**Tectonic evolution of the Sevier and Laramide belts within the North American Cordillera
orogenic system**

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8   **Abstract**

9   The thin-skin Sevier and thick-skin Laramide belts of the North American Cordillera provide a
10   long-term record of the interrelations between evolving styles of mountain building and plate
11   dynamics over a complete tectonic cycle, from onset of rapid subduction, to protracted growth of a
12   composite orogenic system, to final collapse. Primary architecture of basement and sedimentary
13   cover rocks, which included a thick passive margin section deposited along the western continental
14   margin, influenced patterns of subsequent deformation. The Cordilleran orogenic system,
15   comprised of an interrelated forearc accretionary complex, magmatic arc, retroarc hinterland,
16   Sevier fold-thrust belt, and foreland basin locally deformed by Laramide arches, developed during
17   Jurassic to Paleogene Andean-style subduction and terrane accretion. The Sevier belt formed as a
18   foreland-propagating (west to east) wedge mostly during Cretaceous to Paleogene time, and
19   included a western thrust system with aerially extensive thrust sheets that carried thick passive
20   margin strata, and an eastern thrust system that carried thinner strata. Within the Wyoming salient
21   of the Sevier belt, major thrust and fold traces display systematic map-view curvature about an
22   average N-S structural trend, reflecting a component of primary curvature related to sedimentary
23   prism architecture, followed by 60-80% vertical-axis rotation of thrust sheets related to curved
24   fault slip and interaction with Laramide arches at the salient ends. Internal deformation in the
25   western thrust sheets was limited within strong shallower levels, whereas deeper levels underwent
26   shear and vertical flattening near a weak basal fault zone. Internal deformation in the eastern thrust
27   sheets included widespread early layer-parallel shortening (LPS), followed by concentration of slip
28   onto weak fault zones. Approximately 200 km of thin-skin shortening in the Sevier belt was
29   transferred into lower crustal thickening and uplift of an orogenic plateau in the hinterland to the
30   west. Synorogenic strata were deposited in an evolving foreland basin to the east that formed

during flexural loading of thrust sheets and regional dynamic subsidence during subduction.

Overall E-oriented shortening in the Sevier belt is interpreted to reflect increased gravitational potential energy and evolving topographic slopes from a hinterland plateau through a growing thrust wedge. Laramide basement-cored arches and intervening basins developed during later Cretaceous to Paleogene time, overlapping with younger stages of Sevier deformation. Arches and associated reverse faults display a wide range of trends within an overall NW-SE oriented, anastomosing network, partly reflecting primary basement heterogeneities. Limited vertical-axis rotations were localized along obliquely trending arch forelimbs and near arch intersections.

Internal deformation in the foreland included limited LPS that refracted near variably trending arches. Laramide deformation was spatially and temporally correlated with a region of flat-slab subduction, recorded by changes in magmatic and subsidence patterns. Overall ENE-oriented shortening in the Laramide belt was at low angles to relative plate motion, likely reflecting increased plate coupling near a cratonic lithosphere keel. An integrated model for tectonic evolution of the Sevier and Laramide belts includes: influence of primary sedimentary architecture and basement weaknesses; enhanced plate coupling from increased overriding plate motion, fast convergence rates, and development of a flat-slab segment; linkage of upper crustal shortening in the Sevier belt with lower crustal thickening and uplift of a hinterland plateau; interaction between frontal thrusts in the Sevier belt and Laramide arches with differently oriented stress fields; concentrated slip along weak fault zones; redistribution of mass by erosion and deposition of synorogenic strata; and a switch to orogenic collapse during decreased convergence rates and slab removal.

1. Introduction

The North American Cordillera, including the Sevier and Laramide belts, forms one of Earth's classic orogenic systems, providing a natural laboratory for seminal studies that have advanced our understanding of thrust kinematics and mechanics (e.g., Rubey and Hubbert, 1959; Dahlstrom, 1970; Price and Mountjoy, 1970; Burchfiel and Davis, 1972; Boyer and Elliott, 1982), synconvergent extension of tectonically thickened crust (e.g., Hodges and Walker, 1992; Wells et al., 2012), and foreland basin subsidence mechanisms (e.g. Jordan, 1981; DeCelles and Giles, 1996; Catuneanu et al., 1997). This archetype orogenic system formed during protracted Jurassic to Paleogene subduction and terrane accretion that spanned over 4000 km from Alaska to Mexico (Fig. 1; Oldow et al., 1989; Dickinson, 2004; DeCelles, 2004). While the modern-day South American Andes provide opportunities to study links between active deformation, landscape evolution, climate, and geophysically imaged lithospheric structure (e.g. Allmendinger et al., 1997; Ramos et al., 2004; Oncken et al., 2006; Alvarado et al., 2007; Anderson et al., 2007; Strecker et al., 2007; Hilley and Coutand, 2010; Giambiagi et al., 2012; DeCelles et al., 2015), the North American Cordillera provides a long-term geologic record of interrelations between upper crustal shortening, lower crustal thickening and metamorphism, and changing plate margin dynamics through a complete tectonic cycle, from initiation of enhanced plate coupling, to growth of a composite orogenic system, to slab removal and final collapse.

Previous studies of the Sevier belt and associated synorogenic strata provide one of the best records of large-scale geometry and long-term timing of any retroarc fold-thrust belt in the world (Armstrong and Oriel, 1965; Armstrong, 1968; Royse et al., 1975; Dorr et al., 1977; Lamerson, 1982; Wiltschko and Dorr, 1983; Lawton, 1985; Royse, 1993; DeCelles, 1994, 2004; DeCelles and Coogan, 2006). Additionally, recent studies of the primary sedimentary prism (Gehrels and Pecha,

2014; Yonkee et al., 2014), 3-D kinematic evolution of curved thrust systems (Weil et al., 2010; Yonkee and Weil, 2010), tectono-metamorphic history of the hinterland (Wells et al., 2012; Hallett and Spear, 2013), nature of Laramide foreland deformation (Erslev and Koenig, 2009; Weil and Yonkee, 2012; Weil et al., 2014; Yeck et al., 2014), patterns of forearc accretion (Dumitru et al., 2015; Wakabayashi, 2015), tempos of arc magmatism and deformation (Cao et al., 2015; Paterson and Ducea, 2015), and geophysically imaged slab remnants (Liu et al., 2010; Jones et al., 2011; Sigloch, 2011; Porritt et al., 2014) have advanced our understanding of relationships between crustal architecture, deformation patterns, igneous processes, and plate margin dynamics.

Here we combine results from multiple studies to develop an integrated model for tectonic evolution of the thin-skin Sevier and thick-skin Laramide belts within the North American Cordilleran orogenic system, focusing along a tectonic transect from present-day California to Wyoming (Fig. 1). We address the following questions:

1. How did primary crustal architecture influence development of the North American Cordilleran orogenic system?
2. What were the interrelations between plate dynamics and evolution of major tectonic components of the Cordilleran system?
3. What were the spatial-temporal patterns of deformation in the Sevier and Laramide belts and how did they relate to each other?
4. How did mechanical factors, topography, and surface processes influence Sevier and Laramide deformation patterns?
5. Given currently available observations, what model best explains the tectonic evolution of the Sevier and Laramide belts within the overall Cordilleran system?

Our intention herein is not to produce a comprehensive geologic description for the entire North American Cordillera, but rather to explore how interconnected processes operated to form thin-skin (Sevier) and thick-skin (Laramide) mountain belts within an evolving orogenic system. Note, in this paper Sevier and Laramide are used to denote different structural styles, not time intervals.

2. Primary crustal architecture

The Cordilleran orogenic system developed partly on North American basement and sedimentary cover rocks, and partly on allochthonous accreted terranes. North American sedimentary cover rocks included thick passive margin strata deposited above extended basement along the previously rifted western margin of Laurentia (present-day coordinates), and thin strata deposited above cratonic basement to the east (Stewart, 1972; Bond and Kominz, 1984; Bond et al., 1985; Levy and Christie-Blick, 1989; Poole et al., 1992; Link et al., 1993; Yonkee et al., 2014). Importantly, characteristics of the North American basement and sedimentary cover partly controlled subsequent structural development of the Sevier and Laramide belts (Burchfiel and Davis, 1972; Lawton et al., 1994; Mitra, 1997; DeCelles, 2004; Weil and Yonkee, 2012). West of the $^{87}\text{Sr}/^{86}\text{Sr} = .706$ line (Kistler and Peterman, 1973), North American basement is absent and the orogenic system included terranes accreted over a protracted period of time (Fig. 1). Characteristics of basement, sedimentary cover, and accreted terranes that comprised the building materials for the orogenic system are summarized below.

2.1. North American basement rocks

Basement rocks in the western U.S. include the Archean Wyoming province and Grouse Creek block, Paleoproterozoic Farmington zone and Mojave province with inherited Archean components, and juvenile Paleoproterozoic Yavapai and Mazatzal provinces (Fig. 2; Foster et al.,

2006; Whitmeyer and Karlstrom, 2007). The Wyoming province, exposed in the cores of Laramide foreland uplifts, consists mostly of Neoarchean (2.8-2.6 Ga) granitoid intrusions, supracrustal belts, and older gneiss (Frost et al., 1998, 2006; Mueller and Frost, 2006). The Grouse Creek block, locally exposed in the hinterland where it was overprinted by Jurassic to Paleogene tectonism, consists of Neoarchean (2.7-2.5 Ga) granitoids, amphibolite, and schist (Wright and Snoke, 1993; Strickland et al., 2011). The Farmington zone, which separates the Grouse Creek block and Wyoming province, contains mixtures of paragneiss and orthogneiss that record early Paleoproterozoic deposition of strata with inherited Archean components, intrusion of granitoid plutons, and late Paleoproterozoic (1.8–1.7 Ga) metamorphism and deformation (Bryant, 1988; Mueller et al., 2011). The Mojave province to the south also contains mixtures of paragneiss and orthogneiss with inherited Archean components that underwent 1.8–1.7 Ga metamorphism (Wooden and Miller, 1990; Shufeldt et al., 2010). The Yavapai and Mazatzal provinces are dominated by 1.8–1.6 Ga juvenile volcanic arc rocks that were accreted together and sutured onto the Wyoming and Mojave provinces and subsequently intruded by 1.3-1.5 Ga granitic plutons (Bennett and DePaolo, 1987; Karlstrom and Bowring, 1988).

Geophysical imaging and isotopic signatures of Neogene volcanic rocks indicate that thick, Archean to Paleoproterozoic mantle lithosphere is preserved beneath much of the foreland east of the Sevier belt (Livaccari and Perry, 1993; Obreski et al., 2011; Porritt et al., 2014). Xenoliths brought up in pre- and post-Laramide volcanic centers also record preservation of Archean to Paleoproterozoic mantle lithosphere beneath the foreland, along with evidence of localized Late Cretaceous to Paleogene metasomatism and hydration likely associated with flat-slab subduction (Humphreys et al., 2003; Glebovitsky et al., 2004; Carlson et al., 2004; Smith and Griffin, 2005). Mantle lithosphere was subsequently modified and thinned in some regions due to asthenosphere

convection associated with development of the Neogene Basin and Range province, Rio Grande Rift, and Yellowstone hotspot (Schmandt and Humphries, 2010; Obreski et al., 2011).

2.2. North American sedimentary cover

The sedimentary cover includes: (1) local Mesoproterozoic strata of the intracratonic Belt basin; (2) Neoproterozoic to Early Cambrian siliciclastic strata and volcanic rocks that accumulated during protracted rifting; (3) Middle Cambrian to Devonian carbonate-rich strata deposited along a passive margin; and (4) Mississippian to Jurassic mixed siliciclastic and carbonate strata deposited during disruption of the passive margin and development of volcanic arcs and basins to the west (Fig. 3A; Poole et al., 1992; Link et al., 1993; Dickinson, 2004; Gehrels and Pecha, 2014). The Belt basin of western Montana to northern Idaho contains up to 15 km of siliciclastic, carbonate, and mafic igneous rocks, which acted as a relatively strong block during subsequent rifting and Sevier thrusting (Sears, 2001). Thick Neoproterozoic and Paleozoic passive margin strata were deposited regionally on rifted crust of western Laurentia, with the Wasatch hinge-zone marking the transition to thin Paleozoic strata deposited on cratonic crust to the east (Fig. 3B; Yonkee et al., 2014). The passive margin trended overall NNW-SSE (present-day coordinates) in Canada, to N-S in the western U.S., and also curved outward (eastward) along the Alberta, Wyoming, and central Utah salients, which partly controlled primary curvature of the future Sevier belt (Figs. 1, 3B; Stewart, 1972; Levy and Christie-Blick, 1989; Lawton et al., 1994; Lund, 2008). Sinistral offset of the passive margin from SE California to Mexico also influenced subsequent deformation patterns (Dickinson and Lawton, 2001a; Stevens et al., 2005).

Earlier Neoproterozoic (~750 Ma) strata (Uinta Mountain and Chuar Groups) accumulated in a basin system that presaged onset of regional rifting (Dehler et al., 2010). These strata are absent across most of the region, but are up to 5 km thick in an E-W trending basin in NE Utah,

which was subsequently inverted to form the Uinta-Cottonwood arch of the Laramide foreland (Bradley and Bruhn, 1988). Middle Neoproterozoic (~720–660 Ma) diamictite and volcanic-bearing strata (Perry Canyon Formation and correlatives) accumulated during an early phase of rifting (Balgord et al., 2013), and included micaceous strata that subsequently formed the basal decollement for western thrusts of the Sevier belt. Later Neoproterozoic (~660–580 Ma) quartzite-rich strata (lower Brigham Group and correlatives) accumulated during regional subsidence following early rifting (Crittenden et al., 1971), and formed a relatively strong beam during subsequent deformation. Latest Neoproterozoic to Early Cambrian (~580–530 Ma) arkosic and basalt-bearing strata (upper Brigham Group and correlatives) record final rifting on previously thinned lithosphere (Yonkee et al., 2014). The thickness of middle to late Neoproterozoic strata increases westward to >3 km (Fig. 3B). Similar patterns occur in the Windermere and Hamill groups of Canada, which contain diamictite and volcanic-bearing strata, mature siliciclastic strata, and overlying arkosic strata with 580 Ma basalts (Ross et al., 1995; Colpron et al., 2002), and in northern Mexico (Stewart, 2005). Final rifting was followed by development of a thermally subsiding passive margin (Bond and Kominz, 1984; Bond et al., 1985).

Middle Cambrian to Devonian carbonate-rich strata accumulated along the margin, ranging in total thickness from ~4-6 km in western areas to <2 km east of the Wasatch hinge zone that marked the transition to cratonic crust (Fig. 3B; Poole et al. 1992; Hintze and Kowallis, 2009). Strata were deposited in mostly shallow marine waters that extended onto the craton during the Sauk and Tippecanoe transgressions. Deeper marine strata accumulated farther west along the continental slope. Middle Cambrian shale and limestone subsequently formed the basal decollement for eastern thrusts of the Sevier belt.

During later Devonian to Permian time, the passive margin was disrupted and sedimentary patterns became varied (Fig. 3B; Speed and Sleep, 1982; Poole et al., 1992; Burchfiel et al., 1992). The Roberts Mountain allochthon, comprising deep marine strata and mafic volcanics that were deposited in a western basin, was thrust eastward over shelf strata as the basin closed during the Late Devonian-Mississippian Antler orogeny (Burchfiel et al., 1992; Siberling et al., 1997). By later Mississippian time, shallow marine carbonates and sandstone were deposited across much of the region (Sandberg et al., 1982). During Pennsylvanian time, the Oquirrh basin subsided in NW Utah with deposition of up to 6 km of marine strata, simultaneous with uplift of the Ancestral Rockies during collision of Laurentia with Gondwana (Dickerson, 2003). Permian deep marine sedimentary and volcanic rocks accumulated in another western basin, and were thrust eastward in the Golconda allochthon during the Triassic Sonoma orogeny (Speed and Sleep, 1982; Oldow et al., 1989). The southwestern part of the plate margin and sedimentary cover were truncated during Late Paleozoic to Triassic sinistral strike-slip faulting (Dickinson and Lawton, 2001a; Stevens et al., 2005), with subsequent modification during Jurassic sinistral transtension along the Mojave-Sonora megashear (Anderson and Nourse, 2005).

Triassic to Jurassic strata were deposited in a mix of marine and continental environments during reorganization of the Cordilleran margin (Dickinson and Gehrels, 2008; DeCelles, 2004). During Early Triassic time, marine strata accumulated to the west, while red beds of the Chugwater Group accumulated to the east, northward of remnants of the Ancestral Rockies. During the Late Triassic, a volcanic arc developed to the west (Barth et al., 2011) and marine strata were deposited in a back-arc basin in NW Nevada (Wyld, 2002), while red beds with tuffaceous layers were deposited in a broad basin to the east, followed by deposition of Early Jurassic eolian strata. Subsidence increased across the Utah-Idaho trough in the retroarc during the Middle

Jurassic with deposition of marine carbonates and evaporites in the Twin Creek Formation and correlative strata (Imlay, 1967), synchronous with early terrane accretion along the western plate margin (Dickinson, 2008). These evaporites subsequently formed detachments during Sevier thrusting (Coogan and Yonkee, 1985). Late Jurassic and Cretaceous to Paleogene strata deposited in the foreland are described in section 4.5.

2.3. *Accreted terranes*

The western part of the Cordilleran orogenic system was constructed on accreted terranes that included: (1) peri-cratonic rocks with fossil and detrital zircon (DZ) signatures typical of North America (Roberts Mountain and Golconda allochthons, Kootenay terrane); (2) Paleozoic to Mesozoic arc and ophiolite rocks of the Intermontane terrane group in Canada, and the Sierra Foothills, Klamath Mountains, and Blue Mountains provinces; and (3) Paleozoic to Mesozoic distal arc and ophiolite rocks of the Insular terrane group exposed mostly in Canada (Fig. 1; Monger et al., 1982; Oldow et al., 1989; Dickinson, 2004).

The Sierra Foothills, Klamath Mountains, and Blue Mountain provinces comprise accreted materials that can be broadly divided into eastern, central, and western belts (Edelman and Sharp, 1989; Dickinson, 2008; Ernst et al., 2008; LaMaskin et al., 2011). The eastern belt contains upper Neoproterozoic to Cambrian mafic igneous rocks and siliciclastic strata, Lower Paleozoic deep water siliciclastic and alkali mafic volcanic rocks, Devonian mafic to calc-alkaline rocks of an immature arc, and Triassic volcanic and sedimentary rocks of a subsequent fringing arc (Edelman and Sharp, 1989; Ernst et al., 2008). The central belt contains ophiolitic mélanges, interpreted to record alternating episodes of subduction and strike-slip motion along the paleo-plate margin, and the western belt contains mostly Triassic to Early Jurassic volcanic rocks of another arc (Dickinson, 2008). The Intermontane terrane group in Canada contains broadly similar belts that

include the Quesnelia terrane with Devonian and Triassic arc-related rocks, the Cache Creek terrane with ophiolitic mélange, and the Stikinia terrane with other arc-related rocks (Monger et al., 1982; Riddell, 2011). These terranes were consolidated and accreted onto the North American margin during the Early to Middle Jurassic (Rusmore et al., 1988; Ernst et al., 2008). The Insular terrane group comprises the Wrangellia, Alexander, and Peninsular terranes, which contain exotic Paleozoic arc rocks and Triassic volcanic and rift deposits (Jones et al., 1977; Monger et al., 1982; McClelland et al., 1992). The Insular terrane group was consolidated and accreted onto the Cordilleran margin by *ca* 100 Ma, followed by crustal thickening along a W-vergent thrust system (Rubin et al., 1990) and dextral offset along the Coast shear zone (McClelland et al., 2000).

3. Plate dynamic setting of the North American Cordillera

The North American Cordillera orogenic system developed in response to subduction of oceanic (Farallon and Kula) plates and accretion of terranes along the North American plate margin during Jurassic to Paleogene time (Fig. 4; Oldow et al., 1989; Burchfiel et al., 1992; Saleeby and Busby-Spera, 1992; Dickinson, 2004; DeCelles, 2004). The style and distribution of deformation within the Cordilleran system changed over time in response to variations in absolute motion of the overriding North American plate, relative motions between the Farallon, Kula, and North American plates, and nature of subducting lithosphere, including varying aged oceanic crust, oceanic plateaus, and accreting terranes. Paleo-plate motions have been estimated from a combination of plate circuit reconstructions using sea floor magnetic stripe data and hot spot tracks (Engebretson et al., 1985; Doubrovine and Tarduno, 2008; Gurnis et al., 2012; Seton et al., 2012), but have significant uncertainties during Jurassic to Early Cretaceous time due to limited preservation of older oceanic crust and potential for additional unrecognized oceanic plates. Absolute plate motion models that incorporate movement of hot spot plumes (Steinberger et al.,

2004), paleomagnetic data tied to reference longitudes (Torsvik et al., 2008), and distribution of slab remnants (van der Meer et al., 2010), provide a framework to relate plate motion with mantle flow (Dobrovine et al., 2012; Williams et al., 2015). The APW path for North American records northward drift during the Jurassic, followed by latitudinal standstill during much of the Cretaceous, while sea-floor stripe data record westward drift of North American starting in the Middle Jurassic and becoming more rapid from ~120–70 Ma during opening of the Atlantic Ocean basin (Torsvik et al., 2008). Rates of absolute plate motion for the overriding North American plate were slow during the earliest Cretaceous and increased to an estimated ~3–6 cm/yr from ~120–70 Ma (Fig. 4B), with motion at overall high angles to the plate margin. Net convergence rates between the North American and Farallon plates were likely high during the Middle to Late Jurassic, but presence of additional marginal oceanic plates and terranes complicates interpretation. Relative convergence rates decreased to an estimated ~5 cm/yr from ~145–130 Ma, increased to 8–12 cm/yr from ~120–50 Ma, and then decreased after 50 Ma (Fig. 4B), along with a change in subduction geometry and development of slab-free windows during the Eocene (Humphreys, 1995; Madsen et al., 2006). Directions of relative motion between the Farallon and North American plates were about orthogonal to much of the overall NNW-trending plate margin during the Middle Jurassic, changing to sinistral oblique convergence during latest Jurassic to earlier Cretaceous time, with sinistral transtension along the Mojave-Sonora shear and Bisbee basin system (Dickinson and Lawton, 2001b). The direction of relative motion changed to SW-NE during the Late Cretaceous and Paleogene with dextral oblique convergence along much of the margin, which was greatest to the north where the Kula plate developed at ~80 Ma (Fig. 4A). The dip of the subducting Farallon plate, interpreted from location of the magmatic arc (Dickinson and Snyder, 1978; Lipman, 1992; Gehrels et al., 2009), decreased along most of the margin during the

Cretaceous, and a flat-slab segment developed in the Late Cretaceous along the Mojave sector, likely related to subduction of an oceanic plateau (Livaccari et al., 1981), interpreted to be the conjugate to the Shatsky Rise (Fig. 4A; Saleeby, 2003). Seismic tomography anomalies in the mantle beneath central to eastern North America are interpreted to represent remnants of the Farallon and Kula slabs, including a broad area that may correspond to the flat slab-segment (Liu et al., 2010; Sigloch, 2011). A combination of increased overriding North American plate motion, increased convergence rates, and subduction of oceanic plateaus and terrane accretion are interpreted to have increased plate coupling, leading to intraplate shortening. An alternative interpretation of slab remnants, subduction polarity, and terrane accretion was given by Sigloch and Mihalynuk (2013).

Oblique plate convergence was partly partitioned into strike-slip transport and transpression of accreted terrane, forearc, and magmatic arc rocks (Oldow et al., 1989; Tikoff and de Saint Blanquat, 1997). Models for magnitudes and timing of strike-slip transport, however, remain controversial (e.g., Irving et al., 1996; Housen and Beck, 1999; Butler et al., 2001; Enkin et al., 2002, 2006), including the Baja-BC model that predicts large offsets of terranes (Cowan et al., 1997). Fault offsets (Price and Carmichael, 1986; Wyld et al., 2006) and DZ patterns of overlapping sedimentary strata (Mahoney et al., 1999; LaMaskin et al., 2011) suggest moderate (<1000 km) northward transport of terranes relative to stable North America during Late Cretaceous to Paleogene time. Paleomagnetic data for volcanic rocks are also consistent with moderate transport, whereas paleomagnetic data for many sedimentary rocks are suggestive of larger transport. However, sedimentary rocks typically preserve shallower paleomagnetic inclinations (I_s) than corresponding paleo-field inclinations (I_0), reflecting inclination shallowing during compaction that results in too low of estimated paleolatitude (Kent and Irving, 2010). The

corrected paleolatitude, (ϕ_{cor}), is given by $2*\tan(\phi_{cor}) = \tan(I_0) = (1/f)*\tan(I_s)$, where f is the flattening factor (Tauxe and Kent, 2004). Inclination shallowing can be evaluated using anisotropy of anhysteretic remanence (AAR) and secular dispersion of individual paleomagnetic vectors for large data sets (E/I analysis; Tauxe and Kent, 2004). Timing of remanence acquisition is also critical for determining reference pole positions and evaluating patterns of latitudinal transport.

Paleomagnetic data for volcanic rocks and hematite-bearing chert of the ~170 Ma Coast Range Ophiolite in California indicate limited (300 ± 700 km) northward transport (Hagstrum and Murchey, 1996; Hagstrum and Jones, 1998). Paleomagnetic data for Jurassic volcanoclastic rocks in the Klamath Mountains are also consistent with limited transport (Mankinen et al., 1989). Paleomagnetic data for Cretaceous forearc sedimentary rocks in southern California have shallow inclinations that have been interpreted to record ~1000–1500 km north transport (Lund et al., 1991). However, Tan and Kodama (1998) and Kodama and Davi (1995) evaluated inclination shallowing using AAR and obtained $f \sim 0.7$ and $\sim 10^\circ$ corrections to paleolatitude, with an estimated 200–800 ± 800 km north transport for ~80–70 Ma forearc strata (after restoring Neogene slip along the San Andreas system). Paleomagnetic data for ~80 Ma red beds, which tend to undergo less flattening, indicate 500 ± 600 km north transport (Whidden et al., 1998). Within the Franciscan Complex, paleomagnetic data for Early Jurassic cherts that were deposited on oceanic basalts record equatorial latitudes (Hagstrum and Murchey, 1993); these oceanic basalts and cherts were subsequently transported to the NE within the Farallon plate and accreted along the Cordilleran plate margin during the mid Cretaceous. Paleomagnetic data for Early Cretaceous forearc strata in the Ochoco Basin of central Oregon have been interpreted to record 1800 ± 500 km northward transport (Housen and Dorsey, 2005), though inclination shallowing has not been quantified.

To the north in British Columbia, paleomagnetic data for the 100-104 Ma Spences Bridge volcanics that overlap parts of the Insular and Intermontane terrane groups indicate moderate (1000 ± 500 km) north transport (Irving et al., 1996; Haskin et al., 2003). In contrast, shallower paleomagnetic inclinations in the overlying ~95 Ma Silverquick sedimentary sequence have been interpreted to record up to 3000 km north transport (Wynne et al., 1995; Enkin et al., 2003). Krijgman and Tauxe (2005) used E/I analysis to evaluate inclination flattening and obtained values for f from 0.75 to 1.0 at a 95% confidence interval; using f of 0.75 gives an estimated 1500 ± 500 km north transport, consistent with the underlying volcanic rocks. Sedimentary rocks of the Upper Cretaceous (~80-65 Ma) Nanaimo Group, which overlaps the Insular terrane group, contain shallow inclinations that have been interpreted to record up to 3000 km north transport (Ward et al., 1997; Enkin et al., 2001). However, Kim and Kodama (2004) estimated $f \sim 0.7$ using AAR, and Krijgman and Tauxe (2005) estimated $f \sim 0.7$ using E/I analysis, giving a $\sim 10^\circ$ correction to paleolatitude and 1600 ± 800 km north transport. Farther north, data for 70 Ma Carmacks volcanics indicate 1900 ± 600 km north transport (Enkin et al., 2006), and data corrected for inclination shallowing from Late Cretaceous strata that overlap Wrangellia indicate 1700 ± 900 km north transport (Stamatakis et al., 2001). Paleomagnetic studies of some plutons have suggested large (>3000 km) northward transport (e.g. Housen and Beck, 1999), whereas studies of other plutons indicate limited (<1000 km) transport (Butler et al., 2001), partly reflecting uncertainties in determining paleo-horizontal in intrusive rocks. Although magnitudes of forearc and terrane displacements remain debated, paleomagnetic and structural data indicate oblique convergence was partly partitioned into important strike-slip transport and transpression along the plate margin. Data from volcanic rocks and sedimentary rocks in which inclination flattening has been evaluated suggest limited ($\sim 200\text{--}800$ km, ± 800 km) north transport in the California area (Kodama and

Ward, 2001), and moderate (~1000–1800 km, ±800 km) transport in the British Columbia area, slightly greater than, but within combined uncertainties of strike-slip fault and shear zone offsets (Wyld et al., 2006). Greater strike-slip transport in the north may reflect more oblique convergence with the Kula plate from ~80–50 Ma, and requires extension along the plate margin, interpreted to have been partly accommodated by development of the Columbia River embayment.

The current configuration of western North America resulted from a long tectonic history with: Jurassic to Paleogene subduction, terrane accretion, strike-slip partitioning, and intraplate shortening; late Paleogene to early Neogene extensional collapse of thickened crust; and Neogene Basin and Range regional extension and strike-slip motion along the San Andreas Fault system. This protracted deformation history progressively modified tectonic features, which have been palinspastically restored based on McQuarrie and Wernicke (2005), Wyld et al. (2006), Dickinson (2013), and our own work. These restorations, along with plate reconstructions and positions of mantle tomographic anomalies that likely represent remnants of subducted oceanic slabs (Liu et al., 2010; Jones et al., 2011), are used to develop a series of paleogeographic maps (Fig. 5) and corresponding schematic crustal cross sections (Fig. 6) along the tectonic transect from California to Wyoming, which illustrate overall evolution of the orogenic system during Jurassic to Paleogene time. The preceding overview represents our efforts to synthesize available data and provide a context for exploring evolution of the Sevier and Laramide belts within the Cordilleran orogenic system. For an overview of the Canadian Cordillera see Evenchick et al. (2007).

4. Tectonic Components of the North American Cordilleran Orogenic System

Interrelated components of the Cordilleran orogenic system included a forearc accretionary complex and basin, magmatic arc, retroarc hinterland, Sevier fold-thrust belt, foreland basin, and Laramide foreland belt, as described below. Emphasis in the following description of components

is placed along the tectonic transect from California to Wyoming (Fig. 1), with particular focus on evolution of the Sevier and Laramide belts.

4.1. Forearc accretionary complex and basin

The forearc along the tectonic transect in California began developing during Early to Middle Jurassic accretion of the Sierra Foothills-Klamath Mountains terranes onto North America, followed by formation of the E-dipping Franciscan subduction zone (Fig. 7; Dickinson, 2008). Although details have been obscured by subsequent deformation, two volcanic arcs, separated by an oceanic basin that may have been part of the Mezcalera plate (Dickinson and Lawton, 2001a), appear to have been active during the Triassic to Early Jurassic (Dickinson, 2008; Ernst et al., 2008). The intervening basin closed by ~175 Ma, culminating in an early phase of deformation within the Sierra Foothills and Klamath Mountains provinces (Wright and Fashan, 1988; Girty et al., 1995; Ernst, 2011). Reorganization of the plate margin from ~175-160 Ma was marked by initiation of an E-dipping subduction zone, sinking of the older ocean basin slab, formation of the Coast Range and Great Valley ophiolites, and deposition of the Mariposa and Galice formations that overlapped the accreted terranes (Godfrey et al., 1997; Ingersoll, 2000; Shervais et al., 2005; Wakabayashi et al., 2010). Convergence along the new subduction zone was associated with intrusion of calc-alkaline plutons within a developing magmatic arc, and Late Jurassic contractional deformation (Nevadan orogeny) that locally reactivated terrane boundary faults (Schweickert et al., 1984; Wright and Fahan, 1988).

The Franciscan Complex formed along the new E-dipping subduction zone during a protracted history of offscraping, underplating, and episodes of subduction erosion (Figs. 6, 7; Hamilton, 1969; Ernst, 1970; Wakabayashi, 1992, 2015; Snow et al., 2010). Accreted materials consisted mostly of clastic sediments eroded from the magmatic arc and deposited at the trench

front, along with lesser amounts of basalt, chert, and limestone derived from subducting oceanic plates, and rare high-grade blocks of metabasite (Shervais and Kimbrough, 1985; Wakabayashi, 1992; Dumitru et al., 2007). Accreted materials formed a series of stacked sheets bounded by high-strain zones interpreted to represent paleo-megathrusts, with individual sheets composed of varying combinations of *mélange*, broken formation, and coherent bedded intervals (Wakabayashi, 2015). Within northern California, the Franciscan Complex is divided into eastern, central, and coastal belts, with clastic sedimentary protolith ages, deformation ages, and metamorphic grade decreasing overall westward and structurally downward (Dumitru et al., 2010; Wakabayashi, 2015). Clastic sediments have similar detrital zircon and geochemical signatures compared to temporally correlative forearc strata of the Great Valley Group, with Early Cretaceous sediment sources mostly from accreted terranes and Jurassic volcanics, and Late Cretaceous sources mostly from more deeply eroded parts of the magmatic arc (Dumitru et al., 2015).

The eastern belt includes from east to west, the South Fork Mountain schist, Valentine Formation, and Yolla Bolly terrane. The eastern belt was juxtaposed against the structurally overlying Coast Range ophiolite and Great Valley Group of the forearc basin along the Coast Range fault zone, which is interpreted to have initiated along the subduction zone at ~160 Ma and been subsequently reactivated as a normal fault that helped exhume the accretionary complex and generate accommodation space for deposition of forearc strata (Unruh et al., 2007; Ernst, 2011). Protoliths of the South Fork Mountain schist include ~135 Ma oceanic basalt and chert overlain by coarsening upwards siliciclastics, which were subsequently subducted to estimated depths of ~25–35 km and metamorphosed at blueschist facies at 123 Ma (Dumitru et al., 2007). This history records sea floor spreading with increasing clastic input as an oceanic plate approached the trench. The Valentine Formation contains coarser volcanoclastics with maximum depositional ages as

young as 123 Ma, which underwent blueschist metamorphism at 117 Ma. The Yolla Bolly terrane consists mostly of clastics with maximum depositional ages of ~120–110 Ma, along with a slab of ~170 Ma oceanic basalt and chert, which underwent lower blueschist metamorphism at ~110 Ma (Dumitru et al., 2010, 2015). Limited accretion prior to 125 Ma may reflect thin trench fill during early arc growth, whereas rapid accretion after ~125 Ma appears associated with thicker trench fill derived from the growing arc. The central belt consists mostly of *mélange* comprised of disrupted volcaniclastic mudstone, sandstone, and exotic blocks, which likely formed by a combination of sedimentary (Wakabayashi, 2015) and tectonic (Cloos, 1982) processes, and subsequently underwent low-grade metamorphism. Strata contain rare Late Cretaceous fossils and have DZ maximum depositional ages of ~100–80 Ma (Dumitru et al., 2010, 2105; Ernst, 2011).

The eastern and central belts also contain small, high-grade (upper blueschist, amphibolite, and eclogite facies) metabasite blocks with geochemical signatures similar to the Coast Range ophiolite, which were metamorphosed from ~170–135 Ma at pressures of 1.6–2.4 GPa (estimated depths of ~50–80 km assuming lithostatic pressure, or ~20–50 km depths assuming tectonic overpressure up to 1 GPa that can be reached in subduction channels; Manctelov, 2008) during early underplating (Wakabayashi and Dumitru, 2007). These blocks formed at higher T/P gradients compared to typical subduction zones with depressed thermal gradients, consistent with metamorphism during early subduction of young oceanic lithosphere (Anczkiewicz et al., 2004). The high-grade blocks were subsequently exhumed, likely during extension above deeper zones of continued underplating (Platt, 1986) or return flow along the subduction channel (Cloos, 1982), and recycled into accreting materials.

Formation ages of basalt (with MORB and OIB geochemical signatures) and chert increase going from the South Fork Mountain schist (~135 Ma) to the Yolla Bolly terrane (~170 Ma) that

were accreted from 125 to 110 Ma, reaching a maximum formation age of ~190 Ma and then decreasing at successively younger levels within the central belt (Fig. 4C; Shervais et al., 2005; Wakabayashi, 2015). The age pattern of basalts is interpreted to record subduction of crust that had formed in a marginal ocean basin prior to ~110 Ma, followed by subduction of older Farallon plate crust generated at an active midocean ridge (Fig. 7B).

The coastal belt, which only developed in the northern part of the Franciscan Complex, contains imbricate thrust sheets with coherent sandstone-bearing intervals that underwent very low-grade metamorphism (Snow et al., 2010). Abundant Late Cretaceous to Paleogene DZ grains and detrital muscovite bracket maximum depositional ages and indicate sources from the Idaho batholith region that was exhumed during extension within Paleogene metamorphic core complexes (Dumitru et al., 2013).

The forearc basin along the tectonic transect is represented by the Great Valley Group, which comprises a ~2- to 7-km-thick sedimentary prism deposited on the Coast Range and Great Valley ophiolites to the west, and on previously accreted terranes to the east (Figs. 6, 7; Dickinson and Rich, 1972; Williams and Graham, 2013). The Great Valley Group consists of mudstone, sandstone, and conglomerate that were derived mostly from erosion of the growing magmatic arc and deposited mostly as submarine fan and slope turbidites, with local bypass of sediments along submarine canyons to the trench (Ingersoll, 1978, 1979). The latest Jurassic(?) to earlier Cretaceous (~145–130 Ma) lower part of the group (Stony Creek Formation) was deposited in the western part of the basin, which underwent local tilting and syndepositional faulting (Constenius et al., 2000), and was sourced mostly from erosion of accreted terranes and volcanics on the west side of the magmatic arc (Sharman et al., 2015). A change in basin structure and angular unconformity mark the base of the overlying Lodoga Formation that was deposited from ~125–100 Ma

(Ingersoll, 1978; Constenius et al., 2000). Mid Cretaceous (~100–80 Ma) strata progressively onlapped eastward during increased tectonic subsidence, which was followed by another change in basin structure during later Cretaceous time, with uplift of the western part of the basin and progradation of deltaic strata along the eastern part during decreasing subsidence, interpreted to record subduction shallowing and thermal contraction of the now inactive Sierra arc (Moxon and Graham, 1987; Williams and Graham, 2013). Petrographic (Ingersoll, 1978), DZ (DeGraaff-Surpless et al., 2002; Sharman et al., 2015), and geochemical (Surpless, 2015) data from mid Cretaceous to Paleogene strata record increasing input from more deeply eroded plutonic rocks within northern to central California, with a drainage divide located along the arc. The drainage divide, however, breached across the arc in the Mojave sector into the Mogollon Highlands, interpreted to record forearc disruption and interior uplift during passage of a flat-slab segment (Sharman et al., 2015). The Great Valley Group was little deformed despite its proximity to the subduction margin, possibly because of the relatively strong nature of the underlying ophiolitic basement. Alternatively, Wright and Wyld (2007) proposed that the Jurassic-Early Cretaceous lower part of the group was originally deposited farther south and then transported northward during later dextral strike-slip faulting.

Forearc characteristics vary south and north of the tectonic transect, related to differences in subduction histories. To the south, the Catalina schist was deposited in a forearc to accretionary wedge setting from ~115–95 Ma and metamorphosed at moderate T/P conditions during mid Cretaceous underthrusting beneath the Peninsular plutonic complex (Grove et al., 2008). In the Mojave sector, the Pelona-Orocopia-Rand schists were underplated during the Late Cretaceous, with deformation and cooling ages decreasing from ~85 to 55 Ma toward the ESE (Fig. 5; Jacobsen et al., 2011). The schists consist mostly of metagreywacke, along with lesser amounts of

metabasalt and metachert that represent offscraped trench sediments and subducted oceanic crust, similar to the Franciscan Complex. However, the schists were metamorphosed at higher grade (T ~300–600 °C) and are in tectonic contact with the base of the Sierra Nevada magmatic arc, recording removal of the arc root and mantle lithosphere, likely during flat-slab subduction of an oceanic plateau (Saleeby, 2003). DZ data bracket Late Cretaceous depositional ages and indicate sources from the magmatic arc and Mogollon Highlands as the arc was erosionally breached (Jacobsen et al., 2011). Extension and intrusion of two-mica, peraluminous granites across the Mojave sector from ~75–65 Ma may record decreased plate coupling and partial melting of underplated sediments and crust during passage of the trailing edge of the oceanic plateau and asthenospheric influx (Miller et al., 1996; Saleeby, 2003), or during mantle lithosphere delamination (Wells et al., 2012).

North of the transect, Cretaceous marine turbidites of the Ochoco basin in central Oregon were deposited along the forearc (Housen and Dorsey, 2005). Farther north, strata of the Methow-Tyughton basin (Fig. 5) record: (1) Late Jurassic to Early Cretaceous deposition of marine turbidites in a forearc basin with sediment sources from a magmatic arc to the east; and (2) Late Cretaceous deposition of fluvial strata in a retroarc basin, following accretion of the Insular terrane group at ~100 Ma (DeGraaff-Surples et al., 2003).

4.2. *Magmatic arc*

The Cordilleran magmatic arc comprises a belt of calc-alkaline intrusive and volcanic rocks, including the Peninsular, Sierra Nevada, Idaho, and Coast Mountain plutonic complexes (Fig. 5; Hamilton and Myers, 1967; Hyndman, 1983; Saleeby and Busby-Spera, 1992; Armstrong and Ward, 1993; Saleeby et al., 2008; Gehrels et al., 2009). Geochemical characteristics of arc rocks record contributions of primary mantle melts, remelting of mantle-derived gabbro and

diorite, and recycling of crustal components (DePaolo, 1981; Wenner and Coleman, 2004; Giardi et al., 2012). The Sierra Nevada complex along the tectonic transect has a current crustal thickness of ~30-35 km and consists mostly of tonalite to granodiorite, along with lesser amounts of wall rock and volcanic pendants (Ducea, 2001; Gray et al., 2008). Petrologic constraints and characteristics of xenoliths indicate that the complex previously had an ~70-km-thick eclogitic root composed of mafic to ultramafic cumulates and restites left over from partial melting of lower continental crust and gabbroic intrusions (Saleeby et al., 2003). The dense eclogitic root was subsequently delaminated due to its negative buoyancy (Zandt et al., 2004). Although igneous rock types are similar along the length of the Sierra Nevada complex, isotopic and trace element patterns record input of varying crustal components, from juvenile accreted terranes in the north (Cecil et al., 2012), to North American basement in the south (Ducea, 2001).

Magma addition rates, estimated from geochronology and outcrop areas of igneous rocks, varied over time from high rates (flare-ups) to low rates (lulls), reflecting changes in subduction rates and associated corner flow of fertile asthenosphere, and rates of partial melting of thickened crust and underplated trench sediments (Fig. 8A; Ducea and Barton, 2007; Paterson et al., 2011; Chapman et al., 2013; Paterson and Ducea, 2015). Estimated addition rates, however, have large uncertainties as the nature of eroded volcanic material and depth extent of underlying plutons must be inferred. Detailed geochronologic data suggest plutonic complexes were incrementally assembled by intrusion of numerous, smaller sheet-like bodies (Coleman et al., 2004). Age distributions of volcanic DZ grains in forearc and retroarc basin strata, which provide proxies for magmatism, display broadly similar regional patterns (Fig. 8B; DeGraff-Surpless et al., 2002; Barth et al., 2013; Laskowski et al., 2013; Dumitru et al., 2015; Sharman et al., 2015). The Sierra Nevada complex had flare-ups from 170–150 Ma and 100–85 Ma, a lull from 140–125 Ma, and

531 ceased to have calc-alkaline activity after ~80 Ma (Fig. 8A; Chen and Moore, 1982; Saleeby et al.,
532 1987; Ducea, 2001). The early flare-up may reflect influx of fertile asthenosphere along the newly
533 initiated subduction zone, whereas the 140–125 Ma lull correlates with a period of reduced
534 convergence rates, similar to patterns along other sectors of the arc. The 100–85 Ma flare-up partly
535 resulted from thickening and partial melting of lower crustal material beneath the arc based on Nd,
536 Sr, and O isotopic data (Ducea and Barton, 2007). Arc magmatism along the Sierra Nevada
537 complex migrated eastward from 120 to 80 Ma, likely in response to decreasing subduction angle
538 (Fig. 8D; Coney and Reynolds, 1977; Dickenson and Snyder, 1978; Lipman, 1992). Cessation of
539 calc-alkaline igneous activity after 80 Ma is interpreted to reflect development of a flat-slab
540 segment along the Mojave sector and disruption of asthenosphere corner flow (Saleeby, 2003).
541 Limited later Cretaceous magmatism, however, propagated far inboard into the Colorado Mineral
542 Belt above the flat slab (Chapin, 2012). A fundamental switch in magmatic style occurred across
543 the western U.S. during Middle Eocene to Oligocene time, marked by widespread eruption of calc-
544 alkaline volcanic rocks with subduction-related geochemical signatures (Armstrong and Ward,
545 1991; Best et al., 2009). Ages of volcanic rocks young overall from north to south, which is
546 interpreted to record progressive removal of the flattened slab and influx of asthenosphere that
547 interacted with previously hydrated mantle lithosphere (Humphreys et al., 2003).

548   Plutons and wall rocks within the Sierra Nevadan arc underwent punctuated deformation,
549 with intervals of increased strain rates ($\sim 10^{-15} \text{ s}^{-1}$ with locally higher rates in shear zones) during
550 Middle to Late Jurassic and mid Cretaceous flare-ups, along with intervals of decreased strain rates
551 ($< 10^{-16} \text{ s}^{-1}$) during magmatic lulls and little deformation after ~80 Ma cessation of magmatism
552 (Cao et al., 2105). The style of deformation also changed over time, partly in response to obliquity
553 of plate convergence. During Middle to Late Jurassic near orthogonal convergence, wall rocks

experienced bulk coaxial strain and shortening along WSW- and ENE-vergent thrust faults, and NNW-striking foliations developed in plutons. During ~95–80 Ma dextral oblique convergence, wall rocks experienced transpressive strain and development of dextral shear zones (Tikoff and de Saint Blanquat, 1997), and NW- to WNW-striking foliations developed in progressively younger plutons (Cao et al., 2015). Shear zones have steep foliations, with widespread steep lineations to locally gently plunging lineations in areas with greater simple shear (Tikoff and Green, 1997). Shear zones comprise a network that developed along the magmatic arc that was likely weak, with locations of elongate plutons partly related to releasing bends in the network (Tikoff and de Saint Blanquat, 1997). Patterns are broadly consistent with models of transpressive strain and strike-slip partitioning during oblique convergence (Dewey et al., 1998; Fossen and Tikoff, 1998).

To the south of the transect, the Peninsular complex includes a western belt of volcanic rocks and plutons emplaced from 140–105 Ma, an eastern belt of voluminous plutons with crustal melt components emplaced during a 100–85 Ma flare-up, and a diffuse arc farther east that developed after 80 Ma (Silver and Chappell, 1988; Kimbrough et al., 2001). To the north, the Idaho batholith includes a western border zone of 120–90 Ma calc-alkaline tonalites deformed during dextral transpression along the West Idaho shear zone (McClelland et al., 2000; Giorgis et al., 2005), a main lobe consisting of metaluminous granitoid plutons intruded from ~100–85 Ma and voluminous peraluminous granodiorite and two-mica granite that incorporated crustal melts intruded from ~80–65 Ma, and the smaller Bitterroot lobe with abundant 66–54 Ma peraluminous granites (Hyndman, 1983; Foster and Fanning, 2001; Gaschnig et al., 2010). Late Cretaceous magmatism also extended eastward and interacted with the Sevier belt in western Montana (Lageson et al., 2001). Farther north, the Coast Mountain complex of Canada had protracted Jurassic-Paleogene activity that overlapped with terrane accretion (Friedman and Armstrong, 1995;

Gehrels et al., 2009). An outboard arc in the Insular terrane group was active from ~160–140 Ma, with a lull from ~140–120 Ma, followed by accretion onto the North America margin at ~100 Ma, crustal thickening (Rubin et al., 1990), and development of a single arc, which migrated eastward and had flare-ups at ~100–80 Ma and ~60–50 Ma (Friedman and Armstrong, 1995). A mid Cretaceous dextral transpressional to later Cretaceous and Paleogene transtensional crustal shear zone, the Coast shear zone, developed along the arc, and was marked by intrusion of regionally extensive tonalite sills (Gehrels et al., 1992; Klepeis et al., 1998; McClelland et al., 2000).

Regional patterns suggest interrelations between magmatism, intraplate shortening, and subduction dynamics. Eastward migration of the arc and shallowing of subduction during the mid Cretaceous correlate with increasing rates of overriding plate motion and relative convergence, upper crustal shortening in the Sevier belt, and crustal thickening in the hinterland to magmatic arc. Flare-ups along the arc had order of magnitude greater fluxes compared to background rates and were associated with increased $^{87}\text{Sr}/^{86}\text{Sr}$, decreased ϵ_{Nd} , and increased $\delta^{18}\text{O}$ that record crustal melt components, as well as increased La/Yb ratios that likely record partial melting in the presence of garnet within an eclogitic arc root (Fig. 8C; Giardi et al., 2012; Paterson and Ducea, 2015). DeCelles et al. (2009) proposed that this cyclic magmatism was tied to lower crustal thickening beneath the arc during retroarc shortening. In this model, crustal thickening leads to eclogite generation followed by foundering of the dense arc root, influx of hot asthenosphere, and partial melting of remaining crust that ignites a flare-up (Ducea and Barton, 2007), followed by a magmatic lull as a new arc root forms and the cycle repeats. The concept of a Cordilleran tectonic-magmatic cycle is further explored in the Discussion section.

4.3. Retroarc hinterland

The hinterland, spanning the region from the magmatic arc to the Sevier fold-thrust belt, developed during a complex tectonic history of middle- to lower-crustal thickening and metamorphism, multiple episodes of igneous intrusion, and subsequent extension (Armstrong, 1968; Snoke and Miller, 1988; Wells et al., 2012). Along the tectonic transect in Nevada to NW Utah, the hinterland includes the western Luning-Fencemaker fold-thrust belt (Wyld, 2002), central Nevada fold-thrust belt (Taylor et al., 2000), poly-deformed mid-crustal rocks exhumed in the Albion-Raft River-Grouse Creek and Ruby-East Humboldt metamorphic core complexes, and gently folded, little metamorphosed Paleozoic strata to the east (Fig. 5; Snoke and Miller, 1988; Miller and Gans, 1989; Miller and Hoisch, 1995; Wells, 1997; McGrew et al., 2000; Long, 2015). Paleozoic strata are unconformably overlain by Oligocene volcanic rocks over much of the hinterland, recording limited pre-Neogene erosional exhumation (Armstrong, 1968; Long, 2013). Locally preserved synorogenic strata include the Aptian-Albian Newark Canyon Formation that was deposited in piggy back basins east of the Central Nevada belt, and the Upper Cretaceous to Paleogene Sheep Pass Formation deposited in internally drained basins during upper crustal extension (Vandervoot and Schmitt, 1990; Druschke et al., 2009, 2011).

During Jurassic time, backarc basin strata in the western part of the hinterland were shortened by ~100 km in the Luning-Fencemaker belt (Fig. 6A; Wyld, 2002; Wyld et al., 2003). Unconformably overlying strata of the ~135–125 Ma King Lear Formation were deposited following exhumation of the belt (Martin et al., 2010). Middle to Late Jurassic plutons and dikes of gabbro and metaluminous granitoids (mostly monzonite and granodiorite) were intruded at shallow levels across parts of the hinterland (Fig. 8D; Miller and Hoisch, 1995). Juvenile isotopic signatures of these igneous bodies record partial melting of mantle lithosphere (Wright and Wooden, 1991), possibly associated with enhanced asthenosphere convection during sinking of an

older subducted slab and initiation of the E-dipping Franciscan subduction zone. Parts of future metamorphic core complexes experienced initial tectonic burial (Cruz-Urbe et al., 2015), and the eastern hinterland experienced limited Late Jurassic shortening and contact metamorphism near igneous intrusions (Miller and Hoisch, 1995; Allmendinger and Jordan, 1981).

During Cretaceous to Paleogene time, lower- to mid-crustal levels of the hinterland experienced Barrovian metamorphism with overall clockwise P-T-t paths that record tectonic burial to depths of 20–30 km in excess of primary stratigraphic thickness (Fig. 9A; Hodges et al., 1992; Smith et al., 1993; Camilleri and Chamberlain, 1997; McGrew et al., 2000; Hoisch et al., 2002; Harris et al., 2007; Wells and Hoisch, 2008; Hallett and Spear, 2013; Kelly et al., 2015), and intrusion of peraluminous two-mica granites that incorporated crustal melts (Barton, 1990; Lee et al., 2003). Within the Albion-Raft River-Grouse Creek complex, relations record: (1) Early Cretaceous tectonic burial with an episode of early exhumation at ~135 Ma (Kelly et al., 2015); (2) mid Cretaceous crustal thickening, emplacement of an extensive thrust sheet (Quartzite allochthon) and vertical flattening of its tectonically buried footwall to the east (Wells et al., 2008), and metamorphism at maximum depths to ~35 km and T ~550 °C by ~85 Ma (Wells and Hoisch, 2008; Wells et al., 2012); (3) later Cretaceous to Paleocene alternating local extension along detachment faults and shortening by refolding with maximum T ~650 °C by ~65 Ma (Wells, 1997; Wells et al., 2012); and (4) Eocene to Oligocene collapse, late-stage igneous intrusion, and rapid extension along the Middle Mountain shear zone, followed by Miocene extension along the Raft River detachment (Wells, 2001). Within the Ruby-East Humboldt complex, relations record: (1) likely Early Cretaceous prograde metamorphism; (2) mid Cretaceous crustal thickening, emplacement of nappes and an extensive thrust sheet (Windemere sheet), which buried the footwall to the east that underwent flattening (Camilleri, 1998), and metamorphism at maximum depths to ~40 km and T

645 ~600 °C by ~85–75 Ma (Hallett and Spear, 2013); (3) later Cretaceous to Paleocene local
646 extension to shortening with maximum T ~700–800 °C, migmatization, and intrusion of
647 peraluminous granites; and (4) Eocene to Oligocene collapse, late-stage igneous intrusion, and
648 rapid exhumation along extensional shear zones (McGrew et al., 2000). Early Cretaceous
649 shortening in the Central Nevada belt may have transferred along strike into early shortening at
650 deeper levels in the Ruby to Albion metamorphic complexes (Taylor et al., 2000).

651   The metamorphic core complexes within the tectonic transect lie along a belt that stretches
652 from Canada to Arizona, which is characterized by poly-deformation of lower- to mid-crustal
653 rocks and exhumation along major extensional shear zones (Fig. 1; Davis and Coney, 1979; Coney
654 and Harms, 1984). These complexes also displays important along strike variations. To the north in
655 Canada, the Monashee and related core complexes lie in the hinterland Omineca belt that
656 experienced: Jurassic crustal thickening and igneous intrusion that overlapped with initiation of
657 faulting in the western part of the retroarc fold-thrust belt (Pana and van der Pluijm, 2015);
658 Cretaceous polyphase metamorphism, igneous intrusion that incorporated crustal melts, and
659 development of shear zones, some of which linked eastward into the fold-thrust belt; Paleocene
660 migmatization and onset of extension in thermally weakened crust; and Eocene regional extension
661 with development of transtensional shear and fault zones (Norlander et al., 2002; Simony and Carr,
662 2011; Gervais and Brown, 2011; Gervais and Hynes, 2013). To the south in Arizona, core
663 complexes underwent Miocene extension that followed crustal thickening mostly related to prior
664 igneous intrusions (Davy et al., 1989). Localized extension in the hinterland and enhanced igneous
665 activity with eruption of large ignimbrite sheets progressed southward from Eocene to Miocene
666 time, likely recording progressive slab removal and influx of hot asthenosphere (Armstrong and
667 Ward, 1991; Best et al., 2009).

Multiple lines of evidence imply that crustal thickening in the hinterland led to formation of a broad orogenic plateau (DeCelles, 2004). Evidence includes restoration of current crustal thickness for Tertiary extension (Coney and Harms, 1984), balancing of mid- to lower-crustal thickening in the hinterland with upper-crustal shortening in the Sevier belt (DeCelles and Coogan, 2006), elevated pressures recorded by metamorphic rocks (Wells et al., 2012; Hallett and Spear, 2014), stratigraphic relations (Druschke et al., 2009, 2011), fossil assemblages (Wolfe et al., 1998; Bonde, 2014), and stable isotopic compositions of primary carbonate, hydrated volcanic glass, and authigenic clay (Chamberlain and Poage, 2000; Horton et al., 2004; Cassel et al., 2009; Mix et al., 2011; Chamberlain et al., 2012; Snell et al., 2014). However, the timing, magnitude, and spatial extent of surface uplift across the hinterland, and relations to regional tectonics, remain debated. Models for paleo-elevation include Late Cretaceous uplift of a broad plateau partly in response to crustal thickening (Coney and Hams, 1984; Jones et al., 1998; DeCelles, 2004), and early Cenozoic uplift partly related to slab removal and mantle lithosphere delamination (Humphries, 1995; Liu et al., 2010).

A schematic model for tectono-topographic evolution of the hinterland along the tectonic transect is shown in Figure 9B. Fossil assemblages in the Lower Cretaceous Newark Canyon Formation in Nevada are similar to assemblages in the correlative Cedar Mountain Formation of the foreland basin in central Utah, suggesting limited elevation differences at ~120 Ma (Bonde, 2014). Comparison of clumped C-O isotopic values of primary carbonates in the Upper Cretaceous to Paleocene Sheep Pass Formation with correlative strata to the east indicate that the hinterland was at ~2-3 km higher elevation than the foreland by 70 Ma (Snell et al., 2014), which is consistent with crustal thickening to ~50-60 km beneath a broad hinterland plateau that balanced ~200 km of thin-skin shortening in the Sevier belt. Following thickening and plateau uplift, mid- to

upper crustal levels experienced alternating episodes of later Cretaceous to Paleocene synconvergent extension to shortening, reflecting a dynamic interplay between tectonic compressive forces, gravitational potential energy of elevated crust, mantle lithosphere thinning, heating and weakening of the crust, and surface processes (Hodges and Walker, 1992; Camilleri and Chamberlain, 1997; Wells et al., 2012). Erosional exhumation of the hinterland at this time appears limited with strata of the Sheep Pass Formation deposited in internally drained basins, interpreted to reflect topographic isolation of a high plateau (Druschke et al., 2009, 2011). O and H isotopic values become more negative and varied in Paleocene to Oligocene strata, suggestive of further increases in elevation and relief during a late-stage (~50-35 Ma) switch to local extension and flare-up of igneous activity (Mix et al., 2011; Chamberlain et al., 2012). Although different studies listed above have proposed varying interpretations, in general, evidence suggests the hinterland experienced mid Cretaceous crustal thickening and regional uplift, later Cretaceous to Paleocene synconvergent upper crustal extension within an internally drained plateau, and Eocene to Oligocene extension and igneous activity during slab removal.

4.4 Retroarc Sevier fold-thrust belt

The thin-skin Sevier belt extends >3000 km from the Front and Main Ranges of Canada (Dahlstrom, 1970; Monger and Price, 1979) to southeastern California (Burchfiel and Davis, 1972) and is characterized by thrust faults and folds that shortened and translated sedimentary cover rocks overall eastward above a regional decollement (Figs. 5, 6; Armstrong and Oriel, 1965; Armstrong, 1968; DeCelles, 2004). The Sevier belt trends overall NNW to N, subparallel to the overall plate margin, and displays curvature over a range of scales, partly related to architecture of the primary sedimentary prism, including the Alberta, Wyoming, and central Utah salients (Figs. 1, 3; Lawton et al., 1994; Mitra, 1997). The sedimentary cover in the belt underwent ~150–300 km of

shortening, which was matched by mid- to lower-crustal thickening in the hinterland (Price and Mountjoy, 1970; DeCelles and Coogan, 2006). Shortening in the Sevier belt occurred mostly during the Early Cretaceous to Paleogene, but locally began during Middle Jurassic time in western Alberta (Pana and van der Pluijm, 2015) during terrane accretion and early development of the Selkirk fan (Colpron et al., 1996; Gibson et al., 2008), and in the Mojave sector where the belt intersected the magmatic arc (Burchfiel and Davis, 1972).

The Wyoming salient of the Sevier belt along the tectonic transect is bound on the north and south ends by the Gros Ventre and Uinta Laramide foreland arches, and displays $\sim 90^\circ$ of regional curvature with trends of large-scale fold and thrust traces ranging from NW in the north part to NE in the south part (Fig. 10). Major thrusts in the salient are divided into: (1) a western system comprising the Willard, Paris, and Meade thrusts that carried thick Neoproterozoic to Paleozoic passive margin strata; and (2) an eastern system comprising the Crawford, Absaroka, and Hogsback-Darby-Prospect thrusts that carried thinner Paleozoic strata that were transitional with much thinner cratonic strata farther east (Fig. 3). Cross sections of the salient show classic thin-skinned structures that are well constrained by seismic and drill hole data (Fig. 11; Royse et al., 1975; Dixon, 1982; Lamerson, 1982; Coogan, 1992; Royse, 1993). Thrusts have “stair-step” geometries, with ramps in more competent intervals (thick-bedded carbonate and sandstone) and flats in incompetent intervals (shale and evaporites), and are associated with fault-bend, fault-propagation, and detachment folds. The basal decollement lies in micaceous Neoproterozoic strata for the western thrust system, and in Cambrian shale for the eastern system. Underlying basement is not incorporated into thrust sheets, except along a basement ramp where the basal decollement stepped up along the edge of the former passive margin onto the eastern thrust system, forming the Wasatch anticlinorium (Yonkee, 1992). In three dimensions, thrusts display oblique ramps,

transfer zones, and tear faults that accommodated along-strike gradients in displacement (Apotria, 1995; Yonkee and Weil, 2010). Fault slip was overall eastward, but slip directions varied around the salient and were oblique to structural trend near the salient ends, with additional complications where frontal thrusts interacted with basement-cored Laramide arches (Schwartz and Van der Voo, 1984; Hunter, 1988; Bradley and Bruhn, 1988; Paulsen and Marshak, 1999). Internal strain and vertical axis-rotations accompanied thrusting and contributed to total shortening and development of curvature (Mitra, 1994; Weil et al., 2010; Yonkee and Weil, 2010).

Thermochronologic data, along with age and provenance of synorogenic deposits, provide a long-term record of fault timing and erosional unroofing of thrust sheets in the salient (Fig. 12). Data record overall forward (west to east) propagation of thrusts, with slip on the Willard-Paris-Meade thrusts from ~125–90 Ma, Crawford to Early Absaroka thrusts from ~90–80 Ma, and Late Absaroka to Hogsback-Darby-Prospect thrusts from ~70–50 Ma (Dorr et al., 1977; Wiltschko and Dorr, 1983; Burtner and Nigrini, 1994; DeCelles, 1994, 2004; Solum and van der Pluijm, 2007). Limited reactivation of internal thrusts, local out-of-sequence thrusting, and uplift over deep ramps accompanied development of the thrust wedge and helped maintain critical taper (Coogan, 1992; DeCelles and Mitra, 1995). Additional details on exhumation patterns are recorded by detrital thermochronologic studies of synorogenic strata. Minimum lag times for apatite fission track ages (the time between source rock cooling below a closure $T \sim 80\text{--}110^\circ\text{C}$ and the depositional age of synorogenic strata) are < 5 Ma, recording rapid exhumation ($\sim 0.5\text{--}1.0$ km/m.y.) from depths of $\sim 3\text{--}4$ km near the leading edge of the Sevier thrust wedge (Painter et al., 2014). U/Th-He ages of detrital zircon grains, however, are much older and indicate that most sediment source rocks were located above the He partial retention zone ($T \sim 160\text{--}180^\circ\text{C}$) at the onset of thrusting. Consequently, total exhumation was less than $\sim 6\text{--}7$ km, implying rapid exhumation was

progressively localized along the propagating wedge front. A switch to local extensional collapse of the Sevier belt started by 50–45 Ma and continued into the Oligocene from Montana to Utah, marked by development of listric normal faults that reactivated thrust ramps and deposition of volcanoclastic strata in half grabens (Constenius, 1996). The following sections, 4.4.1 to 4.4.3, describe more detailed structural patterns of the western and eastern thrust systems of the Wyoming salient that help frame the tectonic model presented in the Discussion.

4.4.1 Wyoming salient- western thrust system

The western thrust system of the Wyoming salient includes the aerially extensive Willard, Paris, and Meade thrust sheets that were emplaced long distances during the Early Cretaceous (Crittenden, 1972; Yonkee, 2005). In the southern part of the salient, the Willard thrust has ~50 km of ESE slip, in the central part the system branches into the Paris and Meade thrusts with ~20 km and 40 km of E slip respectively, and in the northern part ~60 km of ENE slip is distributed along several imbricate thrusts and associated folds. $^{40}\text{Ar}/^{39}\text{Ar}$ ages of pre- to syn-kinematic mica from lower parts of the sheet vary mostly from 140–110 Ma, recording early alteration and strain that overlapped with onset of thrust slip (Yonkee et al., 1989). Zircon U/Th-He ages range mostly from 125–90 Ma in lower parts of the sheet (Eleogram, 2014), recording protracted cooling and exhumation during thrust slip. The western Sheeprock and Canyon Range thrust sheets in central Utah south of the transect have similar structural style and timing (DeCelles and Coogan, 2006).

The Willard sheet is well exposed in northern Utah where younger uplift and erosion above the Wasatch anticlinorium exhumed a wide range of structural levels (Fig. 13). The Willard thrust has a stair-step geometry with a western flat in micaceous Neoproterozoic strata, a ramp through later Neoproterozoic to Early Cambrian quartzite-rich strata, a central flat in middle Cambrian shale, a central ramp through upper Cambrian to Jurassic carbonate-rich strata, and an eastern flat

in Jurassic evaporite-bearing strata. Imbricate thrusts and large-scale folds accommodated additional shortening within the sheet.

Lithologic characteristics, environmental conditions (temperature, fluids), and structural position (proximity to major thrusts) controlled styles and mechanisms of internal deformation in the Willard sheet. Triassic to Jurassic strata have been eroded from much of the sheet. Paleozoic strata at upper levels experienced temperatures (T) < 200 °C, and include strong dolostone with fracturing, and relatively weaker limestone and shale with minor faults, minor folds, veins, and rare spaced cleavage, which are best developed near imbricate faults. Early Cambrian to later Neoproterozoic strata at middle levels of the thrust sheet experienced T of ~ 200 – 300 °C, and consist mostly of strong quartzite with minor faults, and open minor folds (Fig. 14A). Microscopically, quartz grains are typically little deformed, with limited undulose extinction and microcracks (Fig. 14B). Micaceous Neoproterozoic strata at lower levels underwent greenschist-facies metamorphism at maximum T of ~ 300 – 400 °C, and display increasing deformation intensity with increasing mica content and proximity to the main thrust. Diamictite, greywacke, and slate at lower levels display moderately to strongly developed cleavage, inclined minor folds, ductile shear zones, and vein arrays (Fig. 14C). Quartz grains are deformed into elliptical shapes with widespread undulose extinction, subgrains, and local recrystallization, along with dissolution of grain edges and precipitation of quartz-mica fibers at grain ends, recording a combination of crystal plastic and mass transfer deformation mechanisms (Fig. 14D).

The Willard thrust fault zone includes the tectonized base of the hanging wall, a fault core with concentrated slip, and an imbricated footwall (Fig. 14E). The basal hanging wall displays intense thrust-parallel, locally refolded foliation and multiple vein sets. The fault core consists of mixed mylonite, phyllonite, and cataclasite cut by discrete slip surfaces. Mylonite contains highly

elongate, recrystallized quartz ribbons. Phyllonite contains fine-grained, recrystallized to comminuted matrix cut by variably deformed microveins (Fig. 14F). Cataclasite contains brecciated mylonite and phyllonite fragments. Preserved microstructures record overlapping crystal plastic, mass transfer, and cataclastic deformation, with repeated episodes of fluid influx and alteration.

Mesosopic structural relations vary with structural level in the thrust sheet (Fig. 15A). Cleavage varies from absent in quartzite at middle levels, to well-developed and at acute angles to bedding in micaceous strata at lower levels, to intense and subparallel to bedding in the basal part of the sheet. Veins comprise three diffuse sets: steep-dipping, ESE-striking veins subperpendicular to fold axes that are locally grouped in *en echelon* arrays and accommodated minor tangential (strike-parallel) extension and local wrench shear; bed-subparallel veins associated with top-to-ESE shear; and veins subperpendicular to cleavage at lower levels. Widely spaced minor faults at upper levels include reverse-slip wedge faults at acute angles to bedding and strike-slip tear faults at high angles to bedding. Mostly top-to-E shear zones are found at lower levels. Mesoscopic structures record limited layer-parallel shortening (LPS) at upper levels, layer-parallel extension and vertical flattening in micaceous strata at lower levels, top-to-E shear that increases toward the base of the sheet, minor tangential extension at all levels, and localized wrench shear.

Strain data for the Willard Peak area illustrate kinematic and rheologic relations of different rock types and levels during thrusting. The area includes a northern domain with an oblique section across little deformed quartzite of the Caddy Canyon Formation to more deformed greywacke and diamictite of the Perry Canyon Formation and local slices of basement rock (Facer Formation) at lower levels above a central imbricate, and a southern domain with highly deformed and folded strata near the main Willard thrust (Fig. 16). Basement slices were locally incorporated

into the sheet where the thrust propagated across basement highs that likely formed during Neoproterozoic rifting. In the northern domain, bedding is overall gently dipping, subparallel to the central imbricate, and cleavage defines an asymptotic pattern. Strain ellipses (in cross section) for micaceous clasts within diamictite at lower levels have moderate axial ratios of ~ 2 – 3 and long axes at 10 – 30° angles to the thrust. Micaceous clasts with plastically deformed quartz grains lack strain shadows and approximate matrix strain, whereas less altered, stronger granitic and large quartzite clasts display strain shadows and have lower axial ratios, broadly consistent with experimentally determined flow laws that predict relative strengths of dry feldspar $>$ dry quartz $>$ wet quartz $>$ mica at greenschist facies conditions (Yonkee et al., 2013). Strain ellipses for greywacke have ratios of ~ 1.5 – 2 and angles of 20 – 50° , and ellipses for quartzite farther from the imbricate have ratios typically < 1.2 . The southern domain exposes highly deformed slate and greywacke of the Perry Canyon Formation that are folded into a recumbent syncline and overturned anticline related to thrust propagation and shear. In the basal part of the sheet, strain ratios are ~ 3 – 6 and long axes are subparallel to the thrust. At lower levels, greywacke has strain ratios of ~ 1.5 – 2.5 , whereas pebbly to silty slate has higher ratios of ~ 2.5 – 4 (Fig. 16). In detail, strain intensity is also higher in outcrops with widespread veining, interpreted to record hydrolytic weakening and elevated fluid pressure.

The above described structural and microtextural patterns record kinematic and rheological layering of the Willard thrust sheet. A kinematic model of combined shear, vertical flattening, and minor layer-parallel shortening to extension predicts an asymptotic pattern of cleavage and increasing strain ratios toward the base of the sheet that match observed patterns (Fig. 15A). The vertical gradient in layer-parallel extension results in a horizontal gradient in simple shear to maintain compatibility, which is partly accommodated by slip between layers. Increasing vertical

flattening and shear are interpreted to reflect a relatively weak lower part of the thrust sheet. In comparison, a model with only thrust-parallel simple shear predicts moderate-dipping cleavage for moderate strain ratios (Sanderson, 1982), inconsistent with observed patterns in the Willard sheet.

Microstructural and fluid inclusion data constrain a simplified mechanical model for the lower Willard sheet and basal fault zone (Fig. 15B). Internal deformation at lower levels occurred at strain rates on the order of 10^{-14} to 10^{-15} s^{-1} in mica- to quartz-richer rocks respectively. Veins at high angles to the thrust, boudinaged quartzite beds, and minor normal faults in quartz-cemented pods developed synchronously with concentrated thrust slip in the fault core, consistent with elevated fluid pressure, low deviatoric stress ($\sim 10 \text{ MPa}$), and rotation of maximum stress into high angles with a weak, low-friction fault core. Furthermore, fluid inclusion data indicate elevated fluid pressures, which may have episodically approached lithostatic pressure (Yonkee et al., 1989). A weak fault zone and micaceous lower level, combined with an overlying interval of strong quartzite, favored low wedge taper, large fault slip, and limited internal deformation at middle to upper levels in the western system.

4.4.2 Wyoming salient- Wasatch anticlinorium

The Wasatch anticlinorium marks the transition from the western to eastern thrust systems, where the basal decollement propagated from the base of Neoproterozoic passive margin strata to the west, across a basement step (the Wasatch hinge-zone), up to middle Cambrian shale to the east (Fig. 17; Yonkee, 1992). The anticlinorium has a culmination in northern Utah where imbricate slices of basement were repeated along the Ogden thrust system (Schirmer, 1988; Yonkee, 1992; DeCelles, 1994). Fission track ages of $\sim 90\text{--}75 \text{ Ma}$ and $65\text{--}50 \text{ Ma}$ in basement rocks (Naeser et al., 1983) are interpreted to record progressive uplift of the anticlinorium during slip on the Crawford, Absaroka, and Hogsback thrusts to the east (DeCelles, 1994). Uplift in the

anticlinorium was accompanied by deposition of foredeep strata east of the propagating wedge front and local deposition of wedge-top conglomeratic strata with angular unconformities. Uplift along the primary basement step, along with imbrication of basement slices, helped maintain wedge taper and drive slip on the eastern thrust system (DeCelles and Mitra, 1995). Basement in the anticlinorium experienced ~10–15% early internal shortening accommodated by shear zones, which were fanned about the anticlinorium during subsequent large-scale imbrication and folding (Yonkee, 1992). Shear zones are characterized by widespread alteration of feldspar to muscovite, plastic deformation of quartz, and mylonitic to phyllonitic foliation. Concentration of deformation into shear and fault zones reflects reaction softening, hydrolytic weakening of quartz, and grain size reduction (Yonkee et al., 2003). The anticlinorium developed above a basal decollement and thus is an example of basement-involved thin-skin structural style, as defined by Pfiffner (2006).

4.4.3 Wyoming salient- eastern thrust system

The eastern thrust system comprises the Crawford, Absaroka, and Hogsback-Darby-Prospect thrusts, which share a basal decollement in Cambrian shale and carry thinner Paleozoic to Mesozoic strata (Figs. 10, 11). Thrusts display curved structural trends related to changes in thrust slip direction and magnitude, and to interaction of frontal thrusts with Laramide arches at the north and south ends of the salient. The Crawford thrust and associated fold train accommodated ~25 km to 30 km of shortening in the north to central parts of the salient, with multiple detachment levels in Mississippian, Triassic, and Jurassic strata (Banarjee and Mitra, 2005). In the south part of the salient, the Crawford thrust ramps down westward and merges with the Ogden thrust system in the Wasatch anticlinorium (Yonkee, 1992). Growth structures in the synorogenic Weber Canyon and Echo Canyon conglomerates record major slip on the Crawford thrust from ~90-80 Ma (DeCelles, 1994). The Absaroka thrust and related imbricates accommodated ~40 km of shortening in the

898 north part of the salient (Woodward, 1986). In the north-central part, >40 km of shortening is
899 accommodated on the main Absaroka thrust along with imbricate faults and folds that vary with
900 stratigraphic level (Chester, 2003). The main thrust branches southward into the Absaroka,
901 Commissary Ridge, and imbricate thrusts that accommodated ~35 km of shortening in the south-
902 central part of the salient. Farther south, ~20 km of shortening was accommodated by the Mt.
903 Raymond-Crandall Canyon thrusts, which are interpreted to link with the Charleston thrust south
904 of the Uinta-Cottonwood arch (Bradley and Bruhn, 1988; Paulsen and Marshak, 1999). An early
905 phase of slip on the Absaroka thrust occurred during deposition of the late Santonian to early
906 Campanian (~80 Ma) Little Muddy Creek Conglomerate. This conglomerate was subsequently
907 tilted during a later phase of thrust slip and deposition of the Maastrichtian (~70-65 Ma) Hams
908 Fork Conglomerate, which contains basement clasts eroded during renewed uplift of the Wasatch
909 anticlinorium (DeCelles, 1994). The Darby-Prospect and subsurface Granite Creek thrusts
910 accommodated ~30 km of shortening in the north part of the salient. Out-of-sequence imbricate
911 thrusts developed here simultaneous with uplift of the Laramide Gros Ventre arch (Fig. 11D;
912 Hunter, 1988). The Paleocene Hoback Formation lies with angular unconformity over the Granite
913 Creek thrust, bracketing early slip (Dorr et al., 1977). The Hoback Formation is in turn offset by
914 the Prospect thrust, which is locally overlain with angular uniformity by the early Eocene Lookout
915 Mountain Conglomerate deposited during final stages of thrusting. The Darby-Prospect thrusts
916 merge southward into the Hogsback thrust across a transfer zone above the basement-cored Moxa
917 arch (Fig. 10; Kraig et al., 1988). In the south-central part of the salient, ~25 km of slip is
918 concentrated along the Hogsback thrust. In the south part of the salient, the Hogsback thrust is
919 covered by younger strata and likely intersects the North Flank thrust of the Uinta arch in the
920 subsurface (Royse, 1993).

Systematic suites of mesoscopic to microscopic structures, including spaced cleavage, fracture and vein sets, minor folds, and minor faults, accommodated internal strain in the eastern thrust system (Crosby, 1969; Mitra and Yonkee, 1985; Craddock et al., 1988; Mitra, 1994; Weil and Yonkee, 2009; Yonkee and Weil, 2010). Styles of internal deformation are described below for two rock types, limestone of the Jurassic Twin Creek Formation and red beds of the Triassic Ankareh Formation.

Limestone in the Twin Creek Formation commonly displays cleavage defined by close-spaced partings at high angles to bedding (Fig. 18A), veins (Fig. 18B), minor faults, and deformed fossils in bioclastic layers. Microscopically, cleavage is represented by seams enriched in clay and quartz, which formed mostly by dissolution of calcite and bound less deformed microlithons (Fig. 18D). Calcite was partly re-precipitated in veins, but limestone underwent ~5–20% volume loss (Mitra and Yonkee, 1985). The Ankareh Formation locally displays cleavage defined by anastomosing partings at high angles to bedding, which is best developed in calcareous mudstone (Fig. 18C). Microscopically, cleavage is represented by discontinuous seams enriched in clay that formed mostly by dissolution of calcite and quartz. Other red beds lack cleavage, but display fractures at high-angles to bedding that formed along weak LPS fabrics during unloading and folding. Red beds also contain kinked and rotated micas that formed during LPS and contribute to anisotropy of magnetic susceptibility (AMS), which provides a proxy for strain directions in weakly deformed rocks (Weil and Yonkee, 2009).

Cleavage is commonly at high angles to bedding and perpendicular to the shortening direction defined by shapes of reduction spots in red beds and deformed crinoids in limestone, and formed during early LPS (Fig. 19). Local cleavage refraction records limited bed-parallel shear. Veins include a dominant set subperpendicular to structural trend (cross-strike set) that

accommodated tangential extension. Cross-strike veins are locally grouped into *en echelon* arrays associated with wrench shear (Fig. 18B). Other veins are parallel to bedding and accommodated minor extension down the dip of cleavage. Some outcrops display subsets of tectonic stylolites, high-angle fractures, and cross-strike veins at acute (10 to 30°) angles to each other, recording local block rotations and/or stress changes during progressive deformation (Yonkee and Weil, 2010). Minor faults in both formations include: (i) early wedge faults at acute angles to bedding with reverse slip that accommodated LPS; and (ii) early tear faults at high angles to bedding with strike-slip that accommodated wrench shear, small-scale block rotation, and minor tangential extension. Cross-cutting and geometric relationships of later fault sets along fold limbs record protracted internal deformation. Early wedge and tear faults have slip lineations that systematically change plunge around fold limbs, whereas later, gently dipping reverse faults cut steep fold limbs and accommodated fold tightening. Later strike-slip faults have gently plunging slip lineations and accommodated differential shortening and development of map-view curvature. Bed-parallel faults related to flexural slip, shear veins along some cleavage seams, breccia zones, and locally developed, weak second cleavage also accommodated additional internal deformation during large-scale folding (Fig. 19).

Major fault zones in the eastern thrust system typically include a damage zone along the base of the hanging wall comprised of variably brecciated strata cut by multiple vein and minor fault sets, and a poorly exposed fault core comprised of cataclasite cut by discrete slip surfaces (Fig. 18E). Cataclasite contains comminuted fragments that are locally re-cemented and re-brecciated, clay-rich seams, and multiple sets of microveins (Fig. 18F). Microstructures record concentrated cataclasis, grain size reduction, alteration, and repeated episodes of fluid influx within relatively weak fault cores that experienced episodically high fluid pressures.

Timing of internal deformation is constrained by structural relations and characteristics of clasts within synorogenic conglomerates. Within the Crawford sheet, cleavage is strongly fanned about large-scale folds, remaining subperpendicular to bedding around fold limbs (Fig. 20A), indicating cleavage formed before large-scale folding and slip on the Crawford thrust. Cleavage becomes more intense westward and is at acute angles to bedding and the Meade thrust within a footwall syncline (Fig. 20B), indicating cleavage initially developed during Meade footwall deformation. Late Cretaceous strata, deposited in local wedge-top basins synchronous with slip on the Absaroka thrust to the east, lie with angular unconformity over folded rocks of the Crawford sheet and display only limited internal deformation (Coogan, 1992). Thus, cleavage in the Crawford sheet initially developed during Early Cretaceous LPS in front of the Willard-Meade thrusts, followed by mid Cretaceous large-scale folding and slip on the Crawford thrust, and then by mostly passive transport during the later Cretaceous as the thrust front propagated eastward. Emplacement of the Willard-Meade sheet and deposition of thick foredeep sediments increased burial depths, temperatures, fluid flow, and stress, favoring cleavage development within the future Crawford sheet (Mitra and Yonkee, 1985). Cleavage intensity decreased eastward into the future Absaroka thrust sheet where LPS was partly accommodated by wedge faults, followed by major thrust slip, large-scale folding, and development of late reverse faults and local reactivation of cleavage along steep limbs during fold tightening (Fig. 20C). Relations record an overall transition from more homogeneous LPS and cleavage development early in the deformation history of the Crawford and Absaroka sheets, to more discrete deformation during large-scale folding and faulting, partly related to development of weak fault zones (Fig. 19). Within the Hogsback-Darby-Prospect sheets, cleavage intensity is weak in central frontal parts, but intensity increases westward where cleavage is at acute angles to bedding along footwall folds and imbricates of the Absaroka

thrust. Thus, cleavage in the frontal sheets developed mostly during later Cretaceous deformation in front of the Absaroka thrust. Cleavage intensity also increases toward the ends of the salient, likely related to buttressing by foreland arches that enhanced differential stress. Synorogenic conglomerates derived from erosion of the Crawford, Absaroka, and Hogsback-Darby-Prospect sheets contain limestone clasts with pre-existing tectonic stylolites, cleavage seams, and veins that do not continue into the conglomerate matrix, confirming that cleavage formed during early LPS and prior to erosion of each thrust sheet, and that the LPS deformation front propagated progressively east of the growing thrust wedge (Mitra et al., 1984).

LPS directions, estimated from orientations of cleavage, tectonic stylolites, and high-angle fracture sets in the Twin Creek and Ankareh formations (restored for bedding untilted to horizontal), are at high angles to structural trend and define a radial pattern around the salient (Fig. 21A; Mitra, 1994; Yonkee and Weil, 2010). Principal extension directions estimated from cross-strike veins and fractures in both the Twin Creek and Ankareh formations are subparallel to structural trend and define a tangential pattern (Fig. 21B). In detail, LPS and extension directions display minor local variations, reflecting complications from small-scale block rotations and stress-strain refraction. Patterns are also more complex along oblique thrust ramps and near the salient ends where frontal thrusts interacted with Laramide arches. Similar patterns in the Twin Creek and Ankareh formations, combined with additional minor fault and calcite twin strain data from Cretaceous to Paleozoic strata (Crosby, 1969; Craddock et al., 1988; Mitra, 1994; Fig. 21A), record consistent kinematic patterns through the stratigraphic section, although deformation mechanisms vary with lithology.

Strain, estimated from deformed crinoids in bioclastic limestone of the Twin Creek Formation and from reduction spots and AMS fabrics in the Ankareh Formation, also displays

1013 consistent regional patterns with radial shortening directions and tangential extension directions
1014 (Fig. 21C; Mitra, 1994; Weil and Yonkee, 2010). Axial ratios of bed-parallel strain ellipses in the
1015 Twin Creek Formation range from <1.05 in the central Hogsback sheet, to $1.15\text{--}1.3$ in the
1016 Crawford sheet, and increase toward the salient ends. Cleavage intensity and shortening estimated
1017 from cleavage seam widths also increase westward and toward the salient ends (Mitra and Yonkee,
1018 1985). Axial ratios of bed-parallel strain ellipses in the Ankareh Formation range from <1.1 in the
1019 central Hogsback sheet, to $1.2\text{--}1.5$ in the Crawford sheet. Estimated strain ratios are slightly higher
1020 for red beds compared to limestone, which may reflect less penetrative strain in bioclastic
1021 limestone that also shortened by wedge faulting. Differences in ratios between red beds and
1022 limestone, however, are small and closely bracket strain magnitude, with $10\text{--}30\%$ LPS in the
1023 Crawford sheet, $5\text{--}10\%$ LPS in the central Absaroka sheet, and $<5\%$ to 15% LPS in the central
1024 Hogsback to northern Prospect sheet (Yonkee and Weil, 2010). Mitra (1994) reported broadly
1025 similar average LPS values of $\sim 25\%$, 15% , and 10% respectively for parts of the Crawford,
1026 Absaroka, and Hogsback-Prospect sheets. Importantly, LPS adds to total shortening across a fold-
1027 thrust belt. Restoration of section B (Fig. 11B) gives an estimated: 36 km of large-scale shortening
1028 for the Hogsback sheet mostly from slip on the main thrust, with an additional 4 km of internal
1029 shortening from 5% LPS over a bed length of 80 km ; 42 km of large-scale shortening for the
1030 Absaroka sheet mostly from slip on the main thrust, with an additional 9 km of internal shortening
1031 from 10% LPS over a bed length of 90 km ; and an estimated 33 km of large-scale shortening for
1032 the Crawford sheet from slip on the main thrust, imbricates, and major folds, with an additional 14
1033 km of internal shortening from 20% LPS over a bed length of 70 km . Thus, the Sevier belt
1034 experienced $\sim 200\text{ km}$ of total upper crustal shortening (including $\sim 60\text{ km}$ of slip on the Willard-

Paris-Meade thrusts) across the central Wyoming salient, which was balanced by mid- to lower-crustal thickening in the hinterland.

Curvature of thrust systems and radial patterns of LPS directions in the Wyoming salient, and in other fold-thrust belts, have been variously interpreted to represent: (1) primary curvature of thrusts and folds along with uniform thrust slip (Royse, 1993); (2) primary curvature along with radial thrust slip (Crosby, 1969); (3) progressive curvature in which structural trends and LPS directions begin with a component of primary curvature and then undergo varying secondary rotation during curved thrust slip (Yonkee and Weil, 2010); and (4) secondary rotation of initial linear structural trends and prior LPS directions during subsequent deformation (Fig. 22A; see review by Weil and Sussman, 2004). These varying interpretations reflect different assumptions on initial orientations of LPS and structural trends, which are not directly known. Paleomagnetic analysis provides the most robust way to quantify secondary rotations and test these models.

Red beds of the Ankareh Formation carry two remanent magnetic components: a near-primary (early diagenetic) Triassic magnetization carried by hematite (Tr component); and a Cretaceous syntectonic chemical remagnetization carried partly by magnetite (K component; Weil et al., 2010). *In situ* site mean vectors for the Tr and K components show a high degree of scatter from expected Triassic and Cretaceous reference pole directions, indicating significant tilt and rotation subsequent to acquisition of magnetization. Restoration of tilt around folds for individual site means gives well clustered shallow and moderate inclinations respectively for 100% unfolding of the pre-thrusting Tr component and 80–90% unfolding of the syntectonic K component, consistent with Triassic and Cretaceous paleo-latitudes of North America (Weil et al., 2010). Restored declinations for both components are systematically rotated counterclockwise in the north part of the salient and clockwise in the south part (Fig. 21D). Declinations, however, are not

rotated within gentle limbs of adjacent Laramide foreland arches beyond the salient ends, indicating systematic secondary rotations were confined to the Sevier thrust wedge. Additional localized rotations along the leading, north part of the Prospect sheet record shearing of the thrust wedge where it was overridden and buttressed by the Gros Ventre foreland uplift (Grubbs and Vander Voo, 1976).

Statistical analysis of correlations between Tr and K declinations and structural trend, relative to reference declination and trend, yields strike-test slopes of ~ 0.75 , indicating that, on average, $\sim 75\%$ of present-day salient curvature resulted from secondary vertical-axis rotation and $\sim 25\%$ was related to primary curvature (Fig. 22B). Analysis of Tr and K declinations within individual thrust sheets shows only minor differences, with an average of $\sim 80\%$ secondary rotation in the Crawford sheet to $\sim 60\%$ rotation in the frontal Hogsback-Darby-Prospect sheet, indicating rotations were concentrated near the leading edge of the propagating thrust wedge, with only limited additional rotation during overall eastward, passive transport of older thrust sheets in the wedge interior (Weil et al., 2010). Analysis of mesoscopic structural and strain data indicates early LPS directions were about orthogonal to structural trend, with strike-test slopes of ~ 0.9 (Fig. 22B). In summary, structural trends started with slight initial curvature about a regional N-S trend and early LPS directions were slightly dispersed about a regional W-E direction, followed by systematic secondary rotation during curved thrust slip that produced progressive fold-thrust curvature and radial LPS patterns (Fig. 22C), broadly consistent with model 3 above.

4.5 Foreland sedimentary basin

A broad foreland basin developed along and east of the Sevier belt in response to flexural loading from thrust sheets, redistribution of load by erosion and sediment deposition, dynamic subsidence related to subduction, and global sea level changes (Fig. 23A; Williams and Stelck,

1081 1975; Kauffman and Caldwell, 1993; Gardner, 1995; DeCelles and Giles, 1996; Catuneanu et al.,
1082 1997; Liu and Nummedal, 2004; DeCelles, 2004; Liu et al., 2008). The basin was variably
1083 inundated by marine waters of the Western Interior Seaway, which during mid Cretaceous
1084 transgression connected the Arctic Ocean and Gulf of Mexico. The basin included: a wedge-top
1085 zone with local deposition of coarse-grained deposits separated by angular unconformities; a
1086 foredeep zone with deposition of thick, mostly fluvial strata; a low-amplitude forebulge with
1087 episodic deposition of thinner strata and local development of paleosols; and a broad backbulge
1088 zone with deposition of fine-grained, mostly marine strata (DeCelles and Giles, 1996). As the
1089 Sevier belt propagated eastward, these depozones also migrated eastward, resulting in deposition
1090 of overall upward-coarsening strata that were progressively incorporated into the growing thrust
1091 wedge, and in changing deposition rates that were also partly related to dynamic subsidence (Fig.
1092 23B; DeCelles and Currie, 1996; DeCelles, 2004; Liu et al., 2005). The sedimentary basin became
1093 disrupted during later Cretaceous to Eocene uplift of Laramide arches.

1094   The classic Sevier foreland basin was presaged by Middle Jurassic subsidence and
1095 deposition of marine limestone and evaporites of the Twin Creek and Sundance formations in the
1096 Utah-Idaho trough (Imlay, 1967; Heller et al., 1986), Late Jurassic deposition of mostly fluvial
1097 strata of the Morrison Formation in a broad basin (DeCelles and Burden, 1992), and earliest
1098 Cretaceous development of a regional unconformity. Twin Creek-Sundance and Morrison
1099 deposition overlapped with Jurassic terrane accretion and closing of a marginal ocean basin,
1100 initiation of the Franciscan subduction zone, and shortening along the developing Sierra Nevada
1101 arc and Luning-Fencemaker fold-thrust belt (Fig. 7B; Dickinson, 2008). Development of the Utah-
1102 Idaho trough and Morrison basins, followed by regional erosion, is interpreted to partly reflect a
1103 wave of dynamic subsidence and subsequent rebound associated with sinking of older slabs

1104 following terrane accretion and during westward drift of the North American plate. Alternatively,
1105 Bjerrum and Dorsey (1995) proposed that the Utah-Idaho trough developed as a flexural foredeep
1106 during hinterland thickening, and DeCelles and Currie (1996) proposed that the Morrison basin
1107 formed as a backbulge depozone during early hinterland shortening, followed by eastward
1108 propagation of a forebulge marked by earliest Cretaceous regional erosion. A similar pattern
1109 occurs in NW Montana to Alberta with deposition of Middle Jurassic marine strata and Late
1110 Jurassic fluvial strata, followed by development of an earliest Cretaceous unconformity, which
1111 Fuentes et al. (2009) interpreted to record deposition in a backbulge depozone during initial
1112 shortening in the Sevier belt, followed by passage of a forebulge.

1113   Following earliest Cretaceous erosion, a regionally extensive conglomeratic interval was
1114 deposited across the foreland from Alberta to Utah during Aptian time, with DZ patterns recording
1115 western sources from the magmatic arc and developing Sevier belt (Leier and Gehrels, 2011).
1116 During Albian time, up to 3 km of synorogenic strata were deposited in a foredeep near the leading
1117 edge of the western thrust system, including the Gannett Group along the Willard-Meade thrust in
1118 the Wyoming salient (Fig. 24A, B; DeCelles et al., 1993). Other aerially extensive, thick
1119 (dominant) thrust sheets of the western system were also active during this time, including the
1120 Sheeprock, Canyon Range, and Wah Wah thrusts in central to southern Utah (DeCelles, 2004).
1121 Initially, clastic strata were deposited mostly in fluvial environments, with locally developed lakes
1122 possibly controlled by a forebulge. By later Albian time, the Interior Seaway transgressed into
1123 eastern Wyoming, likely related to dynamic subsidence as the subducting slab shallowed and
1124 extended farther eastward. The Newark Canyon Formation was deposited in a wedge top basin
1125 during hinterland shortening along the Central Nevada thrust belt, but elevation differences
1126 between the hinterland and foreland were small during Aptian-early Albian time (Bonde, 2014).

1127   During Cenomanian to Turonian time, up to 2 km of additional synorogenic strata,
1128 including the Bear River to Frontier formations, were deposited during continued slip on the
1129 western thrust system (Fig. 24C, D). To the south, the wedge propagated eastward, with slip on the
1130 Pavant and Keystone thrusts in central Utah to Nevada, and early uplift of the Sevier culmination
1131 in west-central Utah (DeCelles, 2004). The Interior Seaway expanded during sea level rise, with a
1132 mix of proximal alluvial fan to coastal environments near the front of the thrust wedge. By
1133 Turonian time, subsidence increased in NE Arizona, interpreted to reflect enhanced asthenosphere
1134 flow during initiation of a flat-slab segment. DZ patterns in Aptian to Turonian strata record
1135 progressive erosional exhumation of the western thrust system (Lawton et al., 2010).

1136   During Coniacian-Santonian time, the foredeep migrated eastward in response to early
1137 propagation of the eastern thrust system, including deposition of Weber Canyon-Echo Canyon
1138 Conglomerate during slip on the Crawford thrust (Fig. 12; DeCelles, 1994), along with slip on the
1139 Paxton and Charleston thrusts in central Utah (Fig. 24E). The Sevier and Wasatch basement-cored
1140 culminations grew in west-central and northern Utah as the basal decollement propagated upward
1141 along the transition from the western to eastern thrust systems (Yonkee, 1992; DeCelles, and
1142 Coogan, 2006). Thick marine strata were deposited over a broad region to the east, interpreted to
1143 reflect increased dynamic subsidence in front of the NE-underthrusting flat-slab segment.

1144   During early Campanian time, the foredeep continued to migrate eastward in response to
1145 propagation of the eastern thrust system, with deposition of Little Muddy Creek Conglomerate
1146 during early slip on the Absaroka thrust (Fig. 12; DeCelles, 1994), along with early slip on the
1147 Lewis-El Dorado thrusts in Montana (Fig. 24F). Regional subsidence was enhanced in NE Utah to
1148 southern Wyoming during continued NE-underthrusting of the flat slab. Volcanism shut down in
1149 the Sierra Nevada arc, and shifted eastward into the Colorado Mineral belt and western Montana .

1150 Middle- to lower-crustal thickening in the hinterland, which balanced upper crustal shortening in
1151 the Sevier belt, likely led to uplift of a broad hinterland plateau (DeCelles, 2004).

1152 By late Campanian time, regional dynamic subsidence was widespread across eastern
1153 Wyoming to Colorado as the flat slab continued to the NE (Fig. 24G). The Wyoming salient
1154 appears to have experienced a lull in thrust slip with removal of topographic load by erosion,
1155 flexural rebound, and development of a regional unconformity across western Wyoming. Thrusting
1156 and flexural subsidence, however, increased to the north in the Alberta salient, and slip on the
1157 Lewis-El Dorado thrust and volcanism continued in western Montana (Lageson and Schmitt,
1158 1994). Early uplift of the San Rafael arch in the Colorado Plateau (Lawton, 1983) and local
1159 sedimentary thickness variations may record initiation of Laramide foreland deformation.

1160 During Maastrichtian time, late slip on the Absaroka thrust, renewed uplift of the Wasatch
1161 anticlinorium, and deposition of the Hams Fork Conglomerate occurred in the Wyoming salient
1162 (Fig. 12), while slip on the Gunnison thrust and continued growth of the Sevier culmination
1163 occurred in central Utah (Fig. 24H). The foreland became partly partitioned into smaller basins
1164 during early growth of some Laramide arches. Broad uplift and deposition of thin conglomerate
1165 sheets in the Four Corners area may record regional uplift and tilting during NE movement of the
1166 buoyant flat slab beneath this area (Heller et al., 2013). A flare-up of igneous activity in the
1167 Mojave sector may record asthenosphere influx along the trailing, sagging margin of the flat slab
1168 (Miller et al., 1996; Saleeby, 2003). Strata of the Sheep Pass Formation were deposited within
1169 internally drained wedge-top basins associated with normal faulting during local upper crustal
1170 extension across the hinterland plateau (Druschke et al., 2011), which likely had elevations >2-3
1171 km (Snell et al., 2014). By Paleocene to Early Eocene time the foreland was further
1172 compartmentalized during continued growth of Laramide arches (Beck et al., 1988; Dickinson et

al., 1988), simultaneous with emplacement of frontal thrusts of the Sevier belt (Bradley and Bruhn, 1988; Yonkee and Weil, 2010). A switch to extension and deposition of variably volcaniclastic strata in local half grabens occurred during the Middle Eocene to Oligocene (see section 4.6.1).

Characteristics of synorogenic strata and stable isotope values of carbonates in non-marine fossils and paleosols from Cretaceous to Paleogene strata provide additional insights into paleoclimatic, topographic, and sediment dispersal patterns from the Sevier belt to the foreland. $\delta^{18}\text{O}$ values for aragonite shells recovered from trunk river deposits are distinctly more negative than values for shells recovered from local stream and pond deposits, indicating major drainages were connected to higher elevation (~2-3 km), interior parts of the Sevier belt during the later Cretaceous (Fricke et al., 2010; Foreman et al., 2011), similar to paleoelevation estimates for the hinterland at that time (Snell et al., 2014). Paleofloral studies and interpretations of depositional environments indicate areas east of the Sevier belt comprised moist, lowlands adjacent to the Western Interior Seaway, whereas areas along the wedge front comprised seasonally wet/dry fans that drained western highlands (Wolfe and Upchurch, 1987). Global circulation models of the North American Cordillera during middle to later Cretaceous time also predict a monsoonal climate pattern related to paleogeography of the Western Interior Seaway and a high elevation magmatic arc and hinterland, with relatively warm, wet summers marked by overall SE to NW airflow (Poulsen et al., 2007; Fricke et al., 2010). Upslope flow and rain out of precipitation is interpreted to have concentrated exhumation along the Sevier wedge front, consistent with detrital thermochronologic data that indicate focused high exhumation rates there (Painter et al., 2014), whereas drier climates developed in the hinterland plateau, which experienced lower exhumation rates within internally drained basins during later Cretaceous time (Druschke et al., 2011).

4.6.1 Laramide foreland belt- large-scale structural style and timing

The Laramide belt is characterized by thick-skin, basement-cored arches separated by broad basins, which developed within cratonic lithosphere >1000 km inboard from the plate margin, spanning a region from southern Montana to New Mexico (Figs. 1, 25; Berg, 1962; Erslev, 1993). Arches are bound by overall moderate-dipping (typically ~30-45°) reverse faults that have net slips of ~10-30 km and continue upward into folds with steep, sheared limbs in the sedimentary cover (Brown, 1988; Stone, 1993). Major reverse faults are imaged geophysically to mid-crustal (~20 to 30 km) depths (Smithson et al., 1979), but evidence for offset of the MOHO is lacking (Yeck et al., 2014) and faults may flatten in the lower crust (Erslev, 1993). The Laramide belt displays an overall NW-SE structural grain, but individual arches and faults have a range of trends, partly reflecting influence of basement fabrics (Stone, 2002; Erslev and Koenig, 2009; Neely and Erslev, 2009). Cover folds range in trend from N-S to E-W and form both left-stepping and right-stepping *en echelon* zones. Shortening from reverse faults and folds is ~50 km (or ~10%) across the Wyoming foreland (Stone, 1993). The Colorado Plateau also displays thick-skin deformation, but shortening is less and fold trends are more variable, indicating the area acted as a relatively rigid block, possibly related to a strong, mafic lower crust (Shen et al., 2013), with local deformation concentrated along basement weaknesses (Huntoon, 1993). The plateau underwent minor regional rotation (Wawryzniec et al., 2002) and experienced stresses partly related to the Sevier belt along its western margin (Bump, 2004).

General timing of Laramide deformation is constrained by thermochronologic and sedimentologic data (Fig. 26). Apatite fission track (FT) and U/Th-He data indicate that Laramide deformation spanned ~70–50 Ma along the tectonic transect (Cerveny and Steidtmann, 1993; Omar et al., 1994; Crowley et al., 2002; Peyton et al., 2012), temporally overlapping with late Sevier deformation. Maastrichtian to Early Eocene synorogenic strata display regional thickness

variations, preserve progressive unconformities near arch-bounding faults, and contain proximal conglomerates with clast compositions related to arch unroofing, which also indicate major Laramide deformation spanned ~70–50 Ma (Keefer, 1970; Beck et al., 1988; Dickinson et al., 1988; DeCelles et al., 1991; Steidtmann and Middleton, 1991; Hoy and Ridgway, 1997). However, in detail, timing of arch uplift and basin subsidence varied regionally and questions remain on varying models for directions of deformation propagation and changes in shortening directions over time (Chapin and Cather, 1983; Gries, 1983; Bergh and Snoke, 1992; Wawrzyniec et al., 2002; Heller et al., 2013; Fan and Carrapa, 2014). Along the Wind River arch in Wyoming, Maastrichtian strata thin across the NE flank and thicken along the SW margin into the Green River Basin, recording early uplift, followed by deposition of thick Paleocene to Early Eocene strata in the basin during continued arch growth, followed by onlap of Middle Eocene strata that bracket an end to reverse faulting (Figs. 25B, 26B; Steidtmann and Middleton, 1991). Along the Owl Creek and Big Horn arches to the NE, Maastrichtian strata maintain consistent thicknesses, whereas Paleocene to Early Eocene strata thicken markedly in adjacent basins, recording younger onset of major uplift (Figs. 25B, 26B; Keefer, 1970; Hoy and Ridgway, 1997). A switch to local extension and collapse of Laramide arches was underway by the Middle Eocene, contemporaneous with increased igneous activity in the Absaroka volcanic complex and isolation of lake systems (Smith et al., 2008), with continued local extension into the Oligocene (Hall and Chase, 1989).

Laramide deformation roughly coincided spatially and temporally with Late Cretaceous to Paleogene development of a flat-slab segment, interpreted to have developed during subduction of the conjugate to the Shatsky Rise (Saleeby, 2003), a broad oceanic plateau that formed along the ancestral Pacific ridge (Fig. 6D; Mahoney et al., 2005). Evidence of a flat-slab segment includes development of a magmatic gap along the Sierra Nevada and Mojave areas (Coney and Reynolds,

1242 1977; Dickinson and Snyder, 1978; Miller et al., 1992), onset of magmatic activity along the
 1243 Colorado Mineral belt (Chapin, 2012), and forearc uplift and underplating of the Pelona-Orocopia-
 1244 Rand schists (Fig. 24F to H; Dumitru et al., 1991; Smith, 2000; Saleeby, 2003). Further evidence
 1245 comes from xenoliths brought up in Cenozoic volcanic centers in the Four Corners area, including
 1246 eclogite and garnetite xenoliths that are interpreted to represent subducted oceanic crust (Usui et
 1247 al., 2003) and mantle lithosphere metasomatized by fluids derived from the subducted plate (Smith
 1248 and Griffin, 2005). These xenoliths contain ~80 to 60 Ma zircon grains that likely record timing of
 1249 alteration during flat-slab subduction. Mantle xenoliths brought up in Eocene volcanic centers in
 1250 central Montana display disturbed isotopic and trace element patterns interpreted to reflect
 1251 metasomatism from interaction with slab-related melts where the slab steepened (Carlson et al.,
 1252 2004). Additionally, fast seismic anomalies in the mantle beneath the central to eastern U.S.,
 1253 including a broad area now at depths >1000 km, are interpreted to represent remnants of the
 1254 subducted Farallon plate (Fig. 4A; van der Meer et al., 2010; Liu et al., 2010; Sigloch, 2011).
 1255 Inverse mantle convection modeling restores this broad area as a flat-slab segment that entered the
 1256 subduction zone along the Mojave area at ~90 Ma, followed by NE underthrusting beneath
 1257 Arizona to Wyoming during the Late Cretaceous, consistent with patterns of later Cretaceous
 1258 subsidence (Liu et al., 2010). Although multiple lines of evidence indicate former presence of a
 1259 flat slab beneath the Laramide foreland belt, dynamic processes that produced thick-skin crustal
 1260 deformation are debated, including increased coupling from end loading along the plate margin,
 1261 increased basal traction, and enhanced asthenosphere flow near a cratonic keel (Bird, 1984, 1988;
 1262 Livaccari, 1991; Tikoff and Maxson, 2001; English et al., 2003; Jones et al., 2011).
 1263   Basement in the Laramide foreland includes the Archean Wyoming and Paleoproterozoic
 1264 Yavapai-Mazatzal provinces, which were sutured together along the E- to NE-trending Cheyenne

1265 belt (Fig. 2; Karlstrom and Houston, 1984). Basement across much of the foreland displays a
1266 broadly E- to NE-trending regional structural grain recorded by arcuate belts of lower and higher
1267 amplitude magnetic anomalies (Sims et al., 2001), but local fabrics are more variable and include
1268 multiple metamorphic foliations, shear zones, dikes, and Proterozoic normal faults, which partly
1269 influenced localization of deformation within differently oriented Laramide structures (Mitra and
1270 Frost, 1980; Marshak et al., 2000; Stone, 1995, 2002; Neely and Erslev, 2009). Fast seismic
1271 velocities in the lower crust (depths >30 km) beneath the Wyoming province and Colorado Plateau
1272 likely indicate presence of more mafic rocks (Shen et al., 2012). Cratonic basement in the foreland
1273 is overlain by a thin sedimentary cover (Fig. 3).

1274 *4.6.2 Laramide foreland- regional patterns of internal deformation*

1275   Systematic suites of mesoscopic to microscopic structures, including fracture sets, minor
1276 faults, and minor folds, accommodated limited internal strain within the Laramide belt (Yonkee
1277 and Mitra, 1993; Erslev and Koenig, 2009; Neely and Erslev, 2009; Beaudoin et al., 2012; Weil
1278 and Yonkee, 2012; Weil et al., 2014). Styles are described below for red beds and a thin limestone
1279 unit in the Triassic Chugwater Group, and for limestone in the Jurassic Sundance Formation,
1280 which can be compared to styles previously described for correlative strata in the Sevier belt.
1281 Cleavage is absent in the Laramide belt, although rare tectonic stylolites are locally developed in
1282 limestone beds. Fractures include: (1) a set at high angles to bedding and subparallel to structural
1283 trend interpreted to reflect partings along weak LPS fabrics and fracturing during folding; (2) a set
1284 at high angles to bedding and subperpendicular to structural trend that includes thin veins and
1285 accommodated limited tangential (strike-parallel) extension; (3) obliquely striking fractures that
1286 developed during 3-D fold growth; and (4) pre-Laramide fractures (Bellahsen et al., 2006;
1287 Beaudoin et al., 2012). Minor faults with calcite-fiber slip lineations, best developed in limestone

beds, define two main groups: wedge faults at acute angles (20 to 30°) to bedding; and tear faults at high angles to bedding and structural trend (Fig. 27A, B). Minor folds are associated with some wedge faults. Microscopically, limestone displays calcite twinning and widely spaced microveins. Red beds display fracture sets, but otherwise typically appear little deformed in outcrop (Fig. 27C). Microscopically, red beds display micro-kinked and rotated paramagnetic phyllosilicate grains (Fig. 27D), which give rise to subtle, but measurable AMS $K_{\max}$ lineations that are parallel to intersection of bedding and weak LPS fabrics (Weil et al., 2014).

Wedge faults, minor folds, and rare tectonic stylolites in limestone accommodated limited (<5%), early LPS and vertical thickening (Fig. 28). Early tear faults and veins accommodated limited tangential extension and wrench shear. Forelimbs of folds were modified during arch growth with development of gently dipping reverse faults that accommodated limb steepening and shear, strike-slip faults associated with differential limb tilt and shortening, and oblique-slip faults associated with map-view curvature and oblique ramps. Faults that developed during later deformation are wider spaced and have larger displacements compared to early faults, recording a change from more distributed to concentrated deformation along weaker fault zones.

Major fault zones in the Laramide commonly include a damage zone comprised of fractured basement rocks in the hanging wall, a poorly exposed fault core of cataclasite, and overturned, imbricated Paleozoic to Mesozoic strata in the footwall (Fig. 27E). Cataclasite in fault cores displays grain size reduction with clasts of quartz and feldspar sitting in a fine-grained matrix, which locally has crude foliation defined by alignment of clasts and oxide-richer zones, suggestive of macroscopic ductile flow (Fig. 27F; Mitra et al., 1988). Although clay alteration and veins are limited, geochemical differences between basement wall rock and fault rock suggest fluid-rock interaction and volume loss in some fault cores (Goddard and Evans, 1995).

1311   Timing of internal deformation is constrained by structural relations. Most wedge faults
1312 maintain acute angles to tilted bedding around large-scale folds, and thus formed during early LPS,
1313 mostly prior to major arch growth (Fig. 29). Some fold limbs, however, also display cross-cutting
1314 wedge faults with slightly different dips, recording continued internal shortening during early
1315 stages of fold limb tilting. Later, gently dipping reverse faults cut steep fold limbs. Western parts
1316 of the Laramide belt also show local evidence for earlier ~E-W LPS by calcite twinning,
1317 interpreted to record early far-field stresses related to the Sevier belt, followed by ~SW-NE LPS
1318 by wedge faulting typical of the Laramide belt (Beaudoin et al., 2012).

1319   LPS directions for structurally simple sites (excluding steep fold limbs), estimated from
1320 paleo-stress/strain inversion of minor fault data and AMS fabrics in Triassic limestone and red
1321 beds, have an average 240° direction, but display deflections related to changes in structural trend
1322 of major basement arches (Fig. 30A). LPS directions are mostly 230–250° for sites located along
1323 NW- to NNW-trending arches, 200–230° along the more W-trending Sweetwater arch in central
1324 Wyoming, and 250–270° along the N-trending Laramie arch in southeastern Wyoming (Weil and
1325 Yonkee, 2012). Stress magnitude ratios [$\phi = (\sigma_2 - \sigma_3) / (\sigma_1 - \sigma_3)$] are < 0.3 ($\sigma_2 \approx \sigma_3$) for most sites,
1326 consistent with development of both wedge faults (σ_2 horizontal) and tear faults (σ_2 vertical).
1327 Stress ratios are locally higher in areas where multiple fold trends are developed. Deviatoric stress
1328 likely decreased toward the NE margin of the Laramide belt in the Black Hills area where AMS
1329 lineations are very weak to absent and wedge faults are rare. Minor fault data for the Jurassic
1330 Sundance Formation record similar paleo-stress patterns. Erslev and Koenig (2009) estimated a
1331 similar regional average σ_1 trend of 247° across the Laramide belt. More complex stress histories
1332 occur where folds with multiple trends are developed, and near basement faults oriented oblique to
1333 regional WSW-ENE shortening (Stone, 1969; Varga, 1993). Analyses of calcite twins also indicate

1334 local variations in stress directions (Willis and Groshong, 1993; Craddock and Relle, 2003;
1335 Amrouch et al., 2011), along with a decrease in deviatoric stress magnitude toward the continental
1336 interior (Craddock and van der Pluijm, 1999).

1337 Triassic red beds at most sites in the Laramide belt record a near-primary (early diagenetic)
1338 magnetization carried by hematite (Weil and Yonkee, 2012; Weil et al., 2014). Restoration of tilt
1339 results in well clustered, shallow inclinations, consistent with the Triassic paleolatitude of North
1340 America. Paleomagnetic declinations indicate only limited vertical-axis rotations for structurally
1341 simple sites (Fig. 30B). Sites along steep W- and N-trending forelimbs, near differently oriented
1342 faults, and in areas where multiple arch trends intersect, however, display local vertical-axis
1343 rotations. Compared to the Sevier belt, the Laramide belt shows more spatially variable structural
1344 trends and basement fabrics, local refraction of stress-strain directions, and non-systematic
1345 vertical-axis rotations, which are briefly described below for the southern Big Horn arch.

1346 The southern Big Horn arch displays regional curvature from NW- to NNE-trends,
1347 intersects the W-trending Owl Creek arch, is cut by variably oriented faults, and exposes
1348 Precambrian basement with multiple fabrics, providing a microcosm to investigate relationships
1349 between differently oriented structures and basement weaknesses (Fig. 31). The W-striking, steep-
1350 dipping Tensleep fault has sinistral oblique slip and follows a Precambrian shear zone (Hoppin et
1351 al., 1965). The NNE-striking, steep dipping Big Trail fault zone contains dextral oblique-slip
1352 faults, reverse and normal faults in restraining and releasing bends respectively, and formed
1353 subparallel to NNE-striking Precambrian dikes (Ver Ploeg and Greer, 1997). The NNW-striking
1354 Clear Creek reverse fault that bounds the east flank of the arch has ~10 km of top-to-ENE dip slip,
1355 and propagated through a steep fold limb in Paleozoic to Mesozoic strata (Hoy and Ridgway,
1356 1997). This fault loses slip southward and continues beneath a N-trending zone of right-stepping,

en echelon folds that are cored by W-verging imbricate reverse faults. LPS directions estimated from minor fault and AMS data are at moderate to high angles to structural trend, ranging from overall SW-oriented on the western flank of the arch to W-oriented on the eastern flank, along with local refraction (Fig. 31A; Weil et al., 2014). Along the west end of the Tensleep fault, LPS directions estimated from AMS fabrics, early minor faults and tectonic stylolites, and later minor folds change from WSW to SSW to S oriented, oblique to sub-perpendicular with the main fault, likely recording increasing stress refraction near a weakening fault zone. Paleomagnetic declinations record limited, non-systematic vertical-axis rotations, indicating that wrench shear was insignificant during development of the southern Bighorn arch, despite its N-trend that was at acute angles to regional WSW-oriented shortening. The N-trending zone of *en echelon* folds is suggestive of transpression, but statistically significant rotation is not observed, which may reflect refraction of shortening into high angles with the zone, or partitioning of shear onto discrete strike-slip faults. Paleostress trajectories, fully restored for bed tilt and vertical-axis rotation, display deflections related to basement weaknesses and faults (Fig. 31C), with refraction of σ_1 into high angles with the Tensleep and Clear Creek fault zones, and with the W- and NW-trending Owl Creek and Casper arches. The paleostress field records regional WSW-ENE shortening, along with local refraction related to structural trend, proximity to faults, and basement weaknesses.

The variability in orientations of Laramide arches has inspired multiple kinematic models including (Fig. 32A): (1) temporal changes in shortening directions from early E-W to late N-S (Chapin and Cather, 1983; Gries, 1983); (2) bulk shortening with wrench shear concentrated within differently oriented zones (Sales, 1968; Stone, 1995); and (3) regional SW-NE shortening with local refraction of stress directions along anastomosing arch trends (Brown, 1988; Erslev, 1993). Each of these models predicts different patterns of large-scale fault-fold sets, LPS

directions, and vertical-axis rotations that can be evaluated with strike tests. Model 1 predicts intersection of fault-fold sets of different ages, multiple, cross-cutting LPS fabrics, and no systematic vertical-axis rotation. Model 2 predicts left- and right-stepping *en echelon* folds and strike-slip faults within more W- and N-trending zones respectively, progressive rotation of early LPS fabrics, and significant vertical-axis rotations correlated with arch trend. Model 3 predicts anastomosing arches of overlapping age, overall SW-NE LPS fabrics that partly correlate with structural trend, and limited vertical-axis rotation. Arches form subperpendicular to regional compression in model 1, whereas arches form oblique to compression in model 2, and along weak zones partly related to basement fabrics in model 3.

Statistical analysis of paleomagnetic declinations for structurally simple sites indicates rotations are limited and uncorrelated with structural trend (Fig. 32B; Weil and Yonkee, 2012). Thus, regional variations in Laramide arch trends are primary, although some sites underwent limited, non-systematic block rotations. LPS directions estimated from minor fault and AMS data are partly correlated with structural trend with statistically indistinguishable strike-test slopes of 0.55 (± 0.10) and 0.63 (± 0.14), respectively (Fig. 32B), reflecting partial refraction with LPS oblique to subperpendicular to structural trends of anastomosing arches, consistent with model 3. Regional sedimentologic and structural relations also support broadly synchronous development of anastomosing arches, (Dickinson et al., 1988; Erslev, 1993; Molzer and Erslev, 1995; Johnson and Anderson, 2009). Steeper forelimbs of more W- and N-trending arches and areas with multiple fold trends, however, experienced local rotations (Tetreault et al., 2008), and *en echelon* folds and strike-slip faults are locally developed (Paylor and Yin, 1993; Stone, 1969), consistent with wrench shear in some zones. Development of multiple joint sets, variations in calcite twin strain directions (Craddock and van der Pluijm, 1999), and local intersection of multiple fault-fold sets

(Bergh and Snoke, 1992) also support temporal changes in shortening directions in some areas. Thus, a model that combines bulk WSW-ENE shortening with stress refraction related to basement fabrics, local wrench shear along steep, oblique-trending forelimbs, and local temporal changes in stress related to activation of different basement weaknesses is most consistent with available data.

4.7. Final collapse

A fundamental switch in deformation and magmatic styles occurred across the western U.S. during Middle Eocene to Oligocene time. Enhanced calc-alkaline igneous activity, with eruption of large ignimbrite sheets, progressed southward from mid Eocene to Miocene time (Armstrong and Ward, 1991; Best et al., 2009), likely recording progressive slab removal and influx of asthenosphere that interacted with previously hydrated mantle lithosphere (Humphreys et al., 2003). Within the hinterland, localized, rapid exhumation occurred along extensional shear zones in metamorphic core complexes, younging from mid Eocene in Idaho to Miocene in Arizona (Coney and Harms, 1984). Within the Sevier belt, extensional collapse started at ~50 Ma and continued into the Oligocene from Montana to Utah, marked by development of listric normal faults that reactivated thrust ramps and deposition of volcanoclastic strata in half grabens (Constenius, 1996). Within the Laramide belt, local development of normal faults and limited collapse of arches with the greatest structural relief also started at ~50 Ma (Steidtmann and Middleton, 1991), penecontemporaneous with increased igneous activity in the Absaroka volcanic field (Fig. 25), and continued into the Oligocene (Hall and Chase, 1989). Progressive isolation of lake systems and changes in paleo-drainage patterns from ~50 to 45 Ma are interpreted to record southward propagation of a topographic high in the Laramide foreland related to slab removal (Smith et al., 2008), and stable isotopic data in the hinterland are also suggestive of southward propagation of surface uplift (Cassel et al., 2009). This switch to early extension also occurred

1426 during decreased plate convergence rates (Fig. 4B) and likely reduced horizontal compression that
1427 favored collapse of previously thickened, thermally weakened crust (Wenicke et al., 1987).
1428 Subsequently, the San Andreas transform system developed from mid Miocene to Recent time
1429 (Atwater, 1970), with a change to regional, distributed extension and basaltic igneous activity
1430 across the Basin and Range province (Surpless et al., 2002).

1431 **5. Discussion and Conclusions**

1432 Structural, stratigraphic, and geochronologic data along the California to Wyoming tectonic
1433 transect are now synthesized, and used to address the questions raised in the Introduction section,
1434 beginning with a comparison of deformation patterns between the Sevier and Laramide belts,
1435 followed by a discussion of influences of plate dynamics, primary crustal architecture, and
1436 mechanical factors on deformation patterns, and then presentation of an integrated model for
1437 evolution of the Sevier and Laramide belts in the context of an Andean-style orogenic system. We
1438 finish with a list of outstanding questions for future research.

1439 *5.1 Relations between deformation patterns in the Sevier and Laramide belts*

1440 Although parts of the same overall Cordilleran system, the Sevier and Laramide belts
1441 developed with differences in structural styles, shortening directions, curvature, and vertical-axis
1442 rotation patterns related to differences in primary crustal architecture and changing plate dynamics.
1443 The Sevier belt comprises thin-skin thrust sheets that developed within thicker passive margin to
1444 transitional strata. Thermochronologic and stratigraphic data record progressive west-to-east
1445 emplacement and exhumation of thrust sheets from 125-50 Ma (Fig. 12), with synorogenic
1446 foredeep deposits progressively incorporated into a propagating, tapered wedge with an overall E-
1447 sloping topographic surface. A western thrust system with aerially extensive, thick thrust sheets

1448 developed in passive margin strata, and an eastern thrust system with closer spaced thrusts
1449 developed in thinner transitional strata. Structural trends began with slight curvature about an
1450 average N-S direction. Thrust sheets then experienced 60-80% secondary rotation related to curved
1451 thrust slip, along-strike changes in fold-thrust shortening, and interaction of younger, frontal
1452 thrusts with Laramide foreland arches at the salient ends. Internal deformation in the western thrust
1453 system was limited at upper levels above a strong quartzite interval, whereas lower levels
1454 underwent simple shear and vertical flattening above a weak basal fault zone. Internal deformation
1455 in the eastern system included widespread early LPS (~5 to 30%) and minor tangential (strike-
1456 parallel) extension. LPS directions display a radial pattern that reflects initial dispersion about a
1457 regional E-W direction and systematic secondary rotation.

1458   The Laramide belt comprises an anastomosing network of thick-skin arches that developed
1459 in cratonic basement rocks overlain by thin strata. Thermochronologic and stratigraphic data
1460 record uplift and exhumation of arches from ~70-50 Ma, which had variable topographic slopes
1461 and were separated by broad basins (Fig. 26). Structural trends define a crude NW-SE structural
1462 fabric, but individual arches and faults display a wide range of trends and primary curvature partly
1463 related to basement weaknesses. Vertical-axis rotations were mostly limited to steep forelimbs of
1464 obliquely trending arches. Internal deformation was limited with <5% early LPS and minor strike-
1465 parallel extension. LPS directions display broad refractions about a regional WSW-ENE direction.

1466   LPS directions for the Sevier and Laramide belts, fully restored for bed tilt and vertical-axis
1467 rotations, illustrate evolution of a composite paleo-stress/ strain field during Early Cretaceous, to
1468 Paleogene time (Fig. 33). During Early Cretaceous time, the western Willard-Paris-Meade sheet of
1469 the Sevier belt was emplaced and early LPS fabrics developed across the future Crawford and
1470 Absaroka sheets. LPS directions were slightly dispersed about 270°, ranging from 260° to 280° in

the northern to southern parts of the Wyoming salient. During mid to later Cretaceous time, the Crawford and Absaroka sheets were emplaced and early LPS fabrics and structural trends underwent secondary rotations as the thrust wedge propagated toward the foreland. LPS fabrics then developed across the future Hogsback-Darby-Prospect sheet and across the Laramide foreland, but with different orientations. LPS directions in the frontal Sevier belt were slightly dispersed about 270° , whereas directions across the Laramide foreland refracted about a regional 240° , ranging from $\sim 220^\circ$ to $\sim 260^\circ$ along W-trending to N-trending arches, respectively. During Paleocene to Early Eocene time, the Hogsback-Darby-Prospect sheet was emplaced and initial structural trends and LPS fabrics underwent secondary rotation, locally enhanced by buttressing near the Laramide Gros Ventre and Uinta arches. LPS fabrics in the foreland did not experience systematic rotation, but the paleostress/strain field locally refracted into high angles with variably oriented, weaker faults. LPS patterns are broadly consistent with gradients in paleo-velocity vectors for Cretaceous and Paleocene time proposed by Bird (1998), which are also at low angles to the SW-NE direction of relative plate motion.

5.2 Influences of plate dynamics on intraplate deformation

Tectonic evolution of the Cordilleran orogenic system was fundamentally related to changing subduction kinematics and dynamics. Subduction boundaries can be divided into two types: (1) retreating, where the rate of trench roll back is greater than the margin-normal absolute motion of the overriding plate, resulting in backarc extension; and (2) advancing, where overriding plate motion is greater than the rate of trench migration, resulting in retroarc contraction (Royden, 1993; Cawood and Buchan, 2007). Kinematically, relative, and in general oblique, plate motion along an advancing boundary is accommodated by a combination of megathrust slip along the subduction channel, potential strike-slip faulting along the outer rise of the subducting plate (Ishii

et al., 2013), transpression and strike-slip faulting concentrated along a thermally weakened magmatic arc, and retroarc shortening (Fig. 34). These kinematic relations are related to a combination of driving and resisting forces: slab pull from negative buoyancy of the subducting plate that increases with ocean lithosphere age; bending resistance of the subducting plate that increases with ocean lithosphere age; shear resistance along the subduction channel; and viscous resistance to asthenosphere flow, including flow along the dipping slab edges, 3-D flow around slab ends, induced corner flow (trench suction), and plate drag relative to deep mantle flow (Fig. 34; Royden and Husson, 2006; Husson, 2012; Nakakuki and Mura, 2013). In general, the rate of trench migration decreases with increasing subducting plate velocity, subduction of older oceanic lithosphere that resists bending, subduction of more buoyant oceanic plateaus and accreting arcs, and deep mantle flow in the same direction as slab dip, which combined with increasing rates of overriding plate motion, promote intraplate contractional deformation (Heuret and Lallemand, 2005; Cawood and Buch, 2007; Lallemand et al., 2008).

Cretaceous to Paleocene retroarc shortening along the tectonic transect developed during increased westward drift of North America, fast rates of subduction of older Farallon oceanic lithosphere, and decreasing subduction angle (Figs. 4, 6, 8). This geodynamic setting led to development of an advancing type boundary and resultant long-term retroarc shortening in the Sevier belt and hinterland. Doglioni et al. (2007) also showed the importance of net westward drift of plates relative to the deeper mantle during Neogene development of advancing and retreating type boundaries along the eastern and western Pacific, respectively, which may also have applied to the North American Cordillera. Punctuated terrane accretion during the Jurassic, and during the mid Cretaceous along the northern part of the boundary, may have further increased plate coupling and promoted episodes of enhanced crustal thickening near the margin. Relative convergence was

oblique to the boundary during later Cretaceous to Paleogene time with transpression and strike-slip faulting concentrated along active parts of the magmatic arc (Tikoff and de Saint Blanquat, 1997; McClelland et al., 2000). Figure 34B illustrates a simple plate kinematic model for the tectonic transect during the mid Cretaceous, with a relative plate convergence ($V_{sp} - V_{op}$) of 100 mm/yr at a 70° angle to the margin, which is partitioned into ~85 mm/yr of near orthogonal slip along the subduction channel (V_{sc}) and potential strike slip along the outer rise (V_{or}), ~6 mm/yr of distributed transpression (V_{tp}) and ~10 mm/yr of partitioned strike-slip faulting (V_{ss}) across the magmatic arc, and total retroarc shortening rate (V_{ra}) of ~5 mm/yr in the Sevier fold-thrust and hinterland. Relative convergence was mostly taken up along the subduction channel, with <10 mm/yr total intraplate shortening related to the difference between overriding plate motion and trench migration. Flat-slab subduction along the Mojave sector resulted in additional shortening subparallel to relative plate motion across the Laramide belt, likely partly related to increased basal traction. In comparison, shortening in the Sevier belt was nearly orthogonal to the plate margin, consistent with decoupling of the upper crust and influence of gravitational spreading from a hinterland plateau (Royden, 1996).

5.3 Influences of primary crustal architecture on deformation styles

Differences in deformation styles between the Sevier and Laramide belts were partly related to variations in primary crustal architecture, and to evolving topographic, thermal, and mechanical factors. Potential influences on evolution of each belt are briefly described below.

The Sevier belt developed within thicker passive margin to transitional strata that contained weak detachment horizons, forming a tapered fold-thrust wedge above a basal decollement. The basal decollement rooted westward beneath the hinterland where crustal thickening led to development of a plateau. Such wedges evolve in response to multiple factors including: variations

1540 in initial sedimentary thickness (Marshak et al., 1992; Lawton et al., 1994; Boyer, 1995; Paulsen
1541 and Marshak, 1999); variations in material strengths (Davis et al., 1983; Dahlen, 1990; Willett,
1542 1992); buttressing by foreland uplifts (Grubbs and Van der Voo, 1976; Beutner, 1977); variable
1543 stress transmission from a hinterland plateau; and patterns of erosion/ sedimentation (Fig. 35).
1544 Analytical critical taper models provide simple predictions on overall wedge shape that depend on
1545 strengths of the wedge interior and a weak basal decollement (Davis et al., 1983; Dahlen, 1990).
1546 Numerical models provide additional predictions on stress-strain histories of wedges, which may
1547 include parts below failure stresses (Simpson, 2011), areas undergoing synorogenic sedimentation
1548 and erosion (Stockmal et al., 2007; Fillon et al., 2013), and parts with T-sensitive viscous flow at
1549 deeper levels (Jamieson and Beaumont, 2013); see Buiter (2012) for a review of wedge models.
1550 Sedimentary prisms with initially thicker strata develop wider spaced thrusts that propagate farther
1551 outward, leading to thrust curvature. Fully restored isopachs for the Wyoming salient indicate
1552 slight primary curvature of the sedimentary prism with thicker strata in the central part of the
1553 salient (Fig. 3B). Presence of a weak basal decollement, such as in Neoproterozoic micaceous
1554 strata and Cambrian shale in the western and eastern parts of the Sevier belt, favors low taper,
1555 whereas pinching out of weak horizons over basement highs favors higher taper. Temporal
1556 changes in fault strength during progressive deformation also affect wedge evolution. For example,
1557 elevated fluid pressure, alteration, and enhanced plastic deformation at deeper levels may weaken
1558 the basal decollement and favor very low taper and thrust propagation with only minor additional
1559 thickening of the wedge interior. Strain softening processes also favor a transition from early,
1560 distributed LPS to later, concentrated slip along very weak fault zones. Foreland arches may act as
1561 buttresses that resist emplacement of frontal thrust sheets, resulting in decreased slip, break-
1562 backward thrusts, and localized wrench shear, consistent with relations in the northern Darby-

1563 Prospect sheet near the Gros Ventre arch (Fig. 11D; Grubbs and Van der Voo, 1976; Hunter,
1564 1988). Development of a high elevation hinterland transmits stress across the rear of a fold-thrust
1565 wedge, with greater crustal thickening in a wider hinterland plateau enhancing wedge shortening.
1566 Incorporation of synorogenic strata into the wedge, which was important in the Sevier belt, favors
1567 propagation of wider spaced thrusts. Enhanced erosion, in contrast, favors concentrated
1568 exhumation and narrower wedges that expose deeper levels. Enhanced erosion along the front the
1569 Sevier belt, recorded by detrital thermochronology (Painter et al., 2014), may have slowed
1570 eastward propagation of the wedge front, whereas decreased erosion across the hinterland likely
1571 favored development of a broad plateau. In summary, evolution of the Sevier belt was influenced
1572 by a combination of variations in sedimentary prism thickness and lithology, presence of weak
1573 fault zones, buttressing of frontal thrusts by foreland uplifts, stress transmitted from a hinterland
1574 plateau through a tapered wedge, and synorogenic erosion and sedimentation (Fig. 35).

1575   The Laramide belt developed in cratonic lithosphere containing basement overlain by a thin
1576 sedimentary cover. The Laramide formed during flat-slab subduction that likely increased coupling
1577 far inboard from the plate margin, but the nature of stress transfer is uncertain. Models include
1578 increased basal traction (Bird, 1984, 1988), increased coupling along a cratonic lithosphere root
1579 (O'Driscall et al., 2009), hydrodynamic forces related to forced mantle convection (Jones et al.,
1580 2011), and return flow where the slab steepened (Fig. 36A). The nature of stress transfer and
1581 removal of mantle lithosphere depend on mantle and crustal rheology, and thus on thermal state,
1582 fluids, and basement rock types. Thermochronologic data in the Laramide foreland (Peyton et al.,
1583 2012; Fan and Carrapa, 2014) record a near-surface paleo-geothermal gradient of ~ 20 °C/km
1584 (corresponding to a surface heat flow of ~ 50 mWm⁻² for a thermal conductivity ~ 2.5 Wm⁻¹K⁻¹),
1585 and xenolith data (Ritter and Smith, 1996; Glebovitsky et al., 2004) indicate a low mantle paleo-

geothermal gradient of ~ 6 °C/km (corresponding to a basal heat flow of ~ 17 mWm⁻² for peridotite conductivity ~ 2.8 Wm⁻¹K⁻¹). A geotherm for the foreland prior to flat-slab subduction, calculated using these values and internal heat production rates of 1.0 and 0.5 μ Wm⁻³ for upper and lower crustal thicknesses of 20 km each, indicates a cool, thick lithosphere (curve [1] in Fig. 36B). Subsequently, flat-slab subduction brought relatively cold lithosphere in contact with the base of the overriding mantle lithosphere, which led to further cooling (curve [2] in Fig. 36B; Currie and Beaumont, 2011). A simple lithospheric strength model for these thermal conditions indicates brittle failure to >20 km depth and limited plastic flow in the cool crust (Fig. 36C), consistent with geophysical imaging of major reverse faults to mid-crustal levels (Smithson et al., 1979). Upper crust is strong for a typical friction coefficient of 0.6 and hydrostatic fluid pressure, but fault strength may be lower due to basement weaknesses, alteration, and increased fluid pressure. A lower crust comprised of felsic granulite would be moderately weak with potential concentration of plastic flow along a deep detachment, whereas lower crust comprised of mafic granulite would be strong with potential for lithospheric buckling (Tikoff and Maxson, 2001). Strength of the mantle lithosphere likely changed during flat-slab subduction, depending on competing effects of cooling and hydration (Humphreys et al., 2003), which, respectively, favor increased strength and stress transfer, or conversely decreased strength and removal of mantle. Increased plate coupling led to diffuse intraplate shortening along a network of anastomosing arches separated by broad basins, with basement fabrics influencing paleostress and faulting patterns. During increasing loading of heterogeneous basement, the most favorable weaknesses, related to a combination of low friction, high fluid pressure, and orientation relative to regional compression, are first activated as reverse faults. Uplift of arches above early faults increases local mean stress that may

suppress further uplift, leading to activation of other, less favorable weaknesses (Hilley et al., 2005), consistent with protracted development of variable trending arches in the Laramide belt.

5.4 Integrated model for evolution of the Sevier and Laramide belts in the Cordilleran system

Regional structural, paleomagnetic, stratigraphic, and geochronologic data sets are now integrated to develop a model for evolution of the Sevier and Laramide belts along the tectonic transect in the context of the Cordilleran orogenic system. Spatial-temporal relations of tectonic components of the orogenic system are summarized in Figure 37, and a model for evolution of the Sevier and Laramide belts at various stages is illustrated in Figure 38.

During Triassic to Early Jurassic time (>180 Ma), plate convergence was likely slow with an oceanic plate separating marginal arcs from North America. In the area of the future Sevier and Laramide belts, strata were deposited atop an arcuate sedimentary prism that included a thick western passive margin section and a thin eastern cratonic section (Fig. 38A). In the following stages, future thrust sheets of the Sevier belt had greater slip and propagated farther eastward in the center of the Wyoming salient where the initial sedimentary prism was thickest, whereas Laramide arches formed in the cratonic section that only had a thin sedimentary cover underlain by basement with a variety of fabric orientations that partly controlled subsequent foreland deformation.

During Middle to Late Jurassic time (~ 170 -145 Ma), convergence rates increased, westward drift of North America began, marginal arcs (Intermontane and related terranes) were accreted, the E-dipping Franciscan subduction zone developed with underplating of high-grade blocks, and a magmatic arc formed near the plate margin (Fig. 37). Early contractional deformation was concentrated across the magmatic arc and Luning-Fencemaker belt in the western hinterland, forming a relatively narrow orogenic wedge. Limited igneous activity and local deformation associated with thermal weakening occurred in the hinterland, along with early slip on

the Manning Canyon detachment (Allmendinger and Jordan, 1981). Late Jurassic strata were deposited in the Morrison basin across the area of the future Sevier and Laramide belts, and possibly in a subsequently eroded basin across part of the hinterland (Fig. 38B), likely related to dynamic subsidence and flexural loading from early crustal thickening to the west (Currie, 1998).

During earliest Cretaceous time (~145-130 Ma), convergence rates decreased, contemporaneous with local subsidence and deposition of the lower Great Valley Group, a decrease in arc magmatism, and local sinistral transtension to local shortening in the hinterland. Much of the retroarc, including the future Sevier and Laramide belts, experienced erosion, possibly related to dynamic rebound as the subducting slab steepened westward.

During Aptian-Albian time (~125-100 Ma), plate convergence rates increased, contemporaneous with accretion of the eastern Franciscan Complex, renewed subsidence and deposition of the Great Valley Group, and increasing arc magmatism (Fig. 37). The orogenic front propagated rapidly eastward across the hinterland into the western Sevier belt. The hinterland experienced early thickening and metamorphism, but early surface uplift was likely subdued during erosion of non-resistant Mesozoic strata. In the Wyoming salient of the Sevier belt, the aerially extensive Willard-Paris-Meade thrust sheets were emplaced eastward (Fig. 38C). These thrusts had 50-100 km of slip and carried strong, quartzite-rich strata above a weak basal decollement in micaceous strata, which favored low wedge taper, concentrated fault slip, limited internal deformation at upper levels, and vertical flattening at lower levels. Shortening was greatest where the primary sedimentary prism was thicker, leading to structural curvature of the western thrust system and higher elevations in the central rear part of the salient. Emplacement of the western thrust sheets led to isostatic flexure and deposition of thick, mostly continental synorogenic strata in an evolving foredeep. LPS fabrics began developing across the future

1654 Crawford sheet, reflecting burial beneath foredeep deposits, enhanced fluid flow, and increased
1655 stress above propagating footwall imbricate faults. Early LPS directions were slightly dispersed
1656 about E-W, reflecting a combination of tectonic stress and topographically induced stress along the
1657 curved thrust front.

1658   During Cenomanian-Turonian time (~100-90 Ma), rates of dextral oblique plate
1659 convergence increased, contemporaneous with accretion in the central belt of the Franciscan
1660 Complex along the tectonic transect (and accretion of the Insular group farther north), increased
1661 subsidence and deposition of the Great Valley Group across a widening forearc basin, and
1662 increased magmatism in the Sierra Nevada arc that developed a thick eclogitic root and
1663 experienced dextral transpression (Fig. 37). Lower crustal thickening in the hinterland formed a
1664 growing plateau that helped drive upper crustal shortening in the Sevier fold-thrust wedge. Upper
1665 crustal emplacement of thrust sheets in parts of the hinterland, buried, heated, and weakened
1666 footwall rocks that underwent vertical flattening. In the Sevier belt, slip continued on the western
1667 thrust system and LPS fabrics continued to form in front of the wedge. Mixed continental and
1668 marine synorogenic strata were deposited across the foreland basin that was variably inundated by
1669 the Western Interior Seaway, reflecting dynamic subsidence farther eastward during shallowing of
1670 the subducted slab.

1671   During Santonian to early Campanian time (~90-80 Ma), dextral oblique convergence rates
1672 remained high with dextral transpression along the arc and strike-slip motion of accreted terranes,
1673 subduction angles continued to shallow, and a flat-slab segment developed along the Mojave
1674 sector. Accretion continued in the central Franciscan Complex, subsidence and deposition of the
1675 Great Valley Group continued, and the Sierra Nevada arc experienced a high flux of magmas that
1676 incorporated crustal melts (Ducea, 2001), but activity then waned as asthenospheric flow became

1677 disrupted by the developing flat slab (Fig. 37). The hinterland experienced continued crustal
1678 thickening and formed a growing plateau, with metamorphic rocks exposed today in core
1679 complexes reaching maximum depths during this time. In the Sevier belt, the Crawford and early
1680 Absaroka thrust sheets were emplaced as the orogenic front propagated slowly eastward and
1681 incorporated prior synorogenic strata that were thickest toward the central part of the salient (Fig.
1682 38D). Curved fault slip paths led to vertical-axis rotations along the front of the growing thrust
1683 wedge. Crawford shortening was accommodated mostly by large-scale folding with multiple
1684 detachments in the northern part of the salient, and by major thrust slip in the southern part, with
1685 slip partly transferred into early growth of the Wasatch anticlinorium. The western thrust sheets,
1686 now in the internal part of the wedge, were carried overall passively eastward above a very weak
1687 basal decollement. Synorogenic strata were deposited across a broad foreland basin during
1688 continued isostatic flexure from thrust loading and dynamic subsidence.

1689   During late Campanian to Maastrichtian time (~78-66 Ma), convergence rates remained
1690 high, with NE underthrusting of the flat-slab along the Mojave sector, contemporaneous with
1691 underplating of the Rand-Orocopia-Pelona schists, exhumation of the forearc to the south and
1692 deposition of the upper Great Valley Group in shallow waters along the eastern forearc to the
1693 north, and cessation of arc magmatism as asthenospheric flow was disrupted and a diffuse belt of
1694 plutons were intruded along the Colorado Mineral belt (Fig. 37). Within the hinterland, lower
1695 crustal thickening continued, producing a broad, high-elevation (>2-3 km) plateau underlain by
1696 ~50-60 km thick crust. Upper crustal levels experienced episodic extension, reflecting dynamic
1697 equilibrium between tectonic stress, gravitational potential energy, and weakening from crustal
1698 heating. Internally drained basins formed that limited erosional removal of mass from the plateau.
1699 Within the Sevier belt, subsidence and thrust slip rates decreased during the Late Campanian, and

1700 then increased during Maastrichtian slip on the late Absaroka and associated thrusts (Fig. 38E).
1701 Initial curvature of the Absaroka sheet was slightly enhanced due to incorporation of foredeep
1702 synorogenic strata that were wider and thicker in the central part of the Wyoming salient. Curved
1703 thrust slip paths related to stresses induced from curved topographic slopes led to vertical-axis
1704 rotations concentrated along the wedge front. Coarse-grained strata were locally deposited in
1705 wedge-top basins during uplift along deeper ramps and continued growth of the Wasatch
1706 anticlinorium, which helped maintain wedge taper (Coogan, 1992; DeCelles and Mitra, 1995).
1707 Widespread deposition across the foreland basin occurred in response to isostatic flexure from
1708 thrust loading, dynamic subsidence, and early development of foreland arches and basins. LPS
1709 fabrics developed within the future Hogsback-Darby-Prospect sheet, related to Absaroka footwall
1710 deformation, with directions slightly dispersed about E-W. Weak LPS fabrics also developed in the
1711 Laramide foreland, but with a regional WSW-ENE direction related to enhanced stresses above a
1712 flat-slab segment, and early uplift of basement arches began.

1713   During Paleocene to early Eocene time (~65-50 Ma), convergence rates were overall high,
1714 but began decreasing, and the flat-slab segment continued underthrusting to the NE with the
1715 trailing edge passing beneath the Colorado Plateau, contemporaneous with accretion of the coastal
1716 belt of the Franciscan Complex to the north, uplift of the western forearc to slow subsidence of the
1717 eastern part, and exhumation of the inactive Sierra arc (Fig. 37). The hinterland began to undergo
1718 more widespread extension, likely related to heating and weakening of the lower crust, and
1719 possibly to mantle lithosphere removal (Wells et al., 2012). In the Sevier belt, the Hogsback-
1720 Darby-Prospect thrusts were active as the Sevier wedge propagated farther eastward and interacted
1721 with growing Laramide foreland arches (Fig. 38F). Curved thrust slip paths caused regional
1722 rotation near the wedge front, with additional rotation and out-of-sequence thrusting adjacent to

the Gros Ventre and Uinta arches at the salient ends. Continued growth of the Wasatch anticlinorium and uplift over interior ramps helped maintain wedge taper. Sedimentation in the foreland became partitioned into multiple basins between Laramide arches. Regional WSW-ENE thick-skin shortening and uplift of Laramide basement-cored arches overlapped temporally with the final stages of thin-skin shortening in the Sevier belt. This difference in structural style for the Laramide foreland reflected presence of a thin sedimentary cover underlain by overall strong basement and mantle lithosphere, which interacted with the flat-slab segment, with stress transfer by a combination of basal traction, end loading of a lithospheric keel, and enhanced asthenosphere flow. Variations in orientations and timing of arch uplifts and refraction of paleo-stress fields reflected influences of basement fabrics and strain softening along fault zones.

Finally, during middle to late Eocene time (50-35 Ma), plate dynamics changed, resulting in a switch to orogenic collapse across the Cordilleran system, especially in areas with higher elevation and weaker crust. Plate convergence rates decreased, the plate margin stepped outboard from the Columbia embayment to incorporate the Salizia terrane, and the Farallon slab began to roll back and tear, leading to decreased plate coupling and development of slab windows (Humphreys, 1995; Madsen et al., 2006). Igneous activity increased across a broad east-west zone above the northern margin of the flat slab, which then progressed southward during slab roll back and mantle lithosphere delamination (Best et al., 2009). The hinterland experienced concentrated extension in core complexes, and increased paleo-elevation and relief, which progressed southward over time (Chamberlain et al., 2012). The Sevier and Laramide belts underwent limited extension with local development of listric normal faults along thrust ramps.

5.5 Outstanding Questions and Future Research Opportunities

Past and ongoing studies of the North American (NA) Cordillera have provided extensive data bases for developing the tectonic model presented herein that focused on a tectonic transect from California to Wyoming (Figs. 37, 38). However, our understanding of Andean-style systems in general remains incomplete, and outstanding, partly interrelated questions, remain, some of which are briefly addressed below.

1. What are the relations between subduction dynamics, terrane accretion, and intraplate shortening? The absolute rate of the overriding plate motion, relative rate and obliquity of plate convergence, and nature of subducting lithosphere (age and thickness of oceanic crust, plateaus, and arcs) are all thought to be important controls on orogenesis (van Hunen et al., 2002; Schellart, 2005; Lallemand et al., 2008; Martinod et al., 2010). Intraplate shortening in the NA Cordillera was correlated with increased westward motion of NA, increased convergence rates, and decreases in subduction angles, but processes of stress transfer and importance of terrane accretion and strike-slip motion remain debated. Was stress transfer mostly from end loading by terrane accretion, advance of the NA plate relative to the trench, or basal loading by viscous forces during asthenospheric flow? The Sierra Foothills-Klamath Mountains provinces were accreted during the Jurassic with shortening concentrated near the plate margin, whereas Cretaceous to Paleogene Sevier shortening occurred during normal subduction of oceanic lithosphere along much of the margin. Average intraplate shortening rates (~ 5 mm/yr) were much less than plate convergence rates (~ 100 mm./yr; Fig. 37), although shortening rates may have been higher during tectonic pulses (Pana and van der Pluijm, 2015). Active intraplate shortening in the Andes is also related to normal ($\sim 30^\circ$ dip) subduction along much of the Nazca-South America plate boundary, but shortening rates have varied temporally along the length of the margin, indicating multiple factors, including oceanic lithosphere age and flat-slab segments, control orogenesis (Schellart, 2008).

1768 Future geodynamic modeling, including 3-D asthenospheric flow during oblique convergence, and
1769 high-resolution geochronologic studies to estimate deformation rates may provide new insights on
1770 processes of stress transfer, partitioning of strike-slip motion, and controls on variable shortening
1771 rates in Andean-style systems.

1772   2. What is the importance of flat-slab subduction to orogenesis? Flat-slab subduction
1773 appears to be a key process in continental tectonics, influencing styles of intraplate shortening and
1774 seismicity (Dickinson and Snyder, 1978; Jordan et al., 1983; Epsurt et al., 2008), spatial
1775 distribution and composition of magmatism (Coney and Reynolds, 1977; Kay and Mpodozis,
1776 2002), and patterns of surface uplift and subsidence (Davila, et al., 2010; Liu and Gurnis, 2010).
1777 Flat-slab subduction is favored by rapid convergence, absolute trenchward motion of the upper
1778 plate, and increased buoyancy from thicker subducted crust (Gutscher, 2000; van Hunen et al.,
1779 2002; Epsurt et al., 2008). These factors likely contributed to the ancient Laramide flat slab of the
1780 NA Cordillera (Saleeby, 2003; Liu et al., 2008, 2010), and currently contribute to the Andean 27–
1781 33°S flat-slab segment where the thick-skin Sierras Pampeanas are actively uplifting (Martinod et
1782 al., 2010). However, processes of stress transfer to overlying lithosphere, nature of lower crustal
1783 deformation, and influences of basement fabrics remain debated. The nature of slab dehydration
1784 and changes in overriding mantle lithosphere rheology, density, and composition, which contribute
1785 to patterns of surface uplift and magmatism following slab removal, pose additional questions
1786 (Humphreys et al., 2003). Furthermore, the spatial-temporal patterns of transformation of basalt-
1787 gabbro to eclogite, which changes subducted plate buoyancy and thus plate coupling, are poorly
1788 understood. Gabbro-eclogite transformation may require $T > 500^{\circ}\text{C}$ (Schmidt and Poli, 1998), such
1789 that this transformation occurs over long distances inboard from margins in cool subducted plates
1790 (Currie and Beaumont, 2011). Further studies of varying tectonic responses of overriding plates

1791 along different flat-slab segments (Gutscher et al., 2000), along with improved geophysical
1792 imaging, may provide new insights on the nature of stress transfer, slab dehydration, eclogite
1793 transformation, and upper mantle to lower crust rheology.

1794   3. What were the locations, timing, and polarity of terrane accretion and subsequent strike-
1795 slip offsets? The paleo-latitude of accretion and subsequent strike-slip motion of the Intermontane
1796 and Insular terrane groups remain controversial, partly based on interpretations of shallow
1797 paleomagnetic inclinations in sedimentary strata. This paper presented a model with moderate
1798 translation of accreted terranes, following Wyld et al. (2006), and E-dipping subduction during
1799 Middle Jurassic to Paleogene development of the orogenic system. Other models, however, have
1800 been proposed that incorporate major strike-slip offset of terranes and W-dipping subduction
1801 (Johnston, 2008; Hildebrand, 2013). Future paleomagnetic studies, including quantification of
1802 inclination flattening, additional geochronologic and structural studies of overlap strata and
1803 stitching plutons along terrane boundaries, and expanded DZ studies tied to source areas will
1804 provide tests of, and refinements to models of terrane accretion and offset.

1805   4. What is the influence of primary crustal architecture on thin-skin, thick-skin, and hybrid
1806 structural styles? Crustal architecture appears to have been an important control on intraplate
1807 deformation in the NA Cordillera, with the thin-skin Sevier belt accommodating significant
1808 shortening (~50% or 200 km total) in a sedimentary cover with weak decollement horizons, and
1809 the thick-skin Laramide belt accommodating less shortening (~10% or 50 km total) in stronger
1810 basement. The basal decollement of the Sevier belt also locally cut basement steps with
1811 incorporation of basement slices into the wedge, and rooted westward into the lower crust beneath
1812 the hinterland. Primary sedimentary prism architecture partly controlled initial curvature,
1813 subsequent shortening, and vertical-axis rotation patterns in the Sevier belt, and basement fabrics

1814 affected arch trends and paleo-stress directions in the Laramide belt. The active Andean system
1815 also displays distinctive along- and across-strike changes in structural style that correlate with
1816 changes in both crustal architecture and plate dynamics (Ramos et al., 2004), with the Precordillera
1817 and Sierras Pampeanas in western Argentina interpreted as modern analogs to the Sevier and
1818 Laramide belts (Jordan and Allmendinger, 1986; Fielding and Jordan, 1988). Important differences
1819 exist, however, including presence of Paleozoic strata in the Precordillera that underwent prior
1820 metamorphism and have properties intermediate between sedimentary rocks and crystalline
1821 basement, and more widespread reactivation of basement faults with partial detachment of the
1822 sedimentary cover (Giambiagi et al., 2012). Basement in the Sierras Pampeanas contains N- to
1823 NW-trending Paleozoic sutures that are at high angles to the current regional shortening direction
1824 and thus favorably oriented for reactivation, whereas the Laramide contained more variable, E to
1825 NE-trending fabrics at low angles to regional shortening that were less favorably oriented for
1826 reactivation. Characterizing the wide range in structural styles of intraplate shortening and
1827 relations to crustal architecture remains a key area of research (Pfiffner, 2006).

1828   5. What is the nature of hinterland plateau uplift and its relations to crustal shortening and
1829 mantle lithosphere removal? Development of hinterland plateaus is controlled by a balance
1830 between horizontal compressive forces, gravitational potential energy, changing rheology, and
1831 erosional processes, which may be in a state of dynamic equilibrium such that high-elevation parts
1832 of convergent orogens experience episodes of extension and gravitational spreading during overall
1833 crustal thickening (Molnar and Lyon-Caen, 1988; England and Houseman, 1989). Mechanisms to
1834 initiate extension include decreased compressive stress transmitted through the orogen due to
1835 changing plate dynamics, and removal of lithospheric mantle that leads to thermal weakening and
1836 overall surface uplift. Within the NA Cordillera, balancing of Sevier shortening with lower crustal

1837 thickening, P-T-t paths of metamorphic rocks, and stable isotopic data indicate development of a
1838 plateau, but the timing and extent of surface uplift remain debated (Chamberlain et al., 2012).
1839 Removal of mantle lithosphere has been interpreted to have triggered later Cretaceous extension in
1840 the Sevier hinterland (Wells et al., 2012), but flat-slab subduction may have inhibited lithosphere
1841 removal until the Miocene (Zandt et al., 2004). Synconvergent lithospheric drips and delaminated
1842 slabs have been imaged beneath the Altiplano-Puna Plateaus where they appear to have
1843 contributed to early, local basin subsidence and sedimentation followed by broader uplift
1844 (DeCelles et al., 2015). Integrating additional studies of exhumed metamorphic rocks in
1845 hinterlands of ancient systems, with improved geophysical imaging of lower crust to mantle
1846 beneath actively forming plateaus will improve our understanding of plateau formation.

1847   6. What are feedbacks between climate, topography, and tectonic evolution of orogenic
1848 wedges? Topography and related stresses that help drive wedge propagation result from an
1849 interplay between crustal thickening, buoyancy related to mantle processes, and erosion that
1850 depends on climate, rock durability, and physiography (Willett, 1999). Climate, especially rainfall
1851 patterns, in turn, is influenced by topography. In the NA Cordillera, stable isotope data and
1852 sedimentary patterns suggest a monsoonal climate and development of a high elevation plateau
1853 that became, in part, internally drained (Fricke et al., 2010; Druschke et al., 2011). Rainfall and
1854 increased erosion rates along the wedge front, interpreted from characteristics of synorogenic strata
1855 (Painter et al., 2014), may have favored localized shortening to maintain wedge taper and slowed
1856 the rate of wedge propagation. As the Sevier wedge propagated over time, precipitation may have
1857 decreased toward the hinterland, which was also in a rain shadow from the magmatic arc, slowing
1858 erosion and favoring surface uplift and development of a high-elevation plateau with internally
1859 drained basins. The southern Central Andes also comprise a broad wedge with concentrated

1860 rainfall along the retroarc orogenic front and an arid climate over a high-elevation plateau.
1861 Although climate-tectonic coupling partly controls wedge evolution, some models suggest that
1862 convergence rates driven by plate dynamics are most important in shaping orogenic systems (Roe
1863 et al., 2006; Strecker et al., 2007). Developing refined models of climate feedbacks, along with
1864 additional in situ and detrital thermochronologic studies, are key areas of future research.

1865   7. How do feedbacks between mantle/lower crustal delamination, topographic uplift,
1866 shortening, and wedge propagation modulate orogenesis? In the Cordilleran cycle model (DeCelles
1867 et al., 2009; DeCelles and Graham, 2015), upper crustal shortening in the retroarc fold-thrust belt
1868 is balanced by lower crustal thickening beneath the hinterland and magmatic arc, which leads to
1869 increased partial melting and magmatic flux, while depleted lower crust and mafic cumulates form
1870 an eclogitic arc root. The dense eclogitic root and mantle lithosphere beneath the arc and
1871 hinterland are then removed by delamination or drips, leading to early local basin subsidence
1872 (depending on crustal strength), followed by regional uplift and increased topographic slope that
1873 promotes propagation of the orogenic front toward the foreland. Removal of the root provides
1874 space for renewed underthrusting and increased shortening as a new feedback cycle begins.
1875 Magmatic flux varied greatly during evolution of the NA Cordillera, partly related to varying
1876 contributions of crustal melts, consistent with this model. Testing predictions of this model,
1877 however, is hindered by limited information on the topographic evolution of most hinterlands,
1878 difficulties in quantifying deformation rates over shorter (~m.y.) time scales in ancient orogens,
1879 and separating influences of changing plate dynamics on arc magmatic tempos and forearc
1880 accretion patterns (Paterson and Ducea, 2015). Long-term shortening rates in the Sevier belt, based
1881 on sedimentation patterns (Fig. 12), appear to have been broadly constant, rather than tied to
1882 variations in magmatic flux. However, Pana and van der Pluijm (2015) interpreted thrust faulting

to have been highly episodic in the Canadian Cordillera, based on Ar-Ar dating of clay gouges. Although feedbacks, including arc root delamination and surface processes, help modulate orogenesis, changing plate dynamics and primary crustal architecture were also important in shaping the NA Cordillera. Our understanding of interrelations between plate dynamics, crustal architecture, and feedbacks between surface and mantle processes during Andean-style orogenesis will improve with future integrated studies, such as comparative 4-D anatomy of different orogenic systems, combined geochronologic, thermochronologic, and basin analysis studies to more accurately resolve deformation rates, improved geophysical imaging of actively deforming regions, and refined geodynamic modeling.

Acknowledgements. Careful reviews by Adrian Pfiffner, Stephen Johnston, and Carlo Diglioni led to revisions that significantly improved the manuscript. Numerous discussions over the years with our colleagues, including Rick Allmendinger, Jim Coogan, Pete DeCelles, Eric Erslev, Gautam Mitra, Basil Tikoff, and Michael Wells, have helped guide our work. We wish to especially thank current and former students Elizabeth Balgord, Fern Beetle-Moorcroft, Tyler Cluff, Andrea Cutruzzula, Steve Fellows, Jamie Kendall, Melissa Lindholm, Anna Mazzariello, Sarah McCullough, Andriy Mshanetsky, Evan Pugh, Zoe Ruge, Mary Schultz, Zoe Statman-Weil, Cameron Thompson, Matt Tomich, Kira Tushman, Virginia Walker, David Wicks, Peiyeng Wen, and Amelia Lee Zhi Yi, who helped with extensive field and laboratory work. Student summer support was partially funded by the Bryn Mawr College Summer Science Research Fellowship program and its funding agencies and offices. This work was supported by National Science Foundation grants EAR-0409103, EAR-0408653, EAR-0838021, EAR-0948677, and EAR-0948692.

1906 **Figure Captions.**

1907 Figure 1. Generalized map of the North American Cordilleran system shows current distribution of
1908 major tectonic elements, including Jurassic-Eocene accretionary complexes, forearc basins,
1909 magmatic arc, hinterland metamorphic core complexes, Sevier fold-thrust belt, Laramide foreland
1910 belt, and foreland basin. Approximate boundary between North American basement and accreted
1911 terranes indicated by $^{87}\text{Sr}/^{86}\text{Sr} = .706$ line (Kistler and Peterman, 1978). Terrane groups and major
1912 strike-slip and transpressional zones indicated. Neogene tectonic features include the Cascadia
1913 subduction zone (representing a remnant of the more extensive Farallon subduction zone), San
1914 Andreas Fault system, Basin and Range province, Yellowstone hot spot trace, and Rio Grande
1915 Rift. Location of the California to Wyoming tectonic transect is indicated.

1916
1917 Figure 2. Map of basement provinces in the western U.S. (restored for Neogene extension).
1918 General ranges of igneous zircon U-Pb ages (t_{Zr}) and Nd model ages (t_{Nd}) indicated. $^{87}\text{Sr}/^{86}\text{Sr}$
1919 $= .706$ line marks edge of North American basement. Modified from Foster et al. (2006),
1920 Whitmeyer and Karlstrom (2007), and Yonkee et al. (2014), with additional isotopic data from
1921 Farmer and DePaolo (1983). Locations of Mesoproterozoic (~1.5-1.4 Ga) Belt Basin, and restored
1922 leading edge of western and eastern thrust systems of the Sevier belt also indicated.

1923
1924 Figure 3. Primary architecture of the sedimentary cover. **A.** Stratigraphic columns illustrate
1925 regional variations in thickness and lithology for passive margin (Willard thrust sheet, western
1926 Sevier belt), transitional (Crawford thrust sheet, eastern Sevier belt), and cratonic (Wind River
1927 arch, Laramide belt) strata. Sedimentary cover rocks were deposited during Neoproterozoic to
1928 Early Cambrian rifting, Early Paleozoic subsidence along a passive margin, and Late Paleozoic to

1929 Early Mesozoic disruption of the margin. **B.** Palinspastically restored isopach maps for (i) middle-
1930 late Neoproterozoic, (ii) Cambrian, (iii) Ordovician, and (iv) Pennsylvanian strata. General
1931 locations of the continental slope, Wasatch hinge zone (transition to passive margin), and restored
1932 leading edges of western and eastern thrust systems of the Sevier belt indicated. Locations of
1933 Roberts Mountain allochthon, Oquirrh basin, Ancestral Rockies, and sinistral offset of the margin
1934 shown for Pennsylvanian time. Isopachs based on Howell and McDougall (1978), Levy and
1935 Christie-Blick (1989), and Hintze and Kowallis (2009).

1936
1937 Figure 4. Plate motion history. **A.** Reconstructions of North America, Pacific, Farallon, Izanagi,
1938 and Kula plates at various times based on G-plate model (Seton et al., 2012; Gurnis et al., 2012), .
1939 Reconstructed locations of accreting terranes and Shatsky (Sh) oceanic plateau and conjugate,
1940 Intermontane (IM) and Insular (In) terrane groups, Mezcalera plate (Dickinson and Lawton,
1941 2001a), and present day seismic anomaly at 1000-1300 km depth interpreted to be remnants of
1942 subducted Farallon plate (shaded area, from Liu et al., 2010) also shown. **B.** Estimated rates and
1943 directions of North America absolute and relative motion near central California along the tectonic
1944 transect, based on Engebretson et al. (1985), Torsvik et al. (2008), van der Meer et al. (2010),
1945 Doubrovine et al. (2012), and Williams et al. (2015). Rates shown over 10 m.y. time intervals for
1946 comparison. Convergence rates between the North American and Farallon plates are dashed prior to
1947 110 Ma when a marginal plate was likely being subducted along the margin. **C.** Constraints on
1948 subduction ages and formation ages of oceanic crust blocks in the Franciscan Complex based on
1949 summaries in Shervais et al. (2005) and Wakabayashi (2015).

1950

1951 Figure 5. Palinspastically restored regional maps show spatial evolution of tectonic components of
1952 the North American Cordilleran orogenic system for: **A-** Late Jurassic (~150 Ma) consolidation of
1953 the subduction boundary and the Nevadan orogeny; **B-** Early Cretaceous (~100 Ma) increase in
1954 plate convergence and early Sevier deformation; **C-** Late Cretaceous (~80 Ma) initial development
1955 of flat-slab segment, continued Sevier deformation, and initial Laramide deformation; and **D-**
1956 Paleogene (~50 Ma) end of Sevier and Laramide deformation. From west to east, the major
1957 tectonic components are forearc accretionary wedge and basin, magmatic arc, hinterland with
1958 metamorphic core complexes, Sevier fold-thrust belt, Laramide foreland belt, and foreland basin
1959 (see Figure 1 for explanation of map symbols and labels). Relative motion between subducting
1960 oceanic and North American plates was partitioned into slip along the subduction boundary, strike-
1961 slip transport of accreted terranes, and intraplate shortening. Lines for cross sections in Figure 6
1962 indicated. Maps incorporate relations presented by Oldow et al. (1989), Saleeby (2002), and
1963 DeCelles (2004), with restoration of Neogene extension, strike slip along the San Andreas Fault
1964 system, and Cretaceous contraction and dextral transport of accreted terranes based on McQuarrie
1965 and Wernicke (2005), Wyld et al (2007), and this study.

1966
1967 Figure 6. Schematic lithospheric cross sections, shown for: **A-** Late Jurassic (~150 Ma); **B-** Early
1968 Cretaceous (~100 Ma); **C-** Late Cretaceous (~80 Ma); and **D-** Paleogene (~50 Ma) times. The Late
1969 Jurassic was marked by development of an E-dipping subduction zone, increased igneous activity
1970 along the magmatic arc, and initial shortening in the western part of the orogenic system. The
1971 Early Cretaceous included development of an accretionary complex, growth of the magmatic arc,
1972 crustal thickening in the hinterland, thin-skin shortening of the western Sevier belt, and
1973 sedimentation within a foreland basin. The Late Cretaceous included decreasing subduction angle

1974 (along with early development of a flat-slab segment to the south), a flare-up followed by
1975 shutdown of the magmatic arc, lower crustal thickening beneath a hinterland plateau, shortening in
1976 the eastern Sevier belt, and broad dynamic subsidence of the Western Interior Seaway. The
1977 Paleogene included rapid subduction and NE-underthrusting of the flat-slab segment, minor
1978 extension in the thickened hinterland, final shortening in the eastern Sevier belt, and uplift of
1979 Laramide arches that disrupted the foreland basin.

1980
1981 Figure 7. Forearc history. **A.** Time line shows metamorphic and accretion ages of the Franciscan
1982 Complex, sedimentation history of the forearc basin, general location and relative flux in the
1983 magmatic arc, and deformation episodes in the arc to western hinterland. Includes data from
1984 Shervais et al. (2005), Ernst (2011), Williams and Graham (2013), Ducea et al. (2015), and
1985 Wakabayashi (2015). **B.** Schematic cross sections illustrate evolving forearc and magmatic arc
1986 during changing subduction, with: Early Jurassic subduction of an intra-arc oceanic plate (northern
1987 part of the Mezcalera plate; Dickinson and Lawton, 2001a); Middle Jurassic development of the E-
1988 dipping Franciscan subduction zone, formation of the Coast Range ophiolite, and overlap of
1989 Mariposa-Galice sequence; Late Jurassic intraplate shortening (Nevadan), early growth of the
1990 magmatic arc, sinking of the prior oceanic plate with enhanced asthenosphere flow, retroarc
1991 igneous activity, and subsidence of the Morrison basin; earlier Cretaceous slow convergence with
1992 local transtension, decreased magmatism, and exhumation across the retroarc; Early Cretaceous
1993 increasing convergence with accretion, forearc basin sedimentation, and increased magmatism;
1994 and mid Cretaceous rapid convergence and shallowing subduction with continued accretion,
1995 forearc basin sedimentation, magmatic flare-up and removal of an eclogitic root.

1996

1997 Figure 8. Magmatic arc. **A.** Estimated magmatic addition rates for Sierra Nevada (DeCelles et al.,
 1998 2009), Cascade (Paterson et al., 2011), and Coast Mountains (Gehrels et al., 2009) igneous
 1999 complexes. Typical background addition rate for arcs indicated by dashed line. **B.** Detrital zircon
 2000 (DZ) probability age distributions for forearc strata (DeGraff-Surpless et al., 2002; Ernst et al.,
 2001 2009; Sharman et al., 2015), Alberta to Utah retroarc strata (Laskowski et al., 2013), and Mojave
 2002 sector strata (Barth et al., 2013). **C.** Correlations between episodes of high magmatic addition
 2003 rates, increased $^{87}\text{Sr}/^{86}\text{Sr}$, and increased La/Yb ratios interpreted to record partial melting of
 2004 eclogite, based on Giradi et al. (2012). **D.** Age-distance relations of arc rocks for southern and
 2005 northern parts of Sierra Nevada plutonic complex (Cecil et al., 2012) indicate shallowing of
 2006 subduction from 120 to 80 Ma. Relations for retroarc record widespread Middle to Late Jurassic
 2007 magmatism with mantle sources, and Cretaceous magmatism related to crustal partial melting.
 2008

2009 Figure 9. Hinterland history. **A.** P-T-t paths for East Humboldt-Ruby (McGrew et al., 2000; Hallett
 2010 and Spear, 2014) and Albion-Raft River-Grouse Creek (Wells et al., 2012; Kelly et al., 2015) core
 2011 complexes. Overall clockwise paths record: local Late Jurassic and regional Early to mid
 2012 Cretaceous tectonic burial with peak P by ~80 Ma; later Cretaceous heating with local, alternating
 2013 shortening/extension; and Paleocene slower exhumation followed by Eocene to Oligocene rapid
 2014 exhumation and cooling during orogenic collapse. Estimated depths assume lithostatic pressure. **B.**
 2015 Schematic cross sections showing pre-Jurassic crustal architecture and Cretaceous development of
 2016 a hinterland plateau during Sevier shortening. Evidence for a plateau includes restoration of
 2017 current crustal thicknesses for Neogene extension (Coney and Harms, 1984), paleo-elevations
 2018 based on stable isotopic data (Snell et al., 2014), P-T-t paths of metamorphic rocks, balancing of
 2019 Sevier upper crustal shortening (area A2) and hinterland lower crustal thickening (area A1).

2020

2021 Figure 10. Geologic and shaded relief map of the Wyoming salient of the Sevier fold-thrust belt.

2022 Major thrusts (Willard, Paris, Meade, Crawford, Absaroka, Hogsback, Darby, Prospect) and

2023 associated folds display systematic curvature from NW trends in the north part to NE trends in the

2024 south part of the salient. Locations of cross sections A-A', B-B', C-C', and D-D' in Figure 11

2025 indicated. Locations of section lines for Figures 13 and 17 and BE- Big Elk, GC- Georgetown

2026 Canyon, and SC- Swift Creek areas for Figure 20 also indicated. Modified from Coogan (1992).

2027

2028 Figure 11. Cross sections A-A', B-B', and C-C' across southern, central, and north-central parts of

2029 the Wyoming salient illustrate typical thin-skin structural style. Fold-thrust shortening is greater in

2030 the central part of the salient. Section D-D' illustrates interaction of frontal thrusts with the

2031 Laramide foreland Gros Ventre arch. Sections modified from Hunter (1988), Coogan (1992),

2032 Royse (1993), Peyton et al. (2011), and Yonkee and Weil (2010).

2033

2034 Figure 12. Timing of major thrust faults in the Wyoming salient. **A.** Synorogenic strata from the

2035 southern Wyoming salient across the Green River basin record: K1- Early Cretaceous deposition in

2036 a foredeep during Willard-Paris-Meade thrusting; K2- mid Cretaceous foredeep deposition during

2037 Crawford thrusting along with regional (dynamic) subsidence; K3- mid and later Cretaceous

2038 foredeep deposition during early and late Absaroka thrusting, separated by a local unconformity;

2039 T1- Paleocene to Early Eocene wedge top, foredeep, and foreland sedimentation during

2040 overlapping Hogsback thrusting and Laramide arch uplift; and later Eocene deposition in local half

2041 grabens during a switch to extension. Thrusts developed from west to east (forward propagating),

2042 with limited thrust reactivation and interior uplift of the Wasatch anticlinorium to maintain critical

2043 taper. Modified from Royse (1993). **B.** Geochronologic and thermochronologic data, including
2044 apatite fission track ages (Naeser et al., 1983; Burtner and Nigrini, 1994), K-Ar ages of illite
2045 (Burtner and Nigrini, 1994), and $^{40}\text{Ar}/^{39}\text{Ar}$ ages of muscovite (Yonkee et al., 1989) record
2046 progression of internal deformation and thrust exhumation across the Sevier belt during Cretaceous
2047 to Early Eocene time, followed by later Eocene switch to extension.

2048
2049 Figure 13. Map and down-plunge projection of the Willard thrust sheet, with effects of subsequent
2050 deformation (uplift of Wasatch anticlinorium and Neogene extension) removed. The Willard thrust
2051 sheet was emplaced ~60 km eastward with additional shortening from imbricate thrusts and large-
2052 scale folds. Cleavage is best developed in the lower parts of the sheet and rotates into parallelism
2053 with the thrust at the base of the sheet. Position of Willard Peak (WP) study area shown in Figure
2054 16 indicated. Modified from Yonkee (2005).

2055
2056 Fig 14. Structural styles at different levels in the Willard sheet, western thrust system.
2057 **A.** Quartzite at middle levels underwent limited cataclastic deformation with development of
2058 fractures and minor faults. **B.** Photomicrograph of quartzite shows little deformed quartz grains (Q)
2059 with limited undulose extinction (u) and microcracks (mc). **C.** Diamictite at lower levels
2060 underwent widespread ductile deformation with development of cleavage (S1) defined by
2061 preferred orientation of mica and flattened clasts. **D.** Photomicrograph of greywacke at lower
2062 levels shows distinctly flattened quartz grains with widespread undulose extinction, local
2063 recrystallization (rx), dissolved grain margins (d), quartz fibers (f), altered micaceous matrix (M),
2064 and shear bands (sb) related to top-to-E shear. **E.** View to north of Willard thrust that places highly
2065 deformed micaceous strata of the Perry Canyon Formation (Zpc) over imbricated middle Cambrian

2066 Ophir and Maxfield shale and limestone (Com) and lower Cambrian Tintic quartzite (Ct);
 2067 phyllonite and mylonite along the fault core are exposed in the foreground. **F.** Photomicrograph of
 2068 phyllonite from fault core shows fine-grained micaceous matrix with contorted foliation cut by
 2069 multiple sets of variably deformed microveins (v1- thicker tightly folded vein, cut by v2- thin
 2070 planar veins), and slip surfaces (ss).
 2071
 2072 Figure 15. Kinematic and rheologic models for the Willard thrust sheet. **A.** Block diagrams on left
 2073 illustrate kinematic relations of mesoscopic structures (cleavage, veins, minor folds), strain
 2074 ellipses, and strain components (Lsh- layer-parallel simple shear [γ], LPS to LPE- layer-parallel
 2075 shortening to extension [α_1], VT- vertical thinning [α_3], TE- tangential extension [α_2], and Wsh-
 2076 wrench shear [γ_w]) at different levels of the Willard sheet. Diagrams on right illustrate large-scale
 2077 fold-fault relations and map-view curvature. Insets shows idealized kinematic model with
 2078 combined thrust-parallel simple shear, vertical thinning, and thrust-parallel shortening to
 2079 extension. **B.** Simple rheologic model for the Willard fault zone comprised of: tectonized base of
 2080 the hanging wall with intense cleavage in micaceous strata, boudinaged quartzite layers, and
 2081 multiple vein sets; fault core with mylonite, phyllonite, cataclasite, quartz-cemented pod cut by
 2082 minor normal fault, and discrete thrust slip surfaces; and imbricated footwall. Mohr stress circle
 2083 constructed for elevated fluid pressure and σ_1 (maximum stress) at high angle to the fault shows
 2084 thrust slip along a low-friction micaceous fault core, tensile failure of veins, normal shear failure of
 2085 quartz-cemented pod, and power law creep of muscovite (M^* , modified for mica aggregate; Mares
 2086 and Kronenberg, 1993) and wet quartz (Q; Hirth et al., 2001) at strains rates $\sim 10^{-14}$ and 10^{-15} s^{-1} ,
 2087 respectively.
 2088

2089 Figure 16. Strain relations for the Willard Peak area, Willard thrust sheet. Down-plunge
2090 projections of northern and southern domains show bedding, cleavage trajectories, and strain
2091 ellipses of micaceous clasts in diamictite and slate and quartz clasts in greywacke of the Perry
2092 Canyon Formation, micaceous quartzite of the Maple Canyon Formation, and quartzite of the
2093 Caddy Canyon Formation. Inset shows detailed relations for a recumbent syncline. Strain ellipse
2094 ratios increase and long axis angles decrease overall toward the base of the sheet, consistent with
2095 simple shear and vertical flattening at lower levels.

2096
2097 Figure 17. Structural model for sequential development of the Wasatch anticlinorium and eastern
2098 thrust system in the southern Wyoming salient. Major thrusts are labeled from west to east: WT-
2099 Willard thrust, OTU and OTL- upper and lower branches of Ogden thrust system, BT- basal thrust,
2100 CT- Crawford thrust, AT- Absaroka thrust, and HT- Hogsback thrust. Synorogenic deposits for
2101 each thrust correspond to those given in Figure 12. Future thrusts are dashed, active thrusts in
2102 black, and inactive thrusts shown in gray. Modified from Yonkee (1992) and DeCelles (1994).

2103
2104 Figure 18. Structural styles in the eastern thrust system. **A.** Micritic limestone of the Jurassic Twin
2105 Creek Formation displays spaced cleavage (S1) at high angle to bedding (S0) that formed during
2106 early LPS (dark arrows). **B.** Bed-parallel outcrop of limestone shows veins at high angles to bed
2107 strike that accommodated tangential extension (light arrows) and shear along *en echelon* arrays.
2108 Chisel for scale. **C.** Red beds of the Triassic Ankareh Formation display spaced cleavage (S1), best
2109 developed in calcareous mudstone. Shapes of reduction spots record strain. Hammer for scale. **D.**
2110 Photomicrograph of limestone shows cleavage defined by seams enriched in clay and quartz that
2111 formed mostly from dissolution of calcite (dark arrows). Some calcite was reprecipitated in veins

2112 (light arrows), but limestone underwent net volume loss. **E.** View to west of Darby thrust fault
2113 zone that places brecciated Mississippian limestone and sandstone cut by multiple minor faults in
2114 the hanging wall, over Cretaceous shale in the footwall, with slip concentrated along a thin coal
2115 bed (inset). **F.** Photomicrograph of cataclasite from base of hanging wall shows multiple sets of
2116 calcite-sealed microveins (v) and slip surfaces.

2117
2118 Figure 19. Block diagrams illustrate kinematic evolution of mesoscopic structures in the Twin
2119 Creek and Ankareh formations (left) and progressive development of large-scale fold-fault
2120 relations and map-view curvature (right) in the eastern thrust system, from (i) initial undeformed
2121 state, to (ii) early LPS, tangential extension (TE), and wrench shear (Wsh) accommodated by
2122 cleavage, minor wedge and tear faults, minor folds, and veins, to (iii) tilting of bedding and
2123 cleavage, development of weak second cleavage, and slip on later reverse and strike-slip faults
2124 during large-scale folding and thrusting.

2125
2126 Figure 20. Structural and timing relations of internal deformation in the eastern thrust system. **A.**
2127 Down-plunge projection of the Big Elk Mountain anticline (location BE in Fig. 10). Cleavage is
2128 strongly fanned about the fold, remaining subperpendicular to bedding in both the fold crest and
2129 limbs. Stereograms show *in situ* and restored orientation of bedding, cleavage, and vein poles.
2130 Cleavage (S1) formed during early LPS and was then tilted during large-scale folding and
2131 thrusting. Restored LPS and extension directions labeled. **B.** Down-plunge projection of
2132 Georgetown Canyon syncline in footwall of the Meade thrust (GC in Fig. 10) and folds in the
2133 Crawford thrust sheet. Early cleavage (S1) developed at acute angles to bedding and the Meade
2134 thrust within the syncline during synthrust shear. The Meade thrust and footwall syncline were

then tilted during younger slip on the Crawford thrust and a second cleavage (S2) locally developed. **C.** Outcrop sketch along Swift Creek area in Absaroka thrust sheet (SC in Fig. 10) shows relations between bedding (S0), early cleavage (S1), and early wedge faults at acute angles to bedding, along with later tectonic stylolites (S2), reactivated S1 cleavage, and gentle-dipping reverse faults associated with fold tightening. Modified from Mitra and Yonkee (1985).

Figure 21. Regional patterns of internal deformation and vertical-axis rotation for the Twin Creek and Ankareh formations. **A.** LPS directions estimated from cleavage, high-angle fracture sets, and early wedge faults define a radial pattern. **B.** Extension directions estimated from veins define a tangential pattern. **C.** Bed-parallel strain ellipses estimated from crinoids in limestone and reduction spots in red beds record combined radial LPS and minor tangential extension. Cleavage intensity is low and strain ellipses are roughly circular in the central Hogsback sheet, whereas cleavage intensity is higher and strain ratios increase westward in the Crawford sheet and toward the salient ends. Anisotropy of magnetic susceptibility (AMS) K_{max} lineations are oriented parallel to intersection of LPS fabrics with bedding. **D.** Paleomagnetic site-mean declinations record systematic counterclockwise and clockwise rotations in the northern and southern parts of the salient respectively. For purposes of comparison, declinations are plotted with respect to a N-directed reference paleopole. Modified from Weil and Yonkee (2009), Yonkee and Weil (2010) and Weil et al. (2010), with additional LPS directions for Cretaceous strata and additional paleomagnetic data from Grubbs and Van der Voo (1976). Data from overturned fold limbs and transfer zones are not shown.

Figure 22. **A.** General kinematic models of thrust belt curvature include: Model 1- primary arc with uniform thrust slip and no secondary rotation; Model 2- primary arc with radial slip and no secondary rotation; Model 3- progressive secondary rotation with curved thrust slip; and Model 4- later secondary rotation of original linear belt. Each of these models predicts different patterns of rotations recorded by paleomagnetic data and internal deformation, which can be quantitatively evaluated using strike tests. For a strike test, paleomagnetic declinations relative to a reference declination (P-Pr), or LPS directions relative to a reference shortening direction (L-Lr), are correlated with changes in structural trend relative to regional trend (St-Sr). **B.** Strike-test results for the Wyoming salient. Slopes for paleomagnetic declinations of Tr and K components (relative to reference Triassic and Cretaceous paleopoles for North America) are respectively $0.75 (\pm .11)$ at 95% confidence interval) and $0.77 (\pm .26)$, indicating ~75% secondary vertical-axis rotation and 25% primary curvature. Modified from Weil et al. (2010). Slopes for LPS directions estimated from cleavage/high-angle fractures and finite strain directions are respectively $0.88/0.95 (\pm .10)$ and $0.89 (\pm .12)$. Modified from Yonkee and Weil (2010). Example site error bars, number of sites (N), and average residual (σ_r) are listed. **C.** Simplified kinematic model for evolution of the Wyoming salient illustrates idealized paleomagnetic declinations (P), LPS directions (L), thrust slip trajectories (θ), and variations in structural trend (S). Stage 1- slip on Crawford thrust, stage 2- slip on Absaroka thrust, and stage 3- slip on Hogsback-Darby-Prospect thrusts and uplift of Laramide foreland arches. Rotation was concentrated along the propagating front of the thrust wedge.

Figure 23. Foreland basin architecture and subsidence mechanisms. **A.** Schematic lithospheric section and detail of retroarc thrust wedge and foreland basin illustrate long-wavelength dynamic subsidence related to asthenosphere flow induced by subduction, isostatic flexural loading from

emplacement of thrust sheets, and loading from redistribution of sediments in wedge top, foredeep, forebulge, and backbulge depozones. Propagation of the thrust wedge and associated depozones leads to deposition of overall upward coarsening synorogenic strata and increasing foredeep subsidence rates. **B.** Representative sedimentary thickness curves shown for strata within the southern Wyoming salient to foreland (inset map shows locations of sections). Note eastward propagation of subsidence and later Cretaceous to Paleogene disruption by Laramide arches.

Fig. 24. Paleogeographic reconstructions of the foreland basin within the Cordilleran orogenic system illustrate evolving topography (schematic), active fault locations, general depositional environments, and deposition rates (m/m.y.) of synorogenic strata. Reconstructions shown for: A and B- Aptian to Albian early shortening along western thrusts of the Sevier belt and early transgression of Western Interior Seaway; C and D- Cenomanian to Turonian continued thrust shortening and enhanced dynamic subsidence to the east; E and F- Santonian to early Campanian propagation of eastern thrusts of the Sevier belt, increased dynamic subsidence, and initial development of the flat-slab segment; and G and H- late Campanian to Maastrichtian continued slip on eastern thrusts, propagation of the flat-slab segment and shut down of the Sierra Nevada arc, and early deformation in Laramide foreland. Maps incorporate data from Robinson et al. (2005), DeCelles (2004), and our own work.

Figure 25. **A.** Generalized map of Laramide foreland in Wyoming shows structural trends of basement (stipple) arches (from Blackstone, 1993), basins (yellow), major faults and cover folds, and outcrops of Triassic red beds. Locations of section lines X-X' and Y-Y' and the Big Horn arch (BH) study area shown in Figure 31 are indicated. **B.** Cross sections X-X' and Y-Y' across Wind

2203 River and Bighorn areas, respectively, show thick-skin structural style of arches bound by reverse
2204 faults that continue to at least mid-crust depths, but do not appear to cut the MOHO based on
2205 seismic data (Smithson et al., 1979; Yeck et al., 2013).

2206
2207 Figure 26. Timing constraints for Laramide foreland. **A.** Apatite fission track ages for the
2208 Beartooth arch (Omar et al., 1994), Big Horn arch (Peyton et al., 2012), and northern Wind River
2209 arch (Cervený and Steidtmann, 1993; Peyton et al., 2012) are plotted in terms of structural depth
2210 on cross sections constructed through the arches. Estimated depth of basement-sedimentary cover
2211 contact, depth range of PAZ (partial annealing zone), and T-t path from Fan and Carrapa (2014)
2212 based on apatite fission track length distribution for Wind River arch also indicated. Fission track
2213 data and ages for onset of enhanced subsidence and overlap of post-faulting strata (DeCelles et al.,
2214 1991; Hoy and Ridgway, 1997; Beck et al., 1988; Steidtmann and Middleton, 1991) record major
2215 arch uplift from ~70-50 Ma. **B.** Isopach maps and cross sections illustrate evolving syntectonic
2216 sedimentation patterns in the Wind River Basin (thickness data from Keefer, 1970).

2217
2218 Figure 27. Structural styles in the Laramide foreland. **A.** Outcrop of limestone in the Sundance
2219 Formation has wedge fault with calcite-fiber steps that give slip sense. **B.** Hand sample from
2220 Alcova Limestone in Triassic Chugwater Group displays a minor wedge fault (wf) at acute angle
2221 to bedding and minor tear fault (tf) at high angle to bedding (S0). **C.** Red beds in Triassic
2222 Chugwater Group appear little deformed except for fracture sets; inset shows approximately
2223 circular reduction spots on bedding surface. **D.** SEM image of red bed sample with kinked and
2224 rotated phyllosilicates that contribute to AMS K_{max} lineations. **E.** View to southeast of the White
2225 Rock reverse fault that places fractured Archean gneiss cut by minor faults in the hanging wall on

2226 overturned Paleozoic strata in the footwall, with slip concentrated along a core of poorly exposed
2227 cataclasite. **F.** Photomicrograph of cataclasite from fault core shows grain size reduction and
2228 limited alteration.

2229
2230 Figure 28. Block diagrams illustrate relations of mesoscopic structures and AMS fabrics in red
2231 beds and thin limestone of the Triassic Chugwater Group and limestone in the Jurassic Sundance
2232 Formation (left), and progressive development of large-scale fold-fault relations (right) for (i)
2233 initial undeformed state, (ii) early LPS, vertical thickening (VT), tangential extension (TE), and
2234 minor wrench shear (Wsh) accommodated by conjugate wedge and tear faults, high-angle and
2235 cross-strike fracture sets, thin veins, and limited minor folds; and (iii) tilting of bedding and early
2236 faults along steep forelimbs with younger reverse- and strike-slip faults related to limb steepening
2237 and rotation during large-scale arch uplift.

2238
2239 Figure 29. Structural and timing relations for internal deformation in the Laramide belt, Garner
2240 Mountain anticline, southern Bighorn arch. Photos show anticline viewed looking NW and
2241 example of minor wedge faults. Cross section shows typical basement-cored fold with steep
2242 forelimb, and insets show *in situ* and restored minor fault geometries. Equal-area stereograms for
2243 sites CH160 and CH154 along the SW and NE fold limbs show: *in situ* and restored orientations of
2244 minor faults and slip lineations with best-fit σ_1 vectors (stars); *in situ* and restored AMS $K_{\max}$, K_{int} ,
2245 and $K_{\min}$ axes for individual cores and site mean directions; and restored paleomagnetic vectors for
2246 cores and site mean vectors with 95% confidence cones. Minor faults formed during early LPS and
2247 were tilted during large-scale folding. Modified from Weil and Yonkee (2012).

2248

Figure 30. **A.** Regional LPS directions estimated from minor fault data and AMS fabrics for structurally simple sites in the Laramide foreland range from SSW to WSW along varying trends of anastomosing basement-cored arches, with an average orientation of 240° . Data from the Wyoming salient also shown for comparison. **B.** Paleomagnetic site mean declinations relative to reference declination for the North America Triassic paleopole (with bedding tilt restored to horizontal) in the Laramide foreland record only limited, non-systematic block rotations.

Figure 31. Structural relations for the southern Big Horn arch. **A.** Simplified geologic map shows major fold and fault traces, regional arch trends, and tilt-corrected LPS (bedding restored to horizontal) directions estimated from AMS fabrics and minor fault data. **B.** Tilt-corrected site-mean paleomagnetic declinations (plotted relative to the reference North America Triassic paleopole, cones indicate 95% confidence intervals) indicate only limited rotations. **C.** Fully restored paleostress trajectories, based on pooled minor fault and AMS data corrected for bed tilting and vertical-axis rotations, range from overall SW trending on the western flank to overall WSW trending on the eastern flank, along with local refraction near the Tensleep fault and northern Casper arch. General E to NE trending patterns of Archean granite and gneiss are based on outcrop and geophysical data. Modified from Weil et al. (2014).

Figure 32. **A.** Idealized kinematic models for development of Laramide arches: (1) temporal changes in shortening directions from early E-W (blue arrows) to late N-S (red arrows); (2) bulk shortening with wrench shear along differently oriented zones; and (3) regional SW-NE shortening with local stress refraction and fault strike variation partly related to basement fabrics. Each model predicts different patterns of map-scale fault-fold systems and outcrop-scale internal deformation,

including LPS fabrics and vertical-axis rotations, which can be tested with paleomagnetic and LPS strike tests. **B.** Strike-test slope for red bed paleomagnetic declinations (relative to reference North America Triassic paleopole) is $-0.09 (\pm .11)$ at a 95% confidence interval), indicating no systematic vertical-axis rotation. Strike test slope for LPS directions estimated from minor fault data (relative to a reference direction of 240°) is $0.55 (\pm .10)$. Example site error bars, number of sites (N), and average residual (σ_r) listed. Modified from Weil and Yonkee (2012).

Figure 33. Regional maps of fully restored (corrected for estimated vertical-axis rotation) initial LPS and final (rotated) LPS directions in the Laramide and Sevier belts. Mesoscopic structural and AMS data for multiple sites in structural domains are combined to estimate a vector mean LPS direction. Stage 1- Early to mid Cretaceous emplacement of Willard-Paris-Meade sheet and initial LPS in future Crawford and Absaroka sheets. Stage 2- Late Cretaceous emplacement of Crawford and Absaroka sheets and initial LPS in future Hogsback-Darby-Prospect sheet of Sevier belt and early LPS in Laramide belt. Stage 3- Paleocene major uplift of Laramide arches and emplacement of Hogsback-Darby-Prospect sheet. Initial LPS directions in the Sevier belt were slightly dispersed about a regional W-E direction, and then underwent rotation during eastward propagation of the thrust wedge. Initial LPS directions in the Laramide refracted about a regional WSW-ENE direction, ranging from $\sim$ SSW-NNE to $\sim$ W-E along W- to N-trending arches. Mid Cretaceous to Paleocene paleo-velocity models of Bird (1998) also shown.

Figure 34. A. General model for an advancing type of subduction boundary where motion of the overriding plate is greater than the rate of trench retreat, resulting in intraplate shortening. Driving and resisting forces for subduction and kinematic relations of different velocity components

2295 schematically indicated (Royden and Husson, 2006; Doglioni et al., 2007; Lallemand et al., 2008).

2296 B. Model for partitioning of relative convergence between the Farallon and North American plates
2297 along the tectonic transect during the mid Cretaceous. Convergence is partitioned into slip along
2298 the subduction channel, transpression and strike-slip faulting across the magmatic arc, and retroarc
2299 shortening in the Sevier fold-thrust belt to hinterland.

2300

2301 Figure 35. Schematic relations between thrust wedge slope, curvature, and mechanics. **A.** Simple
2302 critical wedge model gives relations between surface slope (α), basal dip (β), and critical taper
2303 (ϕ_c), which depends on the relative strengths of the basal fault and wedge interior. **B.** Influence of
2304 stratigraphic thickness on wedge geometry, with thicker basin forming wider wedge. **C.** Influence
2305 of variations in lithology and fault strength, with weak basal fault favoring wider wedge, and
2306 decrease in fault strength in interior favoring very low taper. **D.** Influence of foreland arch that
2307 inhibits thrust slip and favors localized shear and out-of-sequence thrusting. **E.** Transmission of
2308 stress from a hinterland plateau and curved topographic slope help drive wedge shortening and
2309 curved thrust slip. **F.** Increased sedimentation favors propagation of wider spaced thrusts in a
2310 thicker wedge. **G.** Greater erosion favors closer spaced thrusts and deeper exhumation of a
2311 narrower wedge.

2312

2313 Figure 36. Controls of Laramide foreland deformation. **A.** Schematic lithospheric block diagram
2314 illustrates potential stress transfer mechanisms in the Laramide foreland related to basal shear
2315 traction, end loading, and hydrodynamic forces from enhanced asthenospheric flow. Potential for
2316 lower crustal detachment and lithospheric buckling schematically shown. **B.** General geotherms for
2317 pre- and syn-Laramide cratonic lithosphere, constrained by thermochronologic and xenolith data

(see text for details). C. Crustal strength model calculated using general geotherms and flow laws given in Kohlstedt et al. (1995), with strain rate of $\sim 10^{-14} \text{ s}^{-1}$.

Figure 37. Summary of time-space relations between tectonic components across the orogenic system (fore arc, magmatic arc, hinterland, Sevier belt, foreland). Upper plot shows schematic crustal cross section at 50 Ma near end of orogenesis. Lower plots show: plate motion rates (see Fig. 4); generalized forearc subsidence rates (Moxon and Graham, 1987); magma addition rates (Paterson and Ducea, 2015); generalized hinterland topographic evolution; long-term shortening rates for the hinterland, Sevier, and Laramide belts; and representative sedimentary thickness curves from the foreland (see Fig. 23).

Figure 38. Model for progressive evolution of the Wyoming salient of the Sevier belt and Laramide foreland belt. **A.** Early Jurassic ($\sim 180 \text{ Ma}$) primary sedimentary cover with passive margin and cratonic strata above basement with heterogeneous fabrics. **B.** Middle to Late Jurassic ($\sim 150 \text{ Ma}$)- shortening in the Luning-Fencemaker belt, limited deformation and pluton intrusion in the hinterland, and subsidence of the Morrison basin. **C.** Early Cretaceous ($\sim 100 \text{ Ma}$)- shortening and start of crustal thickening in the hinterland, emplacement of the Willard-Paris-Meade sheet (western thrust system) to form a topographically curved wedge front, early LPS in the future Crawford sheet, and sedimentation in the foreland basin. **D.** Mid Cretaceous ($\sim 80 \text{ Ma}$)- crustal thickening and uplift of the hinterland plateau, wedge propagation with emplacement of the Crawford sheet (eastern thrust system) along curved slip paths that produce rotation along the wedge front, and sedimentation in foredeep with increasing regional dynamic subsidence. **E.** Late Cretaceous ($\sim 70 \text{ Ma}$)- high-elevation hinterland plateau with lower crustal thickening and upper

2341 crustal synconvergent extension, emplacement of the Absaroka sheet (eastern thrust system) with
2342 rotation concentrated near the curved wedge front, early ~E-directed LPS in the future Hogsback-
2343 Darby-Prospect sheet along with overall ENE-directed LPS in the Laramide foreland, and
2344 sedimentation in basins partly disrupted by initial Laramide uplifts. **F.** Early Eocene (~50 Ma)-
2345 upper crustal extension and continued lower crustal metamorphism and weakening in the
2346 hinterland, final wedge propagation and emplacement of the Hogsback-Darby-Prospect sheet with
2347 localized tilting and wrench shear where the salient ends interacted with Laramide foreland uplifts,
2348 final rotation of LPS fabrics, sedimentation broken basins.
2349

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Figure

Figure

Figure

B

Figure

A Plate Reconstructions

B Plate Motions

C Subduction Ages

Figure

A Late Jurassic ~150 Ma

B Early Cretaceous ~100 Ma

C Late Cretaceous ~80 Ma

D Paleogene ~50 Ma

Figure

A Late Jurassic (~150 Ma)

B Early Cretaceous (~100 Ma)

C Late Cretaceous (~80 Ma)

D Paleogene (~50 Ma)

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E. Coniacian - Santonian ~ 89 - 83 Ma

F. early Campanian ~ 83 - 78 Ma

G. late Campanian ~ 78 - 72 Ma

H. Maastrichtian ~ 72 - 66 Ma

Figure

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Figure

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Figure

A

(1) Temporal Changes in Shortening Directions

(2) Bulk Shortening with Wrench Shear

(3) Regional SW-NE Shortening with Local Stress Refraction

Regional

Schematic Block

Map-Scale

Outcrop-Scale

Strike Tests

Figure A illustrates the evolution of shortening directions across different scales and scenarios. The diagram is organized into a grid with rows representing scales (Regional, Schematic Block, Map-Scale, Outcrop-Scale) and columns representing scenarios (1, 2, 3). The bottom row shows strike tests for various geological features.

Regional Scale:

- (1) Temporal Changes in Shortening Directions:** Shows early arches (blue) and late arches (red) with a north arrow.
- (2) Bulk Shortening with Wrench Shear:** Shows a complex pattern of shortening directions with purple and black arrows.
- (3) Regional SW-NE Shortening with Local Stress Refraction:** Shows a regional shortening direction (red dashed line) and local stress refraction (black arrows).

Schematic Block:

- (1) Late ~N-S:** Shows a block with early ~E-W shortening and late ~N-S shortening.
- (2) Bulk Shortening:** Shows a block with wrench shear and bulk shortening.
- (3) SW-NE Regional Shortening:** Shows a block with regional SW-NE shortening and local stress refraction.

Map-Scale:

- (1) Late ~N-S:** Shows a map view with early ~E-W shortening and late ~N-S shortening.
- (2) Bulk Shortening:** Shows a map view with wrench shear and bulk shortening, including en echelon folds.
- (3) SW-NE Regional Shortening:** Shows a map view with regional SW-NE shortening and local stress refraction.

Outcrop-Scale:

- (1) Late ~N-S:** Shows a map view with early ~E-W shortening and late ~N-S shortening, including LPS1 and LPS2.
- (2) Bulk Shortening:** Shows a map view with wrench shear and bulk shortening, including LPS1 and LPS2.
- (3) SW-NE Regional Shortening:** Shows a map view with regional SW-NE shortening and local stress refraction, including LPS1 and LPS2.

Strike Tests:

- Pmag strike test:** Shows a plot of Pmag strike test results for P-Pr, LPS, and L-Lr.
- P-Pr:** Shows a plot of P-Pr strike test results for P-Pr, LPS, and L-Lr.
- LPS strike test:** Shows a plot of LPS strike test results for P-Pr, LPS, and L-Lr.
- L-Lr:** Shows a plot of L-Lr strike test results for P-Pr, LPS, and L-Lr.

Figure

Figure

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