

# Geotectonic evolution of the Great Basin

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## ABSTRACT

The crust of the Great Basin has occupied a range of tectonic settings through geologic time. Archean and Paleoproterozoic crustal genesis preceded residence of Laurentia within the Mesoproterozoic supercontinent Rodinia, which rifted in the late Neoproterozoic to delineate the Cordilleran flank of Laurentia. Successive stages of Phanerozoic evolution included (1) early to middle Paleozoic miogeoclinal sedimentation along a passive continental margin, (2) late Paleozoic to earliest Mesozoic thrusting of oceanic Antler and Sonoma allochthons over the continental margin in response to episodic slab rollback beneath an offshore Klamath-Sierran island-arc complex, (3) Mesozoic to mid-Cenozoic arc-rear and backarc thrusting, together with pulses of interior magmatism, associated with development of the Cordilleran magmatic arc to the west where subduction and arc accretion expanded the continental margin, and (4) middle to late Cenozoic crustal extension, which involved initial intra-arc to backarc deformation and later transtensional torsion of the continental block inland from the evolving San Andreas transform system. Potential metallogenic influences on Great Basin tectonic evolution included transfer of substance from mantle to crust by magmatism and associated metasomatism, and reworking of crustal materials by both magmatism and intracrustal fluid flow, the latter of which was induced both by thermal effects of magmatism and by reconfiguration of fluid-bearing rock masses during multiple episodes of Great Basin deformation.

**Keywords:** Basin and Range, geologic history, geotectonics, Great Basin, Nevada.

## INTRODUCTION

The broad outlines of Cordilleran plate tectonics are now well understood (Dickinson, 2000, 2002, 2004), although disputes continue regarding the impetus for each stage of tectonic evolution, and our mental picture grows progressively

dimmer as earlier and earlier geologic time frames are considered. The purpose of this paper is to provide an overview of Great Basin geotectonics through geologic time as background for discussions of Great Basin metallogeny in subsequent papers of this special issue.

## TECTONICS AND METALLOGENY

To understand metallogeny, one must know the geologic sources of the constituents in ore minerals, appreciate the structural preparation of rock masses for ore deposition, and understand the mechanisms for ore transport and precipitation. Tectonics lies at the root of all these issues. Potential sources of metals in the crust and mantle vary with tectonic setting, structural conditions within the crust are a function of tectonic evolution, and fluid flow through the crust is dictated by ambient tectonic environments. Few regions of the world have had as varied a tectonic history as the Great Basin, and its geologic complexity challenges interpretations of metallogeny to the utmost.

The nature of potential metal sources in the deep mantle changed over time as lithospheric plates moved over asthenosphere. The composition of the crust and the immediately subjacent lithospheric mantle were modified over time as mantle magmatism and associated metasomatism added materials that were previously absent, and extraction of crustal melts and leaching by rising fluids removed materials once present. Ground preparation by structural deformation in the Great Basin reflects the effects of multiple tectonic episodes of contrasting structural style, which resulted in superposed structural features and older structures overprinted by younger ones.

Ore transport and deposition involve inherent thermochemical variability difficult to infer because the fluids and thermal conditions that formed ore are recorded in the geologic record only by subtle indicators that are difficult to read without ambiguity. For relations between tectonics and metallogeny, there are two linked but separate facets of ore genesis to keep in mind: (1) mobilization of elements of interest from mantle or crustal reservoirs, and

(2) precipitation or fixation of those elements at the site of an ore deposit. Setting an element in motion from some suitable reservoir is a necessary but insufficient factor for ore genesis, because no ore deposit is formed so long as the element keeps on moving. Thermomechanical conditions for mobilization and for precipitation are diametrically opposed, yet both are required for ore genesis. Many metallic elements may be set in motion through some segment of the crust during a given tectonic regime, but only those induced thermochemically to stop moving will occur in a given ore deposit. The source of metals may be of secondary interest for metallogeny, with site conditions that encouraged ore deposition of primary importance.

## GREAT BASIN TECTONIC HISTORY

The Great Basin forms the widest segment of the vast Basin and Range taphrogen, which extends for >2500 km from the Pacific Northwest to central Mexico (Fig. 1). From the Colorado Plateau on the east to the Sierra Nevada on the west, and from the Snake River Plain on the north to the Garlock fault and the Mojave block on the south, the Great Basin occupies a 600 km by 600 km tract of rugged internal topography. The bulk of the Great Basin lies within the state of Nevada (state outlines are shown on accompanying paleotectonic maps), but it extends also into western Utah and the eastern fringe of California. The Oregon Plateau segment of the Basin and Range taphrogen (Fig. 1) is in part internally drained, and it can be considered an appendage of the Great Basin proper.

The Great Basin evolved along the western fringe of Precambrian Laurentia through diverse chapters of Earth history, each of which had potential but varying implications for metallogeny (Fig. 2), including:

(1) Precambrian emergence of juvenile continental crust from the mantle to form the Archean Wyoming Province and the Paleoproterozoic Mojave Province;

(2) Mesoproterozoic incorporation of the Precambrian basement into the Rodinian supercontinent during an interval punctuated by incipient Belt-age rifting;



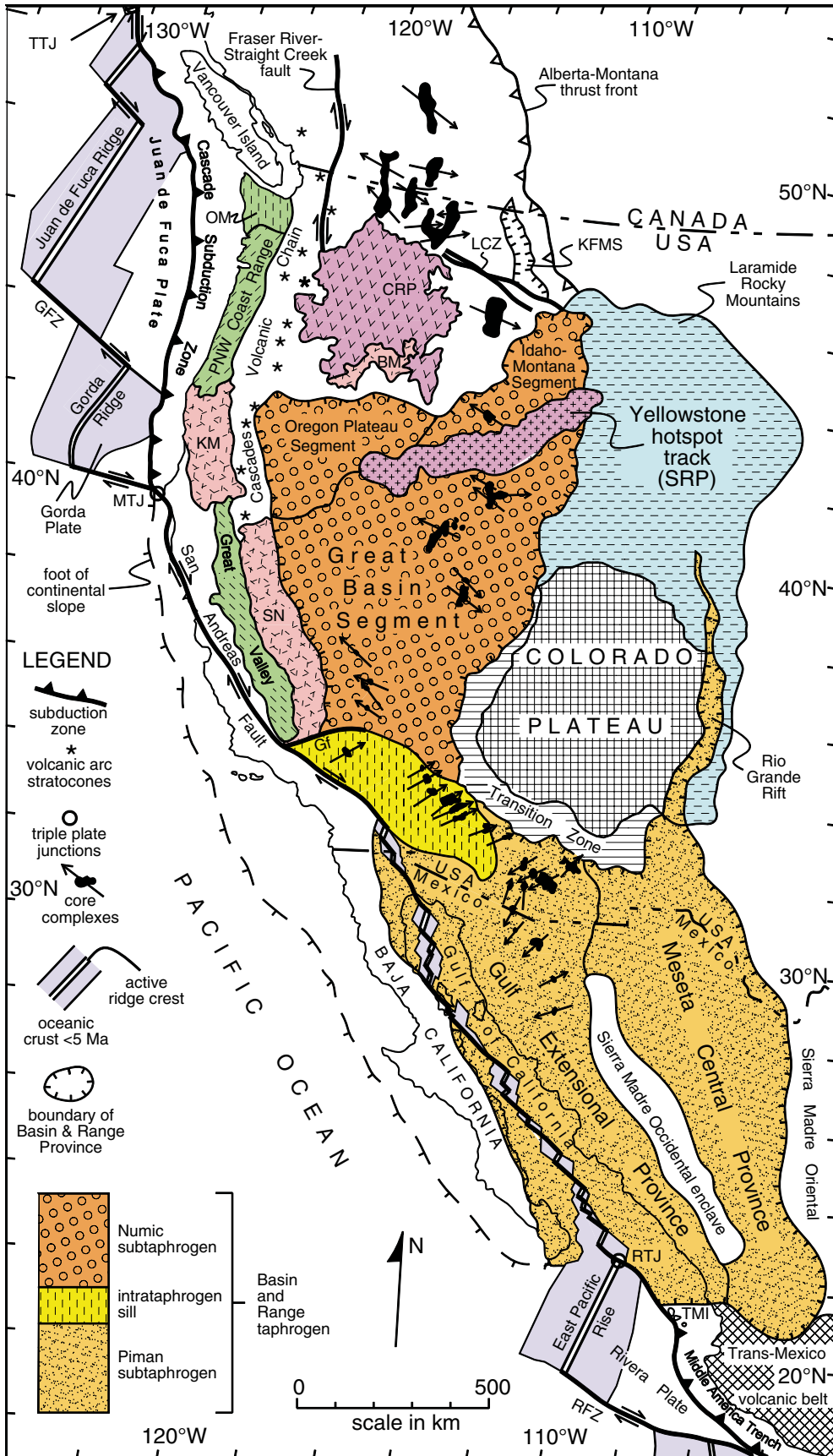


Figure 1. Position of the Great Basin in the western Cordillera (adapted after Dickinson, 2002). Modern triple plate junctions: MTJ—Mendocino; RTJ—Rivera; TTJ—Tofino. Other abbreviations: BM—Blue Mountains; CRP—Columbia River Plateau (check pattern and red color denote extent of Columbia River Basalt lavas); KFMS—Kisenehn-Flathead-Mission-Swan extensional Paleogene basins; KM—Klamath Mountains; LCZ—Lewis and Clark fault zone; PNW—Pacific Northwest; RFZ—Rivera Fracture Zone; SN—Sierra Nevada; SRP—Snake River Plain; TMI—Tres Marias Islands (cross pattern and red color denote extent of bimodal volcanic suite).



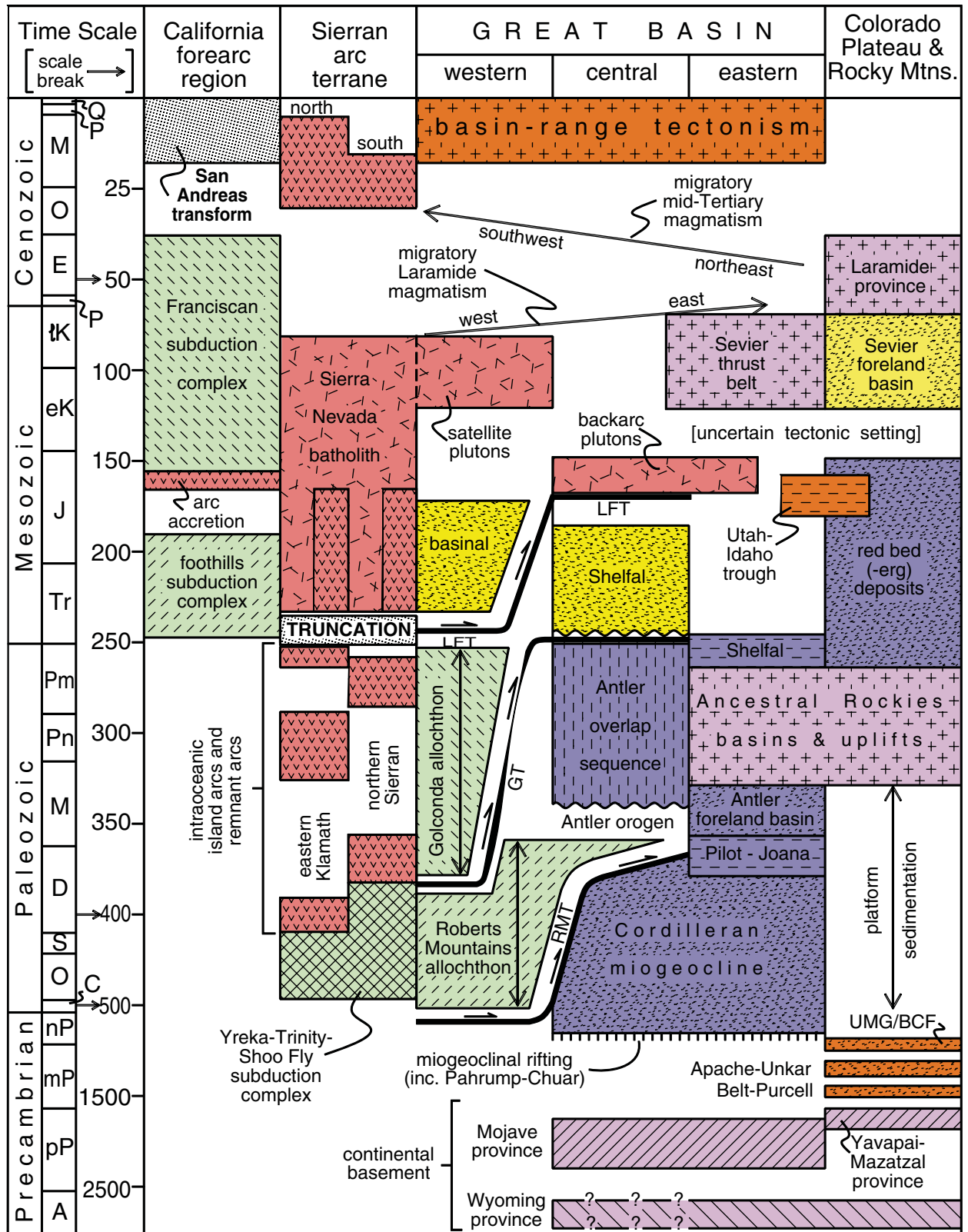


Figure 2. Time-space diagram of lithic assemblages in the Great Basin and adjoining areas (note time-scale breaks at 50 Ma, 400 Ma, and 500 Ma). The rectangle labeled truncation denotes schematically the completion of continental truncation along the Permian-Triassic California-Coahuila transform and subsequent initiation of the Mesozoic-Cenozoic Cordilleran continental-margin magmatic arc. Key thrusts: GT—Golconda; LFT—Luning-Fencemaker; RMT—Roberts Mountains. UMG/BCF—Uinta Mountain Group and Big Cottonwood Formation.



(3) Neoproterozoic rifting that delineated the Cordilleran miogeocline along which passive-margin sedimentation continued until mid–Late Devonian time;

(4) Late Devonian to early Mississippian thrusting (Antler orogeny) of deformed oceanic facies, forming the Roberts Mountains allochthon, over the miogeocline;

(5) late Mississippian to Permian (post-Antler) deposition of the marine-nonmarine Antler overlap sequence atop the Antler orogen, of oceanic strata in the Havallah basin west of the Antler orogen, and of foreland-basin clastic strata east of the Antler orogen, with local tectonic disruption of the Antler foreland related to intracontinental deformation that formed the Ancestral Rocky Mountains;

(6) Late Permian to mid–Early Triassic thrusting (Sonoma orogeny) of deformed Havallah oceanic facies, forming the Golconda allochthon, over the dormant Antler orogen;

(7) Mid–Early Triassic development of the nascent Cordilleran magmatic arc flanked by a Late Triassic to Early Jurassic backarc basin (Auld Lang Syne Group);

(8) Middle Jurassic and mid-Cretaceous development of retroarc thrust belts, with an intervening episode of Late Jurassic backarc magmatism well inland from the continental margin associated with arc accretion along the continental margin;

(9) Late Cretaceous to Paleogene inland migration of Laramide magmatism from the Sierra Nevada on the west toward the Rocky Mountain continental interior;

(10) Late Eocene to Early Miocene re-migration of spatially complex magmatism back across the Great Basin from the interior toward the continental margin;

(11) Oligocene to Miocene intra-arc to back-arc extensional deformation involving tectonic denudation of basement rocks in several Cordilleran core complexes; and

(12) Neogene basin-range transtension and accompanying dispersed magmatism linked geodynamically to evolution of the San Andreas transform system.

## PALINSPASTIC ISSUES

The paleogeography of large tracts of the Great Basin has been repeatedly rearranged, first by overlap of rock masses during episodes of Paleozoic–Mesozoic thrusting and later by separation of rock masses during Cenozoic extension of the intermountain belt. Since our knowledge of tectonic geometry within the Great Basin remains imperfect, palinspastic adjustments to paleogeography for paleotectonic reconstructions are somewhat idiosyncratic and

partly model-driven. Moreover, shifting modern geography into palinspastic frameworks would make the resulting paleotectonic maps difficult to relate to modern geographic features familiar to Great Basin geologists.

Accordingly, the paleotectonic maps of this paper are plotted (base map after Muehlberger, 1992) on modern geography, apart from the palinspastic restoration of crustal elements that have moved laterally, as internally more-or-less intact blocks, for >75 km along discrete structures. The latter are typified by the San Andreas fault but include several other more controversial structures within the arc assemblages that lie mostly to the west of the Great Basin but in part along its western fringe. The only palinspastic adjustments incorporated into the paleotectonic maps are the following, with no attempt made to restore more distributive strain within the Great Basin itself:

(1) reversal of 470 km of net post–mid-Miocene dextral slip along the central California coast across multiple strands of the San Andreas transform system (Dickinson and Butler, 1998);

(2) post-mid-Oligocene reversal of post-mid-Eocene tectonic rotations (clockwise) of the Pacific Northwest (PNW) Coast Range and the Blue Mountains Province by a nominal 45° each with respect to the continental interior, with concomitant eastward shift of the Klamath Mountains block linked depositionally to the Pacific Northwest Coast Range since Paleocene time (Dickinson, 2002, 2004);

(3) pre-Cretaceous shift of Klamath-Sierran rock masses southward to compensate for 210 km of Early Cretaceous dextral slip (Dickinson, 2005) along the Mojave–Snow Lake fault (Lahren and Schweickert, 1989; Schweickert and Lahren, 1990) and related faults longitudinal to the alignment of the Sierra Nevada batholith; and

(4) reversal of 950 km of Permian-Triassic sinistral slip along the California-Coahuila transform, which truncated the Cordilleran continental margin in California and linked Sonoma orogenic trends in the Great Basin to an arc-trench system in eastern Mexico (Dickinson, 2000; Dickinson and Lawton, 2001a).

## Mojave–Snow Lake Conundrum

Mojave–Snow Lake fault displacement along a cryptic structure, which was obliterated by later intrusion of the Sierra Nevada batholith, was detected by recognition of the Cambrian Zabriskie Quartzite and associated stratigraphic units of the southeastern or Death Valley facies of the Cordilleran miogeocline in the Snow Lake roof pendant of the central Sierra Nevada (Grasse et al., 1999). Exposures

of the Death Valley facies in the Snow Lake pendant are separated from stratal counterparts in the Death Valley region by a wide expanse that exposes only the northwestern or Inyo facies of the miogeocline.

Initial analysis of apparent Mojave–Snow Lake offset suggested a displacement of 400–500 km (Lahren and Schweickert, 1989; Schweickert and Lahren, 1990). Because the northward trace of the Mojave–Snow Lake fault is inferred to pass east of the northern Sierra Nevada (Schweickert and Lahren, 1990, 1993a; Wyld and Wright, 2001), reversal of such large dextral slip along the Mojave–Snow Lake fault would restore Paleozoic arc assemblages of the eastern Klamath Mountains and northern Sierra Nevada to an unlikely position athwart the trend of the Cordilleran miogeoclinal belt.

The misalignment of Death Valley and Inyo miogeoclinal facies across the Mojave–Snow Lake fault does not, however, require such large displacement (Saleeby and Busby, 1993). Strata of the Death Valley facies, including the diagnostic Zabriskie Quartzite, extend as far northwest as the vicinity of Lone Pine in Owens Valley between the Inyo Mountains and the Sierra Nevada (Nelson, 1976; Stewart, 1983). Derivation of the Snow Lake pendant from the vicinity of Lone Pine requires net fault offset of only  $210 \pm 15$  km. The Mojave–Snow Lake fault can be viewed as just one of a family of dextral faults offsetting miogeoclinal strata of the eastern Sierra Nevada (Stevens and Greene, 1999, 2000), and the indicated net slip of 210 km can be interpreted as the sum of parallel displacements on multiple structures. The postulated net slip of 210 km adopted here is sufficient to explain the distortion of strontium-isotope and other geochemical isopleths in the central Sierra Nevada (Kistler, 1993), and greater Mojave–Snow Lake offset would seem precluded by the known geographic pattern of the isopleths. Restoration of 210 km of Mojave–Snow Lake displacement additionally places lithic assemblages of the eastern Klamath Mountains along tectonic strike from counterparts in the Pine Forest Range (Wyld, 1990) of northwestern Nevada.

## PREMIOGEOCLINAL PRECAMBRIAN

Precambrian basement in the Great Basin lies along the western flank of Laurentia and was built by successive accretionary episodes of crustal genesis around an Archean continental nucleus. Archean rocks of the Wyoming province (older than 2500 Ma) extend into northernmost Utah and northeastern Nevada (Fig. 3). Farther south in the Great Basin, basement lies within the poorly known Mojave Province of Paleoproterozoic rocks, which are partly older



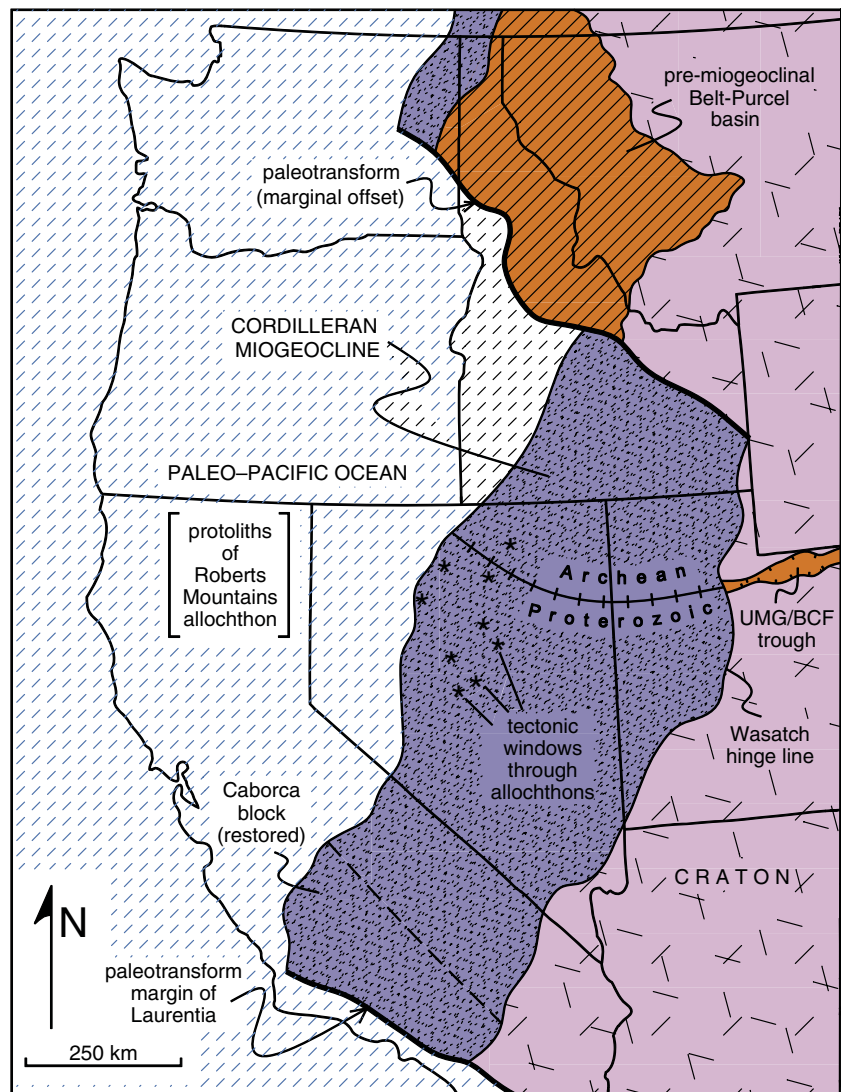
than the conjoined Yavapai-Mazatzal terranes (1800–1600 Ma) farther east (Fig. 2).

Laurentia was incorporated into the Mesoproterozoic supercontinent of Rodinia at the time of the Grenville orogeny (1300–1000 Ma), but the identity of the crustal blocks lying immediately west of the Great Basin within Rodinia remains uncertain. Options include the Precambrian cores of Siberia (Sears and Price, 1978, 2000, 2003), East Antarctica (Hoffman, 1991; Moores, 1991; Dalziel, 1991; Weil et al., 1998; Li, 1999), and Australia (Brookfield, 1993; Powell et al., 1994; Karlstrom et al., 1999, 2001). Postulated geologic ties of Laurentia to Australia or Antarctica are difficult to defend in detail (Wingate et al., 2002), but a close match of Proterozoic lithostratigraphy from the Death Valley region at the southern limit of the Great Basin to the Sette-Daban Range of southeastern Siberia provisionally confirms the Siberian connection (Sears et al., 2005).

The possible significance of Precambrian evolution for later metallogeny in the Great Basin is uncertain, but subjacent Precambrian basement is a potential reservoir for metals mobilized by Phanerozoic tectonic events. Continental crust is commonly viewed in generic terms, as if one continental block were indistinguishable from another, but different crustal blocks around the world harbor different kinds of ore deposits, and differences in the continental basement of the Great Basin might be significant for metallogeny. The Cheyenne suture belt between Archean and Paleoproterozoic terranes (Karlstrom and Houston, 1984; Chamberlain et al., 1993) projects westward into the Great Basin (Fig. 3) and delineates a local contrast in crustal architecture (Wright and Wooden, 1991).

Incipient Precambrian rifting may have locally affected Precambrian basement of the Great Basin by introduction of mantle-derived magmas into the crustal profile or by redistribution of crustal materials through the thermal effects of rifting. Pre-Rodinian intracontinental rifting (1470–1370 Ma) of Laurentian crust produced the extensive Belt-Purcell basin (Fig. 3) of the northern Rocky Mountains (Evans et al., 2000; Luepke and Lyons, 2001), and undetected coeval structures could well be present farther south in the subsurface of the Great Basin (Fig. 2). Somewhat younger rift structures, associated with deposition of the Unkar Group (1255–1105 Ma) in the Grand Canyon south of the Great Basin, were coeval with the Grenville assembly of Rodinia and may have counterparts that extend into the Great Basin (Timmons et al., 2005).

Of special interest is the possibility that the Uinta Mountain–Big Cottonwood trough or aulacogen projects beneath miogeoclinal cover into the northeastern Great Basin along the trend of



**Figure 3.** Precambrian to early Paleozoic (pre-375 Ma, mid-Late Devonian) paleotectonic map of the Cordilleran miogeocline and associated geotectonic features in the Great Basin and adjacent areas. The ancestral eastern Klamath–northern Sierra Nevada intraoceanic island-arc complex is not shown within the paleo-Pacific Ocean because its distance offshore is uncertain for the time frame depicted. UMG/BCF—Uinta Mountain Group and Big Cottonwood Formation.

the Archean-Paleoproterozoic suture (Fig. 3), which may have controlled the position of a local bend in the configuration of the miogeoclinal Paleozoic continental margin (Miller et al., 1991). The Uinta Mountain Group and Big Cottonwood Formation are poorly dated, but the best available geochronology indicates deposition during the pre-miogeoclinal interval of 770–740 Ma (Dehler et al., 2005). This time frame is coeval with deposition of the pre-miogeoclinal Chuar Group (775–735 Ma) of the Grand Canyon in fault-controlled rift basins (Timmons et al., 2001), which may also have counterparts in the subsurface of the Great Basin to the northwest.

### CORDILLERAN MIOGEOCLINE

Neoproterozoic continental separation by rifting delineated the Cordilleran margin not long before the onset of Phanerozoic time and initiated deposition of a westward-thickening prism of miogeoclinal sediment (Fig. 2), including both shelfal and off-shelf slope strata of Neoproterozoic to Devonian age (Poole et al., 1993). Few have speculated about the possible influence of continental rifting on ore genesis because all modern examples of passive continental margins are buried beneath thick sediment cover and are unavailable for direct observation,



but thermal effects of rifting on continental crust may be significant as deeper crustal levels are brought toward the surface by tectonic denudation. Simultaneous injection of mantle melts into thinning crust may further introduce metallic elements not previously present in comparable abundance within the crustal profile. The continental basement beneath the miogeoclinal sediment prism thins westward across the Great

Basin from the Wasatch hinge line, flanking the unrifted craton, to a feather edge beyond which paleo-Pacific oceanic crust once lay west of the miogeoclinal belt (Fig. 3).

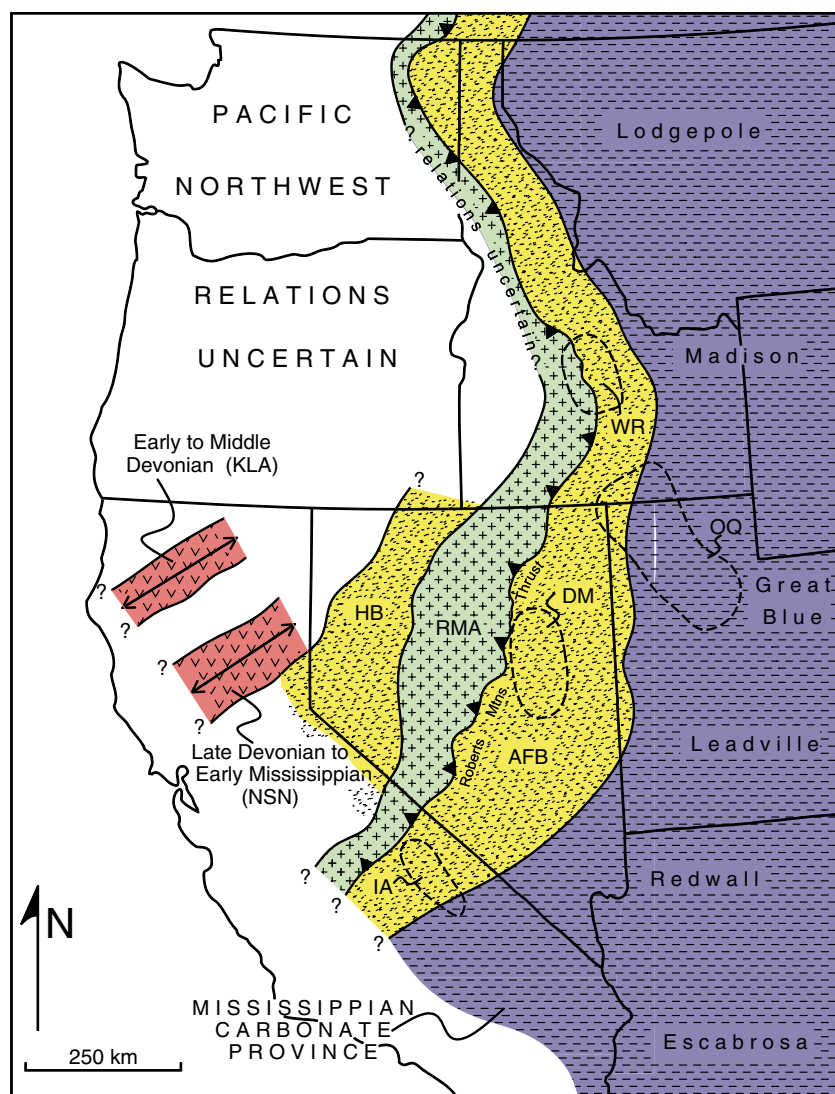
Continental separation and initiation of miogeoclinal sedimentation was apparently diachronous north and south of a paleotransform that defines a prominent marginal offset in the rifted continental margin at the northern limit

of the Great Basin segment of the Cordilleran miogeocline (Fig. 3). In Canada and Washington farther north, basaltic rocks associated with glaciomarine diamictite in basal horizons of the miogeoclinal succession (Ross, 1991) have been dated isotopically at 770–735 Ma (Devlin et al., 1988; Rainbird et al., 1996; Colpron et al., 2002). Correlative strata (see previous sections) exposed marginal to the Great Basin in the Uinta Mountains (Uinta Mountain Group) and the Grand Canyon (Chuar Group) occupy intracontinental rift troughs (Fig. 2) that developed before continental separation, which was delayed in the Death Valley region until after 600 Ma (Prave, 1999).

The onset of postrift thermotectonic subsidence of the miogeoclinal continental margin within the Great Basin occurred in Early Cambrian time (Armin and Mayer, 1983; Levy and Christie-Blick, 1991), at 525–515 Ma as adjusted for modern geologic time scales. By analogy with the modern Atlantic passive continental margin of North America, where ~55 m.y. elapsed between initial development of Triassic rift basins and the earliest emplacement of Jurassic oceanic crust offshore (Manspeizer and Cousminer, 1988), continental separation in the Great Basin can be inferred at ca. 575 Ma in late Neoproterozoic time. Isotopic dating of synrift volcanic rocks in southern British Columbia at  $570 \pm 5$  Ma (Colpron et al., 2002) implies that final continental separation in the Great Basin was accompanied by rejuvenation of rifting along the preexisting passive continental margin farther north in Canada. Given multiple rifting events along the Cordilleran margin, and the long duration (>50 m.y.) of incremental rifting required to achieve full continental separation, basement rocks at depth below the miogeoclinal sediment prism may have experienced a high geotherm for a prolonged interval of Neoproterozoic time.

#### ANTLER-SONOMA OBDUCTION

Miogeoclinal sedimentation was terminated after mid-Paleozoic time by eastward obduction of overthrust subduction complexes forming the Roberts Mountains and Golconda allochthons (Figs. 4 and 5), which were emplaced during the Devonian-Mississippian Antler orogeny and the Permian-Triassic Sonoma orogeny, respectively (Fig. 2). The two migratory subduction complexes approached the Cordilleran continental margin in response to slab rollback beneath a system of active and remnant intraoceanic island arcs now exposed in the eastern Klamath Mountains and northern Sierra Nevada (Dickinson, 2000). The offshore island-arc system faced southeast toward the continental margin



**Figure 4.** Syn-Antler (Devonian-Mississippian) and post-Antler but pre-Sonoma paleotectonic map of the Great Basin and adjacent areas. The relative positions of an intraoceanic frontal arc in the northern Sierra Nevada (NSN) and its paired remnant arc in the eastern Klamath Mountains (KLA) are uncertain. Double-headed arrows denote the NE-SW tectonic trends of Paleozoic island-arc elements oriented at a high angle to the Mesozoic-Cenozoic continental margin of California (Dickinson, 2000). Key tectonic elements in Nevada (west to east): HB—Havallah basin; RMA—Roberts Mountains allochthon (capped by the Antler overlap sequence); AFB—Antler foreland basin. Formational names indicate local components of the Mississippian carbonate province. Dashed outlines denote basins of the Ancestral Rocky Mountains province closest to the Antler thrust front: DM—Dry Mountain; I-A—Inyo-Argus; OQ—Oquirrh; WR—Wood River.



in central Nevada, subducting the miogeoclinal belt downward to the northwest. The miogeoclinal sediment prism was drawn down to depths of 5–15 km beneath the internally deformed overthrust assemblages, and burial depth increased westward as the subducted miogeocline was tilted downward to the west beneath westward-thickening subduction complexes (Speed and Sleep, 1982).

The continuity of the underthrust miogeoclinal prism beneath the overthrust oceanic allochthons is confirmed by exposures in multiple tectonic windows (Fig. 3) distributed from the Antler-Sonoma foreland westward to central Nevada (Stewart and Carlson, 1976; Stewart, 1980). The superposed Roberts Mountains and Golconda allochthons were derived in bulk from an oceanic region (Dickinson, 2000) that lay beyond the offshore limit of the miogeoclinal belt (Rowell et al., 1979). Disparate paleogeographic origins for the two allochthons are confirmed by the ages of detrital zircons in deformed sedimentary assemblages of both allochthons, which generally lack zircons comparable in age to those present in sandstones of the underlying miogeocline derived from the adjacent craton (Gehrels and Dickinson, 1995; Dickinson and Gehrels, 2000). Along the eastern fringe of the Roberts Mountains allochthon, however, selected stratigraphic units contain detrital zircons apparently derived from the adjacent craton and presumably represent stratal increments added from an offshore continental rise to the front of a growing subduction complex as the latter approached the continental margin.

### Antler-Sonoma Events

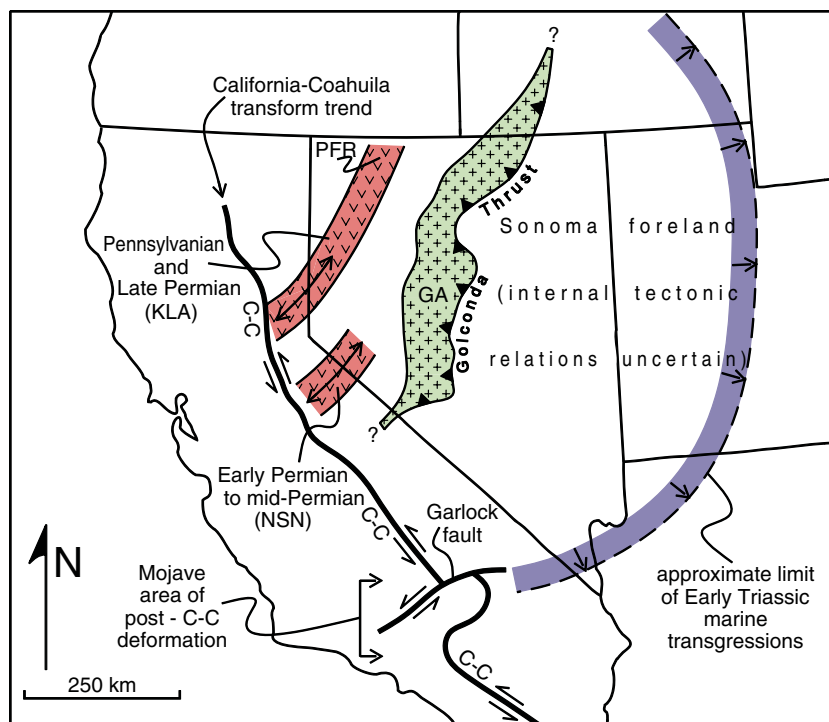
The Antler orogen, formed in central Nevada by thrust emplacement of the Roberts Mountains allochthon in latest Devonian to earliest Mississippian time (Fig. 4), was capped discontinuously by nonmarine to shallow-marine strata of the Antler overlap sequence (Fig. 2). The orogen-capping succession is broken by multiple unconformities, but ranges in age from late Mississippian through Permian, and shelf deposits that form its uppermost horizons locally include lowermost Triassic strata. Lateral equivalents of the Antler overlap sequence to the east include Mississippian clastic strata of the Antler foreland basin (Fig. 3), which was downflexed beyond the Roberts Mountains thrust front of the Antler orogen along an elongate belt flanking extensive Mississippian carbonate platforms of the continental interior (Dickinson et al., 1983). The foreland succession is underlain by Devonian-Mississippian shale and limestone of the migratory Pilot-Joana backbulge-forebulge system (Goebel, 1991; Giles and Dickinson, 1995; Giles,

1996), and it is overlain by Pennsylvanian-Permian limestone with intercalated clastic intervals (Fig. 2). To the west, lateral equivalents of the Antler overlap sequence form the Havallah sequence (Fig. 4), which was deposited within a residual oceanic trough lying offshore from the continental margin from latest Devonian to latest Permian time (Dickinson, 2000).

Late Paleozoic deformation centered in Pennsylvanian time in the Ancestral Rocky Mountains Province of the continental interior (Dickinson and Lawton, 2003) extended westward far enough to affect the Antler foreland region in the interval between Antler and Sonoma events (Fig. 2). Local depocenters (Fig. 4), clastic wedges, and multiple unconformities of Carboniferous to Permian age have been reported from widespread localities lying east of the Antler and Sonoma thrust fronts (Trexler et al., 2004).

The deformed Havallah sequence was thrust over the Antler overlap sequence in latest Permian to earliest Triassic time (Fig. 2), as the

Golconda allochthon of the Sonoma orogen (Fig. 5). The Golconda allochthon was intruded after structural emplacement by an Upper Triassic pluton (219 Ma) of the Mesozoic Sierra Nevada arc assemblage near Mono Lake (Schweickert and Lahren, 1987, 1993b), and an Upper Triassic (Norian) overlap sequence of clastic strata (Auld Lang Syne Group) resting on the Golconda allochthon in central Nevada is contiguous with facies equivalents on the Colorado Plateau to the east (Lupe and Silberling, 1985; Riggs et al., 1996). These intrusive and stratigraphic relationships tie the Golconda allochthon to the continental block by mid-Triassic time, even though no well-developed foreland basin can be discerned beyond the Golconda thrust front (Lawton, 1994). Increasingly marine facies toward the west within the Moenkopi Formation (Early to Middle Triassic) of the Colorado Plateau represent a marine transgression of the continental interior (Fig. 5) and is interpreted here to reflect seaward tilt



**Figure 5.** Syn-Sonoma (ca. 250 Ma) paleotectonic map of the Great Basin and adjacent areas (GA—Golconda allochthon) near the Permian-Triassic time boundary during oblique truncation of the Cordilleran continental margin by the California-Coahuila transform (C-C), which is contorted and offset by younger deformation in the Mojave region south of the Garlock fault. Eastern Klamath Mountains (KLA) and northern Sierra Nevada (NSN) arcs and remnant arcs (double-headed arrows denote NE-SW tectonic trends within the island-arc complex) are shifted SSE by 210 km to reverse Early Cretaceous dextral slip along the Mojave-Snow Lake fault, and the Klamath Mountains block is additionally shifted eastward to align with the Sierra Nevada block prior to Early Cretaceous forearc extension (Constenius et al., 2000) and later translation associated with Paleogene clockwise rotation of the Pacific Northwest (Oregon-Washington) Coast Range (see Fig. 8). PFR—Pine Forest Range.



from Sonoma foreland downflexure east of the Golconda thrust front.

Extensions of Paleozoic miogeoclinal and arc-related tectonic belts to the southwest of the Great Basin are largely unknown (Fig. 4) because of tectonic truncation (Fig. 2) of Laurentia by Permian-Triassic sinistral strike slip in the Sierra Nevada foothills along the Melones fault trend, a segment of the California-Coahuila transform (Fig. 5). The Antler orogen was offset southward from the vicinity of Mono Lake in the central Sierra Nevada to the Kern Plateau in the southern Sierra Nevada and the El Paso Mountains adjacent to the modern Garlock fault, and the miogeocline was offset into Sonora as the Caborca block (Fig. 3). Offset of Sonoma tectonic trends is less well understood but can be inferred as well. Northward from the Great Basin, the Antler thrust system is present in central Idaho (Wust and Link, 1988a, 1988b; Link et al., 1996) and follows the arcuate structural trend of the transnational Kootenai belt of northeastern Washington into Canada (Smith and Gehrels, 1992; Smith et al., 1993). The Antler foreland basin (Fig. 4) also extended through Idaho-Montana into Canada (Nilsen, 1977; Dorobek et al., 1991; Savoy and Mountjoy, 1995).

### Metallogenic Implications

Exhalative ore deposits (Papke, 1984) of syngenetic character within the Paleozoic allochthons of central Nevada did not form above continental basement, but rode over it from the oceanic region to the west during partial subduction of the miogeocline beneath oceanic subduction complexes. The environment of ore genesis for these deposits lay within a remnant or marginal ocean basin underlain by oceanic lithosphere with a thin crustal profile.

The potential impact of partially subducting the miogeoclinal sediment prism on the development of protores in central Nevada has commonly been overlooked, but in my view should receive close attention. Crustal fluids typically migrate upward and away from thrust systems associated with subduction zones (Dickinson, 1974; Oliver, 1992). When the previously undeformed miogeoclinal sediment prism was tilted and drawn beneath the overriding Roberts Mountains allochthon during the Antler event, fluids contained within the miogeoclinal strata were perforce driven up to the east away from the evolving subduction zone. Metals could then have been scavenged by the migratory fluids from large volumes of sediment and transported for long distances to the east into cooler crustal levels of the miogeocline or into the basal part of the structurally overlying allochthon, where

precipitation of metals might have been favored. Migration paths for fluids would have been more disrupted by previous Antler deformation during the subsequent Sonoma event, but with that caveat, somewhat analogous conditions would have prevailed beneath the Golconda allochthon.

Fluid migration within the sediment fill of a compound Antler-Sonoma foreland basin may also have transported metals eastward far beyond the frontal edges of the Roberts Mountains and Golconda allochthons. Multiple studies have shown the combined efficacy of regional dip within foreland basins and the topographic relief of associated thrust highlands for driving fluids toward craton hinge lines from the deep keels of foreland basins (Bethke and Marshak, 1990; Garven et al., 1993; Ge and Garven, 1994). Clastic Paleozoic strata deposited within the foreland region grade eastward into, or inter-tongue laterally with, carbonate assemblages (Fig. 4) to set up an attractive regional geometry for lateral transport and precipitation of metallic elements by migratory fluids.

### CORDILLERAN ARC-BACKARC

Following regional truncation of pre-Mesozoic tectonic belts trending northeast-southwest across the Great Basin (Figs. 3 and 4) by the California-Coahuila transform (Fig. 5), subduction of seafloor downward to the east beneath the truncated continental margin initiated the Cordilleran magmatic arc (Figs. 2 and 6), which trends at a high angle to older tectonic trends (Dickinson, 2000). Along most of the Cordilleran margin, the oldest magmatic components of the Cordilleran arc assemblage are Late Triassic in age (Dickinson, 2004), but the age range of dated plutons in the Mojave region of southern California spans nearly all of Triassic time, with precursors perhaps as old as latest Permian (Barth and Wooden, 2006).

### Backarc Geodynamics

Mid-Triassic to mid-Jurassic evolution of the continental-margin arc to the west was accompanied in west-central Nevada by turbidite sedimentation within a deep backarc basin probably underlain by a thin crustal substratum inherited from Paleozoic slab rollback. Strata along the west flank of the basin interfinger with arc volcanics (Stewart, 1997). Subsidence of the basin floor was enhanced by backarc extension (Wyld, 2000) that persisted into mid-Jurassic time (Oldow and Bartel, 1987). Mid-Jurassic inversion of the backarc basin (Wyld, 2002), recorded by eastward thrusting of basin fill over coeval shelf strata along the Luning-Fencemaker

thrust system (Fig. 6), coincided closely in time with the accretion of an east-facing intraoceanic island-arc complex at the subduction zone along the continental margin in the Sierra Nevada foothills and western Klamath Mountains (Dickinson, 2004, 2005). Collisional tectonism associated with arc accretion may have been linked geodynamically to Luning-Fencemaker thrusting, which probably passed southward along strike into the East Sierran thrust system along the flank of the arc in eastern California (Dunne and Walker, 2004).

Mid-Jurassic arc accretion resulted from consumption of the oceanic Mezcalera plate that had intervened between the west-facing continental-margin arc, built on the edge of Laurentia, and the east-facing offshore intraoceanic arc that was accreted (Dickinson and Lawton, 2001a). Accretion of the intraoceanic arc system expanded the Pacific margin of Laurentia and triggered initial subduction of seafloor on the Farallon plate, which lay beyond the accreting arc that was built on its eastern edge (Fig. 6). Farallon subduction beneath Laurentia was required to continue plate convergence between Laurentia and the Farallon plate once the intervening Mezcalera plate had been consumed along the arc-continent suture in the Sierra Nevada foothills and western Klamath Mountains (Fig. 6).

Late Middle to Late Jurassic (165–145 Ma) backarc magmatism (Fig. 6), which locally overprinted the Luning-Fencemaker thrust system (Smith et al., 1993), spread across the Great Basin well to the east of the continental-margin arc-trench system. The pulse of backarc magmatism can be ascribed provisionally to thermal effects imposed on the mantle by subterranean slab breakoff of the subducted Mezcalera plate after closure of the arc-arc suture to the west (Cloos et al., 2005). Slab breakoff is inferred to have fostered upwelling of asthenosphere to trigger inland magmatism not directly linked to arc activity. Delayed arrival beneath the Great Basin of the leading edge of the Farallon plate subsequently subducted at the western flank of the accreted arc complex (Fig. 6) provided a time window for the episode of backarc magmatism. Isotopic data indicate a stronger mantle influence on Jurassic magmatism within the Great Basin than on younger Cretaceous magmatism (Barton, 1990; Wright and Wooden, 1991), which was associated in time with the foreland Sevier thrust belt farther east (Fig. 7).

West-derived volcanoclastic detritus (Jordan, 1985) that reached the Middle Jurassic Utah-Idaho trough (Fig. 6) in the foreland region was apparently derived from the backarc Jurassic igneous belt of the Great Basin before thrust highlands intervened (Lawton, 1994). Structural relations permissive of synmagmatic crustal



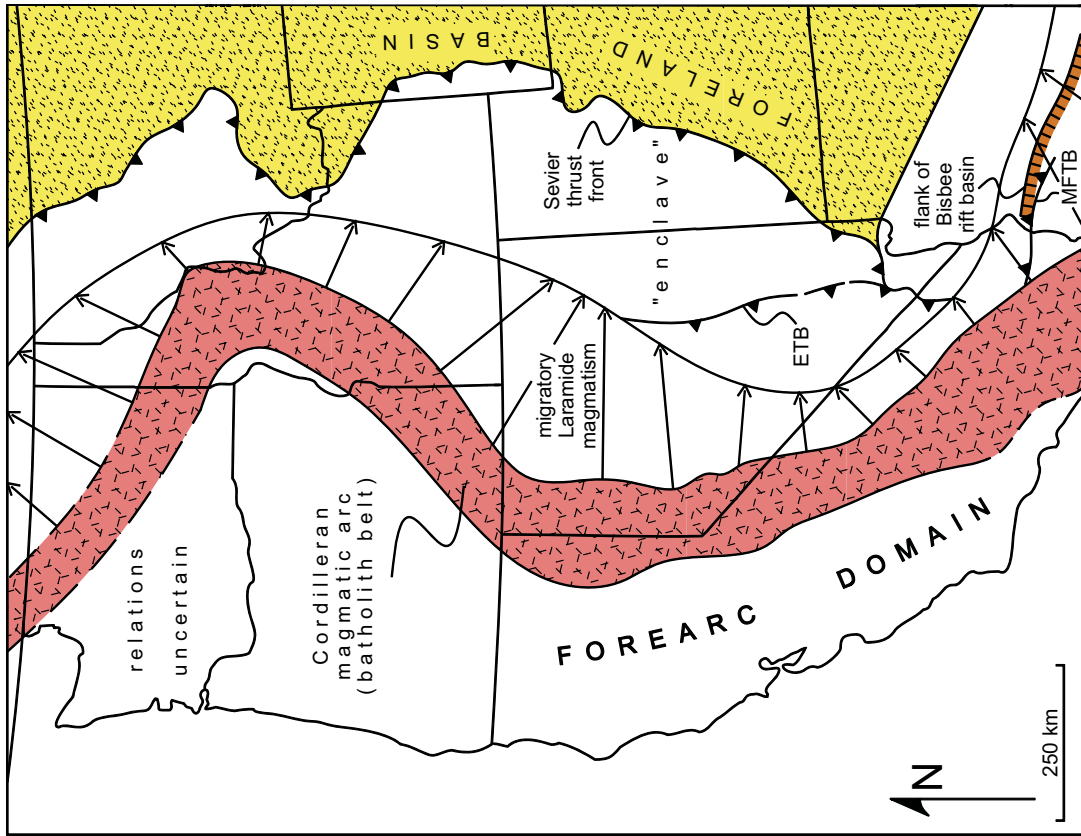


Figure 7. Cretaceous-Paleogene paleotectonic map of the Great Basin and adjacent areas. Migratory Laramide magmatism is modified after Dickinson and Snyder (1978). Flank of Bisbee rift basin is after Dickinson and Lawton (2001b). Subordinate retroarc structures: ETB—Eureka thrust belt; MFTB—Maria fold-and-thrust belt.

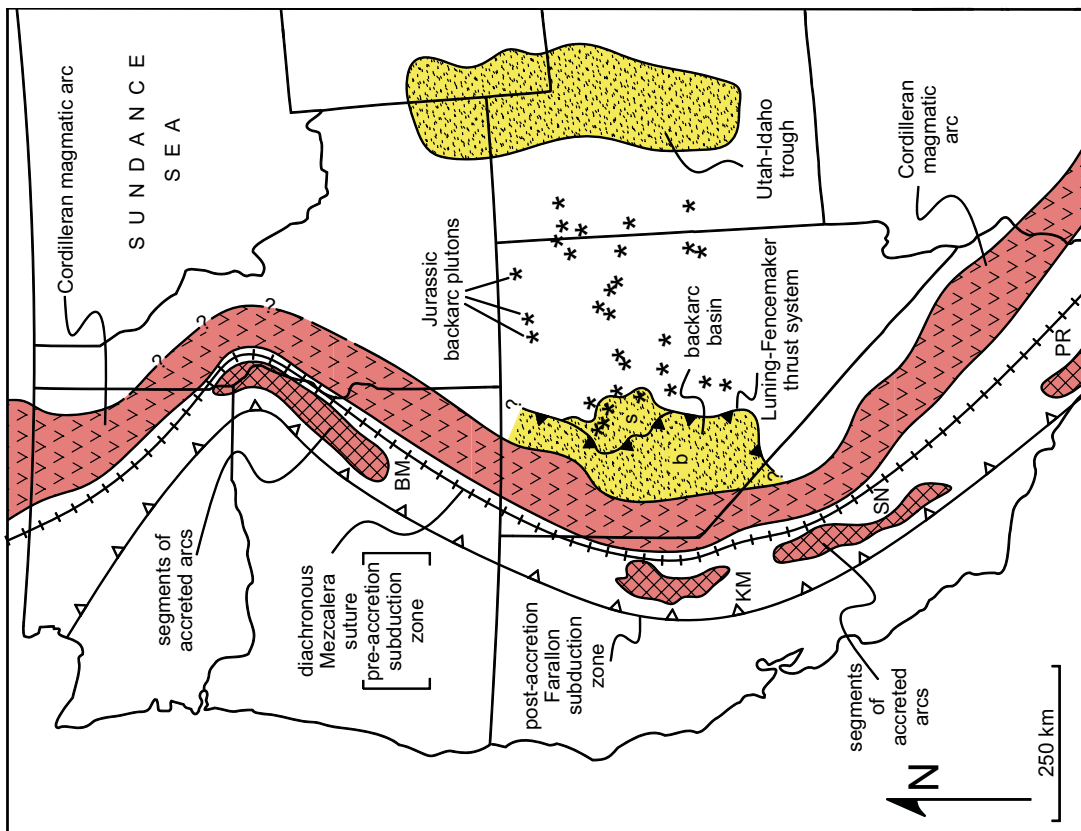


Figure 6. Triassic-Jurassic (post-Sonoma) paleotectonic map of the Great Basin and adjacent areas after mid-Early Triassic initiation of the Cordilleran magmatic arc. Positions of Cordilleran magmatic arc and accreted arc have been adjusted to compensate for younger Cretaceous-Paleogene displacements. Accreted intraoceanic arc segments are shown schematically for the Blue Mountains (BM), western Klamath Mountains (KM), Sierra Nevada foothills (SN), and Peninsular Ranges (PR), Shelfal and basinal (b) facies are shown schematically within the backarc basin of western Nevada. Locations of backarc Jurassic plutons are after Ellison et al. (1990) and Ellison (1995). Configuration of Utah-Idaho trough is after Peterson (1972).



extension across the Great Basin during back-arc Jurassic magmatism (Lawton, 1994) suggest that backarc rifting may have controlled development of the Utah-Idaho trough, which accumulated ~1500 m of strata during the interval 170–160 Ma (Bjerrum and Dorsey, 1995). The restriction of coeval backarc plutons and the Utah-Idaho trough (Fig. 2) to a single broad transect of the Cordilleran orogenic system (Fig. 6) implies some common geodynamic

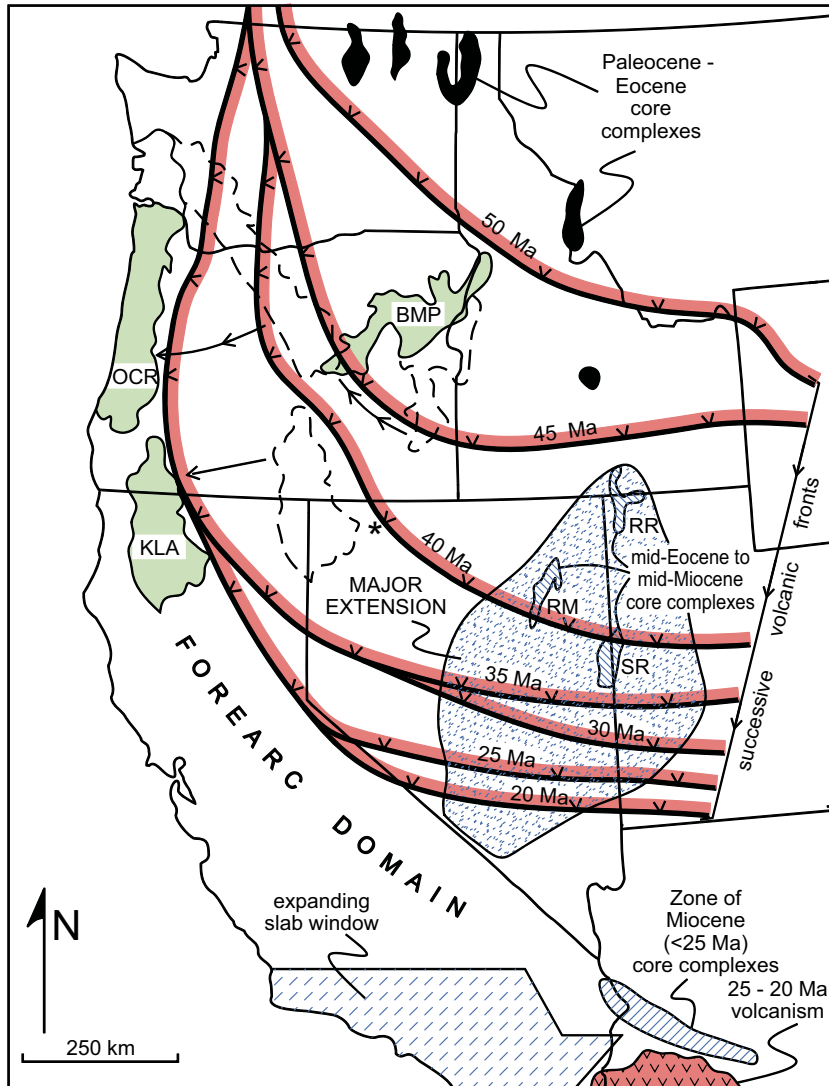
context, and the flanks of the Utah-Idaho trough were at least in part controlled by extensional faulting (Moulton, 1976; Picha and Gibson, 1985). The alternate interpretation (Bjerrum and Dorsey, 1995) that the Utah-Idaho trough was a flexural basin influenced by retroarc thrusting encounters the difficulty that the coeval Luning-Fencemaker thrust system lay too far west for the Utah-Idaho trough to be a foredeep associated with the thrust front (Fig. 6).

Retroarc thrusting along the front of the Sevier belt (Fig. 7) was initiated late in the Early Cretaceous, either during Albian time (Heller et al., 1986; Yingling and Heller, 1992) or perhaps in Aptian time (DeCelles et al., 1995), but in either case not long before the mid-Cretaceous (Cenomanian) Dakota transgression that marked the initial Late Cretaceous flooding of the mid-continent interior seaway. The foredeep depozone of a broad foreland basin (Fig. 7) lay parallel to the Sevier thrust front near the western edge of the Colorado Plateau (DeCelles, 2004). Interpretations that antecedents of the Sevier thrust belt were active in earlier Cretaceous or Jurassic times require the postulate of a “phantom foredeep” (Royse, 1993) that has since been eroded from the Sevier hinterland (DeCelles, 2004). Across Nevada, however, local preservation of volcanic equivalents of Jurassic plutons implies only limited net erosion of the Jurassic magmatic belt in the Sevier hinterland since eruption of the volcanic rocks and emplacement of associated plutons (Miller and Hoisch, 1995).

A divergent branch of the Sevier thrust domain, lying well to the west of the thrust front, extends along the Eureka thrust belt of east-central Nevada (Fig. 7) behind a little-deformed enclave that was markedly distended during Cenozoic time (Bartley and Gleason, 1990). Lower Cretaceous intramontane strata are present along the Eureka thrust belt (Vandervoort and Schmitt, 1990) and also well within the orogen in northwestern Nevada (Quinn et al., 1997). The Sevier thrust belt did not extend farther south than the flank of the pre-mid-Cretaceous Bisbee rift basin (Dickinson and Lawton, 2001b), which extended into the continental block from the opening Gulf of Mexico as far as the inland flank of the Cordilleran magmatic arc (Fig. 7).

### Arc Migrations

Late Cretaceous magmatism was intense along the western side of the Great Basin where the rear flank of the Cordilleran magmatic arc with its batholith belt and outlying satellite plutons encroached upon the intermountain belt (Fig. 7). Post-Jurassic, pre-mid-Cenozoic arc magmatism was much less intense farther east where Laramide magmatism (Fig. 7) associated with the shallowing of slab descent beneath the continental block swept eastward across the Great Basin in latest Cretaceous to earliest Paleogene time (Dickinson and Snyder, 1978). Later mid-Cenozoic (Late Eocene to Early Miocene) magmatism within the Great Basin was associated with a geometrically complex sweep of arc magmatism (Fig. 8) back toward the coast following an amagmatic interval of shallow slab



**Figure 8.** Mid-Cenozoic (Late Eocene to Early Miocene) paleotectonic map of the Great Basin and adjacent areas. Migratory volcanic fronts and core complex ages are after Dickinson (2002). Curved arrows in Pacific Northwest indicate Paleogene rotational displacements of Oregon-Washington (Pacific Northwest) Coast Range (OCR) and Blue Mountains Province (BMP) and translational displacement of Klamath Mountains block (KLA), with former positions shown by dashed outlines. Region of major post-mid-Eocene, pre-mid-Miocene extension is adapted after Seedorff (1991) and Axen et al. (1993). Locations of core complexes are after Wust (1986) and Axen et al. (1993). Selected Great Basin core complexes: RM—Ruby Mountains; RR—Raft River; SR—Snake Range. Zone of southwest Arizona volcanism is after Spencer et al. (1995). Extent of slab window is adapted after Dickinson (1997).



descent (Fig. 2). Over wide areas, mid-Cenozoic plutonism was the most prominent intrusive episode in the Great Basin since backarc Jurassic plutonism (Miller et al., 1987).

In both the Great Basin and the Pacific Northwest, migratory Eocene to Oligocene magmatism has been viewed as unrelated to subduction but instead controlled entirely by intracontinental extension (Seedorff, 1991; Hooper et al., 1995). Patterns of seafloor magnetic anomalies offshore indicate, however, that subduction was under way along the continental margin throughout the evolving magmatic episode. Owing to shallow plate descent, the thermal state of the subducted Paleogene slab beneath the entire Great Basin and Pacific Northwest was comparable to the present thermal state of the more steeply dipping Neogene slab directly beneath the modern Cascades volcanic arc (Severinghaus and Atwater, 1990). Extensional intra-arc and backarc tectonism were a part of the mid-Cenozoic geodynamic picture for the Great Basin but can be related to slab rollback accompanying arc migration. The complex migratory pattern of Cenozoic magmatism across the Pacific Northwest and Great Basin implies slab flexure or rupture during slab rollback, but the areal distribution of volcanic centers defines a coherent array of volcanic fronts embodying components of motion parallel as well as normal to the continental margin (Fig. 8). Southward-migrating volcanic fronts crossing the Great Basin were linked longitudinally, at each successive stage of their progressive evolution, with the continental-margin magmatic arc of the Pacific Northwest and Canada.

### Metallogenic Impacts

Significant ore deposition in the Great Basin accompanied both Mesozoic (Barton, 1996) and mid-Cenozoic (Seedorff, 1991; Henry and Ressel, 2000) arc and backarc magmatism, which typically involves addition of mantle components to various levels in the crust and fosters extensive thermochemical reworking of crustal materials. Advective heat flux is commonly sufficient to generate crustal melts and to stimulate varied hydrothermal and other metasomatic processes. Consequently, Mesozoic-Cenozoic magmatism subjected the Great Basin to superposed episodes of potential metal mobilization and precipitation.

Fluid migration in response to Mesozoic thrusting may also have been complex in both space and time. Jurassic Luning-Fencemaker (Fig. 6) and Cretaceous Eureka (Fig. 7) thrust belts may both have induced fluid migration on a subregional scale. The regional Sevier foreland basin lay largely east of the Great Basin,

but its deformed western fringe and the associated retroarc thrust belt was mostly within the Great Basin, and little is yet known about the migration of fluids within basement and cover that were underthrust to deep crustal levels beneath the Sevier hinterland in the eastern Great Basin (Miller and Gans, 1989; Hudec, 1992; McGrew et al., 2000). By the end of Sevier thrusting, much of the present Great Basin was a broad, highstanding plateau (Dilek and Moores, 1999), similar topographically to the Altiplano of the modern Andes and termed by analogy the "Nevadaplano" (DeCelles, 2004). Inferences about the evolving configuration of the retroarc Sevier foreland system through time are complicated by the need to take into account not only the structural assembly and isostatic effect of telescoped thrust loads in the upper crust, but also the geodynamic effect of an underthrust slab moving at depth into the mantle below (Mitrovica et al., 1989). Surface elevations of both the thrust belt and the foreland basin, with implications for hydrologic conditions in the crust, resulted from these twin crust-mantle influences on isostasy (DeCelles and Giles, 1996).

### COMPOSITE CENOZOIC EXTENSION

The very existence of the Great Basin as an internally drained tract of mountainous topography broken by sedimented valleys stems from a Cenozoic regime of extensional tectonism. The terms Great Basin and Basin and Range Province denote virtually co-extensive domains in Nevada and adjoining Utah, but the Basin and Range taphrogen (Dickinson, 2002) is much larger, embracing an elongate region that extends from the Pacific Northwest to central Mexico (Fig. 1). The composite extensional domain was subject during Cenozoic time to different geodynamic controls that varied in both space and time. In the Great Basin itself, two successive phases of extensional deformation related to different geodynamic settings can be distinguished (Dickinson, 1991, p. 24–25, 33–36).

#### Basin-Range Tectonism

The more recent of the two extensional regimes (Fig. 9) controlled development of the modern basins and ranges beginning in Early Miocene time (ca. 17.5 Ma), after the San Andreas transform system was established along the southern California coast as the boundary between the Pacific and North American plates of lithosphere (Dickinson, 1997). Before then, the two regional plates were largely separated by oceanic microplates that shielded them from direct interaction. Although various geodynamic

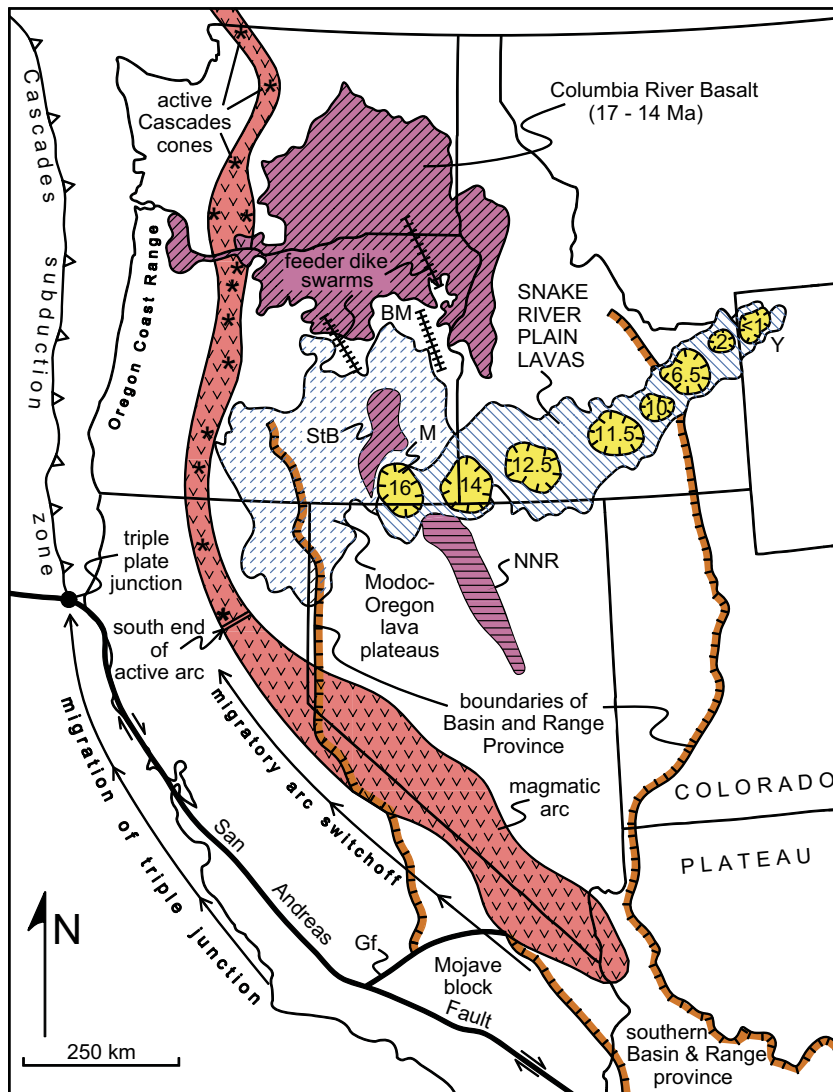
scenarios have been proposed to explain classic basin-range deformation typified by block faulting, transtensional torsion of the continental block under the influence of shear interaction along the San Andreas transform remains the most attractive (Atwater, 1970). Penetration of the Eastern California shear zone strand of the San Andreas system as far east as the Walker Lane belt (Stewart, 1988) near the California-Nevada border demonstrates the distributive style of transform deformation by strike slip in continental crust.

Basaltic magmatism in the backarc of the Pacific Northwest began in mid-Miocene time (17–14 Ma) with voluminous eruptions of Columbia River Basalt (Fig. 9), which may have been triggered by initial deformation of the continental plate under transform shear (Dickinson, 1997). The trend of feeder dike swarms for the Columbia River Basalt is parallel to the coeval Northern Nevada Rift (Zoback et al., 1994) of the Great Basin to the south (Fig. 9), a coincidence that argues for similar mid-Miocene stress orientations throughout the backarc region from the Pacific Northwest down into the Great Basin. An elongate chain of silicic calderas, nestled within or flooded by basalt lavas of the Snake River Plain, was initiated during the same time frame (16–14 Ma) just north of the Northern Nevada Rift but to the south of preserved remnants of the Steens Basalt appendage of the Columbia River Basalt field (Fig. 9). The Snake River Plain, taken here to delimit the Great Basin segment of the Basin and Range taphrogen on the north (Fig. 1), was superimposed across the Basin and Range Province, which continued to evolve both north and south of the Snake River Plain as volcanism proceeded (Dickinson, 2002). Widespread basaltic volcanism along the Snake River Plain succeeded migratory silicic volcanism that was associated with hotspot caldera complexes that young progressively along the plain from the Miocene McDermitt caldera (16 Ma) on the southwest to the modern Yellowstone caldera (younger than 1 Ma) on the northeast (Fig. 9).

#### Core Complex Relations

Pre-mid-Miocene episodes of Cenozoic extension within the Great Basin, including tectonic denudation of core complexes (Fig. 8), cannot be related to evolution of the San Andreas transform system because subduction of the Farallon and derivative oceanic plates was still under way along the continental margin to the west. The timing of pre-basin-range extensional tectonism in relation to migratory intermediate to silicic arc magmatism suggests instead that it can be viewed as intra-arc or backarc





**Figure 9.** Post-18 Ma (basin-range) paleotectonic map of the Great Basin and adjacent areas. Boundaries of Basin and Range Province and extent of magmatic arc are adapted after Dickinson (1997, 2002). Postarc volcanic centers (and local Late Miocene core complexes), formed near California-Nevada line in western Great Basin as Mendocino triple junction migrated northward (offshore arrow), not shown within span of earlier arc magmatism. Chain of Snake River Plain calderas (hachured circles with ages in Ma inside circles) of Yellowstone (Y) hotspot track (initiated with McDermitt caldera [M]) is modified after Pierce and Morgan (1992). Mid-Miocene Columbia River dike swarms (west to east): Monument, Cornucopia, Chief Joseph. Modoc-Oregon lava plateaus are largely Pliocene in age at the surface. Other features: BM—Blue Mountains; Gf—Garlock fault; NNR—Northern Nevada Rift (geophysical extent); StB—Steens Basalt (coeval with Columbia River Basalt).

deformation induced by slab rollback rather than transform shear (Dickinson, 2002).

Local synvolcanic extensional basins of Eocene age are known (Potter et al., 1995), but there is no regional association between initial extension and the onset of migratory Great Basin magmatism. The onset of volcanism generally preceded major extensional deformation (Gans et al., 1989; Seedorff, 1991; Spencer et al., 1995;

Henry and Ressel, 2000), and tilt and offset of the volcanic rocks form a prime geometric control for analysis of mid-Cenozoic extensional features. Detachment faulting and tectonic denudation of core complexes took place for the most part in the wake of the migratory volcanic fronts that swept southward through the Great Basin (Fig. 8) from Eocene time in Idaho to Oligocene time in Nevada, with final exhumation of

core complexes in northeastern Nevada (Fig. 8) delayed until Early Miocene time (Dickinson, 2002). Small Late Miocene core complexes near the California border in southwesternmost Nevada were probably related to superextension linked to strike slip along en echelon subparallel strands of the Walker Lane fault system (Dickinson, 2002).

Multiple reinforcing mechanisms can be invoked for pre-basin-range extensional deformation (Dickinson, 1991, p. 34), including: (1) release of intraplate compressive stress as post-Laramide steepening of slab descent reduced interplate shear at depth, (2) retreat of the offshore trench hinge by slab rollback to allow lateral expansion of the intermountain region, (3) gravitational collapse of an overthickened crustal welt produced by earlier orogenic contraction, and (4) advective heating of the crustal profile by mantle melts to promote intracrustal failure under extensional stresses. The arc-rear or backarc setting of mid-Cenozoic extension contrasted with the “back-transform” setting of currently active basin-range deformation.

Early stages of Neogene basin-range tectonism also developed, however, in a backarc setting. Northward migration of the Mendocino triple junction, which marks the northern end of the San Andreas transform, gradually switched off the magmatic arc west of the Great Basin (Fig. 9), but only after classic basin-range tectonism had begun to the east. Southward migration of the arc trend had largely ceased, however, before the onset of the modern basin-range regime (Figs. 8 and 9); this implies that a geodynamic influence from slab rollback was characteristic of only the pre-basin-range mid-Cenozoic extension. As the development of mid-Cenozoic core complexes involved ductile flow of lower crust (Gans, 1987; Wernicke, 1992; MacCready et al., 1997), the change in structural style from detachment faulting to block faulting may have stemmed in part from a thermomechanical transition in the rheological behavior of crust undergoing extension as cumulative crustal thinning reduced the thickness of ductile lower crust (Harry et al., 1993; Spencer et al., 2001). A rheological stiffening of the crustal profile may also have facilitated transfer of transform shear inland from the San Andreas plate boundary during Neogene basin-range deformation. The scale of Figure 9 precludes illustration of transform strands of the Walker Lane fault system and associated postarc igneous centers that supplanted arc magmatism near the California-Nevada border as progressive arc switchoff swept the southern end of the magmatic arc northward toward its present locus, limited to the Cascades chain of the Pacific Northwest.



## Metallogenic Influences

The metallogenic imprint of Cenozoic extensional deformation on Great Basin ore deposits (Dreier, 1984; John, 2001) may have had two parallel and partly interacting aspects. On the one hand, mantle-derived magmas may have altered the crustal profile over wide areas, both directly by injection of contributions from the mantle and indirectly by stimulating generation of crustal magmas with advective heat flux that accompanied the rise of mantle melts. On the other hand, crustal extension may have generated extensive metal-transporting hydrothermal systems, associated both with igneous centers and with the tectonic exhumation of hot rock masses from the midcrust within core complexes. Either detachment faulting or steep normal faulting, without actually playing any significant role in metallogenesis, might also have brought buried ore bodies closer to the surface where they could be more readily exposed by erosion.

## SUMMARY PERSPECTIVES

The spectrum of Great Basin ore deposits is too broad to permit effective summary in this tectonic overview, but the following recapitulation of metallogenic influences through Great Basin geologic history provides a basis for more detailed discussion:

(1) If any metals were derived from mantle sources at times postdating the crustal genesis that formed Precambrian basement, they could have entered the crust during rift events (Mesoproterozoic locally, Neoproterozoic along the length of the miogeocline, Neogene over much of the Great Basin) or within arc and backarc igneous systems (Late Triassic to Late Cretaceous along the western fringe of the Great Basin, Jurassic and latest Cretaceous through mid-Cenozoic farther east).

(2) If any metals were derived from crustal basement rocks, they could have been scavenged during any of the same rift-related or arc-related igneous episodes, as well as during generation of dominantly crustal Late Cretaceous melts in the eastern Great Basin.

(3) If any metals were derived from strata of sedimentary successions, they could have been set in motion by hydrothermal activity associated with any of the same igneous episodes, or by intracrustal fluid flow triggered by thrusting (Paleozoic and Jurassic in the western Great Basin, Cretaceous in the eastern Great Basin), which tilted basinal strata to form migration pathways and created concomitant topographic relief to provide a hydrologic drive.

(4) Metalliferous fluids set in motion by any of these diverse mechanisms may have followed

either local or subregional migration paths, controlled by bedding or fractures, into a variety of shallower and cooler source rocks.

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