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LATE JURASSIC TO EOCENE EVOLUTION OF THE CORDILLERAN THRUST BELT AND FORELAND BASIN SYSTEM, WESTERN U.S.A.

P. G. DECELLES

Department of Geosciences, University of Arizona, Tucson, Arizona 85721 U.S.A.; decelles@geo.arizona.edu

ABSTRACT. Geochronological, structural, and sedimentological data provide the basis for a regional synthesis of the evolution of the Cordilleran retroarc thrust belt and foreland basin system in the western U.S.A. In this region, the Cordilleran orogenic belt became tectonically consolidated during Late Jurassic time (~155 Ma) with the closure of marginal oceanic basins and accretion of fringing arcs along the western edge of the North American plate. Over the ensuing 100 Myr, contractile deformation propagated approximately 1000 kilometers eastward, culminating in the formation of the Laramide Rocky Mountain ranges. At the peak of its development, the retroarc side of the Cordillera was divided into five tectonomorphic zones, including from west to east the Luning-Fencemaker thrust belt; the central Nevada (or Eureka) thrust belt; a high-elevation plateau (the "Nevadaplano"); the topographically rugged Sevier fold-thrust belt; and the Laramide zone of intraforeland basement uplifts and basins. Mid-crustal rocks beneath the Nevadaplano experienced high-grade metamorphism and shortening during Late Jurassic and mid- to Late Cretaceous time, and the locus of major, upper crustal thrust faulting migrated sporadically eastward. By Late Cretaceous time, the middle crust beneath the Nevadaplano was experiencing decompression and cooling, perhaps in response to large-magnitude ductile extension and isostatic exhumation, concurrent with ongoing thrusting in the frontal Sevier belt. The tectonic history of the Sevier belt was remarkably consistent along strike of the orogenic belt, with emplacement of regional-scale Proterozoic and Paleozoic mega-thrust sheets during Early Cretaceous time and multiple, more closely spaced, Paleozoic and Mesozoic thrust sheets during Late Cretaceous–Paleocene time. Coeval with emplacement of the frontal thrust sheets, large structural culminations in Archean-Proterozoic crystalline basement developed along the basement step formed by Neoproterozoic rifting. A complex foreland basin system evolved in concert with the orogenic wedge. During its early and late history (~155 - 110 Ma and ~70 - 55 Ma) the basin was dominated by nonmarine deposition, whereas marine waters inundated the basin during its midlife (~110 - 70 Ma). Late Jurassic basin development was controlled by both flexural and dynamic subsidence. From Early Cretaceous through early Late Cretaceous time the basin was dominated by flexural subsidence. From Late Cretaceous to mid-Cenozoic time the basin was increasingly partitioned by basement-involved Laramide structures. Linkages between Late Jurassic and Late Cretaceous Cordilleran arc-magmatism and westward underthrusting of North American continental lithosphere beneath the arc are not plainly demonstrable from the geological record in the Cordilleran thrust belt. A significant lag-time (~20 Myr) between shortening and coeval underthrusting, on the one hand, and generation of arc melts, on the other, is required for any linkage to exist. However, inferred Late Jurassic lithospheric delamination may have provided a necessary precondition to allow relatively rapid Early Cretaceous continental underthrusting, which in turn could have catalyzed the Late Cretaceous arc flare-up.

INTRODUCTION

The Cordilleran orogenic belt of North America extends for more than 6,000 kilometers from southern Mexico to the Canadian Arctic and Alaska (fig. 1), forming a

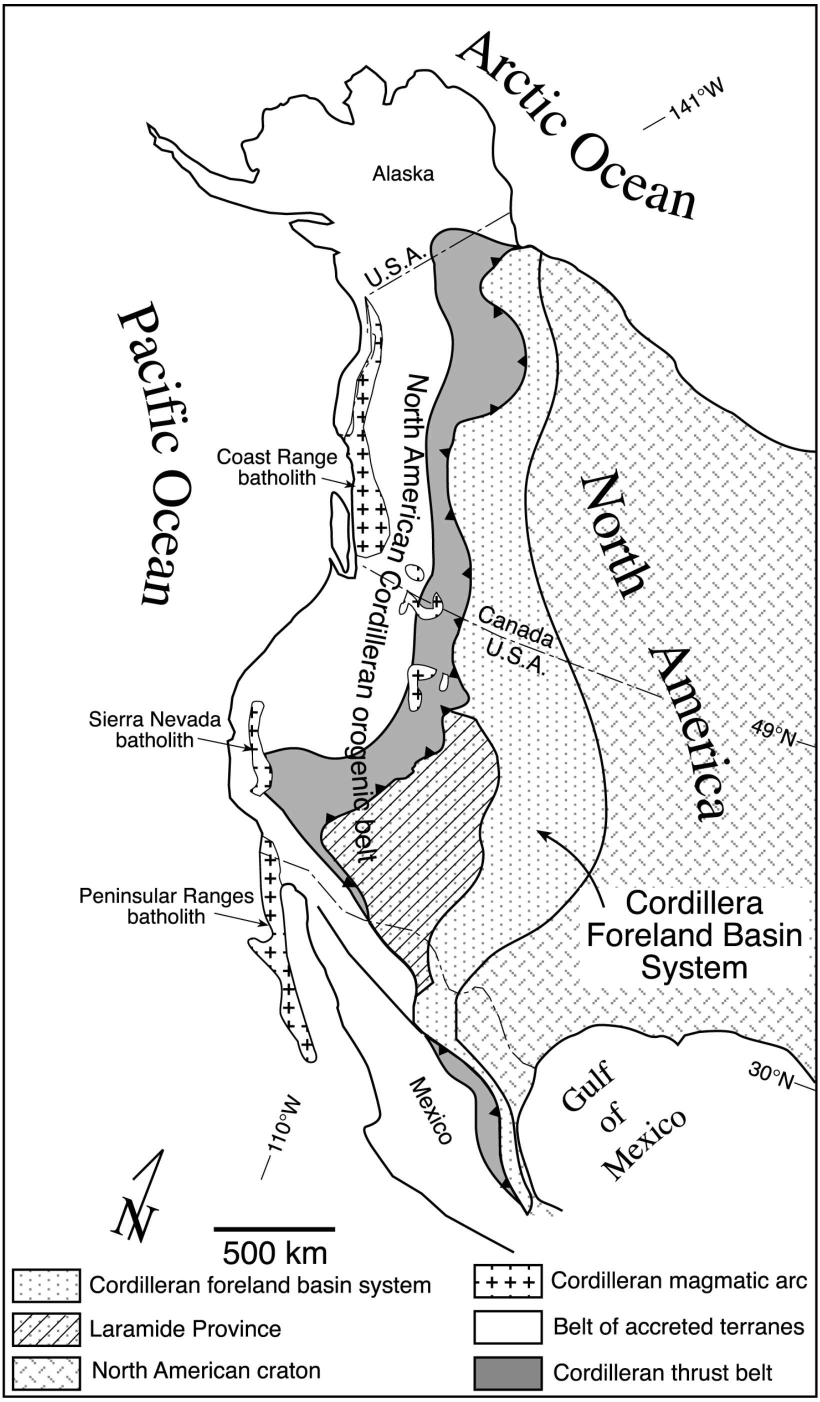

Fig. 1. Generalized tectonic map of western North America, showing the major geologic zones of the Cordilleran orogenic belt and foreland basin system. After Tipper and others (1981) and Coney and Evenchick (1994).

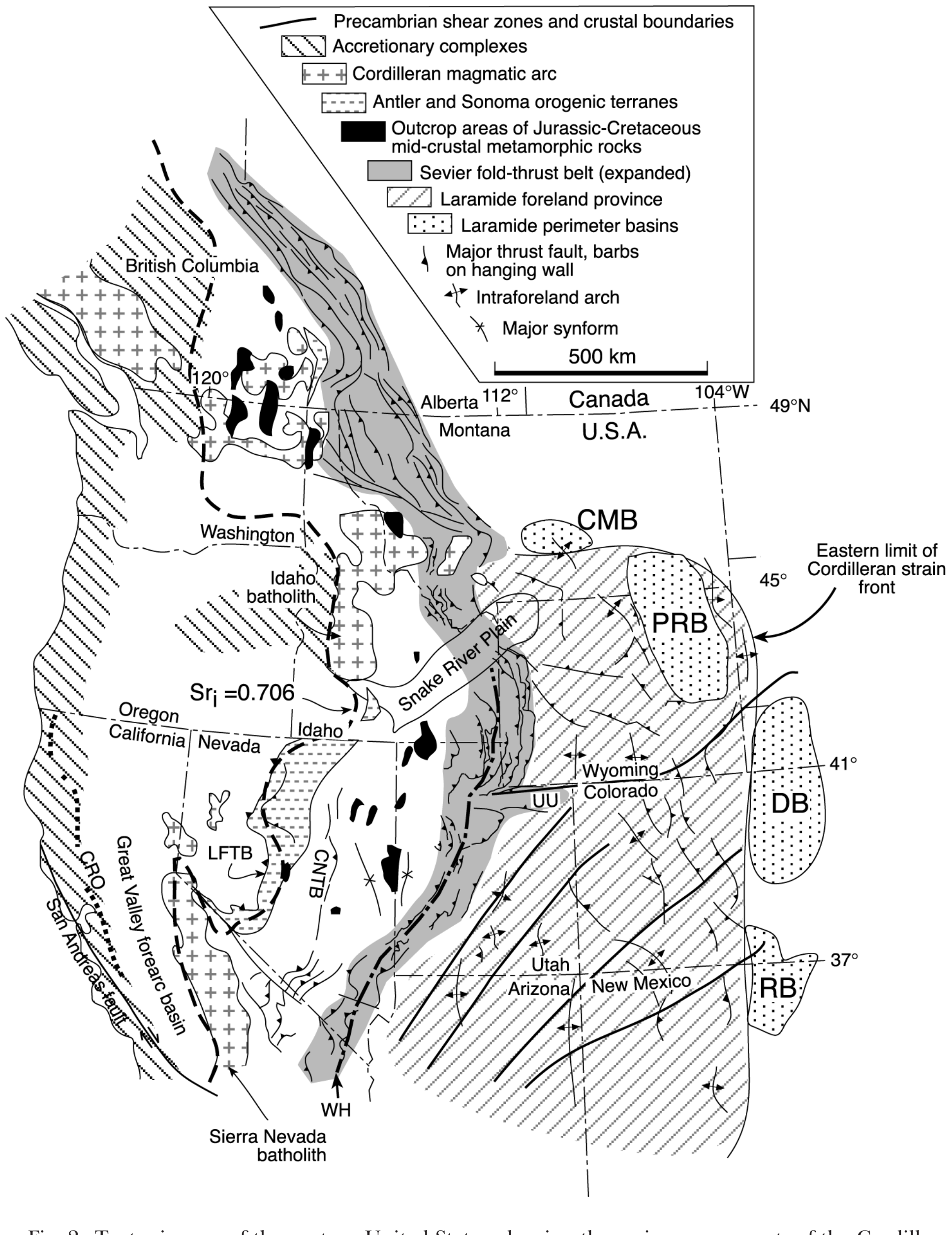

Fig. 2. Tectonic map of the western United States, showing the major components of the Cordilleran orogenic belt. The initial Sr ratio line is taken to represent the approximate western edge of North American cratonic basement (Armstrong and others, 1977; Kistler and Peterman, 1978). Abbreviations as follows: CRO, Coast Range ophiolite; LFTB, Luning-Fencemaker thrust belt; CNTB, Central Nevada thrust belt; WH, Wasatch hinge line; UU, Uinta Mountains uplift; CMB, Crazy Mountains basin; PRB, Powder River basin; DB, Denver basin; RB, Raton basin. Precambrian shear zones after Karlstrom and Williams (1998).

major segment of the Circum-Pacific orogenic belt (Dickinson, 2004). The best known part of this great continental orogen lies in the western interior U.S.A. and southwestern interior Canada, between the latitudes of 36°N and 51°N, where it reaches its maximum width of >1000 kilometers (fig. 2). Constructed for the most part during mid-Mesozoic–Eocene time and subsequently modified by mid- to late Cenozoic extension, the Cordillera is viewed as the archetypal example of an ancient mountain belt formed between converging oceanic and continental plates (Dewey and Bird, 1970; Jordan and Allmendinger, 1986; Coney and Evenchick, 1994). The Cordilleran orogen began to develop as a coherent geodynamic system and topographic edifice during the Late Jurassic in response to subduction of oceanic plates of the Pacific domain beneath the North American continental plate (for example, Monger and Price, 1979; Burchfiel and others, 1992). Contemporaneously, an immense foreland basin that ultimately occupied an area of $>5 \times 10^6$ km^2 began to develop to the east of the orogenic belt (for example, Kauffman and Caldwell, 1993). The orogenic belt and foreland basin evolved together until early Cenozoic time in various types of process-response and feedback relationships. In terms of its area and duration of development ($\sim$100 Myr), no other mountain belt-foreland basin couplet equals the North American Cordilleran system.

Many reviews of the structural, metamorphic, and igneous evolution of various parts of the Cordilleran orogen and the adjacent foreland basin system in the western U.S.A. have been published (for example, Wiltschko and Dorr, 1983; Cross, 1986; Snoke and Miller, 1988; Oldow and others, 1989; Elison, 1991; Burchfiel and others, 1992; Saleeby and Busby-Spera, 1992; Cowan and Bruhn, 1992; D. M. Miller and others, 1992; Allmendinger, 1992; D. L. Smith and others, 1993; Armstrong and Ward, 1993; Lawton, 1994; Lageson and Schmitt, 1994; Ingersoll, 1997; Dickinson, 2001, 2004). These syntheses generally have dealt with either the foreland region or the orogenic belt, leaving many questions regarding potential connections between the two regions unanswered. In addition, most of these reviews are based on the state of knowledge that existed prior to recent breakthroughs in geochronology and thermochronology in the Cordilleran orogen (particularly in its hinterland in Nevada and western Utah), as well as many new contributions on the development of the foreland basin system to the east. Thus, the potential linkages between structural, metamorphic, and erosional processes in the orogenic belt and depositional processes in the foreland basin system have not been recognizable until relatively recently. Important issues remain to be settled in Cordilleran geology, including (1) the timing of onset and regional pattern of crustal shortening in the retroarc region, (2) the nature of the relationship between crustal thickening and extension in the hinterland and frontal (Sevier belt) thrusting, (3) the driving mechanism(s) of subsidence in the retroarc foreland basin system, (4) the degree of temporal and spatial continuity in orogenesis in the Cordilleran thrust belt, and (5) the relationships between retroarc thrusting and arc magmatism. The purpose of this paper is to synthesize the major structural features and kinematic history of the Cordilleran orogenic belt and to assess its paleogeographic development in the context of the large-scale evolution of the Cordilleran foreland basin system. A time-slice, palinspastic, map-view approach is adopted in order to convey the broadest aspects of the paleogeography and regional kinematic and subsidence histories. The picture that emerges is one of broadly continuous, but unsteady, eastward expansion of the orogenic deformation front from western Nevada to central Colorado over an $\sim$100 Myr time period during Late Jurassic to Eocene time. This eastward expansion of the orogenic belt was accomplished by structurally diverse mechanisms that depended in large part on the pre-existing tectonostratigraphic architecture of the western U.S.A. and on the types of rocks involved. Abrupt, long-distance, west-to-east sweeps in the location of the eastern front of the Cordilleran magmatic arc left a strong imprint on upper crustal rocks and serve as a proxy indicator for possible upper mantle processes that influenced the retroarc region. The foreland basin system evolved in response to both crustal and mantle processes.

DATA SOURCES

The data necessary to constrain reconstructions of the Cordilleran orogenic belt are derived from many local and regional, structural and kinematic studies. Ages of important structural features are based on radiometric ages (K-Ar, $^{40}Ar/^{39}Ar$, and U-Pb) of cross-cutting and overlapping igneous rocks; radiometric ages of metamorphism and estimated P-T-t paths; radiometric and biostratigraphic ages of deformed and overlapping sedimentary units; radiometric cooling ages ($^{40}Ar/^{39}Ar$ and fission track) by which tectonic events may be reasonably inferred; datable growth strata; and regional sedimentary provenance data (fig. 3, table 1). The ages of foreland basin stratigraphic units are based on biostratigraphy and, to a lesser degree, radiometric ages of intercalated volcanogenic deposits. These data sets are now substantial in terms of both quantity and quality. Previous local and regional syntheses provide numerous geochronological and thermochronological compilations (for example, Allmendinger and Jordan, 1981; Allmendinger and others, 1984; Perry and others, 1988; Coogan and Royse, 1990; Barton, 1990, 1996; Saleeby and Busby-Spera, 1992; Burchfiel and others, 1992; D. L. Smith and others, 1993; Royse, 1993a; Armstrong and Ward, 1993; Christiansen and others, 1994; Burtner and Nigrini, 1994; Walker and others, 1995; Constenius, 1996; Burchfiel and others, 1998). Thicknesses, paleoflow directions, and provenances of foreland stratigraphic units are compiled from existing syntheses (for example, Suttner, 1969; Furer, 1970; McGookey, 1972; Jordan, 1981; Suttner and others, 1981; Wiltschko and Dorr, 1983; Schwartz and DeCelles, 1988; DeCelles and Burden, 1992; Cobban and others, 1994; Robinson Roberts and Kirschbaum, 1995; Dickinson and Lawton, 2001a; Currie, 2002) and modified to honor newer data.

PLATE TECTONIC SETTING

Evolution of the Cordilleran orogenic belt and foreland basin took place against the backdrop of an opening North Atlantic Ocean and coeval subduction of oceanic plates beneath the western margin of the North American plate. The seafloor spreading rate in the North Atlantic was nearly constant at ~12 to 16 mm/yr during Middle Jurassic through mid-Cretaceous time, but increased to ~30 mm/yr during Late Cretaceous time (~95 Ma, fig. 4A; Sclater and others, 1981; Müller and others, 1997). Although subject to considerable uncertainty, the rate of Farallon-North America convergence generally increased from ~8 mm/yr in Late Jurassic time to 150 mm/yr by Paleocene time, with abrupt increases during Early, mid-, and latest Cretaceous time (fig. 4B; Engebretson and others, 1984, 1985; Page and Engebretson, 1984).

During Late Jurassic time the southwestern margin of North America was nearly tectonically neutral, with regional transtension dominating the plate margin and the proximal retroarc region (Saleeby, 1992; Saleeby and Busby-Spera, 1992; Dickinson and Lawton, 2001a). That the Cordillera initially began to consolidate as a coherent, high-elevation orogenic belt in Late Jurassic time (Coney and Evenchick, 1994) is implied by the following facts:

(1) The forearc region in California evolved from a belt of multiple fringing arcs and interarc oceanic basins in Early–Middle Jurassic time to a single, coherent arc-trench system during Late Jurassic time (Harper and Wright, 1984; Wright and Fahan, 1988; Saleeby and Busby-Spera, 1992; Dickinson and others, 1996). Middle Jurassic (~165 ± 5 Ma) volcanic and sedimentary rocks inferred to have formed in an east-facing intraoceanic arc and westward trailing interarc oceanic ophiolite (Coast Range Ophiolite; fig. 2) (Schweickert and others, 1984; Ingersoll and Schweickert, 1986) or in a backarc basin inboard of an offshore island arc (Harper and Wright, 1984; Wright and Fahan, 1988; Saleeby, 1992) were emplaced in the Sierran foothills and California Coast Ranges by ~155 Ma during the Nevadan orogeny. Although the origin of the Coast Range ophiolite remains debatable (see Saleeby, 1992; Dickinson and others, 1996), most workers seem to agree that it consists of oceanic crust and deep

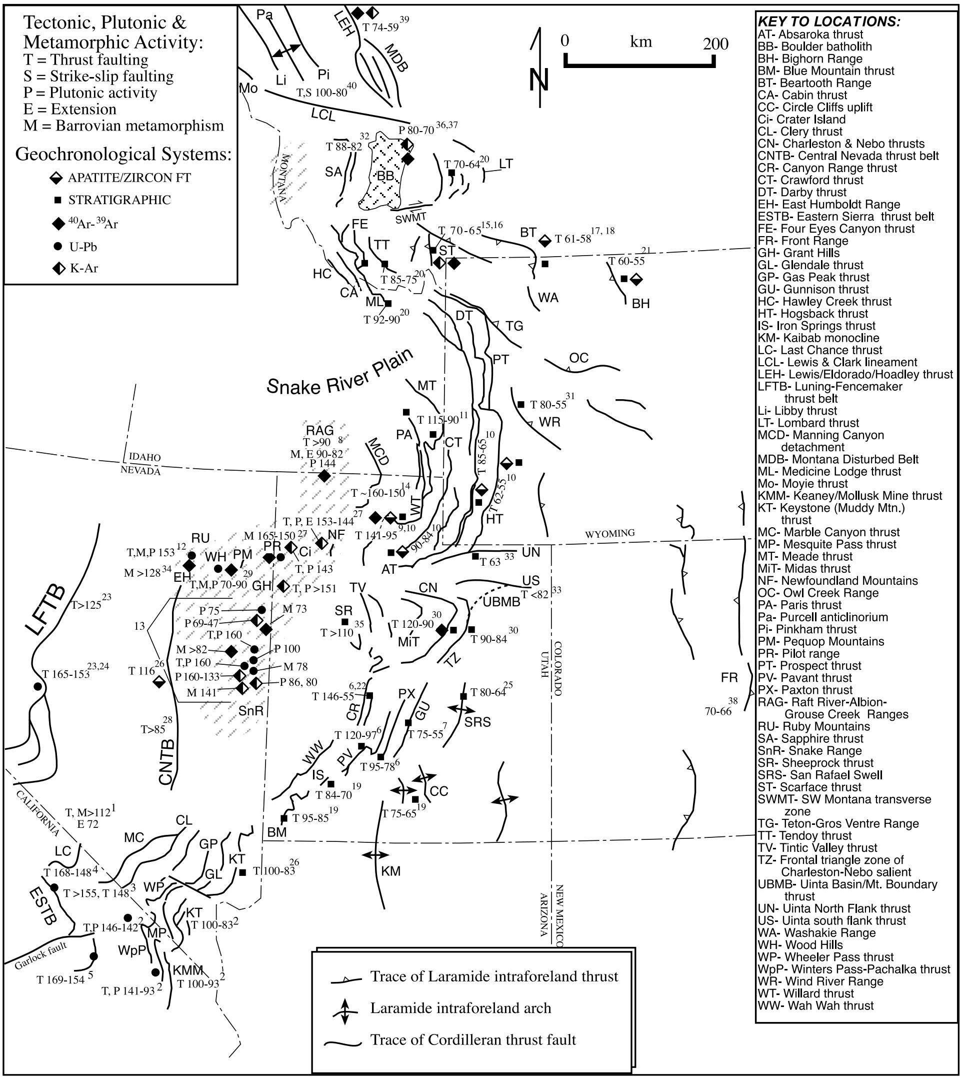

Fig. 3. Map showing major structures of the Cordilleran thrust belt and locations for which geochronologic, thermochronologic, and provenance data indicate timing of tectonic, plutonic, and metamorphic activity. Sources of data are signified by superscripts and listed in table 1. In the hinterland (western Utah and Nevada) only selected representative geochronologic data are shown for purposes of clarity. Cross-hatched areas in Nevada, northwestern Utah, and western Montana are regions of Late Jurassic–Cretaceous high-grade metamorphism.

marine sediments that formed outboard (and possibly south) of the California margin during Middle Jurassic time (165 - 156 Ma). This model requires the existence of at least one additional oceanic (micro?)plate along the California margin during the Middle Jurassic (for example, the "Mezcalera plate" of Dickinson and Lawton, 2001b), or a marginal oceanic basin (Harper and Wright, 1984; Wright and Fahan, 1988; Saleeby, 1992; Saleeby and Busby-Spera, 1992). Thus, the Farallon plate could not have been subducting beneath a coherent western North American plate margin until after the Coast Range ophiolite and its associated intraoceanic arc were accreted onto the continental margin at ~155 Ma. High-pressure metamorphism in the Franciscan

TABLE 1

Key to references in figure 3

1. Applegate and Hodges (1995)
2. Walker and others (1995)
3. Wrucke and others (1995)
4. Dunne and Walker (1993)
5. Walker and others (1990)
6. DeCelles and others (1995)
7. Lawton and others (1993)
8. Wells and others (1990)
9. Yonkee (1992)
10. DeCelles (1994); Burtner and Nigrini (1994)
11. DeCelles and others (1993)
12. Hudec (1992)
13. Miller and others (1988)
14. Allmendinger and Jordan (1981)
15. Tysdal (1986)
16. DeCelles and others (1987)
17. Giegengack and others (1998)
18. DeCelles and others (1991)
19. Goldstrand (1994)
20. Schmitt and others (1995)
21. Hoy and Ridgway (1997)
22. Stockli and others (2001)
23. Wyld (2002)
24. Elison (1991)
25. Lawton (1983, 1986)
26. Carpenter and others (1993)
27. Allmendinger and others (1984)
28. Taylor and others (2000)
29. Camilleri and others (1997)
30. Constenius (ms, 1998); Mitra (1997)
31. Shuster and Steidtmann (1988); Dorr and others (1977)
32. Ruppel and others (1981)
33. Bryant and Nichols (1988)
34. Dallmeyer and others (1986)
35. Mitra (1997)
36. Tilling and others (1968)
37. Robinson and others (1968)
38. Kluth and Nelson (1988)
39. Sears (2001)
40. Wallace and others (1990)

subduction complex was coeval with all of the tectonic activity discussed above, suggesting that the transition from a diffuse to coherent plate margin occurred over a roughly 10 Myr timespan (Dickinson and others, 1996).

(2) The Great Valley forearc basin in central California was initiated at ~150 to 155 Ma and continued to subside and accumulate great thicknesses of arc- and subduction complex-derived sediment until Late Cretaceous time (Dickinson and Seely, 1979; Ingersoll, 1983). Therefore, it may be reasonably inferred that the Cordilleran forearc region was established as a coherent tectonic domain no earlier than Late Jurassic time.

(3) Prior to Late Jurassic time, strata in the retroarc region (mainly Montana, Idaho, Wyoming, Utah, northern Arizona, and western Colorado) contain no evidence of *regional* scale thrusting and erosion of orogenic terranes to the west; stratigraphic

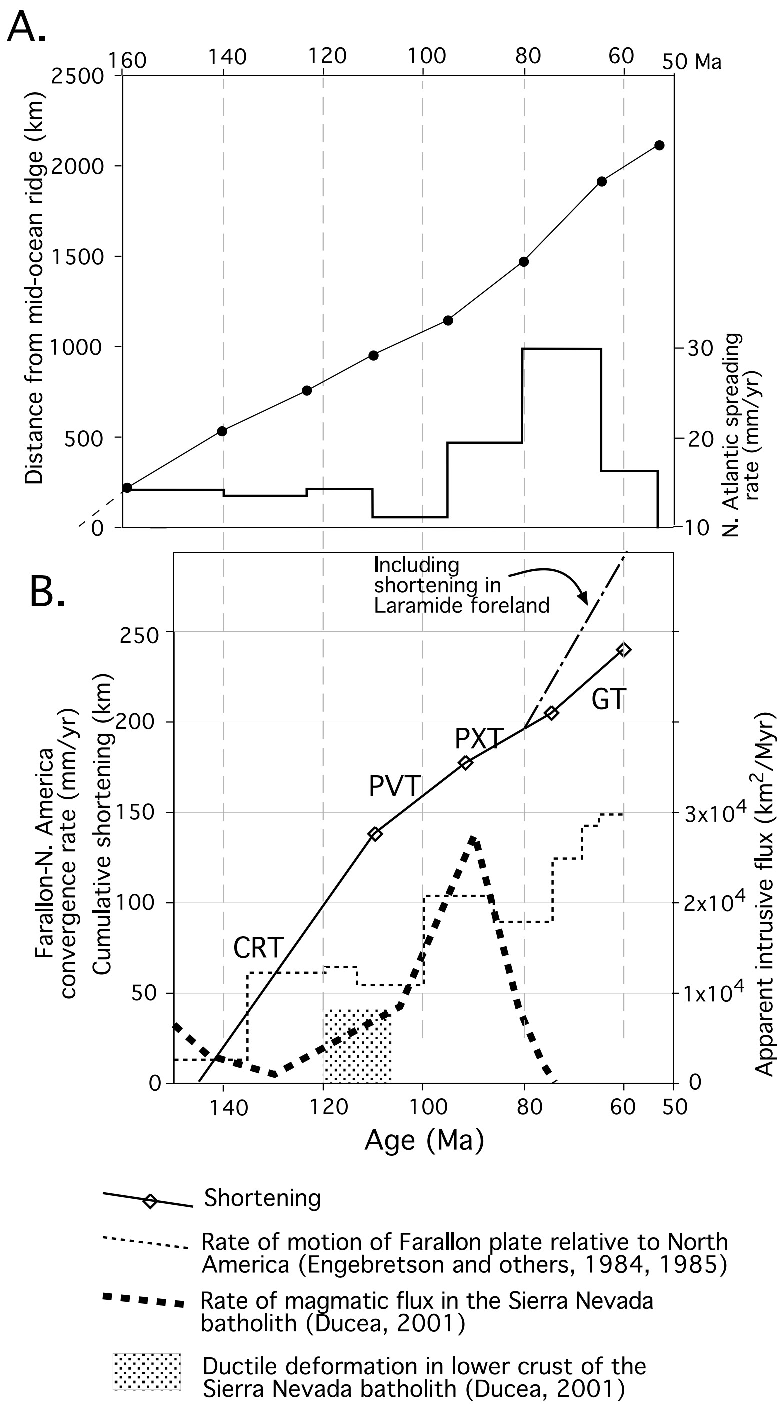

Fig. 4. (A) Time-displacement curve (upper curve) and average rate of seafloor spreading (lower step function) in the central part of the North Atlantic Ocean from Late Jurassic to early Cenozoic time, after Sclater and others (1981) and Müller and others (1997). (B) Comparison of the cumulative shortening in the central Utah-Nevada Sevier belt (DeCelles and others, 1995; Currie, 2002) and Laramide province (Stone, 1993; Erslev, 1993) with rates of Farallon-North America plate convergence (Engebretson and others, 1984, 1985) and rates of magmatic flux in the Sierra Nevada batholith (Ducea, 2001). Major thrust systems in the Sevier belt are abbreviated as follows: CRT, Canyon Range thrust; PVT, Pavant thrust; PXT, Paxton thrust; GT, Gunnison thrust.

thickness trends in the retroarc region are difficult to reconcile with a single, coherent orogenic belt to the west. Nevertheless, thrusting in western Nevada was intense (Wyld, 2002) and Callovian marginal marine rocks in Utah, Idaho, and southwestern Montana locally contain volcaniclastic and sedimentaclastic detritus, indicating that Cordilleran magmatic centers and topographic highs were active and supplying detritus to the retroarc region (for example, Jordan, 1985; Meyers and Schwartz, 1994; Peterson, 1994). Moreover, thickness trends of Early to Middle Jurassic strata in Utah can be reconciled with a flexural subsidence mechanism (Bjerrum and Dorsey, 1995; Allen and others, 2000). However, Early to Middle Jurassic sedimentary rocks in Utah also have been explained as the result of regional extension and/or dynamic subsidence (for example, Marzolf, 1994; Lawton, 1994), which is consistent with a low-standing extensional magmatic arc along the southwestern margin of North America (Busby-Spera, 1988; Saleeby and Busby-Spera, 1992), local evidence for normal faulting in western Utah (Miller and Allmendinger, 1991), and widespread alkaline and mafic magmatism in the retroarc region (Barton, 1996). Plate convergence directions during Early to Middle Jurassic time were highly oblique along the southwestern margin of the continent (for example, Saleeby and Busby-Spera, 1992). In view of the absence of clear-cut evidence for tectonic consolidation of the California margin prior to Late Jurassic time, the Early to Middle Jurassic retroarc strata do not support the existence of a regional-scale, integrated foreland basin system (Lawton, 1994). Once the Farallon-North America subduction system was fully established in Late Jurassic time, it remained active until mid-Cenozoic time, when the East Pacific Rise began to interact with the western margin of North America.

COMPONENTS OF THE CORDILLERAN OROGENIC BELT IN THE WESTERN U.S.A.

The Cordilleran orogenic belt comprises numerous geometrically and stratigraphically distinct components. These components impart diverse characteristics to the geometry and kinematic history of various sectors of the orogenic belt.

Proterozoic Embayments

Middle to early Late Proterozoic strata of the Cordillera in the western U.S.A. can be divided into two successions with different ages and regional distributions (Link and others, 1993). Middle Proterozoic (~1700 - 1200 Ma) strata are confined to basins in western Montana, Idaho, and Arizona. Middle to early Late Proterozoic (~1200 - 780 Ma) strata are preserved in local depocenters in Utah, southern California, and Arizona. At least two of these locally confined accumulations of Precambrian strata were reactivated and structurally inverted during Cordilleran thrusting, such that later thrust geometries were controlled by Precambrian structural-stratigraphic trends. In Montana, southern Alberta, and southeastern British Columbia a thick succession of Middle Proterozoic clastic and carbonate strata in the Belt-Purcell Supergroup defines the paleogeographic extent of the Belt Basin, a pronounced embayment along the northwestern margin of the North American craton (Harrison, 1972; Cressman, 1989). Middle to early Late Proterozoic strata of the Uinta Mountain Group and the Big Cottonwood Formation accumulated in northern Utah along an east-west trending aulacogen (Bryant and Nichols, 1988; Levy and Christie-Blick, 1991). In Montana, the Helena thrust salient formed as a result of inversion of the eastern part of the Belt embayment (Schmidt and O'Neill, 1982), and in Utah the Uinta Mountains uplift resulted from inversion of the Uinta basin fill (fig. 5; Bryant and Nichols, 1988; Constenius, ms 1998, 1999).

Neoproterozoic-Early Paleozoic Miogeocline

The Cordilleran orogenic belt was largely constructed of sedimentary and metasedimentary rocks that accumulated along the rifted western margin of Laurentia during Neoproterozoic– early Paleozoic time (Stewart, 1972, 1976; Christie-Blick, 1982,

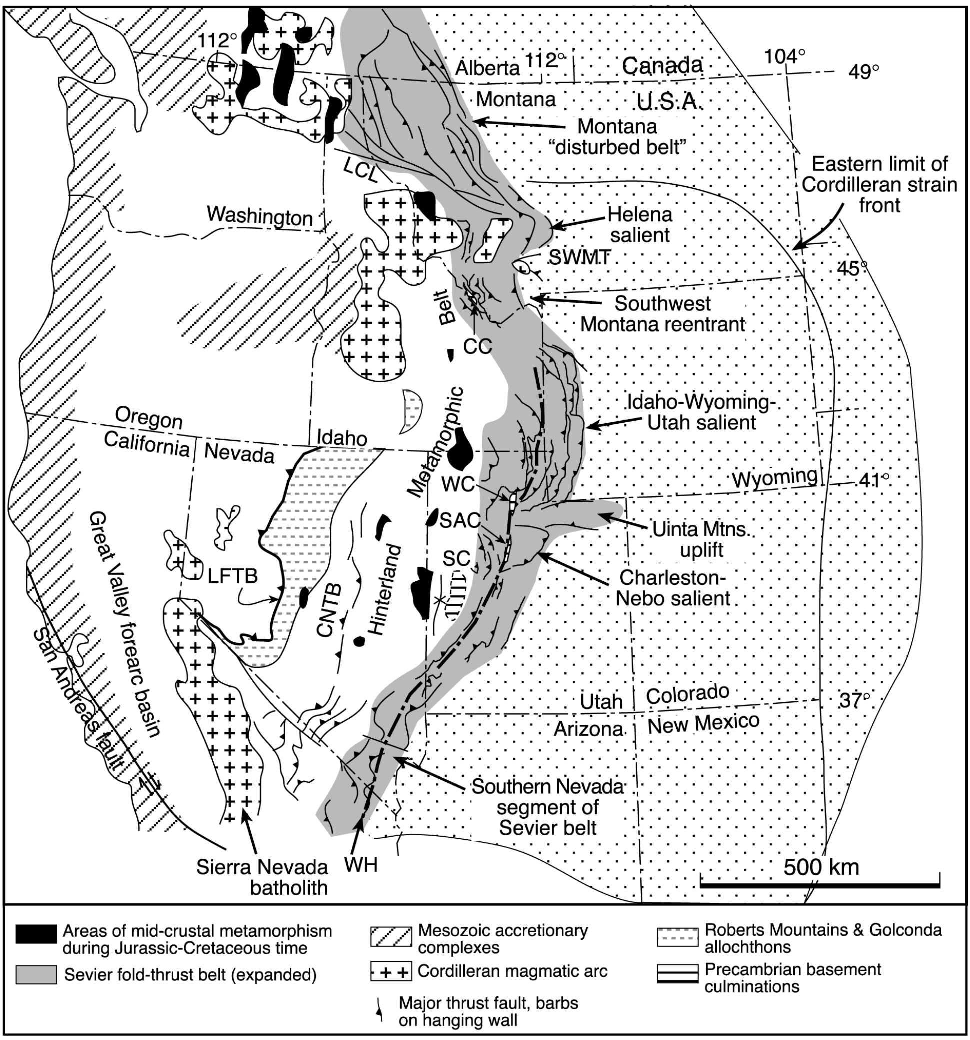

Fig. 5. Simplified version of figure 2, showing some of the major tectonic features in the Cordilleran thrust belt discussed in the text. Abbreviations as follows: LCL, Lewis and Clark line; SWMT, Southwest Montana transverse zone; CC, Cabin culmination; WC, Wasatch culmination; SAC, Santaquin culmination; SC, Sevier culmination; CNTB, Central Nevada thrust belt; LFTB, Luning-Fencemaker thrust belt; WH, Wasatch hinge line. Stippled region represents Cordilleran foreland basin system.

1997; Bond and others, 1985; Levy and Christie-Blick, 1991; Poole and others, 1992; Link and others, 1993; Timmons and others, 2001). Many of the details of Cordilleran thrust belt structural geometry can be attributed to local patterns of thickness and lithofacies in these rocks (for example, Royse, 1993a). The Neoproterozoic–early Paleozoic rocks were deposited in predominantly marine environments, and they become thicker and of deeper marine character westward from the Wasatch hinge line (fig. 5), which marks the approximate eastern limit of significant rifting in Precambrian crystalline basement (Hintze, 1988; Poole and others, 1992). The Proterozoic succession ranges in thickness from ~4 kilometers to >10 kilometers and mainly consists of siliciclastic rocks with subordinate carbonate rocks (Link and others, 1993). In outcrop, these strata are confined to thrust sheets (significantly dismembered by

Cenozoic normal faults) in the frontal Cordilleran thrust belt, and with the exceptions of the embayments discussed above, Precambrian strata were not preserved east of the Wasatch hinge line. The Paleozoic succession comprises up to ~12 kilometers of predominantly carbonate rocks in western Utah and Nevada. East of the Wasatch hinge line the Paleozoic succession is only ~1.5 kilometers thick (Hintze, 1988). The generally north-south oriented prismatic body of Paleozoic strata is disrupted in Utah and southern Idaho by anomalously thick accumulations of Pennsylvanian-Permian carbonate and clastic sediment in basins associated with the Ancestral Rocky Mountains orogenic event. With respect to the Cordilleran orogenic belt, the two most significant of these basins are the Oquirrh and Wood River basins in central and northwestern Utah (E. L. Miller and others, 1992). The Oquirrh basin was structurally inverted during Cordilleran thrusting (Royse, 1993a; Constenius, ms 1998, 1999).

Middle Paleozoic and Early Mesozoic Orogenic Terranes

The nature and kinematic evolution of the Cordilleran retroarc fold-thrust belt were influenced by crustal elements that formed during the late Devonian-early Mississippian Antler and late Permian-early Triassic Sonoma orogenies (Speed, 1977; Oldow, 1984; E. L. Miller and others, 1992; Poole and others, 1992; Burchfiel and others, 1992; Dickinson, 2000). Both orogenic belts can be traced from southwestern Nevada to southern Idaho, and similar rocks are present in northern Washington and British Columbia (fig. 2; M. T. Smith and others, 1993; Dickinson, 2004). Where most extensively exposed in Nevada, the Antler and Sonoma terranes consist of, respectively, the Roberts Mountains and Golconda allochthons. The Roberts Mountains allochthon is composed of Cambrian-Devonian chert, dominantly fine-grained siliciclastic rocks, arkosic sandstone, and local metabasaltic rocks that formed in deep-water, oceanic environments. During the early Mississippian, these rocks were transported eastward at least 200 kilometers on the Roberts Mountains thrust system and juxtaposed with Late Devonian shelf facies and older rocks (Roberts and others, 1958; Nilsen and Stewart, 1980; Poole and others, 1992; Burchfiel and others, 1992). A thick foreland basin system developed coevally to the east of the orogenic belt (Poole and Sandberg, 1977; Harbaugh and Dickinson, 1981; Speed and Sleep, 1982; Silberling and others, 1997). Post-tectonic, Upper Mississippian-Pennsylvanian shallow marine strata overlapped the Roberts Mountains allochthon.

The Sonoman orogenic belt bears many similarities to the Antler orogenic belt. It also involved the eastward transport and emplacement of regional thrust sheets (which form the Golconda allochthon) consisting of fine-grained deep marine sedimentary and volcanic rocks on top of coeval shallow marine facies (Speed, 1978; Oldow, 1983, 1984; Oldow and others, 1984, 1990; Wyld, 2002).

Burchfiel and Royden (1991), Burchfiel and others (1992), and Dickinson (2000) have suggested that the Antler and Sonoma orogenic belts formed by subduction zone rollback, similar to 'retreating' subduction zones of the central Mediterranean region (Royden, 1993). As envisaged by these authors, the hinges of westward dipping subducting oceanic plates rolled eastward toward the North American craton, accompanied by the development of the eastward directed Roberts Mountains and Golconda thrust belts. Coevally, deep marine basins formed along the trailing western sides of each thrust belt, analogous to the Tyrrhenian Sea basin along the western flank of the Apennines thrust belt, which has developed during Neogene southeastward rollback of the Ionian oceanic slab in the central Mediterranean (for example, Malinverno and Ryan, 1986). The Paleozoic orogenic terranes in central Nevada are significant for understanding the younger Cordilleran orogenic belt and foreland basin system because they must have influenced the thickness and structure of the lithosphere prior to the onset of mid-Mesozoic orogeny, and these old orogenic highlands could have provided some of the lithic-rich sediment in the Late Jurassic–Cretaceous foreland basin system to the east. Nevertheless, both the Antler and Sonoma belts were overlapped by post-orogenic, shallow-marine shelf strata (see summary in Dickinson, 2000), indicating that crustal thickness was probably not much greater than normal immediately prior to the onset of regional shortening in the Late Jurassic.

Triassic-Middle Jurassic Succession

From Triassic time onward, depositional environments of Cordilleran strata became dominantly shallow marine to nonmarine. Eolianites, fluvial sandstones and mudrocks, and evaporites are the dominant lithologies in the frontal part of the Cordilleran thrust belt; and shallow marine carbonate is prevalent in eastern Nevada and western Utah (Blakey, 1994; Peterson, 1994). In the western hinterland of the Cordillera, locally deep marine basins developed to the west of the Sonoma orogenic belt (Elison and Speed, 1989; Oldow and others, 1990). Jurassic evaporites are particularly important in controlling local structures in the frontal part of the Cordilleran thrust belt (Royse and others, 1975; Standlee, 1982; Lamerson, 1982; Coogan and Royse, 1990).

Cordilleran Magmatic Arc

The Cordilleran magmatic arc is a defining feature of the Cordilleran orogenic belt (fig. 1). This >4000 kilometer long belt of calc-alkaline granitoid (dominantly granodiorite and tonalite) intrusions formed above the eastward dipping subduction zone along the western margin of the North American continent from Late Triassic to Late Cretaceous time (Hamilton and Myers, 1967; Hyndman, 1983; Bateman, 1983; Saleeby and Busby-Spera, 1992; Armstrong and Ward, 1993; Andronicus and others, 2003). In the western U.S.A. arc-related rocks can be traced almost continuously from the Peninsular Ranges batholith in Baja to the Idaho batholith in north-central Idaho and western Montana (fig. 2). The character of the arc is exemplified by the Sierra Nevada batholith in central and southern California. Present crustal thickness of the batholith is ~35 to 42 kilometers (Fliedner and Ruppert, 1996; Wernicke and others, 1996). Ducea and Saleeby (1998) and Ducea (2001, 2002) argued on the basis of petrologic, geochemical, and geophysical data that at the time of cessation of magmatism (~80 Ma) the Sierra Nevada batholith must have consisted of an upper, ~30 to 35 kilometer thick zone of granitic material and a >70 kilometer thick eclogitic "root," composed of mafic-ultramafic cumulates and residues. This root formed by mafic underplating, crustal underthrusting from the east, or a combination of these processes (Ducea, 2001). The dense eclogitic root is proposed to have delaminated during Late Miocene time (Ducea and Saleeby, 1998). The possibility that westward underthrusting of North American crustal basement beneath the Cordilleran magmatic arc produced much of the eclogitic root provides a rationale to search for physical and temporal linkages between crustal shortening in the retroarc region and magmatic activity in the arc.

Radiometric ages of Sierra Nevada plutonic rocks span the interval ~220 Ma to 80 Ma, but are clustered during two distinct time intervals, 160 to 150 Ma and 100 to 85 Ma, with a major intervening lull in activity (fig. 4; Bateman, 1983, 1992; Barton and others, 1988; Armstrong and Ward, 1993; Barton, 1996; Ducea, 2001). The 100 to 85 Ma event accounts for approximately 78 percent of the exposed area of the Sierra Nevada batholith (Ducea, 2001). Through time, the eastern fringe of arc magmatism migrated eastward and westward over distances of up to ~1000 kilometers, with two particularly large eastward excursions during Late Jurassic (~157 - 148 Ma) and Late Cretaceous-Eocene (~80 - 45 Ma) time (Coney and Reynolds, 1977; Lipman, 1992; Christiansen and others, 1994; Barton, 1996; Constenius, 1996). The Late Jurassic retroarc magmatic event was characterized by local small-volume intrusions of monzonitic to granitoid compositions that may have formed by crustal melting due to addition of heat from mantle-derived mafic melts (Barton, 1990, 1996). In contrast, the Late Cretaceous event was dominated by peraluminous granitic crustal melts that formed in response to thermal relaxation following crustal thickening (Barton, 1990, 1996; Lee and others, 2003).

The Early to Middle Jurassic arc probably had low regional elevations, as indicated by large intra-arc grabens with marine sediments (Busby-Spera, 1988; Saleeby and Busby-Spera, 1992). Substantial amounts of arc-derived volcaniclastic sediment and airborne ash were transported eastward into the Cordilleran retroarc region by Late Jurassic time (for example, Christiansen and others, 1994), suggesting that the arc had achieved regional positive elevation.

MAJOR COMPONENTS OF THE CORDILLERAN RETROARC THRUST BELT

The major components of the Cordilleran thrust belt have been strongly overprinted by mid- to late Cenozoic structural, metamorphic, and igneous processes that obscure pre-existing Cordilleran structure. The most important of these include mid-Cenozoic metamorphism and igneous activity associated with metamorphic core complexes (Gans and Miller, 1983; Wernicke, 1992), Neogene-present brittle extension of the Basin and Range (Miller and others, 1988; Wernicke, 1992; Friedrich and Bartley, 2003), and late Cenozoic igneous activity (Christianson and Yeats, 1992). Late Cenozoic igneous activity associated with the evolution of the Yellowstone-Newberry hotspot totally obscures the pre-existing structure of a large region in southern Idaho and northwestern Nevada (fig. 2). Further damage to Cordilleran structure occurred when the magmatic arc migrated into western Montana and north-central Idaho during late Cretaceous time. Thus, the best-preserved portion of the Cordilleran retroarc is at the latitude of Utah and Nevada, in spite of the fact that Neogene Basin and Range extension has structurally disrupted much of this region (Friedrich and Bartley, 2003). In some cases this later extension has helped to reveal the deeper structure of the orogenic hinterland (for example, Snoke and Miller, 1988; Miller and others, 1988; Miller and Gans, 1989; Hodges and Walker, 1992; Hudec, 1992; Camilleri and others, 1997; Wells, 1997). The major tectonic elements of the Cordilleran retroarc region can be described in terms of six tectonostratigraphic zones including, from west to east: the (1) Luning-Fencemaker thrust belt (LFTB); (2) central Nevada thrust belt (CNTB); (3) hinterland metamorphic belt; (4) Sevier thrust belt; (5) the foreland basin system; and (6) the Laramide intraforeland uplifts and basins (figs. 2 and 5).

Luning-Fencemaker Thrust Belt (LFTB)

The LFTB lies directly east of the Sierra Nevada batholith and comprises an ~500 kilometer long, 100 kilometer wide belt of intensely folded and thrusted Triassic and Jurassic marine sedimentary rocks. Individual thrust systems of note include the Luning, Fencemaker, Pamlico, Boyer, and Wild Horse thrusts (Speed and others, 1988). Thrusts of the LFTB generally carried deep-water, fine-grained turbiditic facies eastward and juxtaposed them against shallow-marine facies of the same age (Speed, 1978; Oldow, 1983; Oldow and others, 1984, 1990; Wyld, 2002). The rocks of the LFTB have been shortened by 50 to 75 percent (corresponding to >100 km), and are characterized by intense cleavage development and local low-grade metamorphism (Wyld, 2002).

Linkages between the LFTB and the Sevier thrust belt have not been documented. Based on the partial overlap in the timing of shortening in the Luning-Fencemaker and Sevier belts, Oldow (1983) and Speed and others (1988) suggested that the two share a common, mid-crustal décollement. Conversely, Wyld (2002) noted that the temporal overlap in thrusting is minor (Cretaceous thrust displacements in the LFTB are minimal) and concluded that the Sevier belt operated independently of the LFTB.

In this paper, all of the thrust systems east of the magmatic arc that were active during Late Jurassic to Eocene time are considered to be manifestations of a coherent Cordilleran orogenic wedge.

Central Nevada Thrust Belt (CNTB)

The CNTB consists of an ~400 kilometer long, north-south striking system of thrust faults and related folds in east-central Nevada (fig. 2; Taylor and others, 2000). The northward continuation of this belt has been referred to as the Eureka belt (Speed, 1983; Speed and others, 1988; Carpenter and others, 1993). These faults generally carry Cambrian to Pennsylvanian strata and the entire region has been beveled by the unconformity beneath Oligocene volcaniclastic and related rocks (Armstrong, 1972). Although the age bracketing on the CNTB is wide (Permian–mid-Cretaceous), Taylor and others (2000) argued that shortening was complete prior to ~85 Ma and Carpenter and others (1993) reported apatite fission track ages that suggest exhumation in the CNTB during Aptian time. Sparsely dated Lower Cretaceous synorogenic sediments are locally associated with the CNTB (Vandervoort and Schmitt, 1990). Maximum total shortening in the CNTB is between 10 kilometers and 15 kilometers (W. J. Taylor, personal communication, 2003).

Armstrong (1972) and Gans and Miller (1983) demonstrated that Tertiary sedimentary and volcanogenic strata generally rest in low-angle unconformity upon middle to upper Paleozoic rocks in east-central Nevada and west-central Utah. Low-amplitude, open folds are the only significant pre-Tertiary structures in this region (Gans and Miller, 1983). The limited amount of pre-Tertiary erosion and absence of significant paleotopography in this part of the thrust belt suggests that, outside of the CNTB, local topographic relief was very low in the hinterland region of eastern Nevada and westernmost Utah.

Hinterland Metamorphic Belt

Extending roughly north-south for several hundred kilometers in eastern Nevada and bending eastward into northwestern Utah and southern Idaho is a belt of high-grade (peak pressures of >9 kbar and temperatures of ~800°C) Barrovian metamorphic rocks that will be referred to as the hinterland metamorphic belt (D. L. Smith and others, 1993; Camilleri and others, 1997; McGrew and others, 2000; Lee and others, 2003). These metamorphic rocks are generally exposed in mid-Cenozoic metamorphic core complexes and therefore were overprinted by Cenozoic tectonics and metamorphism, but they retain a variety of Jurassic, Cretaceous, and early Tertiary metamorphic assemblages, and thermochronologic indicators of substantial Late Cretaceous cooling (for example, Miller and Gans, 1989; Hudec, 1992; Hodges and Walker, 1992; Peters and Wickham, 1994; Camilleri and Chamberlain, 1997; Wells, 1997; McGrew and others, 2000). Coney and Harms (1984) were among the first to recognize this belt of high-grade rocks as the tectonically buried, metamorphic axis of the Cordilleran thrust belt. Much of the high-grade metamorphism in these rocks took place in a contractional tectonic environment (for example, Miller and Gans, 1989; D. L. Smith and others, 1993; Wells, 1997; Camilleri and others, 1997) that overlapped in time with tectonic shortening farther to the east in the frontal part of the thrust belt.

Sevier Thrust Belt

Armstrong (1968) defined the Sevier thrust belt as a narrow zone of regional-scale, thin-skinned thrust faults and related folds in central Utah. Subsequent work has shown that the Sevier thrust belt extends far into the Canadian portion of the Cordillera and as far south as southeastern California (fig. 2; Monger and Price, 1979; Allmendinger, 1992). The southeastward continuation of the Sevier belt into Arizona and northern Mexico remains largely unknown, owing to complexities of Mesozoic paleogeography (principally the Bisbee-McCoy extensional basin) and Cenozoic extension. Although Armstrong (1968) characterized the Sevier thrust belt as entirely thin-skinned, it is now well documented that large slices of Precambrian metamorphic basement rocks were incorporated into the hanging walls of some Sevier thrusts (Burchfiel and Davis, 1972; Royse and others, 1975; Skipp, 1987; Schirmer, 1988; Yonkee, 1992; DeCelles, 1994; DeCelles and others, 1995; Schmitt and others, 1995; Mitra, 1997), and that the well known, brittle thrusts of the classic Sevier thrust belt merge westward and structurally downward with ductile shear zones that are associated with medium- to high-grade metamorphic rocks in the hinterland of eastern Nevada and western Utah (Allmendinger and Jordan, 1981; Snoke and Miller, 1988; Miller and others, 1988; Miller and Gans, 1989; Hudec, 1992; Hodges and Walker, 1992; Camilleri and Chamberlain, 1997; Camilleri and others, 1997; Wells, 1997). Thus, the Sevier thrust belt can be considered to span an east-west distance of ~300 kilometers in central Utah and eastern Nevada and a north-south distance of >2000 kilometers. Restoration of Cenozoic extension (Gans and Miller, 1983; Wernicke, 1992) reduces the maximum width of the Sevier thrust belt to ~200 kilometers.

In northwestern Montana the Sevier belt consists of six major thrust systems: the Moyie, Libby, Pinkham, Whitefish, and Lewis-Eldorado-Hoadley thrusts, and the frontal imbricate belt of the Sawtooth Range or "Montana Disturbed Belt" (fig. 2; Price, 1981; Mudge, 1982; Bally, 1984; Sears, 2001). The western part of the thrust belt is dominated by Proterozoic rocks in the hanging wall of the Lewis-Eldorado-Hoadley thrust system and thrust faults farther west. The Purcell anticlinorium, a regional antiformal culmination in Precambrian sedimentary rocks (Price, 1981; Constenius, 1996; Sears, 2001), occupies the trailing part of this segment of the Sevier belt. The Sawtooth Range consists of closely spaced imbricate thrusts and tight folds in Paleozoic-Mesozoic rocks, including Jurassic–Paleocene foreland basin deposits. This overall structural pattern continues into the Alberta Foothills, Front Ranges, and Main Ranges (Bally and others, 1966; Price and Mountjoy, 1970; Price, 1981; McMechan and Thompson, 1993; Fermor, 1999). Total shortening in the northwest Montana-Alberta segment of the thrust belt exceeds 165 kilometers (Bally 1984; McMechan and Thompson, 1993). However, this figure does not include additional shortening that must have taken place in the hinterland portion of the thrust belt.

To the south of the northwestern Montana segment, the thrust belt forms an eastward-convex salient with a chord length of ~150 kilometers, referred to as the Helena salient (fig. 5). The salient is bounded on the north by the Lewis and Clark line and on the south by the Southwest Montana transverse zone, both of which are complex zones of northwest-southeast and east striking strike-slip faults that have been reactivated several times since Precambrian time (McMannis, 1963; Harrison and others, 1974; Schmidt and O'Neill, 1982; Wallace and others, 1990). The Helena salient is distinguished by the presence of a thick succession of allochthonous Proterozoic clastic strata that form the eastern extremity of the Belt Basin (Harrison, 1972). The salient itself resulted in part from structural inversion of the eastern Belt Basin. The major thrusts of the Helena salient include the Sapphire, Lombard, and Indian Creek thrusts (Lageson and Schmitt, 1994). The center of the Helena salient is occupied by the Late Cretaceous (~80 - 70 Ma) Boulder batholith and related intrusive and volcanic rocks (Robinson and others, 1968).

In southwestern Montana and eastern Idaho north of the Snake River Plain, the Sevier belt forms an eastward concave reentrant between the Helena salient and the Idaho-Wyoming-Utah salient (fig. 5). This part of the thrust belt is significantly disrupted by mid- to late Cenozoic normal faults. Most workers agree that the major thrusts in this portion of the Sevier belt include the Hawley Creek, Fritz Creek, Cabin, Medicine Lodge, Four Eyes Canyon, and Tendoy thrust systems, and the frontal Lima anticline, which is inferred to be cored by a blind thrust (Ruppel and Lopez, 1984; Perry and others, 1988; Skipp, 1988). The western thrusts (Hawley Creek, Fritz Creek, and Cabin thrusts) carry a thick succession of Proterozoic strata; the Cabin thrust also carries Archean crystalline basement rocks (Skipp, 1987, 1988). The Medicine Lodge, Four Eyes Canyon, and Tendoy thrust sheets consist of Paleozoic and lower Mesozoic rocks. Total shortening estimates in the southwestern Montana region are >50 kilometers (Skipp, 1988), but detailed estimates for the entire southwestern Montana segment are not yet available.

Southeast of the Snake River Plain the Sevier belt reemerges and continues southward in an unbroken salient through western Wyoming into northeastern Utah (fig. 5). The Idaho-Wyoming-Utah salient contains eight major thrust systems: the Paris, Willard, Meade-Laketown, Crawford, Medicine Butte, Absaroka, Darby, and Hogsback thrusts (Armstrong and Cressman, 1963; Armstrong and Oriel, 1965; Royse and others, 1975; Dixon, 1982; Lamerson, 1982; Craddock, 1988; Craddock and others, 1988; Coogan and Royse, 1990; Coogan, 1992). In addition, a large duplex formed by the Ogden thrust system is situated structurally beneath the Willard thrust and above the Crawford thrust (Schirmer, 1988; Yonkee and others, 1989; Yonkee, 1992; DeCelles, 1994). The geometry and kinematic history of the thrusts in the Idaho-Wyoming-Utah salient are perhaps better known than in other portions of the Sevier thrust belt, mainly because the region has been extensively explored for hydrocarbons (Dixon, 1982; Lamerson, 1982), but also because a major part of the thrust belt remains partially buried by datable synorogenic foreland basin deposits that help to constrain thrust timing. In addition, well-preserved growth structures are abundant. The Paris, Willard, and Meade thrusts carry a thick succession of Proterozoic and Paleozoic rocks, whereas the Crawford, Medicine Butte, Absaroka, Darby, and Hogsback thrusts branch upward from a regional décollement in Cambrian shale (Armstrong and Cressman, 1963; Royse and others, 1975; Lamerson, 1982; Coogan and Royse, 1990; Royse, 1993a). Archean(?) to Paleoproterozoic crystalline rocks are exposed in the Wasatch culmination (fig. 5), a large antiformal duplex that formed by stacking of basement horses in the Ogden thrust system (Schirmer, 1988; Yonkee, 1992; Yonkee and others, 1989, 2003). Uplift of the Wasatch culmination folded the overlying Willard thrust sheet. Total shortening in the frontal part of the Idaho-Wyoming-Utah salient is at least 100 kilometers (Royse and others, 1975) and probably exceeds 160 kilometers when slip along the Willard thrust is included (Royse, 1993a; Camilleri and others, 1997).

In north-central Utah, the Sevier thrust belt is dominated by another major eastward curving salient, referred to as the Charleston-Nebo (or Provo) salient (fig. 5). Unlike farther north and south, the Charleston-Nebo salient comprises a thick section of Pennsylvanian-Permian marine clastic and carbonate rocks (the Oquirrh Group; Royse, 1993a). The salient is bounded on its north flank by the Charleston thrust, on the south by the Nebo thrust, and on the east by a triangle zone formed by unnamed blind thrusts in Jurassic and Cretaceous rocks. To the west, the Sheeprock, Tintic Valley, Stockton, and Midas thrusts form the trailing part of the Charleston-Nebo salient (figs. 3 and 5; Christie-Blick, 1983; Tooker, 1983; Mitra, 1997; Mukul and Mitra, 1998a). The Sheeprock thrust carries Proterozoic sedimentary rocks; the Tintic Valley, Stockton, and Midas thrusts carry Paleozoic rocks; and the Charleston-Nebo system carries Precambrian basement as well as Paleozoic-Mesozoic strata. The center of the salient is breached by a large graben in the hanging wall of the Wasatch normal fault. The normal fault has west-side-down displacement and truncates the western flank of another enormous structural culmination, the Santaquin culmination, which formed in Precambrian basement through Jurassic rocks in the hanging wall of the Charleston-Nebo thrust system in the central part of the salient (Mitra, 1997; Constenius, ms 1998, 1999). The southern boundary of the salient is marked by the Leamington line (Morris, 1983), which may be a lateral or oblique ramp associated with the southwestern margin of the Oquirrh basin (Mitra, 1997). The internal structure of the frontal Charleston-Nebo salient is a large-scale wedge thrust system, in which slip along the basal décollement and basement duplexing were accommodated by backthrusting at shallow levels, generally above Permian and Jurassic shale and evaporite horizons (Constenius, ms, 1998). Antiformal structures formed above the backthrusts and the frontal triangle zone. A thick blanket of wedge-top foreland basin sediments obscures much of the eastern part of the Charleston-Nebo salient. Shortening in the Charleston-Nebo salient is at least 105 kilometers (Constenius, ms, 1998), but estimates of thrust displacements to the west of the Wasatch normal fault, which are likely to be large (Christie-Blick, 1983; Mukul and Mitra, 1998a, 1998b), are not available.

In south-central Utah four major thrust systems form the Sevier thrust belt: the Canyon Range, Pavant, Paxton, and Gunnison thrusts (figs. 3 and 5; Standlee, 1982; Lawton, 1985; Villien and Kligfield, 1986; Royse, 1993b; DeCelles and others, 1995; Mitra, 1997). The Canyon Range thrust carries a thick section of Proterozoic quartzite and argillite and lower Paleozoic carbonate and shale. In its frontal part, the Canyon Range thrust is folded into a tight antiform-synform pair owing to the growth of an antiformal duplex in its footwall (Mitra and Sussman, 1997). The Pavant thrust carries lower Cambrian through Cretaceous strata. The Paxton and Gunnison thrusts branch upward from a basal décollement in Cambrian shale and form two large antiformal duplexes and a frontal triangle zone (the San Pete Valley antiform, Lawton and others, 1993, 1997; DeCelles and others, 1995). Farther southwest along strike, the Wah Wah, Blue Mountain, and Iron Springs thrusts are roughly equivalent to the Canyon Range, Pavant, and Paxton or Gunnison thrusts, respectively (Goldstrand, 1994; Friedrich and Bartley, 2003). As in the Charleston-Nebo salient, the frontal 50 kilometers of the thrust belt in central Utah are largely buried by thick synorogenic sediments of the wedge-top depozone (DeCelles and others, 1995; Lawton and others, 1997). Total shortening from the Central Nevada thrust belt to the front of the thrust belt in central Utah is at least 240 kilometers (DeCelles and others, 1995; Currie, 2002).

In southern Nevada the Sevier belt strikes northeast-southwest and comprises seven major thrust systems that are offset by the Northern Death Valley and Las Vegas shear zones (Cenozoic strike-slip faults). From northwest to southeast, these consist of the Last Chance, Marble Canyon, Clery, Wheeler Pass, Lee Canyon, Keystone, and Birdspring thrust systems (fig. 3; Burchfiel and Davis, 1972; Bohanon, 1983; Wernicke and others, 1988; Burchfiel and others, 1998). The major thrusts have classic ramp-flat geometries (Wernicke and others, 1988). The Wheeler Pass thrust system carries a thick succession of Proterozoic clastic rocks as well as lower Paleozoic strata, and the eastern thrusts generally branch upward from a décollement in Cambrian rocks and have regional footwall flats in Jurassic strata (Bohannon, 1983; Wernicke and others, 1988; Carpenter and Carpenter, 1994). Total minimum shortening on these thrusts is up to ~75 kilometers (Burchfiel and others, 1974).

In southeastern California, the strike of the thrust belt swings into a northwest–southeast orientation, and the major thrusts cut across the miogeoclinal prism and incorporate Precambrian basement rocks (Burchfiel and Davis, 1975, 1981). The southernmost manifestations of the Sevier belt are exposed in the Clark Mountains and New York Mountains of southeastern California, where Precambrian crystalline basement, Proterozoic and Paleozoic sedimentary rocks, and Mesozoic igneous rocks are involved in thrust sheets (Burchfiel and Davis, 1972, 1975; Walker and others, 1995). From west to east the major thrust systems include the Winters Pass, Pachalka, Mescal, Mesquite Pass, and Keaney/Mollusk Mine systems (Walker and others, 1995). To the west and northwest of the Clark Mountains lies the Eastern Sierra thrust belt (fig. 3), a zone of east-verging Mesozoic thrusts that cut Paleozoic and Mesozoic rocks, including Sierra Nevadan batholithic rocks (Dunne, 1986; Dunne and Walker, 1993; Walker and others, 1995). Because of the intimate association between thrusting and igneous activity in the southern part of the magmatic arc, the southern Nevada-California portion of the Sevier belt is unique in that the timing of thrust displacements is bracketed by radiometric ages on cross-cutting and overlapping igneous rocks (for example, Fleck and Carr, 1990; Fleck and others, 1994; Walker and others, 1995; Wrucke and others, 1995).

The Sevier belt cannot be traced continuously south of the Clark Mountains in southeastern California, in part because of the intensity of Neogene extensional deformation that dominates the regional structure of southern Arizona and northern Sonora, Mexico. In western Arizona the enigmatic Maria fold-thrust belt and Mule Mountains thrust system strike east-west, involve strata that were deposited in the Bisbee-McCoy basin as well as Precambrian basement and Paleozoic cover strata, and have an uncertain relationship with the remainder of the classic Sevier belt to the northwest (Spencer and Reynolds, 1990). Dickinson and Lawton (2001a), suggested that the Maria fold-thrust belt may manifest reactivation of the extensional Bisbee-McCoy boundary rift structures, in a manner analogous to the contractional reactivation of much older structures that occurred in the Helena embayment and Uinta basin.

Foreland Basin System and Laramide Structural Province

To the east of the Sevier thrust belt lies the Western Interior basin (WIB), an immense composite basin that at its apogee spanned an east-west distance of >1000 kilometers and a north-south distance of >5000 kilometers from the Canadian Arctic to the Gulf of Mexico (fig. 1; Williams and Stelck, 1975). Throughout much of its history the WIB was inundated by marine waters of the Western Interior Seaway, which sporadically connected the Gulf of Mexico and the Arctic Ocean (McGookey, 1972; Kauffman and Caldwell, 1993; Robinson Roberts and Kirschbaum, 1995). From Late Jurassic to Late Cretaceous time, the western WIB in Alberta and the western U.S.A. comprised all of the four major depozones (back-bulge, forebulge, foredeep, and wedge-top) of a classic foreland basin system (fig. 6; DeCelles and Giles, 1996). As the Cordilleran thrust belt migrated eastward, the deposits of these four depozones were stacked vertically in an overall upward-coarsening sequence now preserved in western Wyoming, eastern Idaho, and central Utah (DeCelles and Currie, 1996). The Upper Jurassic consists of back-bulge deposits; the Jurassic-Cretaceous unconformity represents the passage of the forebulge; the Lower Cretaceous represents the foredeep depozone; and the Upper Cretaceous in part consists of wedge-top deposits close to the thrust belt and foredeep deposits at more distal locations. Superimposed on the regional flexural subsidence pattern was subsidence (and uplift) driven by dynamic processes in the mantle (fig. 6; Mitrovica and others, 1989; Gurnis, 1992; Catuneanu and others, 1997). The main effect of dynamic subsidence in the WIB was to broaden the wavelength of subsidence into regions far eastward of any plausible surface-load-driven flexural accommodation.

Beginning in late Campanian time, the WIB between latitudes 32° to 46°N became increasingly partitioned into a mosaic of more than 20 smaller basins separated by Precambrian basement uplifts of the Laramide structural province (fig. 2; Brown, 1988; Dickinson and others, 1988; Stone, 1993; Erslev, 1993). The Laramide uplifts are bounded by moderate angle reverse faults that may extend into the lower crust (Smithson and others, 1978; Gries, 1983; Stone, 1993). Many Laramide ranges are flanked by conjugate forethrust-backthrust systems (Brown, 1988; Erslev, 1993). In southwestern Montana, the strongly layered Archean gneissic basement behaved like layered sedimentary rocks and foreland arches with eastward vergence formed above

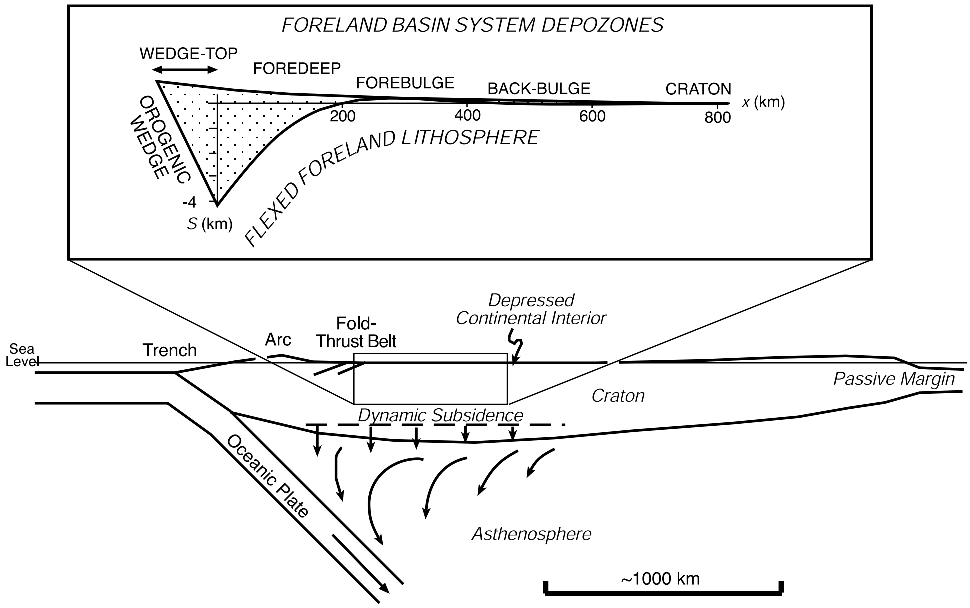

Fig. 6. Schematic cross sections illustrating the principal mechanisms of subsidence in retroarc foreland basins (modified from DeCelles and Giles, 1996). Lower panel shows the effect of long-wavelength dynamic subsidence (Gurnis, 1992), which tilts the craton downward toward the trench. Upper panel depicts shorter-wavelength flexural subsidence (S) versus lateral distance (x) owing to the topographic load of the orogenic wedge, and the four depozones that characterize many foreland basin systems. Note the extreme vertical exaggeration in the upper panel.

Laramide thrust faults (Schmidt and Garihan, 1983; Schmidt and others, 1993). Unlike the region to the west, the Laramide region (including the Colorado Plateau) was not significantly affected by Cenozoic extension, and the present crustal thickness of ~45 to 50 kilometers may be representative of early Cenozoic conditions (Johnson and others, 1984; Karlstrom and others, 2002). Although the Laramide uplifts are generally east of the front of the Sevier fold-thrust belt, those in southwestern Montana and northwestern Wyoming locally interacted with the eastward migrating thrust belt (Schmidt and Garihan, 1983; Craddock and others, 1988). Laramide basins were filled with thick accumulations of fluvial, alluvial, and lacustrine sediments. The location of the main, coherent latest Cretaceous–Eocene depocenter shifted eastward to the perimeter basins of Dickinson and others (1988), including the Crazy Mountains, Powder River, Denver, and Raton basins (fig. 2). These basins were larger than the internal (or axial) Laramide basins (Dickinson and others, 1988), and lay generally east of the Laramide deformation front, with the exception of the Powder River basin, which was partly flanked on the east by the Black Hills uplift.

REGIONAL KINEMATIC AND SUBSIDENCE HISTORIES OF THE CORDILLERAN THRUST BELT AND FORELAND BASIN SYSTEM

The timing of major kinematic events in the Cordilleran orogenic belt is constrained by radiometric ages (geochronological and thermochronological) in the hinterland region, and by a combination of sparse radiometric ages and widespread, datable synorogenic sedimentary rocks in the frontal part of the orogen. The Cretaceous time scale of Obradovich (1993) is employed. Key age data are summarized in figure 3, and regional paleogeographic interpretations are presented in the ensuing figures. Palinspastic maps were restored according to Cenozoic extension and strike-slip estimates of Gans and Miller (1983), Wernicke (1992), Saleeby and Busby-Spera (1992), and Dickinson and Wernicke (1997). Incremental palinspastic restorations during Mesozoic–early Cenozoic shortening are based on documented estimates of shortening during each time frame discussed.

Late Jurassic (fig. 7)

The North American plate acquired a component of northward motion during Late Jurassic time (~155 - 142 Ma; Page and Engebretson, 1984; Engebretson and others, 1985; May and Butler, 1986). Convergence was oriented northwest-southeast, nearly orthogonal to the northern portion of the margin but nearly parallel to the southwestern margin (fig. 7; May and others, 1989). Consequently, the southwestern part of the plate margin (mainly southeastern California, Arizona, and Sonora, Mexico) was dominated by sinistral strike-slip and transtensional tectonics, while the northern retroarc region experienced regional shortening (Saleeby, 1992; Saleeby and Busby-Spera, 1992). Transtension in the southern, inboard part of the magmatic arc in California is indicated by the timing of intrusion of the Independence dike swarm in the southern Sierra Nevada and Mohave regions (~148 Ma; Chen and Moore, 1982; James, 1989) and development of the extensional Bisbee-McCoy basin farther southeast (Dickinson and Lawton, 2001a, 2001b). Coeval sinistral strike-slip along the Pine Nut fault (Oldow and others, 1993) and the Mojave-Sonoran megashear may be considered part of the broader sinistral displacement system along the southwestern margin of North America (Saleeby, 1992).

Crustal shortening, metamorphism, and magmatism occurred locally in the Cordilleran retroarc region during the Late Jurassic in Nevada, western Utah, and southern California (figs. 3 and 7; see summary in D. L. Smith and others, 1993). Diffuse granitic magmatism swept into eastern Nevada and western Utah (Christiansen and others, 1994; Barton, 1996; Lee and others, 2003). High-strain zones in Nevada were localized around magmatic centers (Hudec, 1992; D. L. Smith and others, 1993; Lee and others, 2003). Thrusting in the Eastern Sierra thrust belt is bracketed between ~168 to 148 Ma by U-Pb zircon ages of igneous rocks (Walker and others, 1990; Dunne and Walker, 1993; Wrucke and others, 1995). In the Clark Mountains of southeastern California, displacement on the Pachalka thrust is dated between 146 and 142 Ma by U-Pb zircon ages (Walker and others, 1995). Wyld (2002) summarized the available geochronologic data and structural relationships that bracket the main phase of shortening in the LFTB between mid- and Late Jurassic time (~165 - 150 Ma), with additional but minor mid- to Late Cretaceous deformation. Allmendinger and Jordan (1981) discussed scattered evidence for Late Jurassic shortening in western Utah and eastern Nevada, including possible displacement on the Hansel thrust (or Manning Canyon detachment). Allmendinger and Jordan (1984) presented structural and geochronological evidence for thrusting and plutonism at ~150 Ma in northwestern Utah. Hudec (1992) documented Late Jurassic (~153 Ma) shortening, metamorphism, and granitic intrusion in the central Ruby Mountains of northeastern Nevada. Lee and others (2003) suggested that two-mica granites of inferred Late Jurassic age in the Ruby Mountains-East Humboldt Range core complex were derived from melting of Proterozoic metagraywackes at depths of >30 kilometers. Camilleri and Chamberlain (1997) presented evidence for significant crustal thickening between 153 Ma and 84 Ma owing to displacement on the inferred Windermere thrust in northeastern Nevada. Miller and others (1988) and Miller and Gans (1989) summarized data that indicate Late Jurassic amphibolite-grade metamorphism at mid-crustal depths (7-15 km), crustal anatexis, and local ductile thrusting (top-to-the-west sense of shear) in eastern Nevada and western Utah.

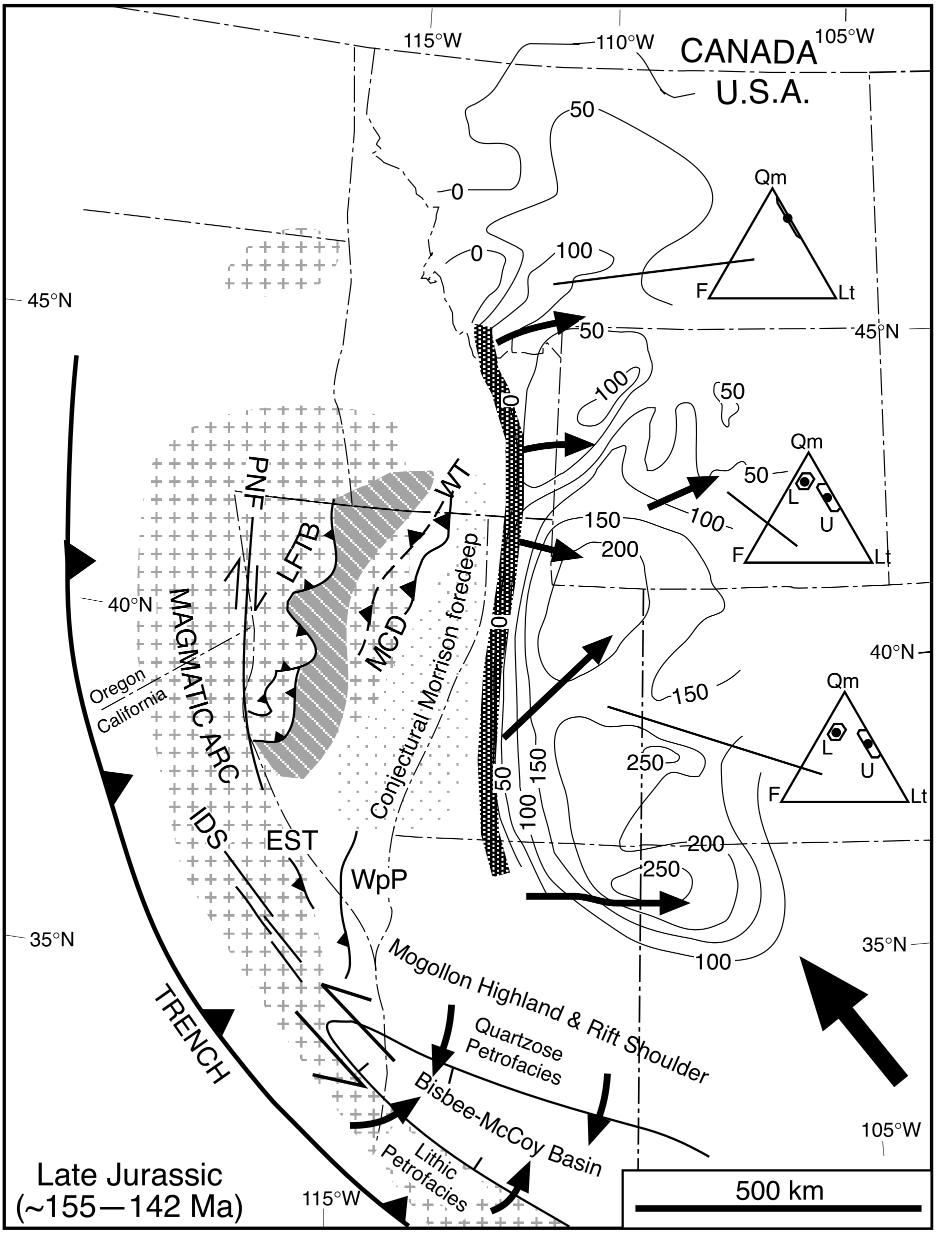

Fig. 7. Palinspastic isopach map (in meters) of the Morrison Formation and reconstructed locations of major active tectonic elements of the Cordilleran thrust belt during Late Jurassic time. Major active structures are abbreviated as follows: LFTB, Luning-Fencemaker thrust; WT, Windermere thrust (which may have been active later); MCD, Manning Canyon detachment; PNF, Pine Nut fault; IDS, Independence dike swarm; EST, Eastern Sierra thrust belt; WpP, Winters Pass-Pachalka thrust. Thin arrows indicate generalized sediment dispersal directions. Large arrow in southeast indicates approximate direction of motion of North America relative to the Farallon plate. Plus sign pattern indicates magmatic arc. Diagonally ruled area in Nevada is tectonically inactive Golconda and Roberts Mountains allochthons. Stippled area is conjectural Morrison foredeep depozone (Royse, 1993a). Shaded thick line in western Utah represents possible forebulge location. Sandstone compositions are illustrated in terms of Qm (monocrystalline quartz), F (total feldspar), and Lt (total lithic fragments). In the two southern triangles, L and U refer to lower and upper Morrison petrofacies, respectively. Based on data from Suttner (1969), Furer (1970), Suttner and others (1981), Allmendinger and others (1984), DeCelles and Burden (1992), Saleeby and Busby-Spera (1992), Malone and Suttner (1992), Camilleri and Chamberlain (1997), Currie (1997), Dickinson and Lawton (2001a), Wyld (2002), and other sources noted in figure 3 and table 1.

The Morrison Formation accumulated during Late Jurassic time in the retroarc region of Montana, Wyoming, Utah, and western Colorado. The Morrison consists of fluvial, eolian, lacustrine, and tidally influenced sandstone, mudstone, and conglomerate (Stokes, 1944; Miall and Turner-Peterson, 1989; DeCelles and Burden, 1992; Currie, 1997, 1998). The Morrison also contains abundant volcanic ash, including radiometrically-dated tuffs (Kowallis and others, 1998). Paleocurrent data indicate sediment dispersal from the southwest and west (fig. 7). Morrison sandstones and conglomerates contain abundant quartz, sedimentary lithic (mainly chert and quartzite), and volcanic lithic grains, with locally significant amounts of feldspar (Suttner, 1969; Furer, 1970; DeCelles and Burden, 1992; Currie, 1997). The Morrison exhibits a regionally complex thickness pattern (fig. 7). South of latitude 42°N, the unit generally thickens radially inward toward a maximum of ~250 meters near the 110th meridian. North of 42°N, the Morrison changes thickness abruptly under the influence of local paleotopographic elements (for example, Suttner, 1969; Malone and Suttner, 1992). Throughout the entire western interior U.S.A. the top of the Morrison is marked by a major unconformity (McGookey, 1972). This unconformity is locally angular in southwestern Montana (DeCelles, 1986). The duration of the hiatus represented by this unconformity is not well constrained, but it may be approximately 20 Myr.

In contrast to the Morrison Formation, Upper Jurassic and lowermost Cretaceous strata in the Bisbee-McCoy basin are on the order of 2 to 3 kilometers thick and consist of locally derived, compositionally and texturally immature, coarse-grained alluvial fan deposits (Bilodeau, 1982; Dickinson and Lawton, 2001a). Bipolar sediment sources shed quartzose sand from the Mogollon highland and rift shoulder located to the northeast and volcanolithic sand from the magmatic arc, which lay to the southwest (fig. 7; Dickinson and Lawton, 2001a). In southeastern Arizona, Bisbee Group strata are interbedded with bimodal volcanics, including tholeitic rocks with a mantle component (Lawton and McMillan, 1999). The Bisbee-McCoy basin generally became broader and deeper toward the southeast, where it accumulated marine deposits and merged with the rifting Chihuahua trough, an arm of the opening oceanic Gulf of Mexico (Dickinson and Lawton, 2001a, 2001b).

Perhaps more than any other time interval in the retroarc region, the Late Jurassic has been the focus of intense debate with respect to tectonic setting and mechanism(s) of basin subsidence. Based mainly on sediment provenance data, early studies concluded that the Morrison Formation was derived from uplifted western thrust belt sources and deposited in a typical foreland setting (for example, Spieker, 1946; Stokes, 1944, 1952; Armstrong and Cressman, 1963; Armstrong and Oriel, 1965; Suttner, 1969; Furer, 1970). Paleocurrent data from the Morrison Formation generally support this interpretation (fig. 7). The abundant volcanic ash in the Morrison is also consistent with western sources in the magmatic arc. However the Morrison Formation does not thicken continuously westward (fig. 7) in the manner of a typical foredeep deposit, suggesting that it was not influenced by flexural loading in a coeval thrust belt to the west (Heller and others, 1986). Heller and others (1986) suggested that the Morrison was deposited under the influence of regional tectonothermal subsidence, rather than flexural subsidence. However, this mechanism is difficult to reconcile with the evidence cited above for active crustal shortening in the retroarc thrust belt. DeCelles and Burden (1992) suggested that the Morrison represents the overfilled, or back-bulge, part of the foreland basin system. This explanation requires an age-equivalent foredeep depozone in western Utah and eastern Nevada, where Upper Jurassic foreland basin deposits are notably absent. In balanced regional cross-sections of the thrust belt, Royse (1993b) showed that ample space was available for an Upper Jurassic foredeep depozone in western Utah and eastern Nevada. He proposed that this "phantom foredeep" was eroded during Cretaceous time as the region was uplifted above structurally lower thrust faults of the frontal Sevier belt. Royse's (1993b) explanation is supported by conodont alteration indices from mid-Mississippian strata in western Utah and eastern Nevada that indicate burial temperatures of 200° to 300°C (Sandberg and Gutschick, 1984). Assuming a normal geothermal gradient and that 3 to 4 kilometers of upper Paleozoic-Triassic strata rested on top of the Mississippian rocks (Hintze, 1988; Miller and others, 1988), the conodont alteration indices place a maximum thickness constraint of ~2 to 5 kilometers on the hypothetical Jurassic foredeep sediments. Given the palinspastic reconstruction of Morrison Formation isopachs and potential orogenic loads in Nevada, an eroded Upper Jurassic foredeep cannot be ruled out (figs. 7 and 8).

DeCelles and Currie (1996) and Currie (1997) suggested that the Morrison was deposited east of the forebulge in the back-bulge depozone of a regional foreland basin system. According to this explanation, which is compatible with Royse (1993b), the eventual eastward migration of the forebulge into eastern Utah would have produced the basal Cretaceous unconformity at the top of the Morrison. However, Currie (1998) noted that the unconformity extends too far eastward to be explained solely by forebulge migration, and proposed that both forebulge migration and dynamic uplift owing to changes in the angle of the subducting plate may have produced the unconformity. Currie's (1998) model combines flexural and regional dynamic subsidence to explain the distribution of the Morrison Formation. This model is similar to those proposed by Mitrovica and others (1989), Beaumont and others (1993), and Catuneanu and others (1997) to explain the widespread distribution of Cretaceous foreland basin deposits in the western interior basin.

Lawton (1994) noted the temporal overlap between Morrison deposition and tectonic events in central California (for example, the Nevadan orogeny, Harper and Wright, 1984; Ingersoll and Schweickert, 1986; Saleeby and Busby-Spera, 1992; Dickinson and others, 1996), and suggested that an oceanic slab (the Mezcalera plate of Dickinson and Lawton, 2001b) was completely subducted beneath western North America during late Middle Jurassic time. This model relies on the foundering of an oceanic slab to drive regional dynamic subsidence that accommodated Morrison sediments. The slab-foundering hypothesis for Morrison deposition is attractive because of the transient nature of the inferred subsidence mechanism. Because the amplitude of viscosity-driven subsidence attenuates rapidly as the foundering slab reaches depths of several hundred kilometers (fig. 8A; Morgan, 1965), regional isostatic rebound is predicted to ensue the subsidence event. Such a rebound event could help to explain the widespread distribution of the basal Cretaceous unconformity. However, three issues render this explanation questionable. First, the existence of the Mezcalera plate is conjectural. Second, the Mezcalera plate would have been subducted roughly 800 to 1000 kilometers outboard of the Morrison depositional region (fig. 8A). Coupled with rapid northward motion of North America during Late Jurassic time (May and Butler, 1986), this means that the Mezcalera slab would have foundered too far outboard to influence subsidence in the western interior region. Third, the subsidence generated by foundering slabs takes place rapidly during the first few hundred kilometers of slab descent, such that most of the subsidence would have been damped out within a few million years of initial slab foundering (fig. 8A). If the Mezcalera plate was completely subducted by ~155 Ma, then the expected viscosity-driven subsidence should have been largely completed by ~150 Ma, yet Morrison sediments continued to accumulate for an additional ~6 to 8 Myr. The Mezcalera slab-foundering hypothesis, therefore, predicts subsidence that is too restricted in both time and space to explain the distribution and thickness of the Morrison Formation. On the other hand, subduction of the Farallon plate (which would have replaced the Mezcalera plate beneath western North America by ~155 Ma;

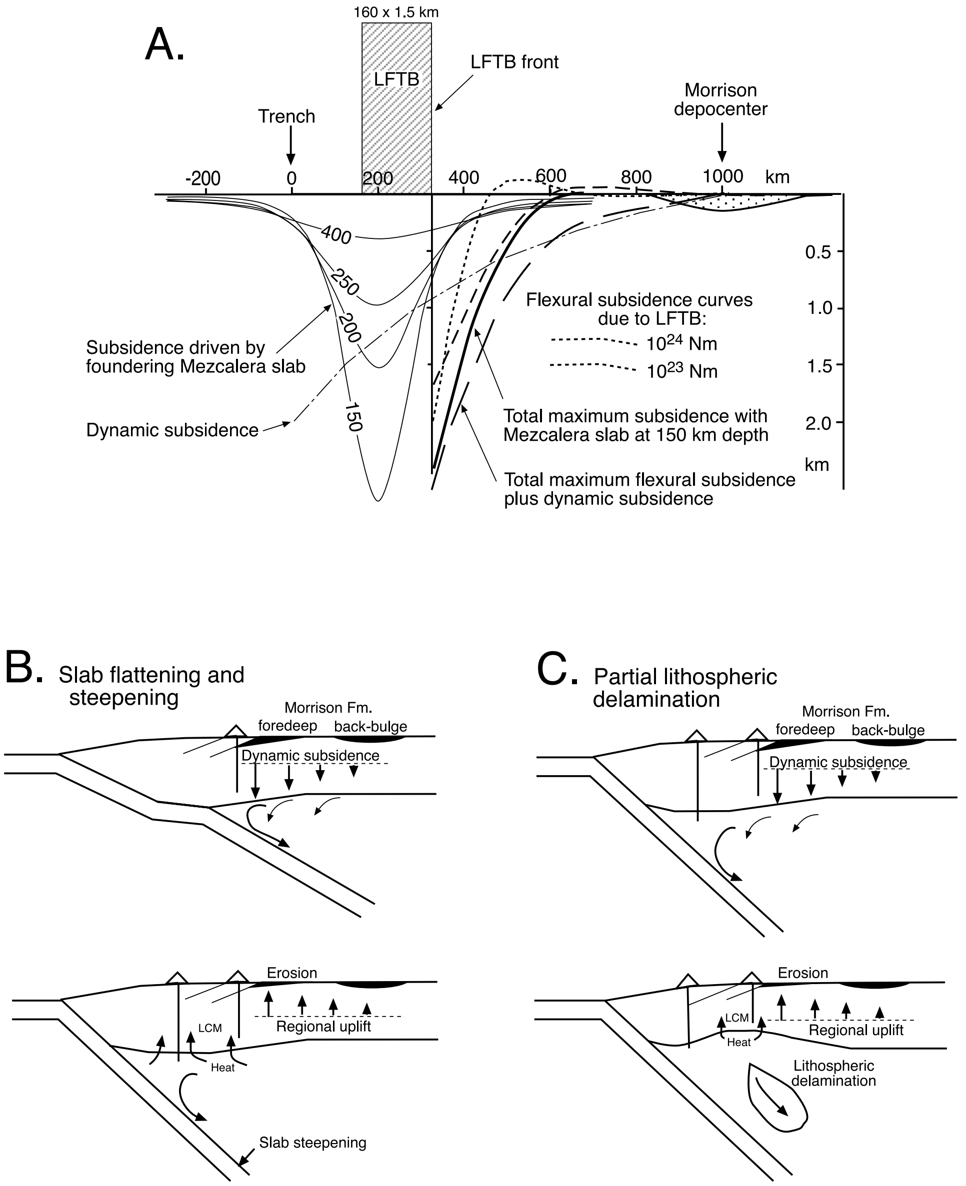

Fig. 8. (A) Comparison of predicted flexural subsidence owing to the surface load of the Luning-Fencemaker thrust belt (LFTB) (dashed and dotted curves, for different flexural rigidities using equations of Turcotte and Schubert, 2002), subsidence driven by viscous foundering of a hypothetical Mezcalera slab at depths of 150 km, 200 km, 250 km, and 400 km (calculated according to Morgan, 1965, assuming a spherical geometry with a diameter of 200 km), and dynamic subsidence driven by viscous coupling between the subducting Farallon slab and North American craton (after Gurnis, 1992). The origin of the horizontal scale is the approximate location of the trench during Late Jurassic time. Thick line represents sum of the subsidence owing to Mezcalera slab foundering (when the slab was at only 150 km depth) and flexure of lithosphere with flexural rigidity of 10^{24} Nm. The distribution of the Morrison Formation is shown at $\sim$1000 $\pm$ 200 km from trench. Note that the Morrison depocenter is situated approximately where a back-bulge depozone is predicted for flexed lithosphere of rigidity = 10^{24} Nm. Predicted subsidence driven by the foundering Mezcalera slab is several hundred kilometers farther west. The long dashed curve is the sum of subsidence for flexed lithosphere of rigidity = 10^{24} Nm and dynamic subsidence. (B) and (C) Schematic east-west cross sections at lithospheric scale, illustrating hypothetical plate tectonic processes beneath western U.S.A. during Late Jurassic–earliest Cretaceous time. LCM indicates lower crustal melting. See text for discussion.

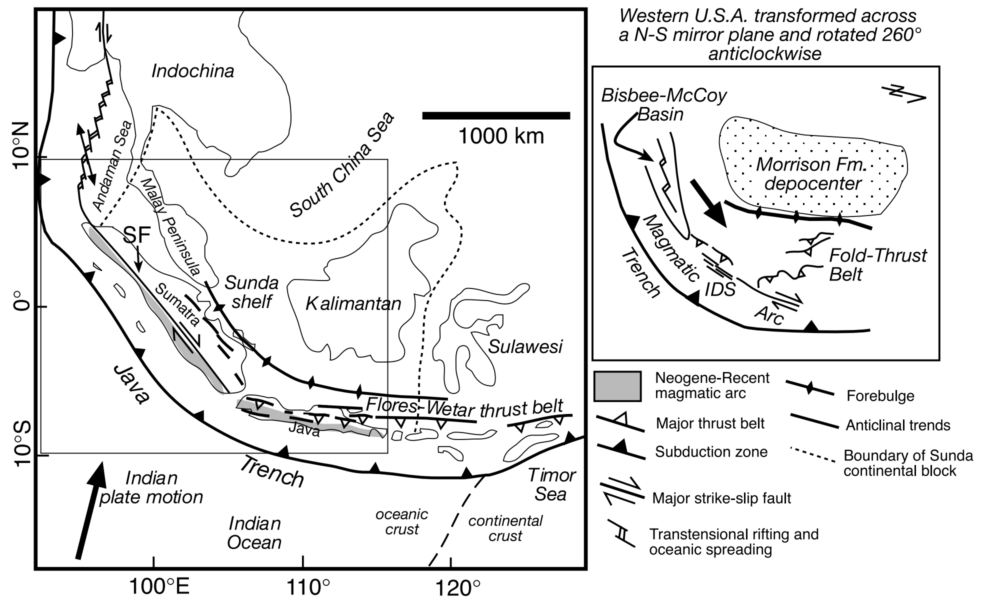

Fig. 9. Generalized tectonic maps (at same scale) of the Indonesian orogenic system (left) and western U.S.A. Cordilleran orogenic system during Late Jurassic time (right). The box in the Indonesian map indicates the same-sized region as depicted in the Late Jurassic map. Abbreviations: SF, Sumatran fault system; IDS, Independence dike swarm. Large arrows indicate relative plate convergence directions. The Late Jurassic map has been simplified from figure 6 and transformed as indicated in title. The Indonesian map is after Hamilton (1979), Silver and others (1983), McCaffrey and Nabelek (1987), and Lee and Lawver (1995).

Dickinson and Lawton, 2001b) could have driven far-field dynamic subsidence in Nevada and Utah during Late Jurassic time, accounting for at least some aspects of the Morrison Formation. The unequivocal evidence for Late Jurassic thrusting in western Nevada and southeastern California and the recycled orogenic composition of Morrison Formation sandstones and conglomerates require the coeval existence of an active retroarc thrust belt and a linked foredeep depozone.

Unlike the controversial state of interpretation for the Morrison Formation, an undivided consensus attributes the Upper Jurassic–Lower Cretaceous strata of the Bisbee-McCoy basin to deposition in a continental rift (for example Bilodeau, 1982; Bilodeau and others, 1987; Mack, 1987; Dickinson and Lawton, 2001a). This interpretation is entirely consistent with the regional plate tectonic apparatus along the southwestern margin of North America (Saleeby and Busby-Spera, 1992).

Thus, the Late Jurassic Cordilleran retroarc region exhibits features typical of both contractional and transtensional tectonics, together with two large sedimentary basins of grossly different character. Perhaps a solution to the Late Jurassic riddle lies in a comparison with a modern continental retroarc system, such as the Indonesian orogen (Lawton, 1994; fig. 9). Along the Java trench, oceanic lithosphere of the Indian plate is subducting beneath the Sunda continental block, which is covered by a shallow sea (<200 m; Ben Avraham and Emery, 1973; Hamilton, 1979). The southwestern edge of the Sunda block is occupied by a Neogene-Recent magmatic arc, inboard of which lie mid-Tertiary to Recent flexural foreland basins related to retroarc crustal shortening (McCaffery and Nabelek, 1987). Because of the convex shape of the Sunda plate margin, the relative convergence direction along the Java trench changes from nearly orthogonal in the southeast to highly oblique (dextral) in the northwest (Lee and Lawver, 1995). Accordingly, the retroarc tectonic environment of the Indonesian orogen is dominated by north-south shortening in its southeastern part (the islands of Java, Bali, Flores, Wetar, and Timor) and dextral strike-slip, transtension, and ultimately oceanic spreading in its northwestern part (Sumatran Fault system, Andaman Sea) (Ben-Avraham and Emery, 1973; Silver and others, 1983; McCaffrey and Nabelek, 1987; McCaffrey, 1988; Bransden and Matthews, 1992; Lee and Lawver, 1995). Subsidence in the southern part of the retroarc region is dominated by flexural loading within a few hundred kilometers of the magmatic arc (McCaffrey and Nabelek, 1987) and long-wavelength dynamic loading as far as 1000 kilometers northeast of the arc (Gurnis, 1993). The Late Jurassic Cordilleran orogen and retroarc basin system shares striking similarities with the modern Indonesian orogenic system (fig. 9), including an active magmatic arc, a comparable along-strike change from orthogonal (in the north) to nearly parallel (in the southeast) plate convergence, an along-strike transition from contractional (in the north) to transtensional (in the southeast) arc and retroarc tectonics, and a corresponding transition from a foreland-style basin in the north (Utah, Wyoming, and Montana) to an extensional (Bisbee-McCoy) basin in the south. As suggested by Lawton (1994) and Currie (1998), Late Jurassic sedimentation in the distal foreland region of central Wyoming and western Colorado can be explained best by dynamic subsidence (fig. 6). Interference between the long-wavelength dynamic subsidence and shorter-wavelength flexural subsidence (for example, Catuneanu and others, 1997) driven by the Luning-Fencemaker thrust belt in western Nevada would have produced a broadly tapered prism of Upper Jurassic sediment, extending from the LFTB into western Colorado (fig. 8A). The forebulge crest (which was probably buried) lay approximately in western Utah, ~250 to 300 kilometers east of the front of the LFTB and ~150 kilometers east of the Hansel thrust sheet (fig. 7; Allmendinger and others, 1984).

The eastward pulse of Late Jurassic magmatism in Nevada and western Utah (for example, Christiansen and others, 1994; Barton, 1996) can be explained in terms of two alternative but not mutually exclusive hypotheses: (1) relatively low-angle subduction of the Farallon plate could have caused Late Jurassic arc magmatism to migrate inboard into eastern Nevada and western Utah (fig. 8B); and/or (2) partial delamination of thickened lithosphere beneath Nevada could have steepened the geotherm and produced melts in the lower crust and remaining lithosphere (fig. 8C; a similar process, based on seismological data, has been proposed to explain magmatism in the central Andes; Beck and Zandt, 2002). Regional isostatic rebound and erosion of the proximal portion of the Morrison Formation could have been driven by subsurface load removal in the aftermath of either scenario–by slab steepening (fig. 8B; Currie, 1998) or decrease in the lithospheric density profile due to partial delamination (fig. 8C; Molnar and others, 1993). Regional uplift above Cretaceous thrusts (for example, Royse, 1993b; Taylor and others, 2000) also would have increased the likelihood of erosion of the proximal Morrison foredeep. Figure 8 illustrates only a few of the plausible combinations of physical parameters and processes in an essentially inscrutable problem. Clearly, a satisfactory resolution of the debate is unlikely until more is known about dynamics in the upper mantle during the Late Jurassic.

Early Cretaceous (fig. 10)

During Early Cretaceous time (~142 - 112 Ma) the volume of arc magmatism diminished significantly and the eastern front of the arc retreated several hundred kilometers to the west (Armstrong and Ward, 1993; Christiansen and others, 1994; Barton, 1996). Dextral strike-slip on the order of 200 kilometers (Dickinson, 2000) to 400 kilometers (Schweickert and Lahren, 1990; Wyld and Wright, 2001) occurred along shear zones within the arc from southern California to Idaho. Evidence for Early

Cretaceous metamorphism and local thrusting is sparsely scattered in the eastern Nevada–western Utah region (figs. 3 and 9; D. L. Smith and others, 1993).

Coupled with the decrease in arc magmatism, the general paucity of evidence for regional Early Cretaceous shortening in the hinterland has led some authors to suggest that the Cordillera experienced a lull in contractional tectonic activity, perhaps in response to extension in the arc and forearc regions (for example, D. L. Smith and others, 1993). Alternatively, deformation could have been strongly partitioned along shear zones in the arc and in the frontal part of the thrust belt, with relatively diminished shortening in the intervening region (Wyld and Wright, 2001). In any case, the foreland basin record indicates that the locus of major crustal shortening during this period propagated several hundred kilometers eastward into the western part of the Sevier belt in central Utah and Idaho. Throughout the central part of the Cordilleran thrust belt (including the western part of the Sevier belt), thick, aerially extensive megathrust sheets containing rheologically competent Proterozoic-Paleozoic sedimentary rocks were displaced on the order of 50 to 100 kilometers eastward during Early Cretaceous time (for example, Mitra, 1997; Yonkee, 1997; Camilleri and others, 1997). From north to south these include the Hawley Creek, Paris, Willard, Sheep Rock, Canyon Range, Wah Wah, Wheeler Pass, and Pachalka thrust sheets (fig. 10). Timing of emplacement of most of these thrust sheets is inferred from sparsely dated fluvial and lacustrine deposits in the foreland basin and thermochronology (for example, Villien and Kligfield, 1986; Kowallis and others, 1991; DeCelles and Burden, 1992; Goldstrand, 1994; Burtner and Nigrini, 1994; DeCelles and others, 1995; Sears, 2001; Currie, 2002). A few radiometric ages are available to constrain thrusting as old as ~140 Ma on the Willard thrust in northeastern Utah ($^{40}Ar/^{39}Ar$ on muscovite, Yonkee and others, 1989, 1997; apatite fission tracks, Burtner and Nigrini, 1994), ~146 Ma on the Canyon Range thrust in central Utah (apatite fission track modeling; Ketcham and others, 1996; Stockli and others, 2001), and as young as ~142 Ma on the Pachalka thrust in southeastern California (U-Pb zircon; Walker and others, 1995). Burtner and Nigrini (1994) documented authigenic illite with K-Ar ages of 140 to 145 Ma in lower Jurassic sandstone beneath the Lower Cretaceous foredeep deposits in Idaho and northern Utah. They interpreted the authigenic ages to represent burial of the foredeep region contemporaneously with exhumation and cooling on the Willard and Paris thrust sheets directly to the west. In Utah these megathrust sheets contain thick sections of strong quartzite and carbonate rocks, and exhibit broad wavelength internal strain with relatively few subsidiary thrust faults (Camilleri and others, 1997; Yonkee, 1997; Mitra, 1997). The high strength of these rocks, coupled with relatively weak fault zone rocks (Yonkee, 1997), allowed stress to be transmitted eastward over a roughly 150 kilometer wide swath of the thrust belt, giving the impression that the region directly to the west was not involved in thrusting (for example, Armstrong, 1972). In northern Utah, the Willard thrust has ~60 kilometers of slip, based on measured offset of well-constrained footwall and hanging-wall cutoffs (Yonkee, 1997). In the central Utah region, the Canyon Range thrust has at least 140 kilometers of slip, accounting for approximately 40 percent of the total shortening in the Cordilleran thrust belt at that latitude (DeCelles and others, 1995; Currie, 2002). To the south of the region of active thrust faulting, extension continued in the Bisbee-McCoy basin during Early to mid-Cretaceous time (Dickinson and Lawton, 2001a).

The oldest Cretaceous sedimentary rocks in the Cordilleran foreland basin of the western U.S.A. consist of sparsely dated Barremian(?)-Aptian fluvial deposits that extend from northern Montana to southern Utah and from the frontal part of the Cordilleran thrust belt to central Colorado (fig. 10). In western Wyoming, eastern Idaho, and eastern Utah, these strata thicken abruptly westward from less than 100 meters to more than 1000 meters (fig. 10). Flexural modeling (Jordan, 1981; Heller

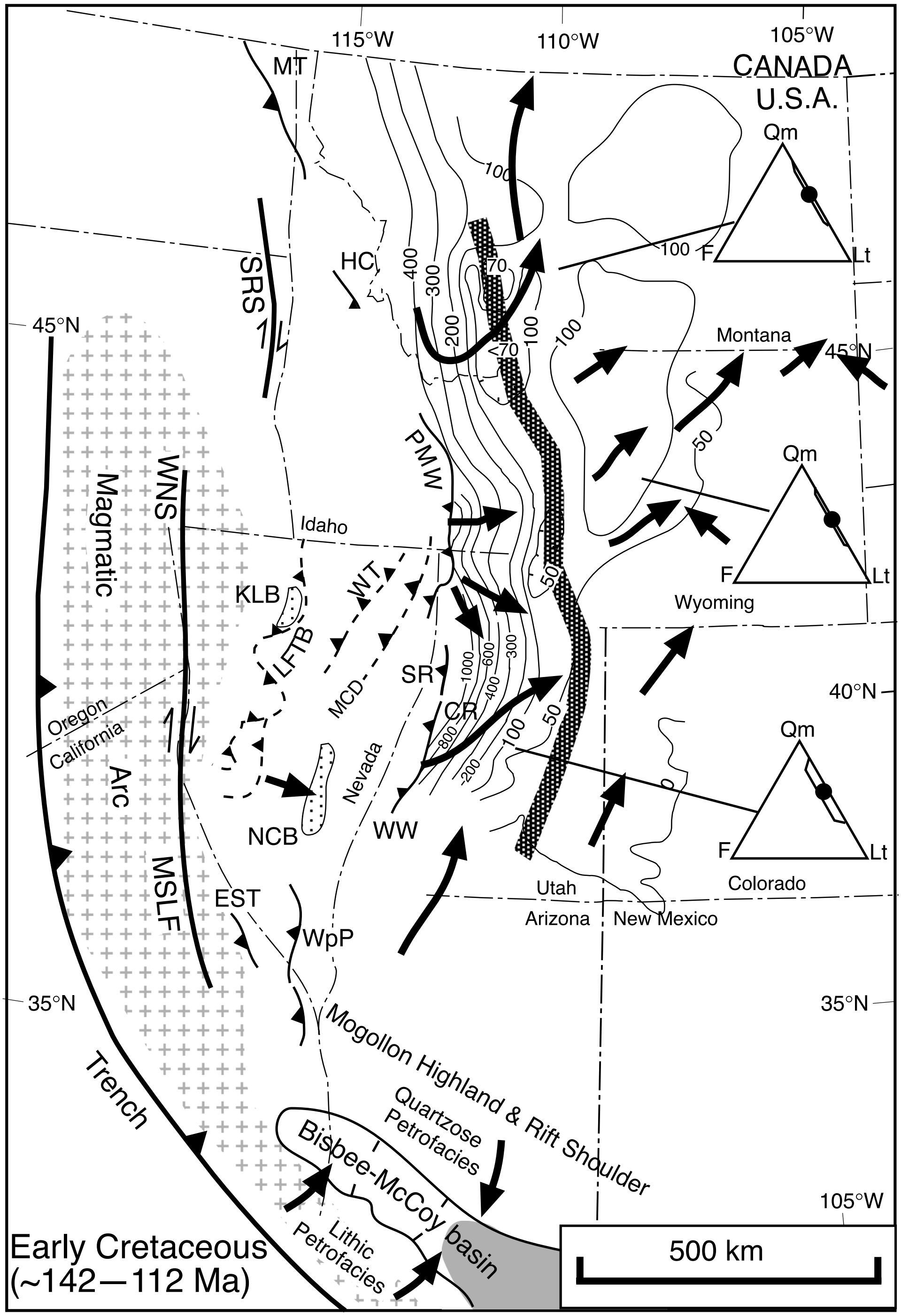

Fig. 10. Palinspastic isopach map (in meters) of Lower Cretaceous strata in the Cordilleran foreland basin system and reconstructed locations of major active tectonic elements of the Cordilleran thrust belt during Early Cretaceous time. This map depicts only the thicknesses of Barremian(?)–Aptian units. Active thrust systems (solid barbed lines) during this time interval are labeled as follows: MT, Moyie thrust (speculative); HC, Hawley Creek thrust; PMW, Paris-Meade-Willard thrust system; CR, Canyon Range thrust;

and others, 1986; Pang and Numedal, 1995; Currie, 1997, 2002) suggests that the thick, wedge-shaped western portion of the Lower Cretaceous basin fill has a shape that is consistent with flexural subsidence driven by the adjacent Cordilleran thrust belt. In addition, local coarse-grained conglomerates can be directly tied to individual thrust sheets in southern Idaho and northeastern Utah (DeCelles and others, 1993; DeCelles and Currie, 1996). Therefore, little doubt exists that the Lower Cretaceous succession was deposited in the foredeep depozone of the Cordilleran foreland basin. The Early Cretaceous flexural forebulge is defined by isopach patterns in eastern Utah, western Wyoming, and southwestern Montana (fig. 10). The forebulge probably had little if any topographic relief because it was shallowly buried by Lower Cretaceous sediment. The foreland basin south of the Canadian border was entirely nonmarine during Early Cretaceous time, with alluvial fans adjacent to the thrust belt, and fluvial systems in the foredeep, forebulge, and back-bulge regions (Schwartz, 1982). Two periods of thick (up to 30 m) lacustrine carbonate deposition took place in the foredeep of Wyoming and Montana (Holm and others, 1977; Suttner and others, 1981). Paleocurrent data from fluvial sandstone and conglomerate indicate that rivers flowed generally north-northeastward in the western part of the basin, toward a marine embayment that extended into southwestern Canada (fig. 10). Sandstone and conglomerate compositions are consistent with this reconstruction, and indicate derivation of sedimentary lithic grains and quartz from thick chert-bearing carbonate and clastic units in the thrust belt. Lower Cretaceous sandstones throughout the foreland basin are typical of the recycled orogen petrofacies of Dickinson and Suczek (1979) (fig. 10). Isopach and paleocurrent data from the distal part of the foreland suggest local reactivation of minor basement faults (Schwartz, 1982; Meyers and others, 1992; Zaleha and others, 2001). Continued fluvial deposition in the Bisbee-McCoy basin buried synrift topography and locally derived coarse-grained alluvial fan facies. By mid-Aptian time marine and marginal-marine conditions prevailed throughout most of the basin, depositing a complex mosaic of carbonate and siliciclastic facies (Mack and others, 1986; Dickinson and Lawton, 2001a).

Albian (fig. 11)

Albian time (~112 - 98.5 Ma) in the Cordilleran thrust belt was a period of widespread plutonism, metamorphism, and local crustal shortening in the hinterland (D. L. Smith and others, 1993) and continued regional scale displacement of the major thrust sheets in the western part of the Sevier belt that contain thick Proterozoic–Paleozoic strata (fig. 11). Although Albian thrust displacements in southwestern Montana have not been directly documented, the Albian synorogenic foreland basin record provides ample evidence for rapid flexural subsidence in front of thrusts older than those that have been dated in the frontal part of the Sevier belt (Schwartz, 1982; DeCelles, 1986; Schwartz and DeCelles, 1988). In the Idaho-Wyoming-Utah salient, displacement on the Willard thrust was transferred eastward onto the Meade-Laketown thrust system, producing the oldest known growth structures in the Cordilleran thrust

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SR, Sheeprock thrust; WW, Wah Wah thrust; EST, Eastern Sierra thrust belt; WpP, Winters Pass-Pachalka thrusts. Windermere thrust (WT) and Manning Canyon detachment (MCD) may have been active during this time frame. LFTB indicates position of Luning-Fencemaker thrust system, which was largely inactive by Early Cretaceous time (Wyld, 2002). KLB is King Lear basin; NCB is the Newark Canyon basin. Thick shaded line indicates approximate location of the forebulge as suggested by isopach patterns and lithofacies distributions. Shaded pattern represents region of marine inundation in Bisbee-McCoy basin. Arrows indicate generalized sediment dispersal directions. See figure 6 caption for explanation of sandstone compositional diagrams and patterns. Based on data from Suttner (1969), Furer (1970), DeCelles (1986), Vandervoort and Schmitt (1990), DeCelles and Burden (1992), Meyers and others (1992), Yingling and Heller (1992), Malone and Suttner (1992), Currie (1997, 1998), Zaleha and others (2001), Dickinson and Lawton (2001a), and other sources noted in figure 3 and table 1.

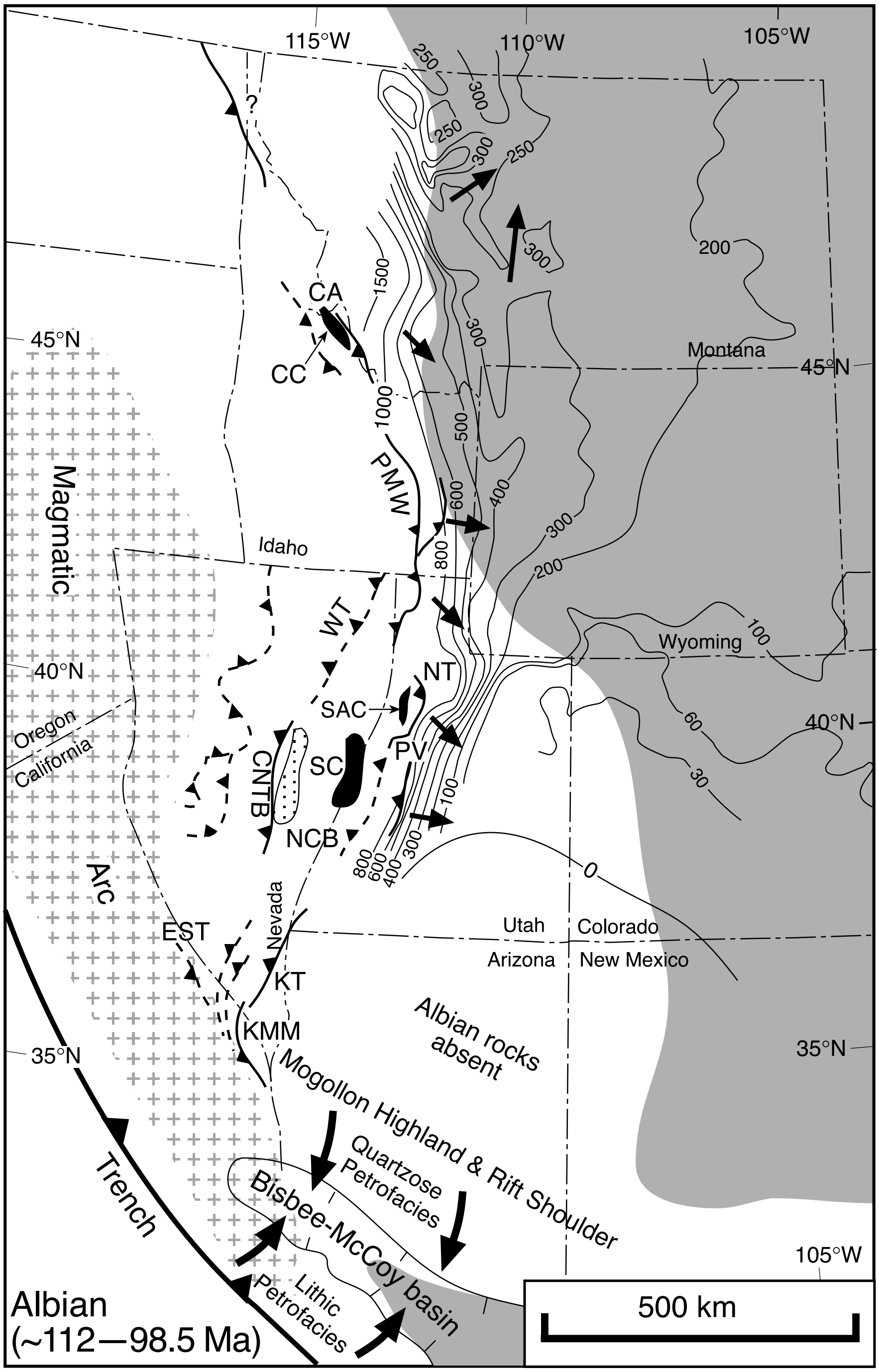

Fig. 11. Palinspastic isopach map (in meters) of Albian strata in the Cordilleran foreland basin system (after McGookey, 1972; Jordan, 1981; Schwans, 1988; Horton and others, unpublished data). Solid barbed lines indicate active thrusts; dashed barbed lines are inactive thrusts. Black areas represent basement structural culminations; shaded pattern represents region of marine inundation; plus signs indicate magmatic arc. Arrows indicate general sediment dispersal directions. Abbreviations as follows: CA, Cabin thrust; CC, Cabin culmination; PMW, Paris-Meade-Willard thrust system; WT, Windermere thrust; NT, Nebo thrust; SAC, Santaquin culmination; CNTB, Central Nevada thrust belt; SC, Sevier culmination; PV, Pavant thrust; NCB, Newark Canyon basin; EST, Eastern Sierra thrust belt; KT, Keystone thrust; KMM, Keaney/Mollusk Mine thrust.

belt (DeCelles and others, 1993). In the Charleston-Nebo salient, initial displacement on the Nebo thrust occurred during late Albian time (Constenius, ms, 1998). Because the trailing portion of the Nebo thrust carries Precambrian crystalline basement rocks that form the core of the Santaquin culmination, the culmination began to grow at this time. In central Utah, initial displacement on the Pavant thrust (Villien and Kligfield, 1986; Schwans, 1988; DeCelles and others, 1995) also gave rise to a large structural culmination, the Sevier culmination. In southern Nevada, the Keystone thrust became active during late Albian time (Fleck and Carr, 1990; Burchfiel and others, 1998), at about the same time that the Keeney/Mollusk Mine thrust system became active in southeastern California (Fleck and others, 1994; Walker and others, 1995). In the hinterland of central Nevada, the CNTB probably became active during late Albian time (Taylor and others, 2000).

Albian deposition in the foreland basin occurred under the influence of the first major marine transgression in the western interior (McGookey, 1972; Williams and Stelck, 1975; Ryer, 1977; Cobban and others, 1994). Marine waters entered the foreland region from the north and ultimately formed a through-going seaway connected to the Gulf of Mexico by mid-Albian time. Fine-grained mudrocks were deposited in offshore areas, particularly in the northern part of the basin, and deltaic, estuarine, coastal plain, and fluvial facies were deposited in the southwestern and western parts of the basin (Cobban and others, 1994). Albian isopachs suggest the presence of a thick foredeep depozone adjacent to the thrust belt, but evidence for a forebulge is equivocal (fig. 11; McGookey, 1972; Jordan, 1981). Sandstone petrofacies data define a typical quartzolithic recycled orogen provenance during Albian time (Schwartz, 1982; Lawton, 1982, 1986; DeCelles, 1986; Schwans, 1988). The Bisbee-McCoy basin was dominated by regional tectonothermal (post-rifting) subsidence, and marine carbonate deposition continued until mid-Albian time (Dickinson and Lawton, 2001a).

Cenomanian–Turonian (figs. 12 and 13)

Igneous intrusion and metamorphism during Cenomanian–Turonian time (~98.5 - 88.7 Ma) were locally intense in the Nevada hinterland (figs. 3, 12, and 13; Miller and Gans, 1989; D. L. Smith and others, 1993; Christiansen and others, 1994). In northwestern Utah and adjacent areas of Idaho and Nevada, magmatic activity was minimal, but metamorphism at temperatures of ~600°C and penetrative deformation in Proterozoic–lower Paleozoic miogeoclinal strata occurred prior to cooling through the $^{40}Ar/^{39}Ar$ blocking temperature at ~90 Ma (Wells, 1997; Hoisch and others, 2002). Wells (1997) presented data from this region that indicate at least two episodes of thrust-related burial at mid-crustal depths, separated by an episode of exhumation-induced cooling. In latest Albian–Cenomanian time, the classic, frontal part of the Sevier belt began to take shape, as the basal décollement along much of the thrust belt climbed several kilometers up to the level of a regionally extensive Cambrian shale and formed a several tens-of-kilometers-long (in the transport direction) structural flat (Royse and others, 1975; Lamerson, 1982; Dixon, 1982; Coogan and Royse, 1990). From the Cambrian level, slip was fed upward and eastward along major frontal ramps to the Permian and Jurassic levels. This transition occurred at slightly different times along the thrust belt. In the Idaho-Wyoming-Utah salient, the transition involved the development of the Laketown imbricate of the Meade thrust as well as continued slip on the Meade system (Coogan, 1992). In the Charleston-Nebo salient, the Nebo thrust continued to construct the Santaquin culmination and fed slip eastward into a frontal triangle zone (Constenius, ms, 1998). In central Utah the Pavant thrust sheet began to imbricate internally and form the Canyon Range duplex, while the trailing Sevier culmination continued to grow (DeCelles and others, 1995; Mitra, 1997; Stockli and others, 2001; Currie, 2002). In southern Nevada and southeastern California, the

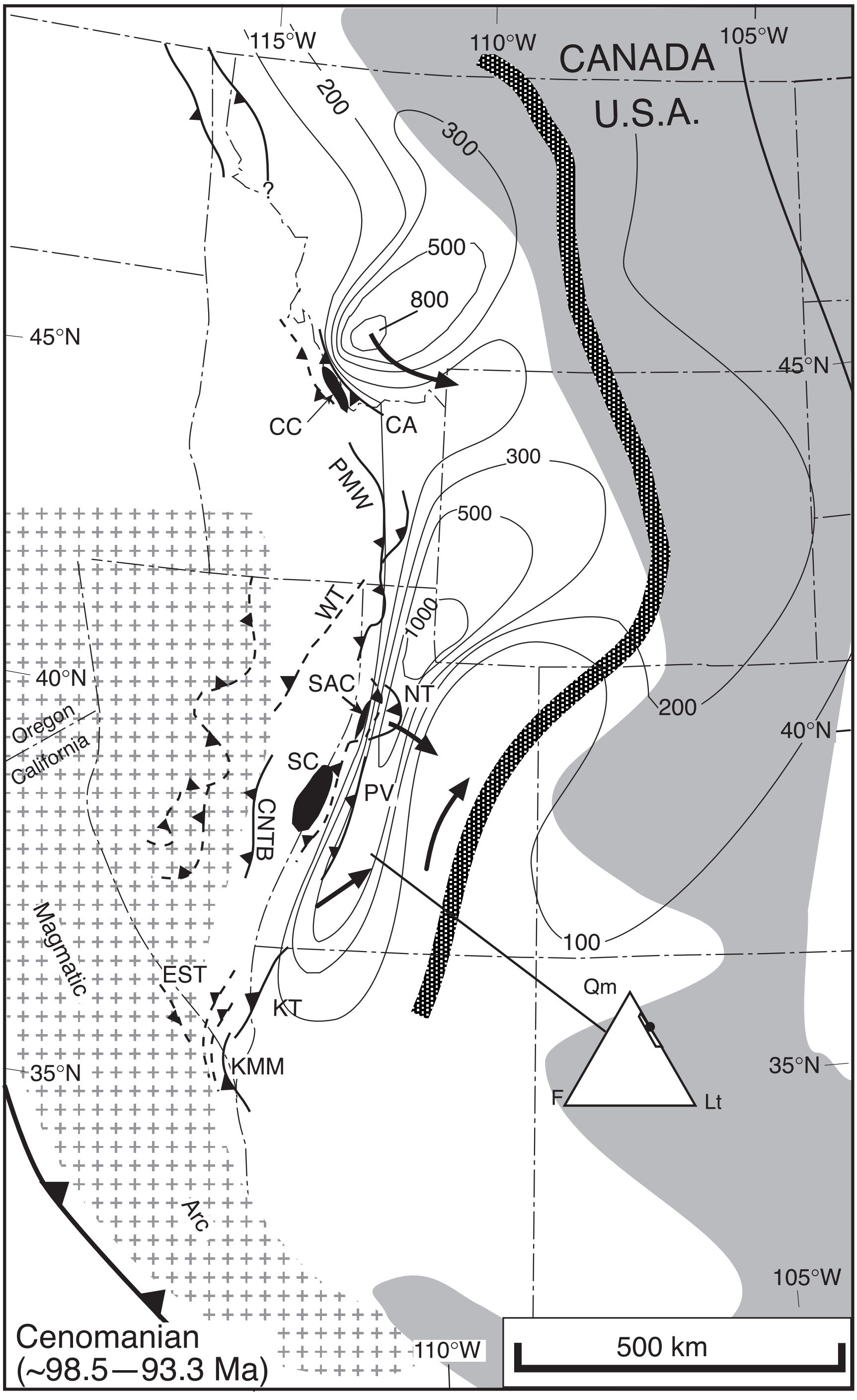

Fig. 12. Palinspastic isopach map (in meters) of lower Cenomanian strata in the Cordilleran foreland basin system (after Robinson Roberts and Kirschbaum, 1995). Solid barbed lines indicate active thrusts; dashed barbed lines are inactive thrusts. Black areas represent basement structural culminations; shaded pattern represents region of marine inundation; plus signs indicate magmatic arc; thick shaded line is possible forebulge location. Abbreviations as follows: CA, Cabin thrust; CC, Cabin culmination; PMW, Paris-Meade-Willard thrust system; WT, Windermere thrust; SAC, Santaquin culmination; NT, Nebo thrust; SC, Sevier culmination; PV, Pavant thrust; KT, Keystone thrust; KMM, Keaney/Mollusk Mine thrust; EST, Eastern Sierra thrust belt; CNTB, Central Nevada thrust belt. Petrographic data from Lawton (1986).

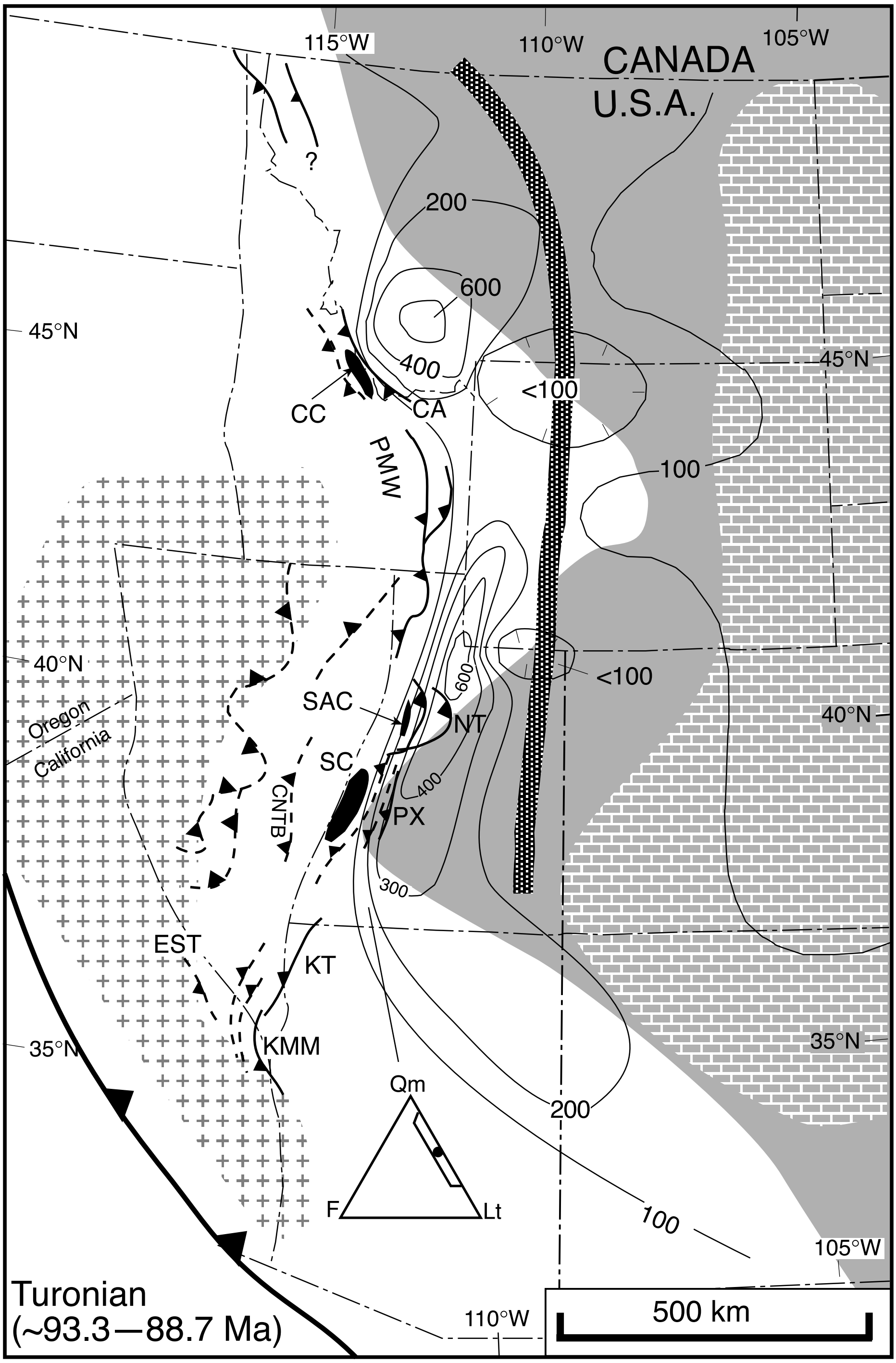

Fig. 13. Palinspastic isopach map (in meters) of upper Turonian strata in the Cordilleran foreland basin system (after Robinson Roberts and Kirschbaum, 1995). Solid barbed lines indicate active thrusts; dashed barbed lines are inactive thrusts. Black areas represent basement structural culminations; shaded pattern represents region of marine inundation; plus signs indicate magmatic arc; brickwork pattern represents region of marine carbonate deposition; thick shaded line is possible forebulge location. Abbreviations as follows: CA, Cabin thrust; CC, Cabin culmination; PMW, Paris-Meade-Willard thrust system; SAC, Santaquin culmination; NT, Nebo thrust; SC, Sevier culmination; PX, Paxton thrust; KT, Keystone thrust; Keaney/Mollusk Mine thrust; EST, Eastern Sierra thrust belt; CNTB, Central Nevada thrust belt. Petrographic data from Lawton and others (2003).

Keystone and Keeney/Mollusk Mine thrust systems probably remained active during Cenomanian time (Fleck and others, 1994; Walker and others, 1995). General timing constraints from the CNTB suggest that moderate shortening in the interior of the Cordilleran thrust belt was also occurring during Cenomanian–Turonian time (Taylor and others, 2000). Because thrusting had already propagated into eastern Utah and Idaho by Albian time, Cenomanian–Turonian thrusting in the CNTB was out-of-sequence with respect to the regional kinematic pattern.

Major marine transgressions occurred in the distal part of the Cordilleran foreland basin during Cenomanian and Turonian time. Accommodation created by rapid flexural subsidence in central Utah, western Wyoming, and southwestern Montana was filled by up to 1,600 meters of fluvial sediment that built a broad coastal plain several hundred kilometers eastward from the topographic front of the Cordilleran thrust belt (figs. 12 and 13). Quartzolithic sandstones and local conglomerates were derived from Paleozoic chert-bearing strata and Proterozoic quartzitic strata to the west and southwest. Farther east, sandy deltaic and nearshore systems developed, and organic-rich muds were deposited offshore (Gardner and Cross, 1994; Gardner, 1995). Isopach patterns suggest that a flexural forebulge was located in easternmost Utah, western Wyoming, and central Montana. White and others (2002) argued that the Turonian forebulge was situated farther east, perhaps in central Colorado. Sediment deposition continued to completely bury the forebulge, however, such that a total of ~300 meters of Cenomanian-Turonian sediment accumulated in the back-bulge region of eastern Wyoming (figs. 12 and 13). Widespread carbonate deposition took place in the eastern part of the basin for the first time (fig. 13; for example, Robinson Roberts and Kirschbaum, 1995). In the Bisbee-McCoy basin, topographic expression was increasingly subdued during Cenomanian time, and the northern rift shoulder was partly overlapped by marine and marginal marine deposits. The prolonged (~30 Myr) phase of thermotectonic subsidence in the Bisbee-McCoy basin came to an end and the Bisbee-McCoy domain became incorporated briefly into the broader retroarc foreland basin system prior to disruption by Laramide structures (Mack, 1987; Dickinson and Lawton, 2001a).

Coniacian-Santonian (fig. 14)

In Coniacian-Santonian time (~88.7 - 83.5 Ma), mid-crustal rocks in the interior of the Cordilleran thrust belt in western Utah and eastern Nevada began to experience intense crustal shortening (top-to-the east sense of shear), metamorphism up to amphibolite grade, and crustal anatexis (figs. 3 and 14; Miller and Gans, 1989; Camilleri and Chamberlain, 1997; Wells, 1997; Camilleri and McGrew, 1997). These processes continued for approximately 20 Myr. Major frontal thrust systems, some with classic 'sledrunner' ramp-flat geometry, transported slabs of Paleozoic and Mesozoic rocks eastward in the frontal Sevier belt (Royse and others, 1975; Lamerson, 1982). Uplift and exhumation of the trailing basement culminations near the Wasatch hinge line continued because the frontal thrusts rooted beneath them. Over the next 35 Myr the frontal thrust belt propagated irregularly eastward into previously deposited foreland basin deposits. In the Idaho-Wyoming-Utah salient the Crawford (during Coniacian) and Absaroka (during late Santonian) thrusts developed; in the Charleston-Nebo salient the Charleston thrust experienced its first major increment of slip; and in the central Utah portion of the thrust belt the Paxton thrust formed. In southern Utah and Nevada, sediment provenance data and structural relationships suggest that the Blue Mountains thrust and the Keystone thrust system were active (Goldstrand, 1994; Carpenter and Carpenter, 1994; Burchfiel and others, 1998). The front of the thrust belt was largely buried by a thick apron of coarse-grained, synorogenic conglomerate, and growth structures were preserved in many locations (Pivnik, 1990; DeCelles, 1994; DeCelles and others, 1995; Talling and others, 1995). To the east, coastal plain and

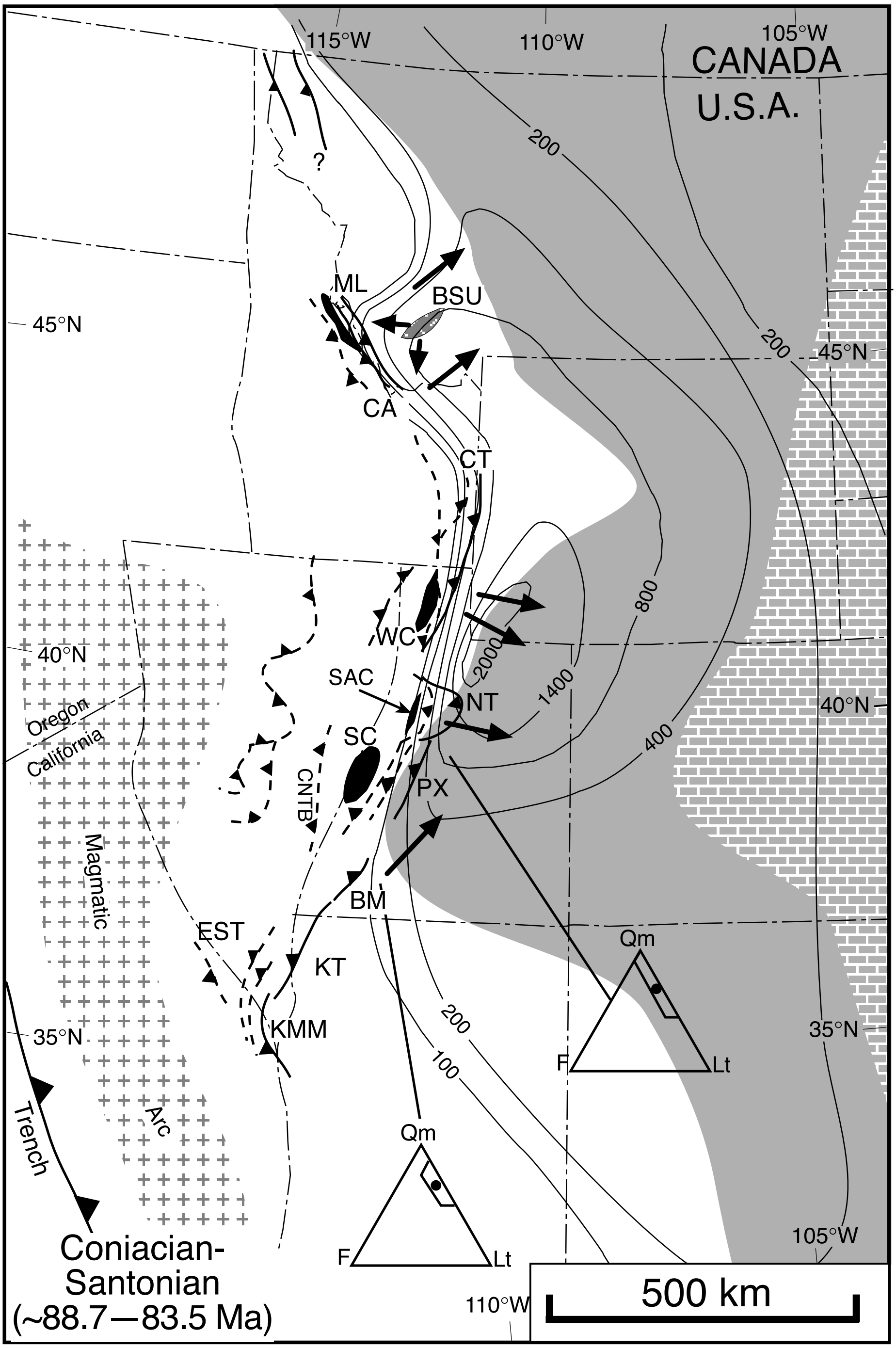

Fig. 14. Palinspastic isopach map (in meters) of middle Santonian strata in the Cordilleran foreland basin system (after DeCelles, 1994; Robinson Roberts and Kirschbaum, 1995; Talling and others, 1995) and Coniacian-Santonian tectonic activity in the thrust belt. Solid barbed lines indicate active thrusts; dashed barbed lines are inactive thrusts. Black areas represent basement structural culminations; shaded pattern represents region of marine inundation; plus signs indicate magmatic arc; brickwork pattern represents region of marine carbonate deposition. Arrows indicate sediment dispersal directions. Abbreviations as follows: BSU, Blacktail-Snowcrest intraforeland uplift; CA, Cabin thrust; ML, Medicine Lodge thrust; CT, Crawford thrust; WC, Wasatch culmination; SAC, Santaquin culmination; NT, Nebo thrust; SC, Sevier culmination; PX, Paxton thrust; KT, Keystone thrust; KMM, Keaney/Mollusk Mine thrust; EST, Eastern Sierra thrust belt; BM, Blue Mountain thrust; CNTB, Central Nevada thrust belt. Petrographic data from Lawton (1986), and Lawton and others (2003), and Horton and others (unpublished data).

deltaic systems fed clastic sediment offshore to form thick marine shales (Fouch and others, 1983). Carbonate deposition continued in the eastern part of the basin. The regional isopach pattern shows the beginning of the demise of the well-organized flexural foreland basin system; the foredeep was poorly defined and no evidence exists for a forebulge in the isopach pattern (fig. 14). The highly asymmetrical pattern of regional subsidence that characterized the Turonian began to shift eastward and become more symmetrical, suggesting that dynamic subsidence (for example, Gurnis, 1992) was beginning to influence the Cordilleran retroarc region (Pang and Nummedal, 1995; D. Nummedal, written communication, 2003).

Campanian (fig. 15)

Peak metamorphic pressures and temperatures were attained in mid-crustal rocks in the interior of the thrust belt during Campanian time (~83.5 - 71.3 Ma; figs. 3 and 15). Although shortening estimates are not available, peak metamorphic conditions surpassed 800°C and 9 kbar at depths of 10 to 30 kilometers (Miller and Gans, 1989; Wells, 1997; Camilleri and Chamberlain, 1997; Camilleri and McGrew, 1997; McGrew and others, 2000; Lee and others, 2003). For example, McGrew and others (2000) reported clockwise metamorphic pressure-temperature paths indicating tectonic burial of pelitic rocks to depths >30 kilometers in the East Humboldt Range in northeast Nevada before ~84 Ma. Magmatism, deformation, and metamorphism were intimately associated spatially and temporally, and igneous intrusions were particularly influential in transferring heat and promoting higher-grade metamorphism at relatively shallow crustal levels (Miller and others, 1988; Miller and Gans, 1989). In northwestern Utah and northeastern Nevada, Barrovian metamorphism was driven mainly by crustal thickening, with only minor magmatism (Wells, 1997; Wells and others, 1997; Hoisch and others, 2002). During early Campanian time, the interior part of the Cordilleran thrust belt also began to experience local exhumation and cooling, possibly in response to tectonic unroofing by large-magnitude extension along low-angle normal faults rooted deep in the middle crust (Hodges and Walker, 1992; Wells, 1997; Wells and others, 1997; Camilleri and others, 1997). Extension and shortening alternated rapidly in time (Wells, 1997; Hoisch and others, 2002).

Frontal thrust systems in the Sevier belt continued to propagate eastward (fig. 15), with the Lewis-Eldorado-Hoadley, Lombard, and Medicine Lodge thrust systems forming in western Montana (Schmitt and others, 1995; Sears, 2001) while the Absaroka thrust remained active in the Idaho-Wyoming-Utah salient (Lamerson, 1982; Pivnik, 1990; DeCelles, 1994). In the Charleston-Nebo salient, the basal thrust linked with the Uinta Basin-Mountain Boundary (Uinta BMB) thrust along the south flank of the Uinta Mountains uplift. Eastward slip on the Uinta block was thus coupled to displacement on thrusts in the frontal Sevier belt (Bruhn and others, 1983; Bradley and Bruhn, 1988; Bryant and Nichols, 1988; Constenius, ms 1998, 1999). In central Utah, a large antiformal duplex formed in mechanically incompetent Mesozoic rocks (the Paxton duplex; DeCelles and others, 1995) and the Gunnison thrust developed along the front of the thrust belt (Lawton and others, 1993, 1997). In southwestern Utah the Iron Springs thrust developed (Goldstrand, 1994; Lawton and others, 2003). The Maria fold-thrust belt and Mule Mountains thrust system in western Arizona were active sometime before ~79 Ma and between ~79 - 70 Ma (Tosdal, 1990), respectively.

During late Campanian time, magmatism began to migrate inboard into the central Rocky Mountain region in response to a decrease in the angle of subduction of the Farallon plate (Dickinson and Snyder, 1978), possibly owing to subduction of an aseismic ridge (Henderson and others, 1984). This second great eastward sweep of magmatic activity was more compositionally coherent and of greater volume than the Late Jurassic event (Armstrong and Ward, 1993; Christiansen and others, 1994; Barton, 1996). For the first time, Laramide intraforeland basement uplifts began to emerge

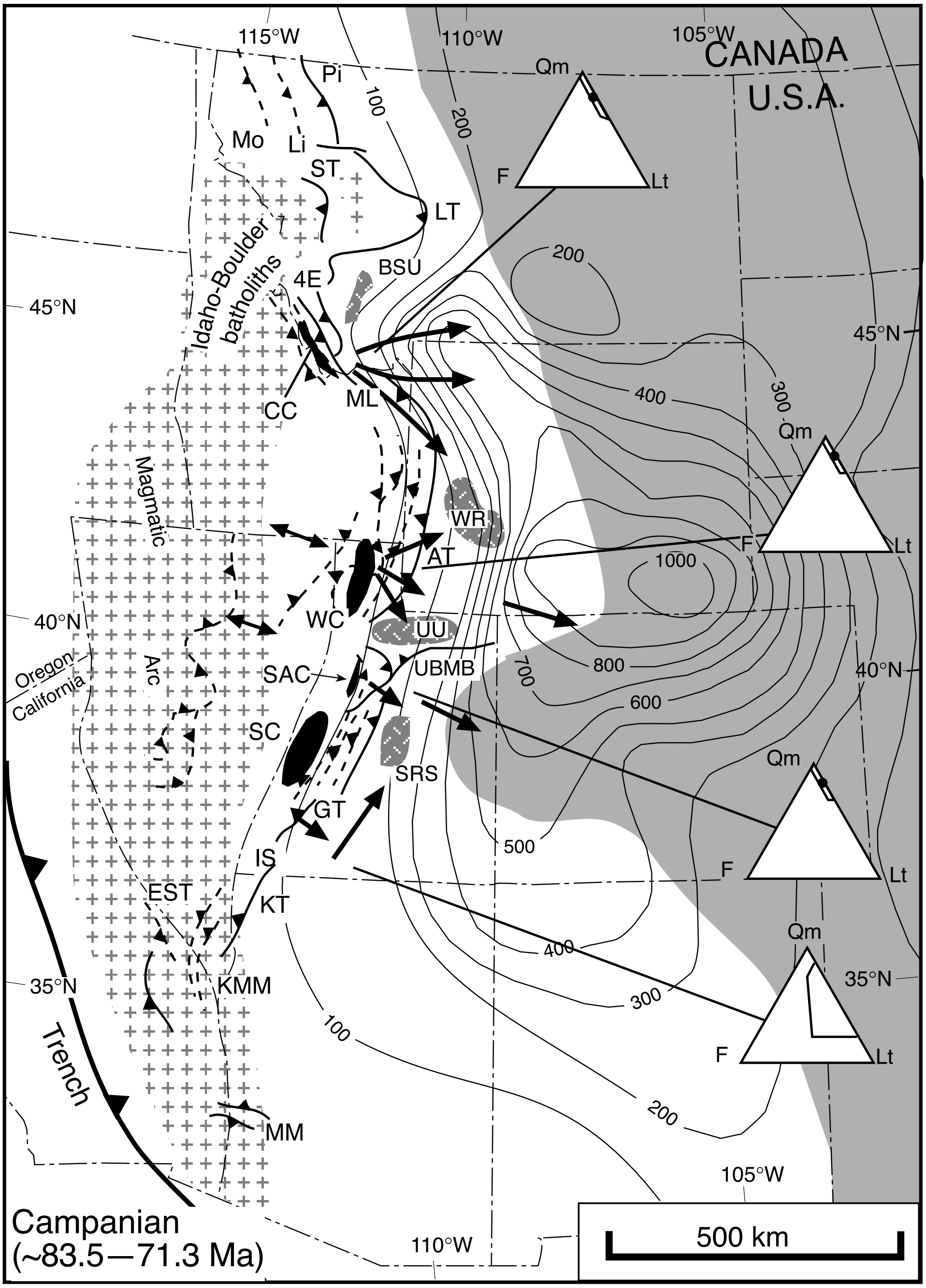

Fig. 15. Palinspastic isopach map (in meters) of lower Campanian strata in the Cordilleran foreland basin system (after Robinson Roberts and Kirschbaum, 1995). Gray jackstraw pattern represents Laramide intraforeland uplifts; solid black areas represent basement structural culminations; shaded pattern represents region of marine inundation; plus signs indicate magmatic arc. Although Laramide basement blocks began to rise during Campanian time, they did not strongly influence isopach patterns until late Campanian time; therefore only the broader pattern of regional thickness trends is discernible in this map. Abbreviations as follows: Pi, Pinkham thrust; Li, Libby thrust; Mo, Moyie thrust; ST, Sapphire thrust; LT, Lombard thrust; 4E, Four-Eyes Canyon thrust; ML, Medicine Lodge thrust; CC, Cabin culmination; AT, Absaroka thrust; UBMB, Uinta Basin-Mountain Boundary thrust; MM, Mule Mountain-Maria thrust belt; WC, Wasatch culmination; SAC, Santaquin culmination; SC, Sevier culmination; GT, Gunnison thrust; IS, Iron Springs thrust; EST, Eastern Sierra thrust belt; KT, Keystone thrust; KMM, Keaney/Mollusk Mine thrust; UU, Uinta Mountains uplift; SRS, San Rafael Swell; WR, Wind River uplift; BSU, Blacktail-Snowcrest uplift. Double-headed arrows indicate locations of mid-crustal extension. Other arrows indicate sediment dispersal directions. Petrographic data from Lawton (1986), Lawton and others (2003), and Horton and others (unpublished data).

and disrupt the regional pattern of subsidence in the foreland region (fig. 15). The oldest of these formed in southwestern Montana, where the thin-skinned thrust belt abutted against the northwestern promontory of Archean foreland basement inherited from the Proterozoic rifting event (Schmidt and Garihan, 1983; Perry and others, 1988). The Wind River uplift in west-central Wyoming (Dorr and others, 1977; Shuster and Steidtmann, 1988) and the San Rafael Swell in central Utah (Lawton, 1983; Guiseppe and Heller, 1998) began to rise, influencing sediment thickness, provenance, and paleocurrent directions.

A major unconformity beveled the frontal part of the Sevier belt during early Campanian time, and this surface was subsequently buried by enormous fluvial megafans composed of quartzite-rich gravels derived from the trailing Proterozoic thrust sheets (Love, 1972; Lindsey, 1972; Schmitt and Steidtmann, 1990; DeCelles, 1994; DeCelles and Cavazza, 1999). With the emergence of intraforeland uplifts, the zone of greatest sediment accumulation shifted eastward. The western flank of the basin fill tapered westward over a distance of ~350 kilometers from a maximum thickness of ~1 kilometer in southeastern Wyoming to less than 100 meters on top of the frontal Sevier belt (fig. 15). Coarse-grained, quartzolithic sediment was derived from Proterozoic–Paleozoic quartzites and carbonates and Mesozoic eolianites exposed in the thrust belt (fig. 15); feldspathic detritus was transported northeastward from the Mogollon highland (Lawton and others, 2003).

The Campanian isopach map is inconsistent with a flexural thrust-loading mechanism of subsidence (Cross, 1986), and suggests that dynamic subsidence began to exert the dominant control on regional sediment distribution (Mitrovica and others, 1989; Pang and Nummedal, 1995).

Maastrichtian–Early Eocene (fig. 16)

Maastrichtian–early Eocene time (~71.3 - 55 Ma) marked the climax of Laramide intraforeland uplift (for example, Cross and Pilger, 1978; Cross, 1986; Dickinson and others, 1988) and the last major phase of thin-skinned shortening in the Cordilleran thrust belt (fig. 16). However, both Cordilleran thrusting and Laramide uplift continued locally into Eocene time. Thus, although thrusting in the Sevier belt commenced earlier, the Laramide orogenic episode overlapped completely in time with Sevier thrusting, a fact that must be incorporated into models explaining the Laramide event (Allmendinger, 1992). Laramide sediments were derived from adjacent basement-cored uplifts and their carapaces of Paleozoic-Mesozoic cover strata, and therefore generally evolved from quartzolithic to feldspatholithic compositions through time (fig. 16). The eastward migration of the front of arc magmatism reached farthest inboard (>1000 km east of the trench) during latest Paleocene and early Eocene time (fig. 17; Coney and Reynolds, 1977; Constenius, 1996).

Although many Laramide uplifts are not well dated, the syntectonic sedimentary rocks associated with several uplifts from southwestern Montana to the Bighorn Mountains have been dated well enough to allow for the recognition of an eastward sweep over a period of ~25 to 30 Myr, beginning in Campanian (or perhaps Santonian) time and ending in Eocene time (DeCelles and others, 1987; Perry and others, 1988; DeCelles and others, 1991; Hoy and Ridgway, 1997). Similarly, Laramide uplifts propagated eastward from central Utah to central Colorado over an ~15 to 20 Myr timespan (Lawton, 1983; Kluth and Nelson, 1988; Raynolds, 2002). The eastward propagation of the deformation front in basement rocks of the foreland is analogous to typical thrust belt behavior, and suggests that the Laramide region was integrated with the Cordilleran orogenic wedge (Livaccari, 1991; Erslev, 1993, 2001). If Laramide intraforeland faults branch upward from a regional mid- or lower-crustal shear zone, as suggested by Erslev (1993), then the Laramide region is simply the frontal part of the Cordilleran orogenic wedge. Alternatively, if the Laramide thrusts were created in response to basal traction on North American lithosphere by the low-angle subducting Farallon plate (Dickinson and Snyder, 1978; Bird, 1998), then it is likely that Laramide and Cordilleran thrust belt structures behaved independently.

Thrusting in the frontal Sevier belt continued with emplacement of the Tendoy, Lima, Lombard, and Indian Creek thrust sheets in southwestern and west-central Montana; the Hogsback thrust in Wyoming; internal back-thrusting and triangle-zone development in the Charleston-Nebo salient; and continued slip on the Gunnison and Iron Springs thrust systems in central and southern Utah. These thrusts involved slip distances on the order of a few tens of kilometers. Thrusting during this interval in northwestern Montana presents an apparent enigma, as the Lewis-Eldorado-Hoadley thrust system transported a >15 kilometer thick slab of Proterozoic strata up to ~140 kilometers eastward. Timing constraints on Lewis-Eldorado-Hoadley thrust emplacement bracket only its youngest interval of slip (between 74 - 59 Ma; Sears, 2001), however, and it is possible that it might have experienced earlier episodes of slip. The frontal part of the northwest Montana thrust belt was also active during Paleocene–early Eocene time (Sears, 2001).

DISCUSSION

The following discussion addresses five fundamental questions in Cordilleran retroarc tectonics: (1) What was the timing of initial shortening in the retroarc region, and how does this earliest shortening fit within the broader pattern of retroarc orogenesis? (2) What was the temporal relationship between hinterland extension and frontal thrusting in the Cordilleran thrust belt? (3) What were the dominant controls on regional subsidence in the foreland basin system, and how did these mechanisms vary in time? (4) Are tectonic events in the western hinterland region related to younger events farther east in a continuum of eastward-propagating deformation, or are the various thrust belts of the retroarc unrelated? (5) Is there a relationship between retroarc thrusting and arc magmatism in the Cordillera?

Regional Patterns of Orogenesis

Figures 17 and 18 illustrate the large-scale temporal and spatial patterns of kinematic, metamorphic, igneous, and sedimentary events in the Cordilleran thrust belt in western interior U.S.A. Thrust faulting propagated eastward through time, such that initial thrust displacements in the frontal Sevier belt lagged ~60 Myr behind the earliest thrust displacements in the LFTB (fig. 17). Thrusting was strongly influenced by pre-existing stratigraphic geometry in the Eocambrian–early Paleozoic miogeocline (for example in western Wyoming; Royse, 1993a), and embayments and promontories in the underlying rifted Precambrian basement (Woodward, 1988) as well as at least one large embayment formed during late Paleozoic Ancestral Rocky Mountains deformation. The most influential basement promontories were located in southwestern Montana and southeastern California, at opposite ends of the classic Sevier belt. In these areas Sevier thrusts cut across the regional trend of the miogeocline and incorporated basement slivers. In southwestern Montana and northwestern Wyoming, Sevier belt thrust faults interfered with Laramide basement uplifts (Kulik and Schmidt, 1988; Craddock and others, 1988).

The character of deformation also changed dramatically depending upon the rocks involved in thrusting: In the fine-grained Triassic strata of the LFTB, thrusts are closely spaced and the rocks developed a penetrative cleavage at low-grade metamorphic conditions (Wyld, 2002). Thrusts, or more likely mid-crustal shear zones, incorporated slices of crystalline North American basement rocks in eastern Nevada and western Utah, near the shelf-to-slope facies transition in the Proterozoic–Paleozoic miogeocline (Camilleri and others, 1997). During Early Cretaceous time, the spacing of thrusts increased dramatically and internal strain was significantly reduced as thrusts

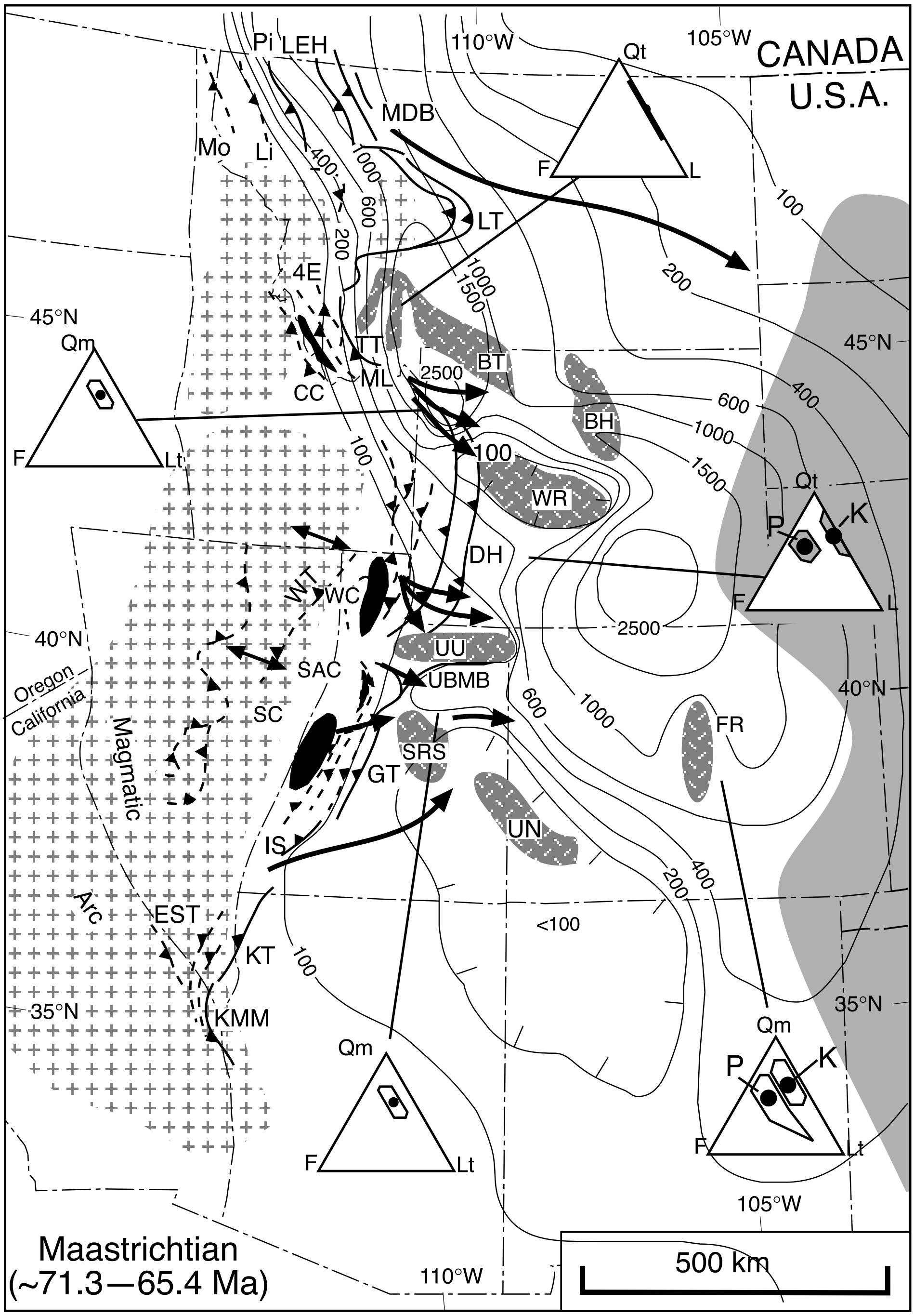

Fig. 16. Palinspastic isopach map (in meters) of Maastrichtian-Paleocene strata in the Cordilleran foreland basin system (after Robinson Roberts and Kirschbaum, 1995). Gray jackstraw pattern represents Laramide intraforeland uplifts; solid black areas represent basement structural culminations; shaded pattern represents region of marine inundation; plus signs indicate magmatic arc. Solid barbed lines indicate active thrusts; dashed barbed lines are inactive thrusts. Large arrows indicate sediment dispersal patterns. Abbreviations as follows: LEH, Lewis-Eldorado-Hoadley thrust; Pi, Pinkham thrust; Li, Libby thrust; Mo, Moyie thrust; LT, Lombard thrust; TT, Tendoy thrust; 4E, Four-Eyes Canyon thrust; CC, Cabin culmination;

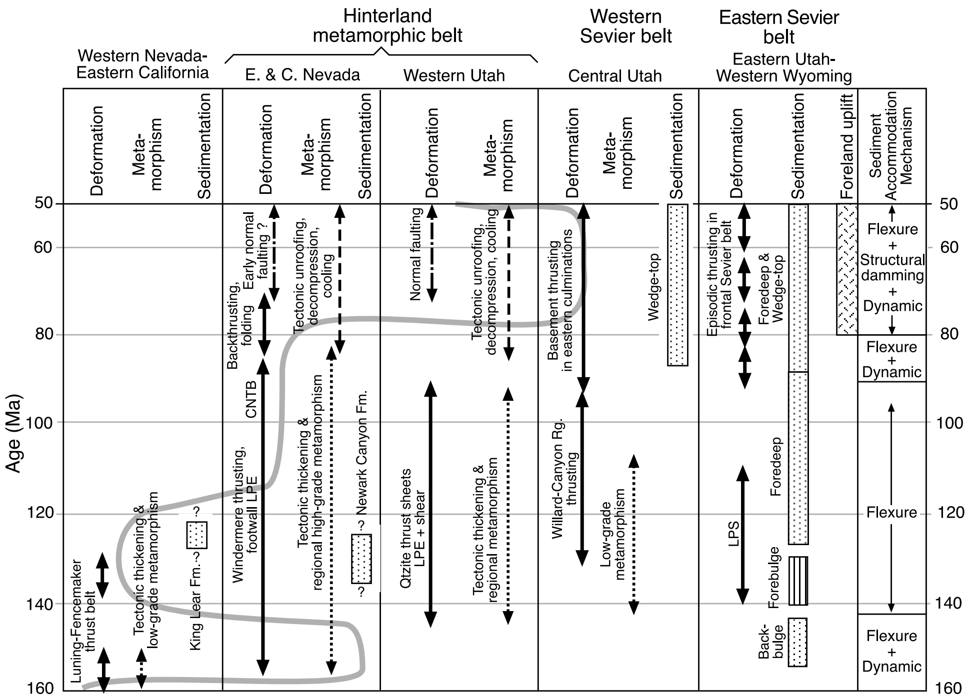

Fig. 17. Diagram illustrating timing of major tectonic, metamorphic, and depositional events in an east-west transect across the Cordilleran orogenic belt and foreland basin system at the approximate latitude of 41°N (northern Utah). Compiled from sources discussed in the text and previous figures. Gray line indicates the approximate eastern front of magmatic activity (after Christiansen and others, 1994).

broke eastward along regional flats below thick Precambrian quartzites and Paleozoic carbonates (figs. 17 and 18). These Precambrian–Paleozoic "megathrust sheets" (Camilleri and others, 1997) behaved coherently over long distances in the transport direction (Mitra, 1997), and displacements were >50 kilometers. A belt of basement culminations (Sevier, Santaquin, Wasatch, Cabin culminations; fig. 18) formed approximately at the basement step associated with the eastern limit of late Precambrian rifting. Eastward propagating thrust faults may have excised basement slices from pre-existing normal fault blocks. To the east of the basement culminations, thrusts generally propagated above the basement-cover interface. By about 80 to 75 Ma the basal thrust (or shear zone) of the orogenic wedge propagated into crystalline basement in the distal foreland region, forming the Laramide uplifts and intervening basins (fig. 17).

DH, Darby-Hogsback thrusts; UBMB, Uinta Basin-Mountain Boundary thrust; GT, Gunnison thrust; IS, Iron Springs thrust; KT, Keystone thrust; KMM, Keaney/Mollusk Mine thrust; EST, Eastern Sierra thrust belt; UN, Uncompahgre uplift; SRS, San Rafael swell; UU, Uinta Mountains uplift; FR, Front Range; WR, Wind River Range; BH, Bighorn Range; BT, Beartooth Range. Double-headed arrows indicate locations of mid-crustal extension. Other arrows show sediment dispersal patterns. Petrographic compositions of selected typical sandstones are from Lindsey (1972), Lawton (1986), Ingersoll and others (1987), Shuster and Steidtmann (1988), and Wilson (2002). Ternary diagrams are in terms of Qm (monocrystalline quartz), F (total feldspar), and Lt (total lithic fragments), Q (total quartzose grains, including chert), and L (non-quartzose lithic fragments). In the two eastern ternary diagrams, compositional groups labeled K and P refer to Cretaceous and Paleocene samples, respectively.

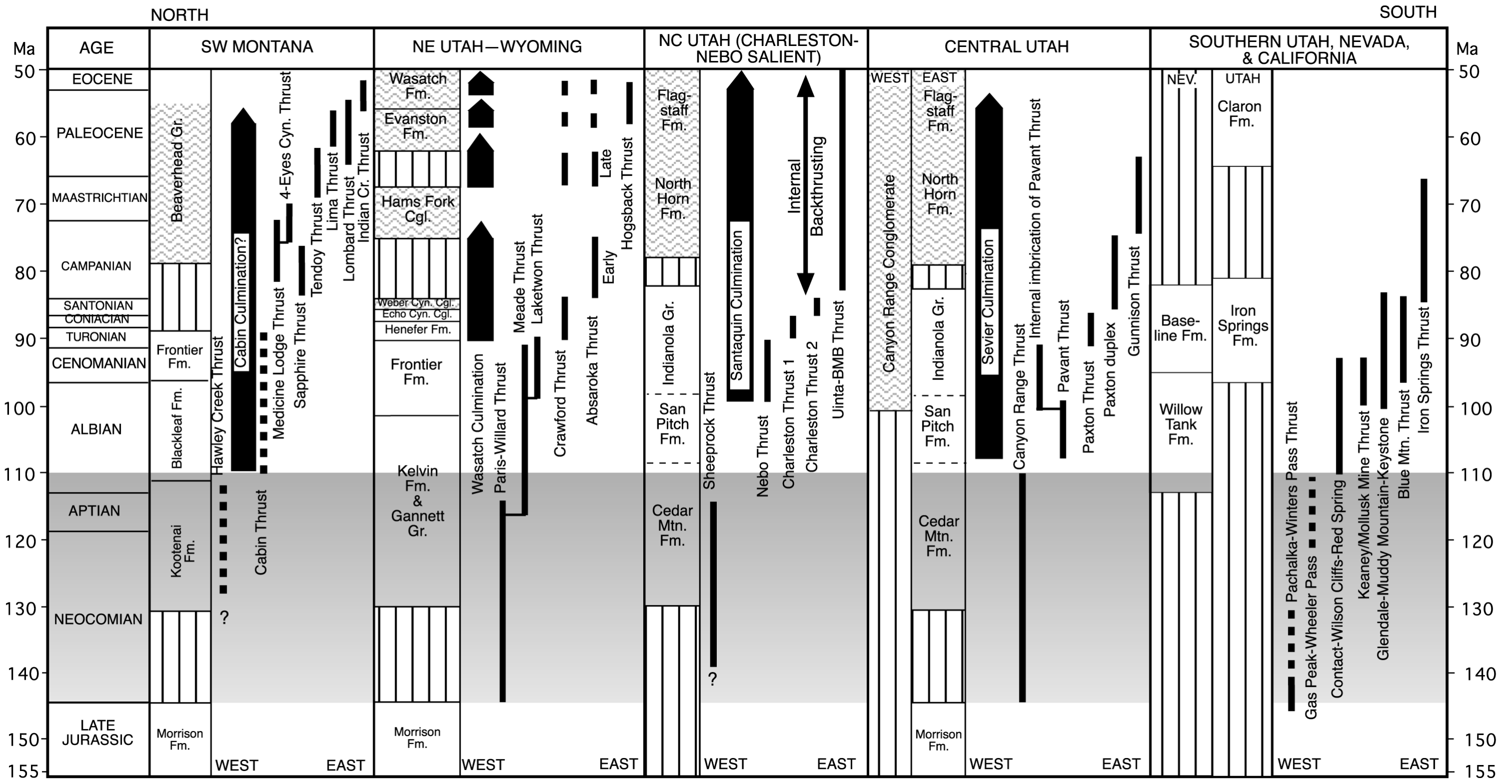

Fig. 18. Diagram illustrating the kinematic history in five selected segments of the frontal (Sevier belt) part of the Cordilleran thrust belt. The gray shaded area in lower part of the diagram indicates the period of emplacement of the Precambrian quartzite megathrust sheets. Black vertical lines indicate general estimates of timing of thrust displacement; dashed vertical lines (in southwest Montana panel) are speculative. Thick vertical arrows represent times of active uplift in basement culminations.

Total shortening estimates along the Cordilleran thrust belt (including shortening in the Luning-Fencemaker, Central Nevada, Sevier, and Laramide systems), though subject to large uncertainties, range from less than 100 kilometers in the south to more than 350 kilometers in central Utah (Elison, 1991; this paper). These estimates are most likely minimum values, however, because significant amounts of internal strain are not included (Mitra, 1994, 1997) and shortening amounts on the western thrust systems are not well known. Because all of this shortening is in upper crustal rocks, an equivalent length of lower crust and lithosphere must have been subducted westward beneath the Cordilleran magmatic arc. The potential implications of this are discussed below.

Barrovian metamorphism reached peak pressures of ~8 to 9 kbar and temperatures of 500° to 800°C in eastern Nevada, suggesting that mid-crustal rocks were buried 10 to 25 kilometers deeper than their original stratigraphic depths (Miller and Gans, 1989; Hodges and others, 1992; D. L. Smith and others, 1993; Wells, 1997; Camilleri and Chamberlain, 1997; McGrew and others, 2000; Lee and others, 2003). Metamorphism in the LFTB to the west and the Proterozoic-Paleozoic megathrust sheets to the east was much lower grade (greenschist facies), and the rocks of the frontal Sevier belt are unmetamorphosed (fig. 17). Thus, it may be inferred that the interior portion of the Cordilleran thrust belt was the locus of maximum crustal thickening (Coney and Harms, 1984). A similar pattern of maximum crustal thickening coincident with maximum metamorphic conditions has been documented in the modern central Andes (McQuarrie, 2002) and Himalayan thrust belts (Hodges, 2000), as well as other large mountain belts.

Magmatic activity associated with the Cordilleran arc swept hundreds of kilometers eastward in two major episodes during Late Jurassic and Late Cretaceous–early Tertiary time (figs. 17, 19B and 19E; Coney and Reynolds, 1977; Barton, 1990, 1996; Christiansen and others, 1994; Constenius, 1996). The younger of these magmatic “sweeps” generally is considered to be related to a decrease in the angle of subduction of the Farallon plate and coeval onset of Laramide crustal shortening (Dickinson and Snyder, 1978; Cross, 1986; Bird, 1998). The Jurassic event is unlikely to have been caused by a similar process because no coherent magmatic front developed (Christiansen and others, 1994). Instead, diffuse crustal melts formed in response to addition of mantle heat, perhaps via mafic intrusions in the lower crust (for example, Barton, 1990; Lee and others, 2003). This process could have been driven by asthenospheric upwelling after a lithospheric delamination event (fig. 8C). Similar transient magmatic sweeps have been documented in the retroarc thrust belt of the central Andes and attributed to decreases in the subduction angle of the Nazca plate and/or lithospheric delamination (Kay and Kay, 1993; Beck and Zandt, 2002).

By Late Cretaceous time, the paleogeography of the Cordilleran thrust belt was characterized by a relatively low-relief hinterland region and a rugged frontal thrust belt (fig. 19D). Although paleoelevation estimates are not available, crustal thicknesses of 50 to 60 kilometers are predicted in western Utah on the basis of shortening estimates in the thrust belt (Elison, 1991; DeCelles and others, 1995; Camilleri and others, 1997; Currie, 2002), restoration of Tertiary extension (Coney and Harms, 1984; Miller and Gans, 1989), and metamorphic mineral assemblages in the hinterland (Wells and others, 1997; Camilleri and others, 1997; McGrew and others, 2000; Lee and others, 2003). Based on compiled paleobotanical estimates of paleoelevation, Chase and others (1998) asserted that high regional elevations (3 - 4 km) existed throughout the western Sevier belt and its hinterland during Eocene time. Therefore, it is plausible that regional paleoelevation in Nevada was >3 kilometers, perhaps forming a “Nevadaplano” (fig. 19) analogous to the central Andean Plateau of Bolivia

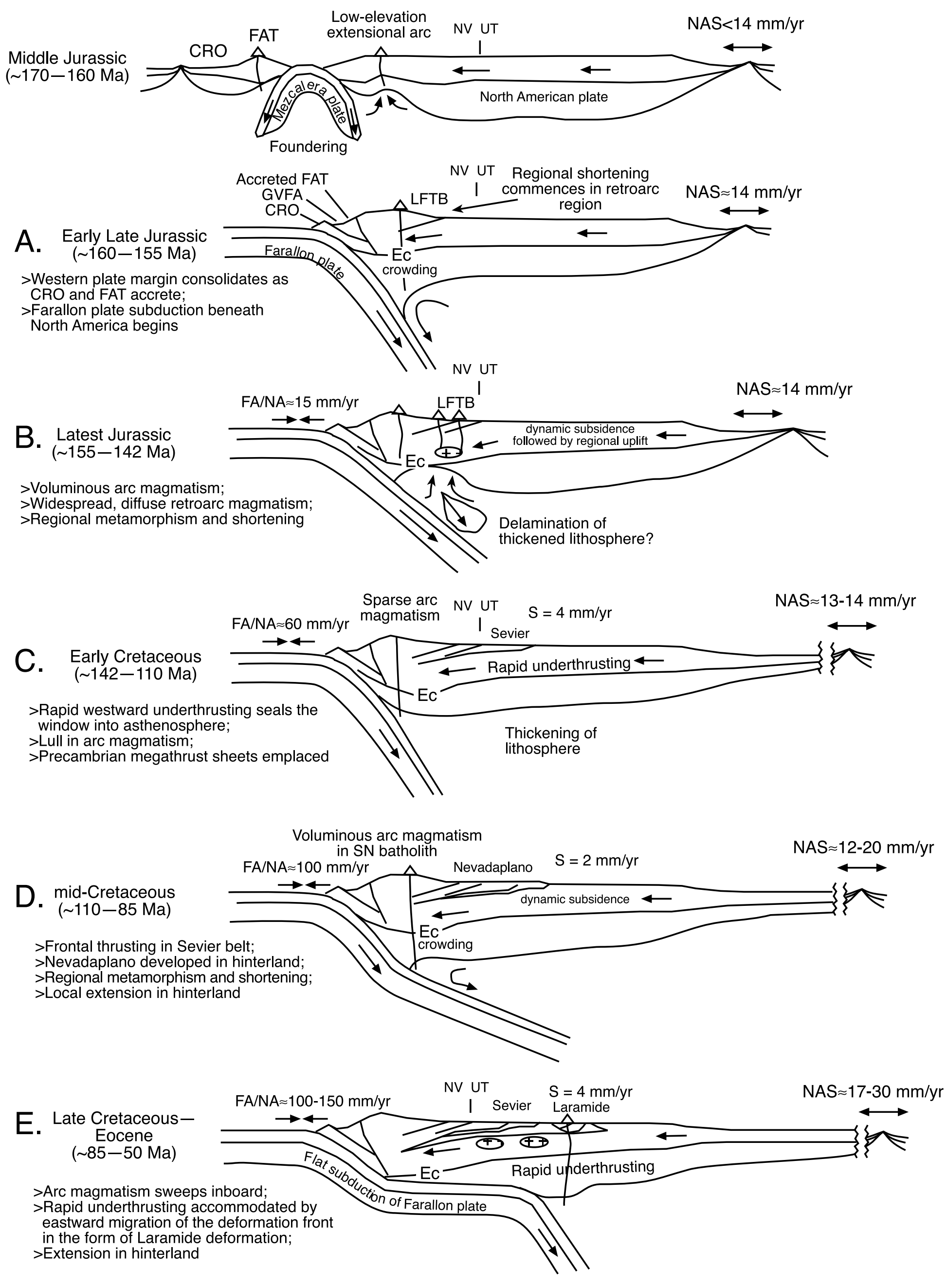

Fig. 19. Sequential plate-scale kinematic reconstruction for the Cordilleran orogenic belt at the latitude of Utah and Nevada. See text for discussion. Abbreviations as follows: CRO, Coast Range ophiolite; FAT, Foothills arc terrane; GVFA, Great Valley forearc basin; SN, Sierra Nevada; Ec, eclogitized lower crust and lithosphere; NAS, North Atlantic spreading rate; S, shortening rate; FA/NA, rate of convergence between Farallon and North American plates; NV|UT, Nevada-Utah border. Although the Middle Jurassic panel incorporates the Mezcalera plate of Dickinson and Lawton (2001b), other models involving marginal oceanic basins and offshore arcs are also viable (for example, Harper and Wright, 1984; Dickinson and others, 1996).

and northern Argentina (for example, Isacks, 1988; Allmendinger and Gubbels, 1996; McQuarrie, 2002).

Causes of Extension in the Interior of the Cordilleran Thrust Belt

Within the limits of available temporal resolution, hinterland extension and frontal thrusting were coeval in the Cordilleran thrust belt during Late Cretaceous time (fig. 17; Wells and others, 1990; Hodges and Walker, 1992; Camilleri and others, 1997; Wells, 1997). Inspired in part by Neogene extensional fault systems in the high, interior parts of the Himalayan thrust belt (for example, Hodges and others, 1996), recent authors have suggested that extension in the Cordilleran hinterland was caused by gravitational collapse in response to the attainment of a limiting elevation (Hodges and Walker, 1992) and/or alternating periods of shortening and extension in response to changes in key physical parameters governing the behavior of the orogenic wedge (Wells, 1997; McGrew and others, 2000). Paradoxically, no surface-breaking large-scale normal faults have been documented that might have facilitated this crustal extension. Hodges and Walker (1992) suggested that extension was confined to the ductile middle and lower crust, and was accommodated by extensional faults that transferred slip to the Moho instead of the surface. Although these authors suggested that extension was fed westward beneath the arc, it seems more plausible that lower (or middle) crustal rocks would have moved eastward (where the crust was presumably thinner), rather than westward, under the influence of gravitational forces. McQuarrie and Chase (1999) suggested that eastward mid-crustal flow from regions of overthickened crust in the Cordilleran thrust belt helped to thicken the crust beneath the Colorado Plateau and Wyoming foreland region. Regardless of how crustal thinning in the Cordilleran interior was structurally accommodated, the fact that frontal thrusting in the Sevier belt continued unabated throughout Late Cretaceous–Paleocene time suggests that the orogenic belt as a whole reached either a limiting elevation or limiting crustal thickness, and that excess potential energy was dissipated by a combination of internal extension and frontal propagation. Hodges and others (2001) proposed a similar mechanism to explain the contemporaneity of frontal thrusting and internal extension in the Neogene Himalayan thrust belt.

Controls on Basin Evolution

The stratigraphic record of Cordilleran orogenesis is preserved in the Late Jurassic–Early Eocene foreland basin system, which is composed of mainly clastic sediment that was deposited under the influence of flexural thrust-loading, longer wavelength dynamic subsidence, and (during Late Cretaceous–Eocene time) local structural damming and loading by Laramide uplifts (fig. 17). Additional, relatively minor sediment accumulations are preserved in the interior of the thrust belt in Nevada. Evolution of the Cordilleran foreland basin system involved four main phases (fig. 17): (1) During Late Jurassic time (~155 - 142 Ma), the basin responded to orogenic thrust loading in the Luning-Fencemaker portion of the thrust belt, and regional dynamic subsidence driven by subduction of the Farallon plate (Lawton, 1994; Currie, 1998). (2) During Early Cretaceous through Turonian time (~142 - 89 Ma), the basin evolved as a classic flexural foreland basin system (Jordan, 1981), with wedge-top, foredeep, forebulge, and back-bulge depozones being well developed (DeCelles and Currie, 1996). (3) Beginning in Coniacian time (~88 Ma), the organized foreland subsidence pattern became more diffuse, with maximum subsidence in eastern Wyoming and Colorado, rather than along a well-defined foredeep trough adjacent to the front of the thrust belt (fig. 14; McGookey, 1972; Jordan, 1981; Cross, 1986; Pang and Nummedal, 1995; Robinson Roberts and Kirschbaum, 1995). The broad pattern of regional subsidence suggests a significant component of dynamic subsidence during this interval (fig. 17; Mitrovica and others, 1989; Pang and

Nummedal, 1995). (4) Beginning ~80 Ma, the foreland region became progressively partitioned into a mosaic of intraforeland uplifts and basins (the Laramide event). By ~75 Ma, individual Laramide uplifts were capable of driving local flexural subsidence and structurally damming adjacent sedimentary basins (fig. 17). Laramide range-bounding faults may merge into a mid-crustal detachment zone that roots westward beneath the Cordilleran thrust belt (Burchfiel and others, 1992, p. 461; Erslev, 1993; Yin and Ingersoll, 1997). In effect, the Laramide structural province may be thought of as the frontal part of the Cordilleran orogenic belt, only lacking a thick sedimentary cover section. If this is a correct way of understanding the regional structure of the Laramide province, then most of the Upper Cretaceous-lower Tertiary rocks between the Utah-Colorado state line and the Colorado Front Range are analogous to "piggy-back" basins that form in the frontal parts of thin-skinned thrust belts. Dickinson and others (1988) referred to these basins in the interior of the Laramide region as axial and ponded basins, and distinguished them from perimeter basins that formed around the periphery of the Laramide region. The perimeter basins, such as the Denver basin (fig. 2; Kluth and Nelson, 1988; Raynolds, 2002) may be analogous to foredeep deposits, and the frontal east-dipping flank of the Colorado Front Range consists of a crustal-scale triangle zone or wedge-fault system (Erslev, 1993), similar to (but much larger than) triangle zones that form along the fronts of many thin-skinned thrust belts (Price, 1986; Vann and others, 1986).

The Late Jurassic–Early Cretaceous depositional history of the foreland basin system was dominated by nonmarine fluvial and lacustrine systems. By mid-Albian time, the foreland basin came under the influence of marine transgressions and regressions (Cobban and others, 1994). The plan view shape of the frontal thrust belt consisted of alternating salients and narrow reentrants (Mitra, 1997). The reentrants served as topographic conduits for long-distance sediment transport from the interior of the thrust belt to the foreland basin (for example, fig. 16; Schmitt and Steidtmann, 1990; Lawton and others, 1994; DeCelles, 1994; Janecke and others, 2000). During high-stands in the Western Interior Seaway, marine waters locally drowned the frontal 10 to 50 kilometers of the orogenic wedge and lapped against steep frontal topography; the tip of the orogenic wedge lay buried beneath a several-kilometer-thick apron of wedge-top sediment (Lamerson, 1982; Lawton and Trexler, 1991; Lawton and others, 1993; Talling and others, 1995; DeCelles and others, 1995). During sea-level lowstands, fluvial megafans emanated from the reentrants in the frontal thrust belt (Lindsey, 1972; Robinson and Slingerland, 1998; DeCelles and Cavazza, 1999), and rivers flowed generally eastward and northeastward across a broad coastal plain to debouch in deltas along the shoreline of the western interior seaway (for example, Goldstrand, 1994; Gardner, 1995; Robinson Roberts and Kirschbaum, 1995). Thick accumulations of coal and lignite developed on the coastal plain (Robinson Roberts and Kirschbaum, 1995), perhaps nurtured by monsoonal rains derived from the Western Interior Seaway and driven westward by seasonal heating of the high elevation hinterland of the Cordilleran thrust belt. Generally eastward progradation of these fluvial, deltaic, and coastal plain systems during mid-Cretaceous through Paleocene time produced an impressive stack of stratigraphic sequences bounded by unconformities of variable duration. Some authors attribute these sequence-bounding unconformities to changes in eustatic sea level (Van Wagoner, 1995; Schwans and Campion, 1997), whereas others attribute them to changes in sediment supply and subsidence rates controlled by tectonics and erosion in the Sevier thrust belt and intraforeland uplift (Yoshida and others, 1996, 1998; Robinson and Slingerland, 1998; Guiseppe and Heller, 1998; McLaurin and Steel, 2000; Miall and Arush, 2001). Resolution of this debate is beyond the scope of this paper, but it is worthwhile to point out that thrusting events in the Sevier belt were mere increments of the larger deformation field that formed the Cordilleran orogenic belt, that regional thrust loading was probably little affected by events in the frontal Sevier belt alone, and that loading was more or less continuous in time, rather than sporadic (figs. 17 and 18). The latter is corroborated by recent geodetic studies of active thrust belts that exhibit continuum displacement behavior (for example, Bilham and others, 1997; Norabuena and others, 1998, 1999; Horton, 1999; Abdrakhmatov and others, 1996, 2001). Thus, it may be unwise to look toward the Sevier thrust belt for individual thrusting and erosional events that could have controlled the development of individual depositional sequences, while excluding the potential effects of changing eustatic sea level, paleoclimate, sediment flux, and intrinsic processes (for example, delta-lobe switching).

Connections Between the Hinterland and the Foreland: Continuity or Not?

Geologists have argued for decades about the degree of connectivity between thrust systems in the western hinterland, the frontal Sevier belt, and the Laramide foreland region (for example, Dickinson and Snyder, 1978; Livaccari, 1991; Bird, 1998). The question of temporal continuity of deformation in the Cordillera has been equally contentious (for example, Heller and others, 1986, 2003; Heller and Paola, 1989; Van Wagoner, 1995; DeCelles and Currie, 1996; Yoshida and others, 1998). The synthesis presented here suggests that treatment of the LFTB, CNTB, hinterland metamorphic belt, Sevier belt, and Laramide foreland as tectonostratigraphically diverse parts of a single, coherent Cordilleran orogenic wedge is consistent with the available radiometric ages and constraints on kinematic, metamorphic, and depositional events throughout the region. Critical taper theory (Davis and others, 1983) provides a useful framework for understanding the structural diversity of the Cordilleran thrust belt in the western U.S.A. (DeCelles and Mitra, 1995; Mitra, 1997). If orogenic belts build taper in order to maintain a least-work configuration (Davis and others, 1983; Willett, 1992), then the rheology and erosional durability of the rocks in the wedge should partly control the spacing and architecture of thrust faults in the wedge. While the front of the Cordilleran orogenic wedge was confined to fine-grained Triassic rocks of the LFTB during late Jurassic time, the thrust belt was probably steeply tapered. Once the front of the thrust wedge had propagated into the strong Proterozoic-Paleozoic section of central Nevada (including previously strain-hardened orogenic crust of the Roberts Mountains and Golconda allochthons) and western Utah, thrusts sheets could transmit stresses much farther eastward, and the orogenic wedge experienced a dramatic decrease of taper. These strong thrust sheets riding on weak fault zones (Yonkee, 1997) were so effective as stress guides that they left the impression that much of the central part of the hinterland was not deformed during the Mesozoic (for example, Armstrong, 1972; Heller and others, 2003). Instead, the record of Mesozoic deformation is preserved at mid-crustal levels now exposed in western Utah and eastern Nevada, where ductile shear zones developed coevally with thrusting farther east (Camilleri and others, 1997). As the thrust belt propagated into relatively thin, Paleozoic and Mesozoic strata of the frontal Sevier belt, slices of crystalline basement rocks were emplaced within antiformal duplexes along the Wasatch hinge line. These structural culminations may have developed as a means of maintaining taper in the frontal part of the thrust belt (DeCelles and Mitra, 1995; Mitra, 1997). The frontal part of the Sevier belt is divided into segments that exhibit a variety of wedge geometries, from extremely low taper (western Wyoming) to relatively high taper (Charleston-Nebo salient), probably as a function of the nature of the rocks at the décollement level, internal strength of the wedge, erosional processes at the surface, and rheological consequences of strain hardening and softening in the décollement zone (DeCelles and Mitra, 1995; Mitra, 1997). Eastward propagation of the basal décollement of the orogenic wedge as a mid-crustal shear zone (Burchfiel and others, 1992; Erlsev, 1993; Yin and Ingersoll, 1997) beneath the Laramide foreland region represents the final phase of Cordilleran shortening. Because the Paleozoic-Mesozoic cover section throughout the Laramide province is thin (<1.5 km) and tabular in shape, this final expression of the orogenic wedge had extremely low initial taper. Therefore, Laramide basement-involved thrusts developed as conjugate systems that facilitated regional shortening with no preferred vergence direction (for example, Erslev, 1993).

Thrusting and Arc Magmatism

Several previous authors have suggested that westward underthrusting of North American crustal basement and lithosphere beneath the Cordilleran magmatic arc may have been the cause of major magmatic flare-ups in the arc (for example, Burchfiel and others, 1992; Ducea, 2001). The general concept provides a link between kinematic processes in the upper crust of the retroarc region and lower crustal/lithospheric processes beneath the magmatic arc. Periods of rapid upper crustal shortening should be correlated with periods of magmatism in the arc as melt-fertile crust from beneath the thrust belt is subducted beneath the arc (Ducea, 2001). As discussed above, the Cordilleran arc experienced two major flare-ups (Late Jurassic, 160 - 150 Ma; and Late Cretaceous, 100 - 85 Ma, but beginning ~120 Ma). Kinematic reconstructions of the entire retroarc thrust belt (including the Laramide province) suggest that the periods of most rapid shortening were during Early Cretaceous (~140 - 110 Ma) and Late Cretaceous-Eocene (~85 - 55 Ma; fig. 4; DeCelles and others, 1995; Currie, 2002). Thus, the linkage, if any, between the timing of thrusting and magmatism in the arc seems weak. However, Ducea (2001) predicted that the onset of melting should post-date thrusting (that is, westward underthrusting of continental basement) by as much as 20 Myr. This prediction may provide an explanation for the lag time between the beginning of the major Late Cretaceous melting event (~120 Ma) and rapid Early Cretaceous shortening. Similarly, the Late Jurassic magmatic flare-up might have been associated with Middle to Late Jurassic thrusting in the Luning-Fencemaker thrust belt. The period of rapid Late Cretaceous–Eocene shortening was associated with a subsequent regional flare-up during early to mid-Cenozoic time (for example, Lipman, 1980; the “ignimbrite flare-up”), but it is widely believed that this event was related to steepening of the subducting Farallon plate after the Laramide flat slab event (for example, Coney and Reynolds, 1977).

Although a direct correlation between rapid westward underthrusting of North American crust and concentrated arc magmatism in the Sierra Nevada is tenuous (figs. 4 and 19), it is interesting to note that the period of rapid upper crustal thrusting during the Early to middle Cretaceous followed a period of diffuse, widespread retroarc magmatism in Late Jurassic time (fig. 17; D. L. Smith and others, 1993; Barton, 1996). In general, westward underthrusting of North American continental lithosphere would have been impeded by crowding between the base of the thickened magmatic arc and the top of the subducting oceanic slab (fig. 19A and 19B). Thus, a mechanism of lithospheric removal is needed to explain the ongoing regional shortening in the retroarc region. Low-angle Farallon plate subduction did not remove North American lithosphere during the well-known Late Cretaceous-early Tertiary flat-slab event (for example, Livaccari and Perry, 1993), so there is no reason to suggest that it might have served this purpose during an earlier episode of slab-flattening. Moreover, as discussed earlier, it is unlikely that the widespread Late Jurassic magmatism was caused by flat-slab subduction because of its diffuse distribution. Alternatively, removal of thickened North American lithosphere from beneath Nevada by delamination would have promoted both widespread melting in the crust (fig. 8C) and the ensuing rapid Early Cretaceous underthrusting (fig. 19B and 19C). Large volumes of North American lithosphere could have been eroded by the subducting Farallon plate or corner flow in the mantle wedge, or by gravitational foundering owing to convective instability (for example, Molnar and others, 1993; Pope and Willett, 1998; Beck and

Zandt, 2002). Crust that was underthrust beneath the arc would have become eclogitized, increasing its gravitational instability and likelihood of removal. Regardless of the exact mechanism, the large amounts of crustal shortening in the retroarc upper crust (>350 km) over the course of Late Jurassic–Eocene time require that a substantial volume of continental lithosphere must have been disposed of beneath the magmatic arc by "A-type" subduction (Bally, 1975). One can easily envisage how a temporal progression of lithospheric thickening by horizontal shortening, eclogitization beneath the arc, delamination or foundering accompanied by widespread partial melting in the lower crust, followed by rapid continental underthrusting, would supply large volumes of continental lower crust to fuel arc flare-ups after a lag-time of 10 to 20 Myr. Thickening and eclogitization of the newly underthrust, melt-fertile continental material would presumably initiate a new cycle. Such a cyclic process may be fundamental to the co-operation of large-scale, long-term continental magmatic arcs and thrust belts.

CONCLUSIONS

1. Shortening began in the Cordilleran retroarc thrust belt during Late Jurassic time, soon after the western margin of the continent became tectonically consolidated into a two-plate, continental margin arc-trench system. Over the ensuing ~100 Myr the front of the Cordilleran thrust belt (including the Laramide province) propagated ~1000 kilometers eastward and shortened supracrustal and basement rocks by >350 kilometers. The overall arcuate shape of the thrust front between latitudes 35°N to 50°N was developed early during the history of thrusting, probably during Early Cretaceous time. Although the bulk of the Cordilleran thrust belt was composed of sedimentary cover rocks, Precambrian basement was thickened along mid-crustal shear zones in the hinterland and in large structural culminations that developed along the relatively high-standing shoulder of rifted basement (the Wasatch hingeline) in central Utah, eastern Idaho, and southwestern Montana.

2. Mid-crustal rocks in the interior of the thrust belt began to experience significant exhumation and cooling, presumably in response to crustal thinning by extension, during Campanian–Maastrichtian time. This extension produced no known surface-breaking faults, and may have been restricted to mid-crustal rocks (Hodges and Walker, 1992). Coeval with extension in the hinterland region, major thin-skinned thrust systems propagated eastward into Wyoming in the frontal Sevier belt. The contemporaneity of frontal thrusting and internal large-magnitude extension is consistent with the idea that the Cordilleran orogenic belt reached a limiting elevation or crustal thickness and began to gravitationally collapse during Late Cretaceous time.

3. Arc magmatism swept hundreds of kilometers eastward twice (Late Jurassic and latest Cretaceous–Eocene). The first event may have been associated with an episode of lithospheric foundering and inferred dynamic uplift (following dynamic subsidence) in the foreland region that destructively interfered with shorter-wavelength, orogenic load-driven flexural subsidence. The second magmatic sweep is well known and was associated with Laramide deformation and low-angle subduction.

4. Predicted linkages between voluminous Cordilleran arc-magmatism during Late Jurassic and Late Cretaceous time and westward underthrusting of North American continental lithosphere beneath the arc are not obvious from the geologic record of regional kinematics in the Cordilleran thrust belt. A significant lag-time (~20 Myr) between periods of rapid shortening (and coeval underthrusting) and generation of arc melts is required to explain the linkage, if any. However, inferred Late Jurassic delamination of North American lithosphere may have produced a necessary precondition to allow relatively rapid Early Cretaceous thrusting (and coeval westward continental underthrusting) by removing a large portion of North American lithosphere from beneath the arc and nascent retroarc thrust belt.

5. By Late Cretaceous time, the paleogeography of the Cordilleran thrust belt in the western U.S.A. was characterized by a broad, low-relief, high-elevation hinterland plateau (the "Nevadaplano") and a topographically rugged frontal thrust belt. Nonmarine and marine depositional systems alternated in time, and marine transgression coupled with high sediment flux sporadically inundated and buried the frontal part of the thrust belt.

6. Sediment accumulation in the Cordilleran thrust belt and foreland basin system occurred in response to several different types of tectonic processes. Flexural subsidence was the predominant mechanism during Early Cretaceous through Turonian time (~142 - 89 Ma), whereas regional dynamic subsidence associated with the subducting Farallon slab may have been important during Late Jurassic and Late Cretaceous time. From Late Cretaceous to Eocene time the foreland region was partitioned into a mosaic of uplifts and basins associated with the Laramide structural province. Although the Laramide uplifts have been considered to be distinct from the Cordilleran thrust belt, it is equally plausible that the Laramide province is merely the frontal part of a composite Cordilleran orogenic wedge.

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