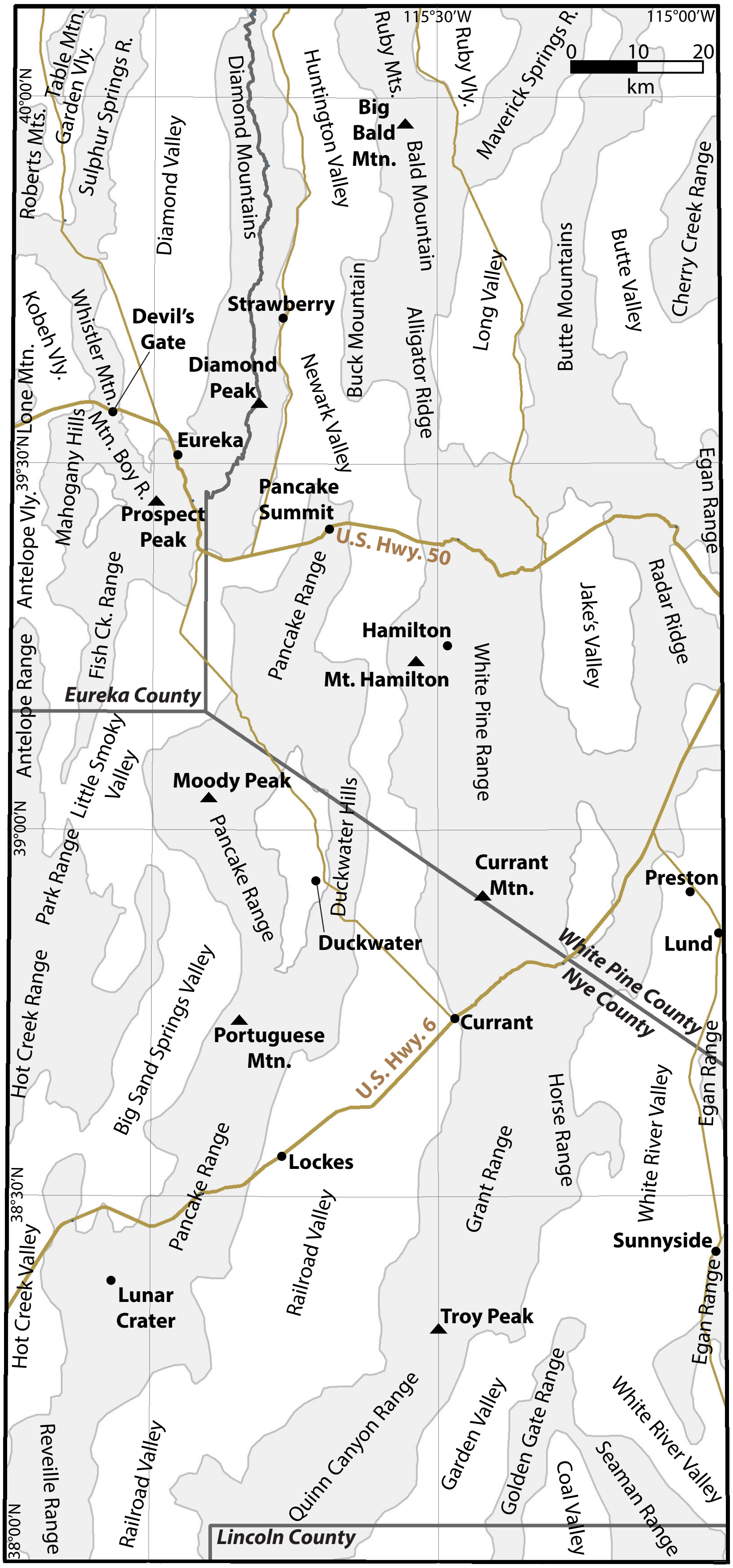


Figure 2. Map of the same area as Plate 1, showing relevant geographic names, including ranges, valleys, towns, peaks, roads, and county boundaries (abbreviations: R.—range, Mtn.—mountain, Mts.—mountains, Vly.—valley).

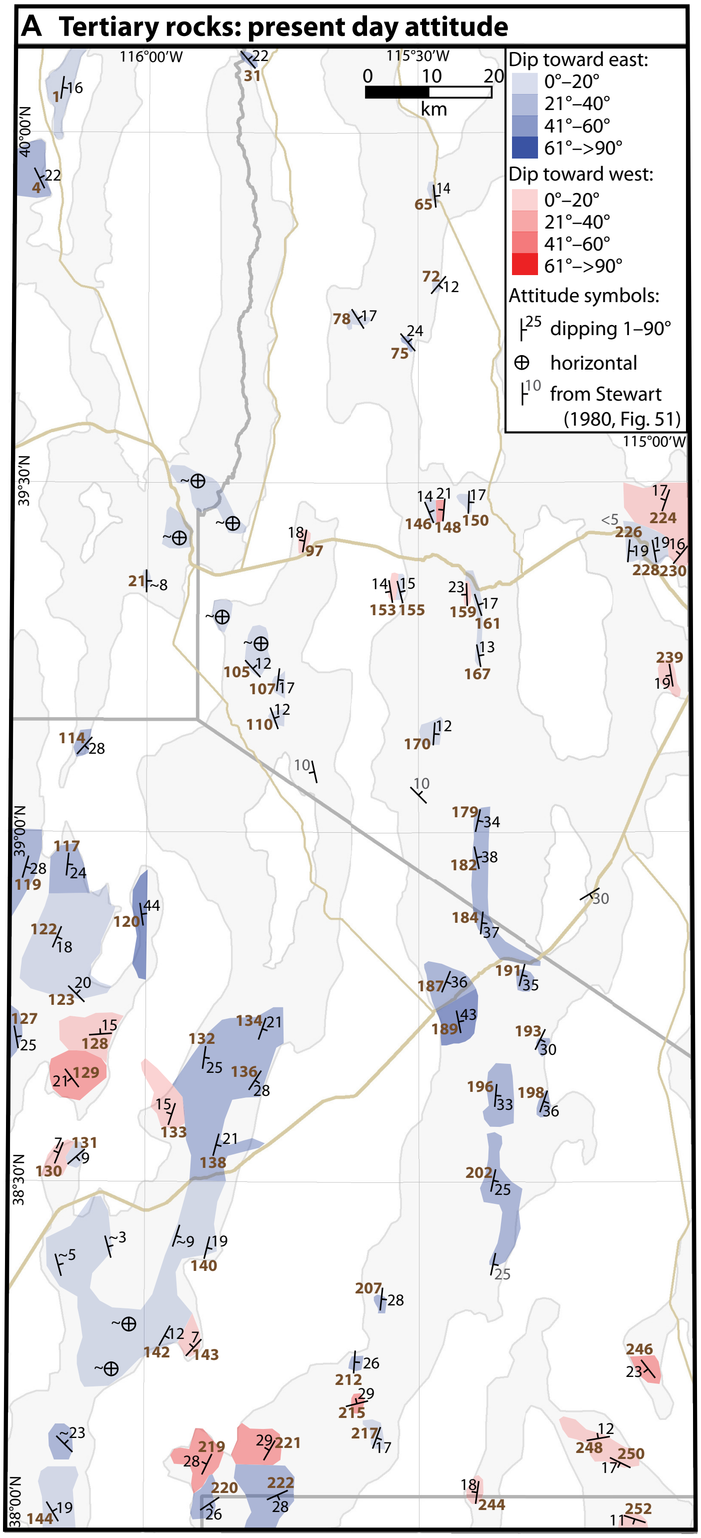
youngest pre-unconformity stratigraphic unit was compiled (data marked with a "Y" on Plate 1), which constrains the lowest possible stratigraphic level of the unconformity. Finally, formation top logs from 116 drill holes (Hess et al., 2004) were also used to locate the exact (n = 77) or lowest possible stratigraphic level (n = 39) of the Paleogene unconformity. All supporting subcrop and drill-hole data are shown on Plate 1, and accompanying ArcGIS data files and a pdf of the paleogeologic map that contains layers that can be turned on and off are included in the Supplemental File.[1]

Structural and stratigraphic contacts on Plate 1 were determined from interpretations and geometries shown in source mapping, and their locations were determined by interpolation between subcrop data points. The spatial density of subcrop data points is highly variable and is a function of multiple factors, including density of bedrock exposure and the degree of preservation of rocks above the Paleogene unconformity. Therefore, except in the places where structural and stratigraphic contacts are precisely located between closely spaced data points, contacts should be interpreted as approximately located. Fold axes shown on Plate 1 were precisely located off of source mapping, where possible, and are based on interpolation of subcrop data points elsewhere. The formation-scale resolution of subcrop data, as well as incorporation of structural data from detailed source maps, allowed mapping of faults and fold axes in many areas where the lower-resolution subcrop map of Long (2012) could not.

Plate 1 differs from a standard geologic map in several ways, and therefore several assumptions must be defined for interpretation of this map: (1) It is assumed that all rocks are assigned to the correct geologic formation on source maps, and that interpretations of stratigraphic and structural contacts on source maps are correct; (2) using stratigraphic thicknesses

[1]Supplemental File. Zipped file containing three supplemental figures, a word document, and a zipped file containing an ArcGIS MXD and accompanying shape files. If you are viewing the PDF of this paper or reading it offline, please visit http://dx.doi.org/10.1130/GES01102.S2 or the full-text article on www.gsapubs.org to view the Supplemental File.

Figure 3 (*on this and following page*). Dip magnitude maps of the same area as Plate 1 (see Plate 1 for structure abbreviations). Colored polygons represent areas on source mapping where multiple attitude measurements demonstrate that rocks consistently have a similar attitude. Areas with variable attitude or insufficient attitude data are not shown. (A) Dip magnitude map showing present-day attitudes of Tertiary rocks. Strike and dip symbols represent the mean attitude of multiple measurements, which are plotted on Figure SM1 (see text footnote 1); brown numbers next to strike and dip symbols correspond to individual stereoplots in Figure SM1 (see text footnote 1).



combined with subcrop patterns to estimate structural relief and fold amplitude requires assuming that topographic relief was small relative to structural relief during the Paleogene; and (3) the subcrop map does not remove the horizontal effects of postunconformity tectonism, including translation that accompanied Neogene extension.

Dip Magnitude Maps

Dip magnitude maps of the same area as Plate 1, which show present-day attitude, color coded by dip direction and dip angle, are shown for Tertiary rocks and Paleozoic–Mesozoic rocks in Figures 3A and 3B, respectively. These maps were constructed by compiling over 4200 bedding measurements from the geologic maps cited on Plate 1, and they highlight areas where multiple attitude measurements demonstrate that Tertiary or Paleozoic–Mesozoic rocks have a consistent attitude. Strike and dip symbols shown on Figure 3A represent the mean attitude of multiple measurements, which are shown in individual stereoplots in Figure SM1 in the supplemental file (strike and dip data for Paleozoic–Mesozoic rocks are shown in Fig. SM2 in the supplemental file [see footnote 1]). The Paleozoic–Mesozoic dip magnitude map (Fig. 3B) corroborates the mapping and regional correlation of folds performed on Plate 1, and it allows quantitative description of present-day fold limb dips, which are summarized in Table 1.

In 71 locations, Paleozoic–Mesozoic rocks were restored to their Paleogene attitude by rotating the mean attitude of Tertiary rocks to horizontal, and rotating Paleozoic–Mesozoic rocks by the same amount. The resulting Paleogene attitudes are shown as strike and dip symbols on Plate 1 and Figure 3C. Though they are distributed across a large region, these data provide quantitative estimates of the local pre-extensional limb dip magnitudes of folds, which are summarized on Table 1. In addition, histograms plotting the difference in dip angle across the Paleogene unconformity within the Eastern Nevada fold belt and Central Nevada thrust belt are shown on Figure 4, and include an additional 17 data points calculated from the subcrop map of Gans and Miller (1983).

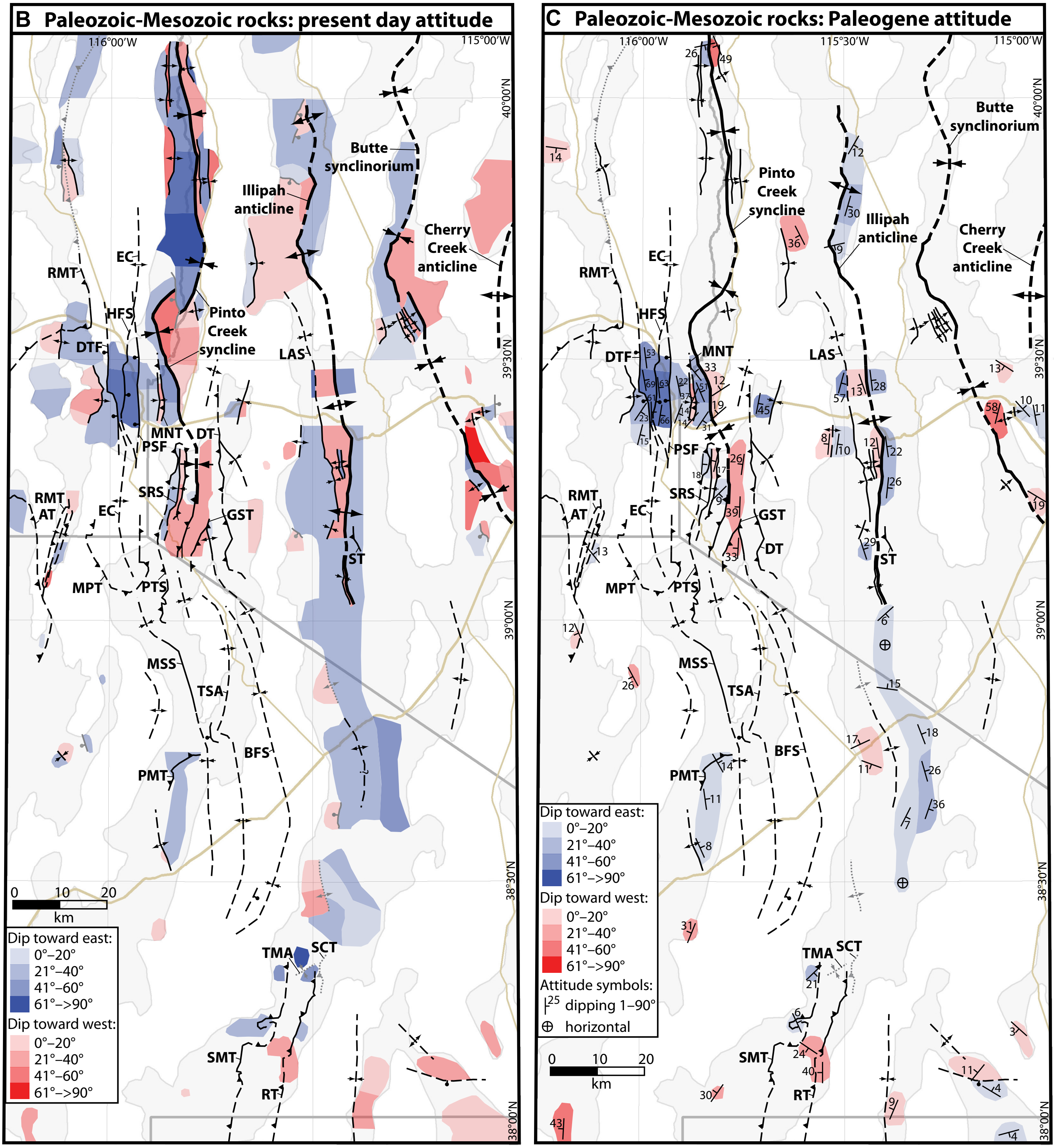


Figure 3 (*continued*). (B) Dip magnitude map showing present-day attitudes of Paleozoic and Mesozoic rocks. Strike and dip symbols are omitted here for simplicity; strike and dip symbols referenced to specific stereoplots are shown in Figure SM2 in the Supplemental File (see text footnote 1). First-order folds of the Eastern Nevada fold belt are emphasized by thick fold symbols. (C) Dip magnitude map showing Paleogene attitudes of Paleozoic and Mesozoic rocks, calculated by rotating Tertiary rocks to horizontal, and rotating Paleozoic–Mesozoic rocks in that locality by the same amount. See Plate 1 for a guide to stereoplots referenced to Figure SM1 (see text footnote 1) for each attitude symbol. First-order folds of the Eastern Nevada fold belt are emphasized by thick fold symbols.

Pre-Extensional Tectonic Reconstruction Base Map

On Figure 5, the locations of thrust faults, fold axes, and the boundaries of structural provinces compiled from Plate 1 and from published subcrop and geologic maps (Gans and Miller, 1983; Colgan and Henry, 2009; Colgan et al., 2010; Long, 2012; Greene, 2014) are superimposed over a base map containing polygons from McQuarrie and Wernicke (2005, their fig. 7A), which shows present-day map dimensions. On Figure 6, these structures and provinces are superimposed over a base map containing polygons from McQuarrie and Wernicke (2005, their Fig. 9E), which shows a regional tectonic reconstruction of extension at 36 Ma. The locations of structures and province boundaries on Figure 6 were taken directly from their locations on the range polygons of Figure 5. Therefore, Figure 6 illustrates the approximate synorogenic dimensions of individual structures and structural provinces, and it facilitates estimation of the pre-extensional wavelength of folds, which are summarized on Table 1. It is acknowledged that estimation of the amount of Cenozoic extension across the Great Basin is an ongoing process and will undoubtedly be refined with future research; however, the McQuarrie and Wernicke (2005) reconstruction is the most recent and comprehensive tectonic reconstruction available to date.

STRATIGRAPHY

Plate 1 shows that rocks ranging between Cambrian and Cretaceous in age were exposed at the surface during the Paleogene. The Cambrian through Triassic rocks are part of a composite sedimentary section that records semi-continuous deposition from the Neoproterozoic to the Triassic (e.g., Stewart, 1980). From Late Neoproterozoic to Devonian time, a westward-thickening section of clastic and carbonate rocks that approaches thicknesses of 10 km was deposited on the rifted western Laurentian continental shelf in western Utah and eastern Nevada (e.g., Stewart and Poole, 1974). The lower part of the section consists of >5 km of Neoproterozoic to Early Cambrian clastic rocks (Stewart, 1980). These rocks are presently exposed in several deeply exhumed ranges in western Utah and eastern Nevada (Stewart and Carlson, 1978; Hintze et al., 2000); however, Paleogene erosion levels were not deep enough to expose these rocks anywhere on Plate 1. The Middle Cambrian through Devonian section represents the upper ~4–5 km of the passive-margin basin (Stewart, 1980) and is dominated by carbonate rocks. On the area of Plate 1, Cambrian rocks

TABLE 1. DATA FOR FOLDS OF THE EASTERN NEVADA FOLD BELT, CENTRAL NEVADA THRUST BELT, AND WESTERN UTAH THRUST BELT

Structure	Paleogene erosion level of western limb	Paleogene erosion level of eastern limb	Deformation timing constraints	Modern western limb dip	Modern eastern limb dip
Central Nevada thrust belt					
Eureka culmination*	Op to Mdc†	Chs to Pcr†	Aptian; syn-Knc*	10–15°W	40–70°E
McClure Spring syncline#	Ddg to Prh, Knc	Ddg to Prh, Knc	folds Prh; cut by undated Knc#	25–40°E (≥90°E#)	unknown (50°W#)
Trap Spring anticline††	Ddg to Prh, Knc	Ddg to IPe	folds Prh#	unknown	unknown
Bacon Flat syncline††	Ddg to IPe	Dn to IPe	folds IPe	unknown	20–35°W
Eastern Nevada fold belt: First-order folds					
Pinto Creek syncline	Mc to Pcr, Knc	Ddg to Pcr, Knc	folds Aptian Knc*	30° to ≥90°E	20–50°W
Illipah anticline	Ddg to Pcr/Prh	Ddg to TRmt	folds TRmt	20–35°W	20–40°E
Butte synclinorium	Ddg to TRmt	Mdp to TRmt	folds TRmt	20–50°E	25°to ≥90°W
Cherry Creek anticline	Mdp to TRmt	Mdp to Lower Jurassic†††	folds Lower Jurassic rocks§§§	25–30°W	unknown
Pequop synclinorium	Mdp to Lower Jurassic†††	Miss. to Lower Jurassic††	folds Lower Jurassic rocks§§§	unknown	unknown
Western Utah thrust belt					
Confusion synclinorium****	Miss. To Lower Triassic†††	Triassic and deeper††††	folds Lower Triassic rocks	50°to ≥90°E††††	15–40°W††††
Eastern Nevada fold belt: Second-order folds					
Location/structure	**Range of Paleogene erosion levels in limbs**	**Age constraints**	**Range of modern limb dips**	**Range of Paleogene limb dips**	**Range of amplitudes (m)**
Northern Diamond Mts.	IPe to Pcr, Knc	fold Pcr and undated Knc	25–80°	unknown	500–1500
Little Antelope syncline	MDpj to Pa	folds Pa	20–65°E	10–55°E	600–1050
Northern White Pine R.	Mdp to IPe	fold IPe	25–40°	unknown	300–600
Central Butte Mountains	Prh to TRmt	fold TRmt	15–30°	unknown	250–700

(*continued*)

TABLE 1. DATA FOR FOLDS OF THE EASTERN NEVADA FOLD BELT, CENTRAL NEVADA THRUST BELT, AND WESTERN UTAH THRUST BELT (*continued*)

Structure	Paleogene western limb dip	Paleogene eastern limb dip	Amplitude (m)	Modern wavelength (km)	Paleogene wavelength (km)	Traceable length (km)	Map data sources on Plate 1
Central Nevada thrust belt							
Eureka culmination*	30–40°W§	20–40°E§	4300–5000*	25–30	19–22*	80	3, 20, 27, 29
McClure Spring syncline#	10–15°E	unknown	1100–2600**	12–19	12–17	65	3, 41, 46
Trap Spring anticline††	unknown	unknown	600–2600**	13–17	13–15	70	3
Bacon Flat syncline††	unknown	10–15°W	600–1500**	6–22	5–14	70	3, 30, 39, 42, 44, 47
Eastern Nevada fold belt: First-order folds							
Pinto Creek syncline	20–30°E; up to 80°E##	10–40°W	1400–3000§§	30–45	25–35	95	6, 11, 20, 27, 29, 34
Illipah anticline	15–35°W	15–30°E	1200–3800**	37–62	35–50	105	2, 12, 13, 29–30, 35, 36
Butte synclinorium	15–30°E	15–60°W	2300–3800**	33–46	25–45	250***	2, 14, 15, 24, 31–32, 37
Cherry Creek anticline	15–20°W	10–35°E###	2300–4000**	26–65	15–40	110†††	16
Pequop synclinorium	10–35°E###	25–30°W###	2500–4000**	≥40	≤25	115†††	–
Western Utah thrust belt							
Confusion synclinorium****	–	–	2500–3000††††	–	–	130****	–
Eastern Nevada fold belt: Second-order folds							

Location/structure	Range of modern wavelengths (km)	Traceable lengths (km)	Map data sources on Plate 1
Northern Diamond Mts.	2–6	10–15	6, 11, 20
Little Antelope syncline	>4–7	50	2, 13, 23, 30, 35, 36
Northern White Pine R.	2–9	8–10	30, 35, 36, 39
Central Butte Mountains	1–4	6–10	2, 24

Note: Unit abbreviations shown on part A of Plate 1.

*Defined in Long et al. (2014). Amplitude and wavelength values given here represent pre-extensional estimate from Long et al. (2014).

†Erosion levels listed here postdate high-throw (2–5 km) normal faulting that predates the Paleogene unconformity (Long et al., 2014).

§Limb dip estimates shown here account for restoration of ~20°–30°of eastward tilting that accompanied pre-unconformity normal faulting, after Long et al. (2014).

#Field relationships of Knc and modern limb dips are described in Perry and Dixon (1993), and correspond only to the part of the fold that preserves Prh and Knc, 10 km west of the town of Duckwater.

**Based on regional stratigraphic thickness data from Stewart (1980), in addition to stratigraphic thicknesses from the cited map sources.

††Originally described by French (1998) based on subcrop patterns in Railroad Valley drill holes, and named here.

§§Estimated based on amplitude of fold measured in cross sections in cited source maps, because of sparse data on the erosion level of the Paleogene unconformity in the Diamond Mountains.

##Approximated based on retrodeformation of ~15°–25°of eastward tilting of the Diamond Mountains, based on tilts of Tertiary rocks in the Sulphur Spring Range, Bald Mountain, and the north end of the Diamond Mountains.

***Total traceable map length estimated from the subcrop map of Long (2012).

†††Map pattern estimated from subcrop maps in this study, Gans and Miller (1983), and Long (2012).

§§§Based on mapping of rocks near Currie, Nevada, that are correlated with the Lower Jurassic Aztec sandstone of southern Nevada (Stewart and Carlson, 1978; Coats, 1987).

###Estimated from the subcrop map of Gans and Miller (1983).

****Interpreted by Greene (2014) as a structural low created by folding above east-vergent thrust faults. Traceable length from Greene (2014, and references therein).

††††Measured from cross sections of Greene (2014). Amplitude and Paleogene erosion levels reported for western limb only, because eastern limb has been modified by growth of the Sevier culmination.

were only exposed in the Fish Creek Range, and include the Secret Canyon Shale, Hamburg Dolomite, Dunderberg Shale, and Windfall Formation (see Plate 1 for a correlation chart of map units). Ordovician rocks were exposed in the Fish Creek, Antelope, Pancake, and Quinn Canyon Ranges and consist of the limestone-dominated Pogonip Group, the Eureka Quartzite, and the Hanson Creek Dolomite. The Silurian section consists of the Lone Mountain Dolomite. Devonian rocks are divided into the lower and upper Nevada Formation, which is dominated by dolomite, the Guilmette Formation/Devil's Gate limestone, and the Pilot Shale.

During the Mississippian Antler orogeny, up to ~2–3 km of conglomerate, shale, and sandstone were deposited in a foreland basin that subsided to the east of an orogenic highland produced from east-vergent contractional deformation (e.g., Poole, 1974; Poole and Sandberg, 1977; Speed and Sleep, 1982). The western third of Plate 1, immediately east of the trace of the Roberts Mountains thrust, was the approximate axis of maximum subsidence in the foreland basin (Stewart, 1980). Mississippian rocks of the Antler foreland basin include the Dale Canyon Formation, Chainman Shale, and Diamond Peak Formation.

After the Antler orogeny, ~2–4 km of Pennsylvanian to Triassic, dominantly shallow-marine carbonate rocks were deposited on the continental shelf (Rich, 1977; Stevens, 1977; Collinson and Hasenmueller, 1978; Stewart, 1980). Eastern Nevada experienced a protracted series of uplift and erosion events during the Pennsylvanian and Permian (Trexler et al., 2004), which are recorded by several unconformities within this part of the section. The Pennsylvanian section consists of the Ely Limestone; the Permian section consists of the Garden Valley Formation, Carbon Ridge Formation, and Rib Hill Sandstone, which are laterally equivalent (Hose and Blake, 1976; Roberts et al., 1967), and the Arcturus Formation and Park City Group. The Triassic section consists of the Thaynes and Moenkopi Formations.

The Early Cretaceous Newark Canyon Formation is preserved in several isolated exposures in the western part of Plate 1, unconformably overlying Mississippian, Pennsylvanian, and Permian rocks. The Newark Canyon Formation has been interpreted to represent deposition contemporary with contractional deformation in the Central Nevada thrust belt (Vandervoort and Schmitt, 1990; Druschke et al., 2010; Long et al., 2014). At its type section in the southern Diamond Mountains, the Newark Canyon Formation has been dated as Aptian (ca. 122–116 Ma) by U-Pb zircon geochronology (Druschke et al.,

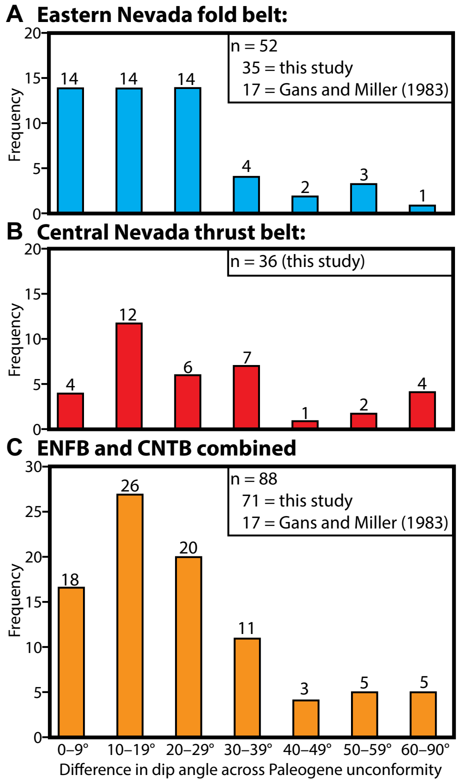


Figure 4. Histograms showing the difference in dip angle between Paleozoic–Mesozoic rocks and Tertiary rocks across the Paleogene unconformity, from (A) the Eastern Nevada fold belt; (B) the Central Nevada thrust belt; and (C) for both of these provinces combined. Strike and dip symbols showing data from this study are on Plate 1 and Figure 3C, and raw data used in calculating these attitudes are shown in stereoplots in Figure SM1 (see text footnote 1). Seventeen data points for the Eastern Nevada fold belt were calculated from the subcrop map of Gans and Miller (1983). These data indicate that the pre-Paleogene dip of fold limbs and other areas of homoclinally dipping rocks within the Eastern Nevada fold belt and Central Nevada thrust belt typically range between 10° and 40°, which is significantly higher than the ≤10°–15° range proposed by early mapping studies (e.g., Young, 1960; Kellogg, 1964; Moores et al., 1968).

2010), and deposition in the Fish Creek Range can be narrowed to the Barremian-Albian (ca. 130–100 Ma) on the basis of biostratigraphy (Fouch et al., 1979; Hose, 1983). However, rocks mapped as Newark Canyon Formation in several localities in the Pancake Range (McDonald, 1989; Carpenter et al., 1993; Perry and Dixon, 1993) and in the northern Diamond Mountains (Stewart and Carlson, 1978) lack precise depositional age constraints.

CONTRACTIONAL STRUCTURES

The following sections discuss the map patterns and geometry of regional-scale folds and thrust faults of the Eastern Nevada fold belt, Central Nevada thrust belt, and Western Utah thrust belt. Several structures discussed here, such as the Butte synclinorium and Eureka culmination, have been described in previous studies; however, many folds and thrust faults are named, described, and regionally correlated for the first time here. Data on Paleogene erosion levels, geometry, and deformation timing constraints for folds and thrust faults are summarized in Tables 1 and 2, and detailed descriptions of how these structures were mapped, correlated, and defined on the basis of source mapping are included in the Supplemental File (see footnote 1). In most cases, the pre-Paleogene amplitudes of folds were determined based on the total range of subcrop units exposed in each limb, combined with regional stratigraphic thicknesses from Stewart (1980). In some cases, the amplitudes of folds were estimated from cross sections accompanying source maps. Modern and pre-Paleogene wavelength ranges for folds were estimated from the distance between the axes of adjacent first-order folds on Figures 5 and 6, respectively. Throw ranges on thrust faults were estimated from the stratigraphic levels of the hanging wall and footwall, combined with stratigraphic thicknesses from Stewart (1980).

Roberts Mountains Thrust

The Roberts Mountains thrust, which emplaced deep-water sedimentary rocks over continental shelf rocks during the Mississippian Antler orogeny (e.g., Speed and Sleep, 1982; Burchfiel and Royden, 1991), is exposed in the western part of Plate 1. North of Devil's Gate, the Roberts Mountains thrust places the Ordovician Vinini Formation over Devonian and Mississippian rocks. During the Paleogene, the Permian Garden Valley Formation, which stratigraphically overlaps the Roberts Mountains thrust, was exposed across much of the northwest corner of the map area. In the southern Fish Creek Range, the Devonian Woodruff Formation is exposed in a klippe above the Roberts Mountains thrust, overlying Mississippian rocks (Stewart and Poole, 1974; Stewart and Carlson, 1978; Stewart, 1980; Hose, 1983).

Eastern Nevada Fold Belt

The Eastern Nevada fold belt is defined as the 100–150-km-wide region between the Central Nevada thrust belt and Western Utah thrust belt (Fig. 5), which restores to a pre-extensional width of 50–100 km (Fig. 6). Five north-trending folds within this province can be traced for map distances between 100 and 250 km, with amplitudes of ~2–4 km and wavelengths of ~20–40 km, and are classified here as first-order folds (Table 1). In several areas, including the northern Diamond Mountains (Haworth, 1979; Larson and Riva, 1963), central White Pine Range (Humphrey, 1960; Guerrero, 1983), and southern Butte Mountains (Douglass, 1960), subsidiary, second-order folds are present, and they can be traced for map distances typically ≤10 km, have amplitudes typically ≤1 km, and have wavelengths of ~1–10 km (Table 1). The five first-order folds of the Eastern Nevada fold belt are described next, from west to east.

Pinto Creek Syncline

A syncline that preserves rocks as young as Permian and Cretaceous in its hinge zone can be traced for ~95 km through the Diamond Mountains and the northern Pancake Range (Plate 1). The portion of this fold in the southern Diamond Mountains was named the Pinto Creek syncline (Nolan et al., 1974), and this name is applied here to its full length. In the Diamond Mountains, the eastern limb dips 20°–40° west, and the dip of the western limb varies between 25°E to 80°W (overturned). In the Pancake Range, the eastern limb dips 20°W, and the western limb dips 15°E. Retrodeformation of tilts of Tertiary rocks indicates original western and eastern limb dips of 20°–30°E and 30°–40°W, respectively, although areas of the western limb may have dipped as steeply as 80°E (Table 1). The amplitude of the Pinto Creek syncline is between 1400 and 3000 m, and the pre-Paleogene wavelength was 25–35 km. The syncline folds the Aptian (ca. 116–122 Ma) Newark Canyon Formation in the southern Diamond Mountains (Long et al., 2014), and it folds undated rocks mapped as Newark Canyon Formation in the northern Diamond Range (Stewart and Carlson, 1978). Long et al. (2014) interpreted the portion of the Pinto Creek syncline in the southern Diamond Mountains as the frontal axis of a hanging-wall ramp above the Ratto Canyon thrust, which grew synchronous with Newark Canyon Formation deposition. However, the construction mechanism for this fold farther

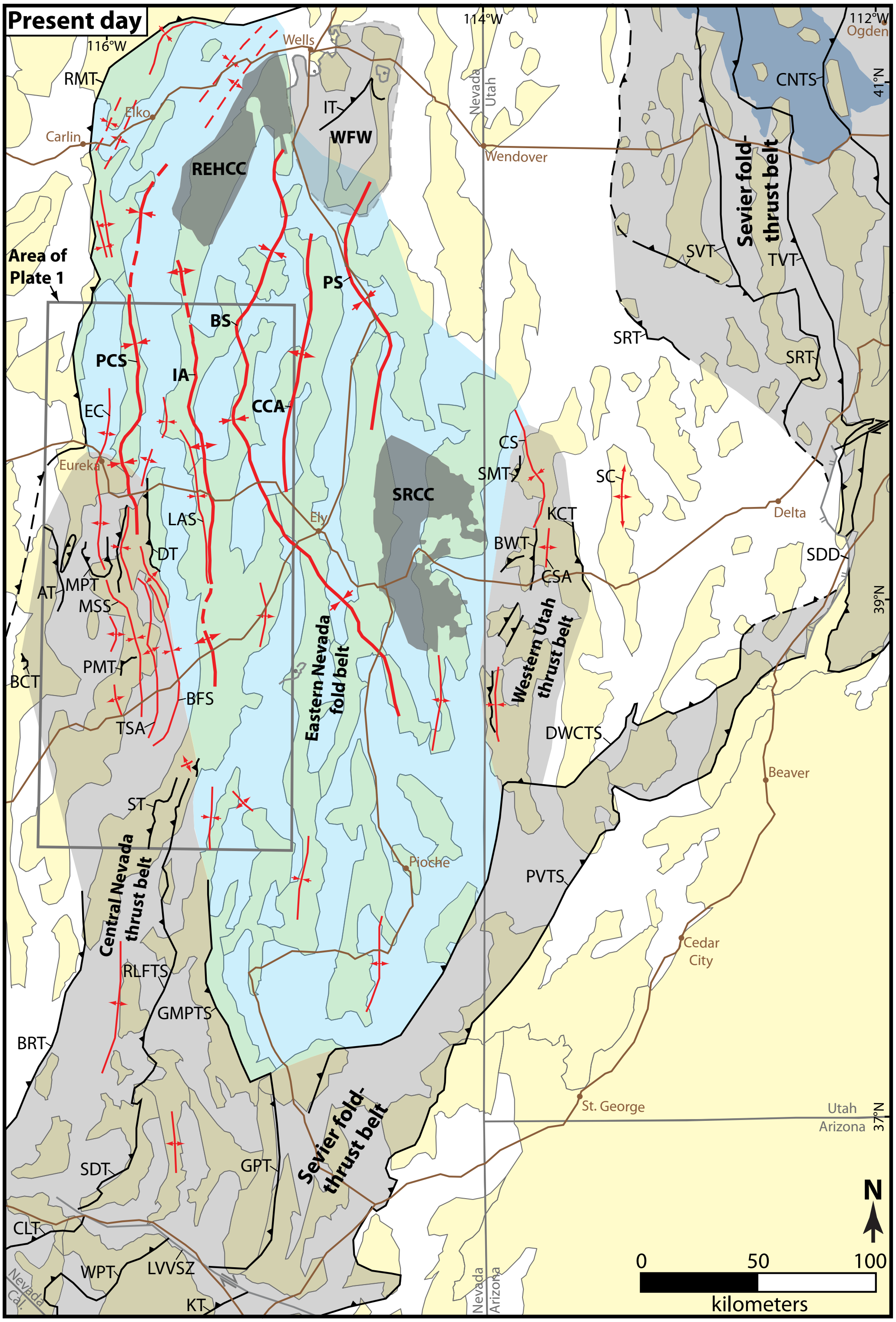


Figure 5. Map of eastern Nevada and western Utah (location shown on Fig. 1A) showing structural provinces of the Sevier hinterland, superimposed over a base map from McQuarrie and Wernicke (2005, their Fig. 7A) that shows present-day dimensions. Structure abbreviations are shown on Figure 6, except for: REHCC—Ruby–East Humboldt core complex, SRCC—Snake Range core complex, and WFW—area interpreted as exhumed footwall of Windermere thrust (Camilleri and Chamberlain, 1997). For areas that fall outside of the area of Plate 1, the locations of structures are compiled from Long (2012) for the Sevier thrust belt, southern part of the Central Nevada thrust belt, and southern part of the Eastern Nevada fold belt; Gans and Miller (1983) for the eastern part of the Eastern Nevada fold belt; Long (2012), Colgan and Henry (2009), and Colgan et al. (2010) for the northern part of the Eastern Nevada fold belt; and Greene (2014) for the Western Utah thrust belt.

north in the Diamond Mountains is unclear, as the geometry changes to a much tighter fold with multiple subsidiary folds, and no thrust faults are exposed through a complete section of Silurian through Permian rocks in its western limb (Nolan et al., 1971).

Illipah Anticline

An anticline that preserves Mississippian and Devonian rocks in its hinge zone can be traced for ~105 km through the White Pine Range, Alligator Ridge, and Bald Mountain (Plate 1). The portion of this fold in the northern White Pine Range was named the Illipah anticline (Humphrey, 1960), and this name is applied here to its full length. The western limb dips 20°–35°W, and the eastern limb dip varies between 15°E and 45°E. Retrodeformation of tilts of Tertiary rocks indicates original limb

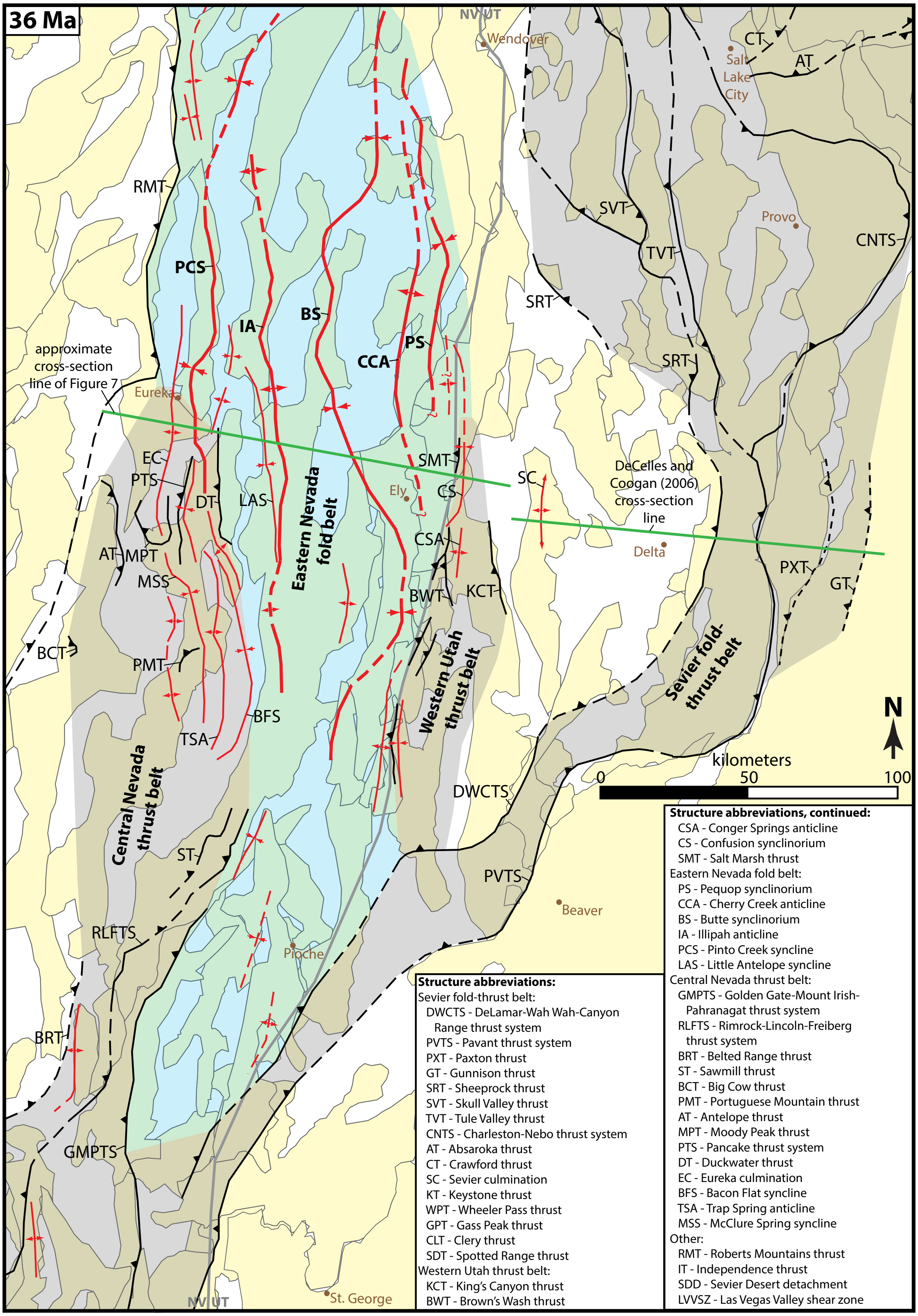


Figure 6. Map of eastern Nevada and western Utah, superimposed over a base map from McQuarrie and Wernicke (2005, their Fig. 9E) that shows a 36 Ma tectonic reconstruction. The map shows the approximate pre-extensional dimensions of structural provinces of the Sevier hinterland. The restored locations of all structures are taken from their locations over polygons on Figure 5.

TABLE 2. DATA FOR THRUST FAULTS OF THE CENTRAL NEVADA THRUST BELT

Structure	Paleogene stratigraphic level of hanging wall	Paleogene stratigraphic level of footwall	Deformation timing constraints	Stratigraphic throw at Paleogene erosion level	Traceable length (km)	Map sources on Plate 1
Pre-Cordilleran structures:						
Roberts Mountains thrust	Ov	Dn, Ddg, MDpjcd	Mississippian*; overlapped by Pgv	unknown	≥75 km	1, 5, 9–10, 18–19
Roberts Mts. thrust (klippe)[†]	Dw	Mdc	Mississippian*; overlapped by IPe	unknown	≥18 km	33
Central Nevada thrust belt:						
Ratto Canyon thrust (blind)[§]	low-mid Cambrian[§]	Slm	Aptian (ca. 116–122 Ma); syn-Knc[§]	2000–3500 m[§]	≥40 km[§§§]	27
Moritz-Nager thrust	Dnl, Dnu, Ddg	Ddg, MDpj, Mdc, Mc, Mdp	Aptian or younger; syn- or post-Knc[§]	1000–1300 m	15 km	27, 29
Antelope thrust	Op, Oe, Slm, Dn, Ddg	Mdc, IPe, IPPu	cuts IPPu	900–2600 m[#]	37 km	33, 38
Moody Peak thrust**	Slm, Oe, Op	Dn	cuts undated rocks mapped as Knc**	700–1750 m	≥40 km[§§§]	3, 33
Pancake thrust system	Ddg, Mc	Mc, Mdp	cuts Mdp; cut by 108±3 Ma pluton[††]	300–1300 m	35 km	3, 29, 34
Green Springs thrust	Ddg, MDpj	Mc, Mdp	cuts Mdp; overlapped by undated Knc[§§]	600–1500 m	10 km	34
Duckwater thrust[##]	Ddg, MDpj	Mdp, IPe, Prh	cuts Prh	1500–2100 m***	40 km	2
Portuguese Mtn. thrust[†††]	Ddg, MDpj	Mc	cuts Mc	250–1350 m	10 km	41
Sawmill thrust	Op, Oe	Dnu, Ddg	cuts Ddg	1250–3050 m	20 km	50–51
Rimrock thrust	Dnu	Ddg, MDpj	cuts Penn. rocks; cut by ca. 90–98 Ma pluton[###]	650–1200 m	185 km****	50, 52
Schofield Canyon thrust	Op, Oe[§§§§]	upper Cambrian[§§§§]	cuts Ord. rocks; cut by 86±5 Ma pluton[††††]	2700 m[§§§§]	5 km	47

Note: Unit abbreviations shown on part A of Plate 1.
*For example, Smith and Ketner (1968), Johnson and Pendergast (1981), Burchfiel and Royden (1991), Speed and Sleep (1982).
[†]Based on interpretation of Dw as part of Roberts Mountains allochthon by Stewart and Poole (1974), Stewart and Carlson (1978), and Hose (1983).
[§]Described in Long et al. (2014); 2000–2500 m throw where drilled; 3500 m throw estimated further to north where rocks as deep as Lower Cambrian are in hanging wall.
[#]Based on stratigraphic thickness data from Stewart (1980), in addition to stratigraphic thicknesses from the cited map sources.
**Named by Carpenter et al. (1993, p. 65), the authors state that the Moody Peak thrust cuts Knc; however, these deposits lack biostratigraphic age control.
[††]The 108 ± 3 Ma (K-Ar biotite) Puma Hill stock cuts structures associated with the Pancake thrust system (Nolan et al., 1974; McDonald, 1989).
[§§]McDonald (1989) states that undated conglomerate mapped as Knc overlaps the Green Springs thrust.
[##]Described by Carpenter et al. (1993), based on reinterpretation of the mapping of Hose and Blake (1976).
***Estimated using stratigraphic thicknesses from Humphrey (1960), Nolan et al. (1974), and Stewart (1980).
[†††]Mapped as an unnamed thrust fault by Quinlivan et al. (1974), and named here. Alternatively interpreted as a normal fault by Perry and Dixon (1993).
[§§§]Based on correlation of the Ratto Canyon thrust with the Moody Peak thrust.
[###]Summarized in Taylor et al. (2000): correlative structures 30–40 km S of Plate 1 cut rocks as young as Pennsylvanian, and are crosscut by the ca. 90–98 Ma Lincoln stock.
****Corresponds to the total traceable length on the subcrop map of Long (2012), based on correlations originally proposed by Taylor et al. (2000).
[††††]Summarized in Taylor et al. (2000): a hanging-wall anticline interpreted to be related to motion on the Schofield Canyon thrust is cut by the 86.4 ± 4.6 Ma Troy granite stock.
[§§§§]The Schofield Canyon thrust was not erosionally breached by the Paleogene. Stratigraphic throw is estimated at 2700 m by Fryxell (1991).

dips between 15°W and 35°W for the western limb, and 15°–30°E for the eastern limb. The amplitude of the Illipah anticline is between 1200 and 3800 m, and its pre-Paleogene wavelength was 35–50 km. Rocks as young as Lower Triassic are folded in the eastern limb, providing a maximum age bound.

Butte Synclinorium

The Butte synclinorium (Hose, 1977; Gans and Miller, 1983) preserves rocks as young as Permian and Triassic in its hinge zone and can be traced for 250 km (Fig. 5; Long, 2012). Subcrop patterns of the central part of the synclinorium are shown on Plate 1. In the Butte Mountains, western and eastern limb dips are 20°–30°E and 25°–30°W, respectively. To the south, along Radar Ridge, the master structure of the synclinorium is referred to as the Radar Ridge syncline (Brokaw and Barosh, 1968), and it has a western limb dip of 35°–50°E and an eastern limb dip that varies between 30°W and overturned. The amplitude of the Butte synclinorium is between 2300 and 3800 m, and its pre-Paleogene wavelength was 25–45 km. Rocks as young as Lower Triassic are folded in the hinge zone.

Cherry Creek Anticline

The subcrop pattern of an anticline that preserves Devonian, Mississippian, and Pennsylvanian rocks in its hinge zone was defined east of the Butte synclinorium on the subcrop map of Long (2012), and it is here named the Cherry Creek anticline (Fig. 5). The fold can be traced for 110 km, from southern Butte Valley through the Cherry Creek Range and northern Egan Range. The southern 30 km of the anticline axis is present on Plate 1, where the western limb dips 25°–30°W. Retrodeformation of Tertiary tilts defines a Paleogene western limb dip of 15°–20°W, and data from the subcrop map of Gans and Miller (1983) define a Paleogene eastern limb dip of 10°–35°E. The amplitude of the Cherry Creek anticline is between 2300 and 4000 m, and the pre-Paleogene wavelength was 15–40 km. Rocks as young as Lower Jurassic (Coats, 1987) are folded in the eastern limb.

Pequop Synclinorium

The Pequop synclinorium (Long, 2012) is defined by a subcrop pattern of Permian, Triassic, and Jurassic rocks preserved for a distance of 115 km, from the northern Schell Creek Range to the southern Pequop Mountains (Fig. 5). Paleogene limb dips for the southern part of the fold in the Schell Creek Range, estimated from retrodeformed attitudes presented in Gans and Miller (1983), are between 10°E and 35°E in the western limb and 25°–30°W in the eastern limb. The amplitude is between 2500 and 4000 m, and the pre-Paleogene wavelength was ≤25 km (Fig. 6). Rocks as young as Lower Jurassic (Coats, 1987) are folded in the hinge zone.

Central Nevada Thrust Belt

The Central Nevada thrust belt is a series of east-vergent thrust faults and folds that accommodated ~10 km of shortening, and which connect southward with the interior part of the Sevier thrust belt in southern Nevada (Fig. 5; Bartley and Gleason, 1990; Taylor et al., 1993, 2000; Long, 2012; Long et al., 2014). Plate 1 facilitates more detailed description and regional correlation of structures in the northern half of the Central Nevada thrust belt than the lower-resolution subcrop map of Long (2012). Next, descriptions of structures defined in source mapping are presented, several of which are named here.

Eureka Culmination and Associated Structures

The Eureka culmination, an anticline with a wavelength of 20 km, an amplitude of 4.5 km, and limb dips of 25°–35°, was defined in retrodeformed cross sections across the northern Fish Creek Range and southern Diamond Mountains (Long et al., 2014). A Cambrian over Silurian relationship defined in drill holes under the anticline crest was interpreted as the blind Ratto Canyon thrust, and the culmination was interpreted as a fault-bend fold constructed by 9 km of eastward motion of the Ratto Canyon thrust sheet over a footwall ramp (Long et al., 2014). The type section of the Early Cretaceous (Aptian, ca. 116–122 Ma) Newark Canyon Formation was deposited in a piggyback basin that developed on the eastern limb of the culmination as it grew (Long et al., 2014). On the basis of deep Paleogene erosion levels, the Eureka culmination can be traced for 80 km.

Within the eastern limb of the Eureka culmination, the east-vergent Moritz-Nager thrust (French, 1993) places Devonian rocks over Mississippian rocks (Plate 1). The Moritz-Nager thrust exhibits 1–2 km of displacement and is interpreted as a subsidiary structure that postdates the majority of motion on the Ratto Canyon thrust (Long et al., 2014).

After its construction, the Eureka culmination was deformed by normal faults that predate late Eocene volcanism (Long et al., 2014), including (from east to west) the Pinto Summit fault, Hoosac fault system, and Dugout Tunnel fault (Plate 1). The youngest prevolcanic normal faulting was accompanied by eastward tilting, which steepened the eastern limb dip to 60°–70°E, and shallowed the western limb dip to 15°–20°W (Fig. 3B).

Antelope Thrust

In the northern Antelope Range, the east-vergent Antelope thrust (Carpenter et al., 1993) places Ordovician rocks over Mississippian rocks (Hose, 1983) and is correlated to the south with a thrust fault mapped in the Park Range that places Devonian rocks over Pennsylvanian–Permian rocks (Dixon et al., 1972). Throw on the Antelope thrust is 2600 m in the Antelope Range, and this decreases to 900–1400 m in the Park Range.

Northern Pancake Range

Different names for folds and thrust faults in the northern Pancake Range have been proposed (Nolan et al., 1974; Carpenter et al., 1993; Ransom and Hansen, 1993; Long, 2012). Here, a new naming scheme is proposed, based on primary map sources (Nolan et al., 1974; McDonald, 1989), with additions from Carpenter et al. (1993).

The Moody Peak thrust (Carpenter et al., 1993) places Ordovician and Silurian rocks over Devonian rocks in the northwest corner of the range (Kleinhampl and Ziony, 1985), and it has an estimated throw of 700–1750 m. Carpenter et al. (1993) stated that the Moody Peak thrust cuts undated rocks mapped as the Newark Canyon Formation. However, given the difficulties in correlating rocks without precise age control that are mapped as Newark Canyon Formation (see discussion in "Stratigraphy" section), the Early Cretaceous maximum motion age constraint that this field relationship implies is considered tentative. The Moody Peak thrust and Ratto Canyon thrust are correlated here, based on their spatial relationships to the deep Paleogene erosion levels of the Eureka culmination, and the relative stratigraphic levels that they deform.

The Pancake thrust system, defined here by correlating thrust faults in the northwest Pancake Range described by Nolan et al. (1974), McDonald (1989), and Carpenter et al. (1993), can be traced for 35 km (Plate 1), and throw on individual structures varies from 300 to 1300 m. Structures of the Pancake thrust system cut rocks as young as Mississippian and are truncated by an Aptian (108 ± 3 Ma, K-Ar biotite; Nolan et al., 1974) dacite stock (McDonald, 1989).

The Green Springs thrust (McDonald, 1989) places Devonian and Lower Mississippian rocks over Upper Mississippian rocks, corresponding to 600–1500 m of throw. Undated conglomerate mapped as the Newark Canyon Formation overlaps the Green Springs thrust (McDonald, 1989); however, similar to the discussion for the Moody Peak thrust, without precise age control on these rocks, this motion age constraint should be considered tentative.

The Duckwater thrust (Carpenter et al., 1993) places Devonian and Lower Mississippian rocks over Upper Mississippian, Pennsylvanian, and Permian rocks (Plate 1), corresponding to 1500–2100 m of throw. The Duckwater thrust can be traced for 30 km, and it is speculatively traced northward into Newark Valley on the basis of subcrop patterns.

Central Pancake Range and Railroad Valley

Subcrop patterns in Railroad Valley were aided by drill-hole data from petroleum exploration (Hess et al., 2004). In most areas, subcrop patterns were interpreted as representing erosionally beveled folds, based on a lack of older over younger structural relationships observed in Railroad Valley drill holes (French, 1998).

A northeast-striking, east-vergent thrust fault mapped in the central Pancake Range (Quinlivan et al., 1974) is here named the Portuguese Mountain thrust. This fault places Devonian and Lower Mississippian rocks over Upper Mississippian rocks, corresponding to a throw range of 250–1350 m.

The McClure Spring syncline (Perry and Dixon, 1993) preserves Mississippian, Pennsylvanian, and Permian rocks in its hinge zone. West of Duckwater, the fold has a vertical to overturned western limb and an eastern limb that dips as steep as 50°W, and it is interpreted as east-vergent (Perry and Dixon, 1993). Here, undated rocks mapped as the Cretaceous Newark Canyon Formation are preserved in its hinge zone (Kleinhampl and Ziony, 1985); however, Perry and Dixon (1993) argued that Permian rocks are the youngest involved in folding. In the southern part of the fold, the western limb dips 25°–40°E and restores to a Paleogene dip of 10°–15°NE. The amplitude of the syncline varies from 1100 to 2600 m along its length.

An anticline can be traced through Railroad Valley and the central Pancake Range on the basis of an elongated subcrop pattern of Devonian rocks (Plate 1). This fold was originally defined in Railroad Valley by French (1998), and it is here named the Trap Spring anticline because of its proximity to the Trap Spring oil field. The amplitude ranges from 600 to 1300 m in Railroad Valley to a maximum of 2600 m where Permian rocks are preserved in its western limb.

A syncline can be traced along the eastern side of Railroad Valley and through the Duckwater Hills, on the basis of an elongated subcrop pattern preserving rocks as young as Pennsylvanian (Plate 1). This syncline was originally defined in Railroad Valley by French (1998), and it is referred to here as the Bacon Flat syncline, after its proximity to the Bacon Flat oil field. Its amplitude varies from 600 to 1500 m.

Quinn Canyon Range and Southern Grant Range

Three thrust faults are mapped in the Quinn Canyon and Grant Ranges, and they are listed here from structurally highest to lowest.

The Sawmill thrust (Bartley and Gleason, 1990) places Ordovician rocks over Devonian rocks, corresponding to a throw range of 1250–3050 m. The Sawmill thrust can be traced for 20 km, and it cuts rocks as young as Upper Devonian.

The Rimrock thrust (Bartley and Gleason, 1990) is the northernmost segment of the Rimrock-Lincoln-Freiberg thrust system (Fig. 5), which connects southward with structures of the Sevier thrust belt (Taylor et al., 2000; Long, 2012). The Rimrock thrust places Lower and Middle Devonian rocks over Upper Devonian rocks (Ekren et al., 2012), corresponding to a throw of 650–1200 m. Thirty kilometers south of Plate 1, structures associated with the correlative Lincoln thrust cut rocks as young as Pennsylvanian, and they are crosscut by a Late Cretaceous (ca. 90–98 Ma; K-Ar biotite) granite pluton (Taylor et al., 2000).

The Schofield Canyon thrust places Cambrian rocks over Ordovician rocks, and it has an estimated throw of 2700 m (Fryxell, 1988, 1991). Based on subcrop data, the Schofield Canyon thrust was not erosionally breached by the Paleogene (Plate 1). The Timber Mountain anticline, a recumbent hanging-wall fold, is interpreted to be genetically related to motion on the Schofield Canyon thrust (Fryxell, 1988). The axis of the anticline is crosscut by the Late Cretaceous (86.4 ± 4.6 Ma; U-Pb zircon) Troy granite stock (Taylor et al., 2000).

Western Utah Thrust Belt

Recent work in the Confusion Range in western Utah (Greene, 2014) has defined the Western Utah thrust belt, a 150-km-long system of surface-breaching, east-vergent thrust faults that branch off of the Sevier thrust belt (Fig. 5), deform Ordovician to Triassic rocks, and collectively accommodated ~10 km of shortening. The Confusion synclinorium (Fig. 5), a 130-km-long, north-trending structural trough that preserves rocks as young as Triassic in its hinge zone (Hose, 1977), and has long been interpreted as a regional-scale syncline (Gans and Miller, 1983; Long, 2012), has recently been shown to represent a structural low formed by a combination of folding above Western Utah thrust belt thrust faults (Greene, 2014) and construction of the Sevier culmination, a structural high to the east (Allmendinger et al., 1986; DeCelles et al., 1995). The Confusion synclinorium has a western limb dip between 50°E and overturned and an eastern limb dip between 15°W and 40°W (Greene, 2014). The western limb of the Confusion synclinorium has an amplitude of 2500–3000 m (Long, 2012; Greene, 2014). Deformation in the Western Utah thrust belt is Triassic or younger, on the basis of the youngest rocks involved in folding, but it is suggested to have been contemporary with Cretaceous to Paleocene shortening in the Sevier thrust belt (Greene, 2014).

DISCUSSION

Eastern Nevada Fold Belt: Absence of Thrust Faults and Deep Décollement Interpretation

The Eastern Nevada fold belt is characterized by five first-order folds that can be traced for distances between 100 and 250 km, with typical limb dips of 10°–30°, amplitudes of 2–4 km, and pre-extensional wavelengths of 20–40 km. For comparison, first-order folds in the Valley and Ridge Province in the north-central Appalachians can be traced for 200–300 km, have amplitudes of 5–7 km, and have wavelengths typically ≥12 km (Nickelsen, 1963; Faill, 1998).

The Eastern Nevada fold belt is differentiated from the Central Nevada and Western Utah thrust belts by an absence of regional-scale thrust faults or décollement horizons at modern levels of exposure. Within the portion of the Eastern Nevada fold belt on Plate 1, several of the older source maps, including Humphrey (1960) in the White Pine Range, Fritz (1968) in the Cherry Creek Range, and Brokaw and Barosh (1968) in the Egan Range, map structures with modern low dip angles that place younger rocks over older rocks as thrust faults. Gans and Miller (1983), in their classic assessment of extensional style in eastern Nevada, reinterpreted these thrust faults in the Cherry Creek and Egan Ranges as series of highly rotated, domino-style normal faults. In this study, this interpretation is similarly applied to the thrust faults mapped by Humphrey (1960) in the White Pine Range, after the pioneering work of Moores et al. (1968) located 12 km to the south.

East- and west-vergent thrust and reverse faults that do exhibit older over younger relationships, with throw, and in some cases offset estimates on accompanying cross sections, on the order of tens to hundreds of meters, are mapped in several areas of the Eastern Nevada fold belt on Plate 1, including the southern Diamond Mountains (Nolan et al., 1971, 1974), Bald Mountain (Nutt, 2000; Nutt and Hart, 2004), the central White Pine Range (Humphrey, 1960), the Grant Range (Lund et al., 1987; Camilleri, 2013), and the Butte Mountains (Otto, 2008). However, these faults can typically only be traced for map distances of 1–5 km, are not traceable across individual source maps or onto adjacent source maps, and can in no cases be correlated regionally. Therefore, on the basis of their modest throw and map extent, they are interpreted as second-order faults that accompanied first-order folding. East of the area of Plate 1, in the Egan, Schell Creek, and Snake Ranges, Gans and Miller (1983) similarly observed that the primary contractional structures are regional-scale folds, and that no regional-scale thrust faults or décollement horizons are present.

The present-day range of stratigraphic levels exposed in the Eastern Nevada fold belt on Plate 1 ranges from Cambrian to Triassic, corresponding to paleodepths up to 8 km below the Paleogene unconformity. Thick sections of Paleozoic rocks that are undisturbed by faults are observed in the limbs of several first-order folds, including straight sections from Silurian to Permian rocks in the western limb of the Pinto Creek syncline (Larson and Riva, 1963; Nolan et al., 1971), from Mississippian to Triassic rocks in the eastern limb of the Butte synclinorium (Douglass, 1960; Brokaw and Barosh, 1968), and from Cambrian to Pennsylvanian rocks in the Cherry Creek Range (Fritz, 1968). East of Plate 1, up to 4 km of Neoproterozoic to Lower Cambrian clastic rocks are exposed in the Cherry Creek, Egan, Schell Creek, Snake, and Deep Creek Ranges (Woodward, 1962; Young, 1960; Stewart, 1980; Gans and Miller, 1983), corresponding to paleodepths up to 12 km below the Paleogene unconformity.

On the basis of their approximately uniform amplitude and wavelength over along-strike distances exceeding 100 km, it is proposed here that the first-order folds of the Eastern Nevada fold belt were produced through décollement-style tectonics above a blind, low-angle fault or shear zone that underlies the entire region. One likely candidate is the basal décollement of the Sevier thrust belt, which must transfer 220 km of displacement westward to deeper structural levels under the hinterland (DeCelles and Coogan, 2006). However, the recent recognition of 10 km of upper-crustal shortening accommodated in the Western Utah thrust belt (Greene, 2014) indicates the possibility of multiple décollement levels at depth (see discussion in section on "Model for Cordilleran Deformation in the Sevier Hinterland at 39°N").

The specific deformation processes at depth that produced the first-order folds of the Eastern Nevada fold belt are difficult to constrain without a more precise analysis of their geometry in cross section. A lack of significant steps in structural level across the Eastern Nevada fold belt implies a regional décollement with relatively low structural relief, which makes fault-bend folding an unlikely formation mechanism for the first-order folds. Also, the possibility that these folds represent fault-bend folds above deep-seated duplexes, perhaps constructed from detached basement thrust sheets, as observed in several localities in the Sevier thrust belt in Utah (Allmendinger et al., 1983; DeCelles, 1994; DeCelles et al., 1995; Camilleri et al., 1997; DeCelles and Coogan, 2006), is not favored by the typical 50–100 km along-strike dimensions of these basement-cored culminations (DeCelles, 2004). Instead, the uniform wavelength, amplitude, and map distance of these folds implies that the primary controlling mechanism was the mechanical stratigraphy of the Neoproterozoic to Mesozoic section through which deformation propagated. On this basis, two likely genetic origins for the first-order folds of the Eastern Nevada fold belt are: (1) fault-propagation folds (e.g., Suppe and Medwedeff, 1990) that formed above splays off of the basal décollement that tip out prior to reaching modern levels of exposure, similar to observations in the north-central Appalachian Valley and Ridge Province (e.g., Gwinn, 1970; Kulander and Dean, 1986; Faill, 1998); or (2) detachment folds (e.g., Mitra, 2003) that formed above a deep décollement, such as observed in the Zagros fold-and-thrust belt (e.g., Colman-Sadd, 1978; McQuarrie, 2004; Molinaro et al., 2005). Construction of precise fold geometries in a retrodeformed cross section across the Eastern Nevada fold belt will be necessary to distinguish between these folding mechanisms, to estimate depth to decollement using excess area techniques (e.g., Epard and Groshong, 1993), and to accurately estimate crustal shortening, which are goals of future work.

Model for Cordilleran Deformation in the Sevier Hinterland at 39°N

Figure 7 shows a generalized, pre-extensional cross section through the Sevier thrust belt and hinterland region at ~39°N. The cross section is used to support the following discussion, which presents a general structural model for Cordilleran deformation at this latitude.

In west-central Utah, the Sevier thrust belt accommodated ~220 km of upper-crustal shortening, distributed among a series of thrust faults that breach the surface over an ~50 km across-strike distance (DeCelles and Coogan, 2006). The Canyon Range thrust, the westernmost and structurally highest fault, carries as much as 6 km of Neoproterozoic to Lower Cambrian clastic rocks, as indicated by drill-hole data (DeCelles and Coogan, 2006) and the Consortium for Continental Reflection Profiling (COCORP) seismic line (Allmendinger et al., 1983, 1987). The Canyon Range thrust sheet is broadly folded under the House Range, forming the Sevier culmination, an anticlinal dome defined by Paleozoic subcrop patterns (Harris, 1959; Hintze and Davis, 2003; Long, 2012) and arched reflectors on the COCORP line (Allmendinger et al., 1983), interpreted as a duplex cored by thrust sheets of Precambrian basement rock (Allmendinger et al., 1987; DeCelles and Coogan, 2006). Frontal structures of the Sevier thrust belt, including the Pavant thrust, Paxton thrust, and Gunnison thrust, are interpreted to root beneath the culmination (DeCelles and Coogan, 2006). Therefore, under the western House Range, which is the farthest point that Sevier thrust faults can be confidently traced on the COCORP line (Allmendinger et al., 1983), there are at least two décollement levels that together must accommodate the total 220 km of Sevier displacement (Fig. 7). These structures both carry rocks that are deeper than modern exposure levels in eastern Nevada.

In the Confusion Range, the Western Utah thrust belt accommodated ~10 km of shortening, which is distributed between a basal décollement in Ordovician rocks and subsidiary splaying thrust faults that ramp up section toward the east and deform rocks as young as Triassic (Greene, 2014). The Western Utah thrust belt defines an additional detachment level that roots under the Eastern Nevada fold belt, which must ramp rapidly down section toward the west, through at least 6 km of Neoproterozoic to Cambrian clastic rocks that are presently exposed in ranges in eastern Nevada (Fig. 7; Stewart, 1980; Gans and Miller, 1983).

Modern exposure levels indicate that a regional décollement that could have produced the first-order folds of the Eastern Nevada fold belt must lie at a minimum depth of 12 km below the top of the Triassic section, corresponding to ~10 km below the Paleogene unconformity (Fig. 7). Under the Eastern Nevada fold belt, the geometry and stratigraphic levels of the basal Sevier décollement and Western Utah thrust belt detachment level are unknown; it is possible that they merge into the same master décollement horizon or shear zone. The contact between Neoproterozoic sedimentary rocks and Precambrian crystalline basement rocks, which is the same exploited by the Canyon Range thrust to the east, is a likely mechanical boundary for localizing such a shear zone; however, involvement of Precambrian crystalline basement cannot be ruled out.

During the Late Cretaceous (ca. 70–90 Ma), in the Egan, Schell Creek, and Snake Ranges, Neoproterozoic to Lower Cambrian rocks experienced regional metamorphism, synchronous with intrusion of granite bodies (Miller and Gans, 1989). This metamorphism was accompanied by penetrative, top-to-the-east simple shear in rocks as shallow as Lower Cambrian (paleodepths of ~7–8 km; Fig. 7), which increases in intensity down section into Neoproterozoic rocks (Miller and Gans, 1989). These shear fabrics are synchronous with shortening in the Sevier thrust belt, and they are interpreted as a consequence of thermal weakening of the upper crust that accompanied the rise of anatectic melts (Miller et al., 1988; Miller and Gans, 1989). The temporal relationship between this shallow penetrative deformation and construction of first-order folds of the Eastern Nevada fold belt is unclear, but these observations indicate that at a late stage in the Cordilleran shortening history, the Sevier thrust belt was rooted westward into a diffuse shear zone that affected the base of the upper crust (Miller and Gans, 1989; Speed et al., 1988).

Farther to the hinterland, the Central Nevada thrust belt accommodated ~10 km of shortening, and it deforms rocks as deep as Lower Cambrian. At 39°N, Central Nevada thrust belt deformation consisted of growth of the Eureka culmination, a fault-bend fold that formed above a Cambrian to Silurian footwall ramp of the Ratto Canyon thrust, which is interpreted as the basal structure (Long et al., 2014). The Aptian Newark Canyon Formation was deposited and folded in a piggyback basin that developed on the eastern limb of the Eureka culmination as it grew (Long et al., 2014). To the east in the White Pine Range, stratigraphic levels from Cambrian to Pennsylvanian are exposed, without any mapped décollement horizons or thrust faults. Therefore, the Duckwater thrust is interpreted as the deformation front for the Central Nevada thrust belt at this latitude, where slip from the Ratto Canyon thrust was fed to the surface (Fig. 7). West of the Eureka culmination, the geometry and stratigraphic level of the basal Central Nevada thrust belt décollement are unknown, but must lie within Lower Cambrian or deeper rocks, and may eventually merge at depth with the Eastern Nevada fold belt décollement.

Deformation Timing in the Sevier Thrust Belt and Hinterland Structural Provinces

The timing of initiation of deformation in the Sevier thrust belt has been long debated; most workers agree that initial Early Cretaceous (ca. 130–125 Ma; Barremian) subsidence in the Sevier foreland basin indicates coeval crustal thickening in the thrust belt (e.g., Jordan, 1981; Lawton, 1985; DeCelles and Currie, 1996;

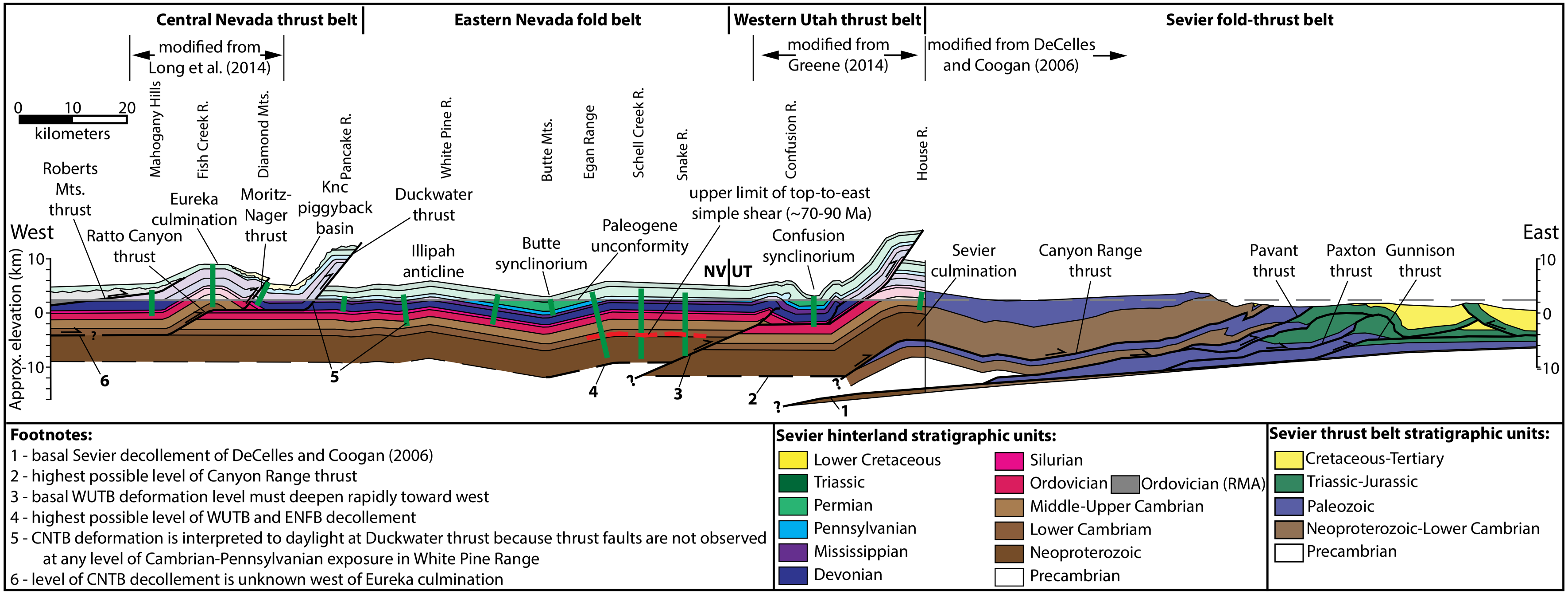


Figure 7. Generalized cross-section through eastern Nevada and western Utah at ~39°N, showing the geometry of the Sevier fold-and-thrust belt (modified from DeCelles and Coogan, 2006, their Fig. 8F), and structural provinces of the Sevier hinterland. This is not a balanced cross section; it is meant to illustrate approximate deformation geometries and constraints on décollement levels in the Sevier hinterland. The hinterland portion is drafted based on horizontal distances on the 36 Ma tectonic reconstruction of McQuarrie and Wernicke (2005) shown on Figure 6, and it uses stratigraphic thickness data from Stewart (1980), Gans and Miller (1983), Greene (2014), and Long et al. (2014). The Paleogene unconformity is drawn as a horizontal datum, at an approximate elevation of 2.5 km, after Snell et al. (2014), and its stratigraphic levels from Plate 1 and Long (2012) were used to constrain approximate geometries for first-order folds of the Eastern Nevada fold belt (note that the specific formation mechanism at depth for these folds is not interpreted on the cross section, and that this transect does not contain all of the first-order folds of the Eastern Nevada fold belt; see Figs. 5 and 6). Translucent areas above the Paleogene unconformity represent eroded rock. Thick green lines show the total range of stratigraphic levels presently exposed in individual ranges (data sources: Douglass, 1960; Humphrey, 1960; Young, 1960; Woodward, 1962; Brokaw, 1967; Dechert, 1967; Nolan et al., 1974; Hose and Blake, 1976; Schalla, 1978; Bentz, 1983; Gans and Miller, 1983; Otto, 2008; Long et al., 2012; Greene, 2014). The thick, dashed red line shows the upper limit of Late Cretaceous, top-to-east simple shear fabrics in easternmost Nevada, approximated from Miller and Gans (1989). Abbreviations: RMA—Roberts Mountains allochthon; CNTB—Central Nevada thrust belt; ENFB—Eastern Nevada fold belt; WUTB—Western Utah thrust belt; NV—Nevada; UT—Utah.

Lawton et al., 1997; Currie, 2002). However, on the basis of thermochronology data (Burtner and Nigrini, 1994; Ketcham et al., 1996; Yonkee et al., 1997; Stockli et al., 2001), initiation of slip on the Canyon Range thrust has been argued to be as old as the Jurassic-Cretaceous boundary (ca. 145 Ma; DeCelles, 2004; DeCelles and Coogan, 2006). Deformation in the Sevier thrust belt at this latitude continued through the Late Cretaceous and Paleocene (Lawton and Trexler, 1991; Lawton et al., 1993, 1997; DeCelles et al., 1995; DeCelles and Coogan, 2006), on the basis of geochronology, biostratigraphy, and structural relationships of foreland basin strata.

In the Sevier hinterland, precise timing constraints on deformation are rare, due to sparse preservation and poor geochronology of synorogenic rocks. Thrust faults and folds of the Western Utah thrust belt deform rocks as young as Lower Triassic, providing a maximum age bound. However, the Western Utah thrust belt diverges off of the Sevier thrust belt (Figs. 5 and 6), and on this basis, has been interpreted to be contemporary with the Cretaceous–Paleocene Sevier shortening history (Greene, 2014).

The Central Nevada thrust belt connects southward with the interior part of the Sevier thrust belt in southern Nevada (Figs. 5 and 6), which also implies an overlap in deformation timing (Taylor et al., 2000; Long, 2012). South of the Eureka culmination, rocks as young as Pennsylvanian and Permian are cut and folded by Central Nevada thrust belt structures, and in three places, undated rocks mapped as the Early Cretaceous Newark Canyon Formation are interpreted to either overlap or be cut by Central Nevada thrust belt structures (Tables 1 and 2; McDonald, 1989; Carpenter et al., 1993; Perry and Dixon, 1993). In three places, plutons ranging between ca. 86 Ma and ca. 108 Ma crosscut Central Nevada thrust belt structures (Table 1; McDonald, 1989; Taylor et al., 2000), indicating that deformation in the southern Central Nevada thrust belt was completed by the Late Cretaceous (Albian–Coniacian). In the northern Central Nevada thrust belt, growth of the Eureka culmination is interpreted to have been contemporary with deposition and folding of the Aptian (ca. 122–116 Ma) Newark Canyon Formation (Long et al., 2014).

First-order folds of the Eastern Nevada fold belt deform Lower Triassic rocks in multiple places, and in one locality deform Lower Jurassic rocks (Table 1). In addition, the Pinto Creek syncline folds the Aptian (ca. 122–116 Ma) Newark Canyon Formation in the southern Diamond Mountains, and it folds undated rocks mapped as the Newark Canyon Formation farther north in the range (Plate 1). It is suggested here that this fold province, which is defined by a uniform deformation style consistent with growth of regional-scale folds above an areally extensive décollement, must have been either contemporary with or postdated the Early Cretaceous (ca. 145–125 Ma) initial migration of the master Sevier décollement into Utah, and that construction of the Eastern Nevada fold belt as an integrated tectonic province could potentially span the Cretaceous–Paleocene deformation history recorded in the frontal Sevier thrust belt.

In summary, the Central Nevada thrust belt, Eastern Nevada fold belt, and Western Utah thrust belt represent structural provinces of the Sevier hinterland that collectively record low-magnitude (a few tens of kilometers) shortening, accommodated during the protracted Cretaceous to Paleocene detachment and eastward translation of the entire Cordilleran passive-margin basin over 220 km eastward during Sevier orogenesis. One of the principal controls on the structural style of the Sevier event was the great thickness and rheological competence of the basin through which deformation propagated, in particular the >6-km-thick section of Neoproterozoic–Lower Cambrian clastic rocks at its base (e.g., Armstrong, 1968, 1972; Gans and Miller, 1983; Speed et al., 1988; DeCelles, 2004). The high strength and along- and across-strike uniformity of this mechanical stratigraphy allowed Cordilleran deformation to be translated eastward through a >200-km-wide hinterland region that only experienced low-magnitude (tens of kilometers) internal deformation. The thinning of this basin by nearly an order of magnitude eastward across the narrowly defined Wasatch hinge line in western Utah (e.g., Poole et al., 1992) focused the majority of Cordilleran shortening into the spatially narrow Sevier thrust belt to the east (e.g., Royse, 1993b; DeCelles, 2004). This illustrates the profound role that inherited basin geometry can exert on the location, geometry, and style of later orogenesis.

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